# MINERAL DEPOSITS

BY

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# PREFACE TO THE FIRST EDITION

Mineral deposits are usually classified and described by the metals or the substances which they contain; for instance, deposits of copper are described together, with little or no effort to separate them into genetic groups. Where a genetic treatment has been attempted it appears to me to have failed in not giving due weight to the physical conditions attending the genesis. Furthermore, it is the custom to divide the mineral deposits into two groups—the metallic and the non-metallic—a line of division which can hardly be defended except on the ground of long-established habit.

This book is the outcome of a desire to place the knowledge of mineral deposits on the broader and more comprehensive basis of a consistent genetic classification and thus bring it into a more worthy position as an important branch of geology. Opinions may differ as to whether our present knowledge is sufficient for such an undertaking. Believing that the time has come for a first attempt, I present this volume, in the hope that its shortcomings may be judged leniently.

The impetus of the work came during the preparation of a series of lectures a few years ago, and a course along the general lines followed in this volume has since then been presented annually at the Massachusetts Institute of Technology.

The general plan has been to select a few suitable examples to illustrate each genetic group of deposits. These examples have been chosen regardless of their geographic location, and it was of course necessary to give up any attempt to describe deposits in detail or to present all known examples of any particular type. As the larger part of my experience has been within the United States of America, a considerable number of examples were gathered from this country. This experience I owe to the United States Geological Survey, in which I have had the honor to serve for many years. My indebtedness to my friends and associates in that organization is greater than can be expressed in words.

WALDEMAR LINDGREN.

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# MINERAL DEPOSITS

## CHAPTER I

## INTRODUCTION

### ECONOMIC GEOLOGY

The application of geology to the practical problems of the industries and the arts constitutes economic geology. This branch of the science includes as its most important division the study of deposits of useful minerals and rocks, but it also teaches the occurrence of underground waters, explains the derivation and constitution of soils and applies geologic principles to the planning of important engineering works.

Only a part of the whole field of economic geology will be covered in these chapters. They will be confined to a description, by classes and type examples, of the occurrence, structure, and origin of the principal deposits of metallic and non-metallic minerals of economic importance. The subjects of coals, mineral oils, and structural materials could not be included without unduly increasing the bulk of the volume. Little space has been given to statistics, while the problems of correlation and origin have been treated rather fully. A general part describing principles of universal application precedes the detailed characterization of the various classes.

A complete treatment of the subject should also include discussions of distribution, production, and valuation of deposits, as well as statements of the uses of the materials mined, processes of mining and reduction, and criteria for judging the value of the products. Such a complete presentation is not attempted in this volume. By examining the subject from a scientific rather than from a utilitarian viewpoint, the student will obtain a clearer insight into the geologic relationship of the various deposits.

Throughout its broad domain economic geology stands on the fundamental sciences of chemistry and physics. It is related on one side to theoretical geology, paleontology, mineralogy, and petrography; on another side to mining, metallurgy, and many other technological arts; on still another side to economics and finance. A student who tries to approach the subject without the necessary knowledge of the allied sciences and arts is building on poor foundations. Even with this aid the study offers peculiar difficulties. The alteration of rocks close to many mineral deposits is intense and, as a result, the student who is familiar with only the fresh, unaltered specimens finds himself in the midst of puzzling and strange types that he is unable to classify with certainty. Altered andesites may assume the aspect of quartzites; a question may arise as to whether a silicified rock was once a limestone or a porphyry; diabases may at some places be converted into white fine-grained calcite-sericite-quartz rocks and at other places appear as aggregates consisting mainly of epidote and chlorite. These examples suffice to show that rock alteration is a subject of prime importance for the mining geologist.

## DISTRIBUTION OF ELEMENTS IN THE EARTH[1]

**The Earth as a Whole.**—In recent years a better understanding of the interior of the earth has been obtained, based on the velocity of earthquake waves at different depths, on the analogy with meteorites, on the compressibility of rocks, and on the chemistry of igneous rocks.

The earth is an essentially rigid spheroid of a specific gravity of about 5.52, and a radius of about 6,400 kilometers. The temperature of the core is unknown, but it is probably not so high as the measured increment at the surface would suggest. Local plasticity and fusion may result in the surface shell by various changes in physical conditions or perhaps by radioactive influences.

Washington holds that the core with a radius of 3,400 kilometers consists of nickel-iron, its density being about 10. The central core passes gradually, in about 1,400 kilometers, into a magnesium-iron silicate, a shell of the composition of peridotite, with more or less oxides and sulphides. The peridotite shell is held to have a thickness of 1,600 kilometers and a specific gravity of 4 Above the peridotite shell is the "crust" of the earth, the thickness of which is from 60 to 100 kilometers. The lower part is of basaltic or gabbroitic composition, and it grades upward into the outermost shell, 15 to 20 kilometers thick, of granitic or granodioritic composition and a density of 2.77 to 2.80. This is the only portion of the globe that is more or less open to our inspection. It is this very thin, uppermost shell that has usually been referred to in the

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[1] J. H. L. Vogt, Ueber die relative Verbreitung der Elemente, etc., *Zeitschr. prakt. Geol.*, 1898, pp. 225–238; 314–325.

V. M. Goldschmidt, Der Stoffwechsel der Erde, Videnskapsselsk. 1, Skr. Math.-Nat. Kl. 11, Oslo, 1922, 25 pp.

Geochemische Verteilungsgesetze der Elemente, I-V, *idem*, 1923, 16 pp.

F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 12–44.

F. W. Clarke and H. S. Washington, Composition of the earth's crust, *Prof. Paper* 127, U. S. Geol. Survey, 1924.

H. S. Washington, The chemical composition of the earth, *Am. Jour. Sci.*, 5th ser., 9, 1925, pp. 351–378.

discussions of "the average igneous rock." The thickness of 10 miles or 16 kilometers is often assigned to it, and from it the material for all of our rock analyses is taken. It consists of granitic rocks, gneiss and schist, volcanic flows, intrusives of various kinds, and a thin veneer of sediments.[1]

V. M. Goldschmidt and G. Tamman consider that the original fluid earth separated by cooling into three fluid, immiscible phases: Metallic melt, sulphide melt, and silicate melt, and one gas phase (the atmosphere). Considerable difference of opinion exists as to the importance of the sulphide melt as a separate phase. Influenced by gravity the three melts assumed a concentric disposition. Goldschmidt also assumes the existence of an eclogite shell of garnet and pyroxene, minerals stable under conditions of greatest pressure, and below this a shell of mixed sulphides and oxides. How far in depth crystallinity can exist seems, however, uncertain.

The silicate shell separated by crystallization and gravity into heavier and lighter parts. Mineral deposits were thus formed; most of them, however, are buried beyond our ken. In the solidified crust chemical processes of many kinds produced further differentiations, including the development of many other mineral deposits.

The gas phase was further separated into a liquid and a gaseous phase corresponding to the hydrosphere and the atmosphere. By the action of these two on the silicate shell many separations and concentrations of the elements took place and these processes produced the rest of the mineral deposits.

On the assumption of a nickel-iron core several calculations of the composition of the earth as a whole have been made, the latest by Washington (1925). He finds that such an earth would contain approximately; metallic iron, 32 per cent; nickel, 3 per cent; oxygen, 28 per cent; iron (in silicates), 8 per cent; magnesium, 9 per cent, all others 20 per cent.

On geological grounds it is to be expected that the rocks of the crust increase in basicity with depth. Seismological observations[2] show an increase with depth of velocities of the longitudinal and transverse waves. Comparison with the known compressibility of rocks indicates that the upper layer is granite, that the intermediate layer may be basalt (glassy?), and that the lowest may be peridotite. The granitic shell would have a thickness of 10 kilometers, the intermediate layer 20 kilometers. The velocities increase to a depth of about 1,500 kilometers and then become constant and finally again decrease. This is interpreted by Jeffreys to

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[1] These designations correspond in a general way to Suess' core, *nife* (nickel iron) and to his outer shells of *sima* (magnesium silicate) and *sial* (silica, aluminum).

[2] Harold Jeffreys, The earth, New York, 1929, pp. 116 and 137.

B. Gutenberg, Lehrbuch der Geophysik, Berlin, 1929, p. 262.

R. A. Daly, Igneous rocks and the depths of the earth, New York, 1933, 598 pp.

indicate a central core beginning at a depth of about 2,900 kilometers and having a sharp boundary. This would be the "nickel-iron" core.

It is well to bear in mind that, however plausible, the "nickel-iron core theory" and the various shells cannot yet be accepted as proved facts.

**The Composition of the "Crust."**—The uppermost thin shell of the earth comprises the atmosphere, the hydrosphere and the "crust."

AVERAGE ELEMENTARY COMPOSITION OF IGNEOUS ROCKS IN A TEN-MILE CRUST

| | |
|---|---|
| Oxygen | 46.59 |
| Silicon | 27.72 |
| Aluminum | 8.13 |
| Iron | 5.01 |
| Calcium | 3.63 |
| Sodium | 2.85 |
| Potassium | 2.60 |
| Magnesium | 2.09 |
| Titanium | 0.63 |
| Phosphorus | 0.13 |
| Hydrogen | 0.13 |
| Manganese | 0.10 |
| Sulphur | 0.052 |
| Barium | 0.050 |
| Chlorine | 0.048 |
| Chromium | 0.037 |
| Carbon | 0.032 |
| Fluorine | 0.030 |
| Zirconium | 0.026 |
| Nickel | 0.020 |
| Strontium | 0.019 |
| Vanadium | 0.017 |
| Cerium and Yttrium | 0.015 |
| Copper | 0.010 |
| Uranium | 0.008 |
| Tungsten | 0.005 |
| Lithium | 0.004 |
| Zinc | 0.004 |
| Columbium and Tantalum | 0.003 |
| Hafnium | 0.003 |
| Thorium | 0.002 |
| Lead | 0.002 |
| Cobalt | 0.001 |
| Boron | 0.001 |
| Beryllium | 0.001 |
| | 100.000 |

By the last, rather arbitrary, term we mean the rocks which are directly accessible to us by boring and mining operations or indirectly accessible by faulting, folding, or erosion. The thickness of this crust is assumed to be 10 miles or 16 kilometers.

In calculating the average composition of the crust it is necessary to separate the igneous rocks from the sedimentary rocks. The latter are derived by erosion and sedimentation from the igneous rocks and form but a thin, irregularly distributed veneer on the crust. Clarke calculates that the crust to the depth of 10 miles consists of 95 per cent of igneous rocks, 4 per cent of shales, 0.75 per cent of sandstones, and 0.25 per cent of limestones and dolomites. The sediments average poorer in calcium, magnesium, and especially in sodium than the igneous rocks and thus show the effect of leaching. They also contain more potassium and carbon dioxide, but on the whole they are, as would be expected, similar in composition to the igneous rocks.

Clarke and Washington have calculated the average composition of igneous rocks in the crust from about 5,000 analyses from all parts of the world. Their method of averaging is explained in their paper referred to above.[1] The data show that all of the elements are present in the crust, though not in uniform distribution. Gold, silver, copper, lead, and zinc may be found, for instance, in practically all rocks, though special methods must sometimes be used to detect them. The same authors have also estimated the amount of most of the rare elements probably present though these estimates are not always of the same weight. The data compiled by Clarke and Washington are shown in the table shown on page 4.

On page 6, the rare elements are first stated in exponential form as parts of a gram per gram of rock. The second figure indicates the corresponding percentage. An $n$ is used instead of $x$ to indicate the first significant though unknown integer. The last figure gives the amount in more intelligible form as milligrams per metric ton.

The eight elements first named above make up 98.62 per cent of the igneous rocks.

Among the six principal metals shown in the average composition only iron, magnesium, and aluminum are of economic importance as metals. The lighter elements predominate, the atomic weight of each falling below 56 (Fe 55.9). In the average composition many of the rarer metals are represented; but, except titanium which amounts to 0.63 per cent, all these metals average below 0.1 per cent. Platinum, gold, silver, copper, lead, zinc, antimony, arsenic, tin, quicksilver, molybdenum, tungsten, and others are present in amounts less than 0.01 per cent.[2]

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[1] *Prof. Paper* 127, U. S. Geol. Survey, 1924, pp. 10–12. The method of compilation has been criticised in some quarters, the assertion being that the result places too much emphasis on rarer rocks and that the real composition should be more acidic. However that may be, these are the best data we have at present.

[2] In a paper by G. Berg (*Zeitschr. prakt. Geol.*, 1925, pp. 73–79) the data of Clarke and Washington are subjected to criticism. His estimates are in part as follows: Zn 0.006, Pb 0.000,8, Mo 0.000,6, Sn 0.000,5, As 0.000,45, Sb 0.000,025, Cd 0.000,01, Ag

ESTIMATES OF THE QUANTITIES OF RARE ELEMENTS IN IGNEOUS ROCKS IN A TEN-MILE CRUST

| Element | Parts per gram | Per cent | Milligrams per metric ton |
|---|---|---|---|
| Molybdenum | $n \times 10^{-6}$ | 0.000$n$ | 1,000 to 9,000 |
| Rubidium | | | |
| Arsenic | | | |
| Tin | | | |
| Bromine | | | |
| Cesium | $n \times 10^{-7}$ | 0.000,0$n$ | 100 to 900 |
| Scandium | | | |
| Antimony | | | |
| Cadmium | | | |
| Mercury | | | |
| Iodine | | | |
| Bismuth | $n \times 10^{-8}$ | 0,000,00$n$ | 10 to 90 |
| Silver | | | |
| Selenium | | | |
| Platinum | $n \times 10^{-9}$ | 0.000,000$n$ | 1 to 9 |
| Tellurium | | | |
| Gold | | | |
| Iridium | $n \times 10^{-10}$ | 0.000,000,0$n$ | 0.1 to 0.9 |
| Osmium | | | |
| Thallium | | | |
| Indium | $n \times 10^{-11}$ | 0.000,000,00$n$ | 0.01 to 0.09 |
| Gallium | | | |
| Palladium | | | |
| Rhodium | | | |
| Ruthenium | | | |
| Germanium | | | |
| Radium | $n \times 10^{-12}$ | 0.000,000,000$n$ | 0.001 to 0.009 |

The percentages of the useful metals in the rocks as given above do not by any means indicate the amount available for industrial use. That amount indeed is so infinitesimal in relation to the volume of the crust that it can not be conveniently expressed on the basis of percentages. The metals in the deposits of useful minerals then comprise only a minute fraction of the quantity of metals in the crust—a fraction which has been locally accumulated by this or that process of concentration.

In general, igneous rocks contain more of the heavy metals than do the sedimentary rocks. We are well justified in regarding the former as the original source of these metals. Dissipation by solution accompanies sedimentation and the many metals found in traces in the sea water furnish evidence of this. On the other hand, it is true that certain

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0.000,004, Bi 0.000,003, Hg 0.000,002,5, Au 0.000,000,1, Pt $n \times 10^{-11}$, Ir $n \times 10^{-12}$, Pd $n \times 10^{-13}$, Ra $n \times 10^{-14}$.

I. and W. Noddack (*Die Naturwissenschaften*, 18, Aug. 29, 1930) mixed and analyzed 118 common igneous rocks and obtained slightly different figures. See also E. Troeger, Der Gehalt an Selteneren Elementen der Eruptivgesteine, *Chem. Erde*, 9(3), 1935, 286–310.

kinds of sedimentation will result in a local concentration of metals, such as iron, zinc, cobalt, nickel, and vanadium.

**Minerals of the Crust.**—The bulk of igneous rocks consists of relatively few minerals, mostly silicates and oxides. They comprise in the main quartz, feldspar, leucite, nepheline, pyroxene, amphibole, muscovite, biotite, olivine, garnet, magnetite, ilmenite, chromite to which should be added the ever-present apatite. The crystalline schists contain besides the minerals mentioned such secondary products as garnet, staurolite, cyanite, andalusite, cordierite, chlorite, and sericite. The sedimentary rocks may contain, besides fragments of the minerals mentioned, much calcite and dolomite and differentiated sedimentary products, such as salt, gypsum, iron ores, and phosphates. More than 1,200 mineral species are known but most of them come from mineral deposits where special processes have been active.

Concerning the form in which the rarer elements enter into rocks we do not have full information. Gold is probably always native metal. Many metals like chromium, nickel, vanadium, titanium, and copper may be contained in isomorphous mixture in silicates. But copper may also be present as sulphide; chlorine, in apatite or in silicates; fluorine as a fluoride or a fluosilicate; boron generally as a borosilicate; tin as oxide, sulphide, or silicate. Much remains to be learned about the rarer elements in rocks by the X-ray method or by spectroscopy. According to Goldschmidt the association is largely determined by the isomorphism of the elements.

In the earlier editions of this book many data were given as to the occurrence of gold, silver, copper, lead, and zinc in rocks. As these facts are recited in Clarke's "Geochemistry" and in the *Professional Paper* 127 by Clarke and Washington they may be omitted here. Regarding gold and silver reference should be made to Don's[1] work which showed that these metals are readily introduced into rocks near precious metal deposits. Luther Wagoner[2] using the cyanide method found that many igneous rocks contained gold to the extent of about 100 milligrams per metric ton and silver of about 1,000 to 5,000 of the same units.

It is well known that sea water contains some gold and silver, the various determinations ranging from 5 to 65 milligrams of gold per metric ton of sea water; silver is present in larger amounts ranging up to 1,900 milligrams. Salt formed by evaporation of sea water also, naturally, contains the precious metals. Wagoner found 457 milligrams of gold and 54.4 grams of silver to the metric ton of such salt, and later reported both metals in appreciable quantities in deep-sea dredgings. Latei

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[1] J. R. Don, The genesis of certain auriferous lodes, *Trans.*, Am. Inst. Min. Eng. 27, 1898, p. 564.

[2] *Trans.*, Am. Inst. Min. Eng., 31, 1901, pp. 798–810; 38, 1907, p. 704.

examinations[1] would indicate that the quantity of gold in sea water has been overestimated and is rather about 1 milligram per ton.

Many cases of visible gold in rocks are known. These occurrences though not of economic importance comprise pegmatite, granite, pitchstone, and porphyries of various kinds. The presence of some gold in peridotite is inferred from occurrences in Tasmania. Visible gold has also been observed in some crystalline schists,[2] and, of course, in many fragmental rocks.

The table shows (p. 4) that certain metals, regarded as rare, such as titanium, chromium, zirconium, nickel, vanadium, and cerium are relatively more abundant in the crust than such industrial metals as copper, lead, and zinc.

Washington notes the difference between the "petrogenic" elements or those that make up the bulk of the rocks, and the "metallogenic" elements or "ore elements" which include those of economic importance. The former occur mainly as silicates, oxides, fluorides, and chlorides; the latter as sulphides, arsenides, selenides, and tellurides. The former are abundant in the crust, the latter scarce and localized in mineral deposits.

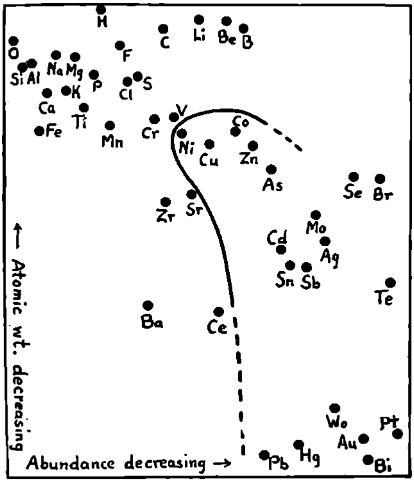

FIG. 1.—Diagram showing atomic weights and relative abundance of the elements. Group to right represents metallogenic elements in the crust. (*After O. H. Erdmannsdörffer.*)

Considering this question from a slightly different standpoint V. M. Goldschmidt divides the elements into three classes: (1) Associated in the core of the earth (siderophile) such as Fe, Ni, Cr, P, C, Pt, (Au?); (2) Associated with sulphur or similar elements: (chalcophile) S, Se, Te, Fe, Ni, Co; Cu, Zn, Cd, Pb; Sn, Ge, Mo; As, Sb, Bi; Ag, Au, Hg; Pd, (Pt); Ga, In, Tl; and (3) Associated with silicate melts (lithophile): O, (S), P, (H); Si, Ti, Zr, Hf, Th; F, Cl, Br, I; B, Al, Cerium metals; Li, Na, K, Rb, Cs; Be, Mg, Ca, Sr, Ba; Fe, V, Cr, Mn, Ni; Nb, Ta, W, U, Sn.

Within the silicate zones the various elements show also a decided preference for certain rocks. In the basic rocks we find phosphorus, sulphur, nickel, iron, chromium, titanium, and platinum. In the acidic

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[1] F. Haber and J. Jaenicke, *Zeitschr. f. anorg. Chemie*, 147, 1925, pp. 156–170.

[2] J. E. Spurr, *Eng. and Min. Jour.-Press*, 76, 1903, p. 500.

A. Lacroix, Sur l'origine de l'or de Madagascar, *Compt. Rend.*, 132, 1901, pp. 180–182.

rocks we find fluorine, boron, lithium, zirconium, tin, tungsten, tantalum, cerium metals, molybdenum, thorium, and beryllium. Barium and strontium are often concentrated in highly alkaline rocks. Gold and silver accompany acidic or intermediate rocks, but in smaller amounts may also follow more basic rocks. It seems that copper occurs in larger quantities in basic than in acidic rocks. In the last differentiates of rocks, namely pegmatites, we find these relations still further accentuated.

## MINERAL DEPOSITS

The thin outermost crust of the earth, directly or indirectly accessible to us, we have defined as a shell having a thickness of 10 miles or 16 kilometers, while the radius of the earth is about 4,000 miles. The deepest shaft is about 8,200 feet (on the Rand), the deepest bore hole 12,786 feet (in Western Texas). Many holes have reached 11,000 feet.

This crust consists, as explained above, mainly of silicate rocks or their derivatives. The rocks consist of manifold mineral aggregates, formed at different times and in various ways. Each individualized mass of mineral aggregates—such as an intrusive mass, a lava flow, a stratum, a dike, a vein, a lenticular mass—is called a "formation," a "member," or in general a "geologic body." Geologic bodies which consist mainly of a single useful mineral—for instance beds of pure gypsum or coal—or which contain, throughout or in places, valuable minerals which can be profitably extracted—for instance veins containing disseminated gold—are called "mineral deposits." Geologic bodies that are not worked for any particular mineral or minerals, but for the aggregate of minerals—the rock itself—are usually designated as deposits of the particular rock. Thus a bed of roofing slate is not spoken of as a mineral deposit but as a slate deposit. Economic geology treats of the occurrence, composition, structure, and origin of those geologic bodies which can be technically utilized; it shows where they may be searched for and how their value may be ascertained.[1]

The mineral deposits which we know are all contained in the "crust" as defined above. Practically all of these were formed within 10 miles and the great majority within 5 miles from the surface. They were formed by many different processes, such as magmatic differentiation, sedimentation, weathering, and, generally, by the action of solutions of many kinds of the solid rocks of the crust.

The mineral deposits are thus local accumulations or concentrations of useful substances. The science of the mineral deposits tries to trace and explain these concentrations. In part they are caused by the preference of certain elements for certain rocks; for instance, tin, tungsten, and molybdenum for acidic rocks; platinum, chromium, nickel, copper

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[1] Stelzner and Bergeat, Die Erzlagerstätten, 1, 1904, p. 1.

for basic rocks. Useless, for instance, to look for tin deposits in basic rocks, or platinum in acidic rocks.

The circulation of the elements is a fascinating study. We trace, for instance, phosphorus from apatite in the igneous rocks to phosphate in organisms and from there to sedimentary phosphate beds. Some elements travel far and wide; others are averse to migration. Naturally, solubility is the main factor.

**Processes of Concentration.**[1] 1. *Differentiation by Fractional Crystallization and Unmixing in Cooling Magmas.*—Fractional crystallization is most important here. The heavy minerals generally separate out first. Thus there may result by settling of crystals, masses of magnetite, chromite, ilmenite, the chromite sometimes carrying with it diamonds and platinum; or the sulphides dissolved in the hot magmas may separate by liquid unmixing carrying with them Ni, Cu, Au, Ag, Pt, and Pd. The deposits are not abundant though often large and important.

Fortunately for humanity there is another process of differentiation in magmas which works in another direction. The salic products are carried upward and with them ascend gaseous products.

2. *Salic Extracts, Mostly Represented in Pegmatite Dikes.*—Alkali feldspar and quartz tend to separate in the upper part of the molten magma carrying with them tin, tungsten, beryllium, columbium, rare earth metals, radium, uranium, phosphorus, fluorine, boron, rarely sulphides, more commonly arsenides. The salic extracts are the first and last deposits for many metals.

3. *Gas-fluxing Components, Reaching Farther from the Magma than the Salic Extracts.*—In these products silica is again well represented carrying with it much water of magmatic origin and other volatile substances. These waters, at first acid, ascend carrying carbon dioxide, nitrogen, sulphur compounds, further fluorides, chlorides, phosphates, arsenic and antimony compounds, selenides, and tellurides. The metals are gold, silver, iron, copper, lead, zinc, bismuth, tin and tungsten, mercury, manganese, nickel, cobalt, radium, and uranium.

The gas-fluxing components appear as sublimates in volcanic eruptions near the surface and as veins and replacement deposits above and near igneous intrusions.

A large proportion of all metal deposits are formed in this way.

4. *Concentration by the Solvent Power of Underground Waters.*—Surface waters descend through the rocks, constantly both depositing and dissolving. They may take up iron on their downward path and deposit it

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[1] W. Lindgren, Concentration and circulation of the elements, *Econ. Geol.*, 18, 1923, pp. 419–442.

W. J. Vernadsky, Geochemie in ausgewählten Kapiteln, Leipzig, 1930.

A. Fersmann, Geochemische Migration der Elemente, *Abhandl. prakt. Geol.* (G. Berg), 18, 1929–1930, 116 pp.

again on their way upward. The effectiveness of the process depends on the solubility of the rocks. Deposits of iron, copper, lead, zinc, and even uranium and vanadium may be thus formed. This class does not as a rule contain gold, silver, arsenic, antimony, tin, and mercury. The lead zinc deposits of the Mississippi Valley are considered by many to have been formed in this way, also the copper-vanadium deposits of the western United States.

5. *Concentration by Surface Waters in the Zone of Oxidation.*—Close to the surface the waters contain much oxygen and therefore tend to decompose the rocks and to convert them to soil. Some difficultly soluble minerals, however, remain and thus important concentrations may result, for example, limonite deposits residual after limestone and serpentine, as in the southern states or in Cuba, or residual after lean ferruginous sediments as in the Lake Superior region. Extensive concentrations are effected in the upper part of other ore deposits to form enriched oxidized lead, zinc, and copper ores, and downward percolating solutions precipitate valuable deposits of secondary sulphides, particularly chalcocite and covellite. Phosphates, and manganese minerals may be concentrated. Colloidal processes are important. The zone of weathering makes vegetable and animal life possible.

Complex cycles of solution and deposition of calcium, iron, silica, nitrogen, sulphur, phosphorus, iodine, sodium, and potassium accompany this process.

6. *Concentration by Sedimentation.*—The processes of erosion and disintegration, followed by transportation, give rise to sedimentation. From the igneous rocks elements like sodium and calcium are leached while other elements are concentrated. For our life these processes are of supreme importance, for during this sorting process chemical and biochemical reactions take place which result in deposits of iron, manganese, phosphates, calcium and magnesium carbonates, gypsum, and alkaline chlorides. One hundred grams of igneous rock contribute 97 grams to the sediments, the remaining 3 grams going to enrich the salts of the sea. Gravity plays a strong part here producing placer deposits of heavy minerals, and beds of kaolin and quartz sand.

7. *Concentration by Biochemical Processes.*—Plants and animals from the largest to the most microscopic play an intense and dramatic part in the concentration of certain elements. We recall that coal is concentrated by plants from the carbon dioxide of the atmosphere, that petroleum is concentrated by the accumulated residues of animals and plants. In about six transformations the phosphorus originally contained in sea water comes to rest in deposits of guano, or phosphate beds. The calcium carbonate of algae, and the silica of diatoms may accumulate to valuable deposits. Iron and manganese bacteria may form bog ores: sulphur bacteria may accumulate sulphur. Copper, zinc, arsenic, iodine,

and vanadium are accumulated by many animals, and even barium and strontium are so segregated. Much of these concentrations may revert to the sea water from which they originated, but if they were suddenly buried by sediments they may well form a source for a second concentration resulting in an ore deposit.

## TECHNICAL CONSIDERATIONS

**Technical Utility.**—We designate as mineral deposits or ore deposits those geological bodies which can be technically utilized. This limitation must, of course, not be taken too literally, especially where questions of origin are concerned, for here, as in many other phases of the subject, applied geology merges into theoretic geology. Moreover, it is no uncommon occurrence that the useless of yesterday becomes the useful of to-day. About 1900 the cupriferous monzonite of Bingham, Utah, which yields an average of 30 cents in gold and 14 cents in silver to the ton and 1.0 per cent of copper, would probably not have been classed as an ore, but with modern methods of treatment it has become an important ore of copper. The zinc minerals of the western states, valueless and even causing loss in the marketing of ores, can now be profitably sold. The tungsten and vanadium ores thrown over the dump not long ago may now be very valuable. Low-grade gold ores—for instance, those of Mercur, Utah—considered as hopelessly refractory before 1890, became rich assets with the introduction of the cyanide process. Many iron ores rich in phosphorus were neglected until the Thomas process provided means for their profitable reduction. Monazite containing thorium acquired importance with the invention of the incandescent mantle for gas burners. New processes of reduction, the rising price of some commodity, inventions calling for rare and unused metals—any of these may suddenly cause a geologic body that has previously been valueless to become of great importance. Titanic iron ores form vast deposits, most of which are now useless because of metallurgical difficulties but which some day will be utilized. This principle also works the other way. Decreasing prices may make a particular deposit unprofitable; that is what happened to many silver mines during the great decline in the price of silver which began in 1880. Great changes, mainly in the direction of rising prices were brought about by the World War beginning in 1914. A large number of metals doubled in price: They include silver, platinum, copper, lead, zinc, tin, antimony, and aluminum. Gold alone, being the standard by which other values are measured, remained apparently stable.

Many radical changes, not easily predicted, have taken place since the war ended, culminating in record low prices in 1932. Continued improvement in technique decreases the cost of production. Differential flotation for instance has rendered profitable many ore-bodies which formerly could not be worked.

**Ore and Gangue.**—These considerations bring us to the terms ore and gangue. "Ore" is a word which has been used in several meanings. An "ore mineral" is a mineral which may be used for the extraction of one or more metals. An "ore," as the term is used here, is that part of a geologic body from which the metal or metals that it contains may be extracted profitably. Thus galena and malachite are ore minerals. An ore is practically always a mixture of minerals. Local usage has adopted several terms as substitutes for "ore." In the lead-zinc district of Missouri crude ore is called "dirt," while concentrates are called "ore." In Michigan the ore is called "rock" and the concentrates are termed "mineral." Gold-bearing gravels are not usually referred to as ore. The use of the term "ore" is not quite consistent. Ordinarily it implies a metal, but the expression "sulphur ore," meaning pyrite, is sometimes seen, and occasionally such a term as "sapphire ore" is found. The useless minerals occurring in the ore are termed "gangue." Thus, a gold ore may consist of quartz, calcite, siderite, native gold, auriferous pyrite, and galena. Here the first three are called "gangue minerals." The terms are not inflexible; for example, siderite may under some circumstances be utilized as an iron ore. Moreover, as stated above, what to-day is useless gangue may prove valuable ore to-morrow. It is therefore safe to make the definition of an ore rather wider than the present technical limits.[1]

It is hardly necessary to call attention to the differences in prices of metals which cause wide disparity in the amounts of different metals necessary to constitute ores. An iron ore must ordinarily contain at least 30 per cent of iron—usually much more. A volcanic rock containing 15 per cent of iron is far from being an iron ore, but quartz containing 0.05 per cent of gold is a rich gold ore, worth $330 a metric ton; in fact, as little as 0.0001 per cent of gold, equivalent to 1 gram to the metric ton, or a value of 66 cents a ton, if occurring in an ore with other useful substances, is ordinarily paid for by smelting works.

**Tenor of Ores.**—While it is not possible to give exact data as to the minimum values which ores of the different metals should have for profitable extraction, some approximate statements may be useful.[2] Local conditions, price of metals, the nature of the ores, and the association of the metals must of course be considered. The following notes refer to average conditions and prices.

**Iron.**—Iron ores from the Lake Superior region usually contain 50 to 60 per cent of iron; but iron ores which contain less than this may be

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[1] For a full discussion of the subject see J. F. Kemp, What is an ore?, *Jour.* Canadian Min. Inst., 12, 1910, pp. 356–367. Also, *Min. and Sci. Press*, March 20, 1909.

[2] J. F. Kemp, Problem of the metalliferous veins. *Econ. Geol.*, 1, 1905, pp. 207–232.

utilized, especially where other conditions are favorable. The Clinton ores of Alabama contain as little as 30 per cent of metallic iron; some types of easily concentrated magnetites may contain as low as 25 per cent and still yield a profit.

**Copper.**—Copper ores of the Lake Superior region may be treated with profit, under favorable circumstances, with as little as 0.5 per cent of metallic copper, though they ordinarily average somewhat higher. Sulphide copper ore of the usual type can rarely be utilized if it contains below 1.5 per cent of copper unless gold and silver are present also, and in many districts the ores must average considerably higher than this.

**Lead.**—In northern Idaho lead ores which contain 5 to 6 per cent of lead and 3 ounces in silver to the ton are profitably mined. Non-argentiferous ores which assay from 5 to 7 per cent of lead are utilized in southeastern Missouri.

**Zinc.**—Zinc ores vary considerably according to locality. At Joplin, Missouri, much of the crude material hoisted yields less than 3 per cent of zinc sulphide and a little lead. This is concentrated to about 60 per cent of zinc. In localities more remote from markets, as in Colorado, Utah, and Idaho, only zinc ores of higher grade can be profitably treated or shipped.

**Silver.**—With average metal prices pure silver ores can be mined if they contain not less than 15 ounces to the ton. The usual ores contain silver in association with lead, copper, or gold, or with all three. In complex ores smelters rarely pay for less than 2 ounces of silver and 0.01 ounce of gold to the ton. Gold and silver are separated from the lead or copper bullion by zinc desilverization or electrolytic refining and the cost of that process, of course, imposes the necessity of a certain minimum tenor of gold and silver for profitable extraction, but at many plants gold and silver, although present in less than these small quantities, are obtained as by-products through the necessity of eliminating some objectionable constituent, like arsenic, from the bullion.

**Gold.**—At the Alaska-Juneau mine gold is now profitably extracted on a large scale from ores yielding somewhat less than $1 per ton, but the ordinary gold quartz ores—for instance, those of California—yield about $5 to the ton; those of Nevada, Colorado, and some other states usually contain more. Gold ores containing from $2.50 to $3 a ton were worked at the Treadwell mines in Alaska. In gold gravels worked by the hydraulic process as little as 4 or 5 cents to the cubic yard may be profitable. By dredging, gravels containing 8 to 15 cents a cubic yard may be utilized in California; in Alaska they should contain from 50 cents to $1 a cubic yard. The costs of gold dredging have lately been brought down to about 4 cents a cubic yard.

**Tin, Etc.**—Tin ores range from 1.5 to 5 per cent in tin, but in tin-bearing gravels a much smaller tenor say 1 pound of cassiterite per cubic yard of gravel is sufficient to yield a profit. Ores of quicksilver contain at least 0.3 per cent of that metal; aluminum ores at least 30 per cent of aluminum. Nickel should be present to the amount of 2 per cent or more to constitute a workable nickel ore. Manganese ore should contain 50 per cent of that metal, but less is required if iron is also present. Chromium ore must contain about 40 per cent of chromic oxide. Poorer ores may be used if amenable to concentration.

## PRICE OF METALS

The prices which the various metals bring express the result of their abundance, of the demand for them, and of the cost of reduction of their ores. The value of gold is fixed by international agreement, hence it constitutes the standard by which the prices of all other commodities are measured. Aluminum, the most common of all metals, brings a high price because it can be produced from only a few of the minerals containing it.

In the following table the first column represents what may be called the "normal" prices for metals. Even in normal times there are, of course, constant fluctuations and some metals like copper, iron, and tin are especially susceptible to economic influences. The World War beginning in 1914 proved to have a potent influence on prices; the prices of most metals had been doubled in 1918. Heavy liquidation and deflation brought extremely low prices in 1931 and 1932.

In June, 1933, prices (in cents) were again moving upwards: Copper 8; Tin 41; Lead 4.1; Zinc 4.3; Silver 35.

COMPARATIVE VALUES OF METALS

| | March, 1914 | February, 1918 | May, 1927 | December, 1932 |
|---|---|---|---|---|
| Platinum..... | $44.00 per troy ounce | $106.00 per troy ounce | $86.00 per troy ounce | $30.00 per troy ounce |
| Gold......... | 20.67 " " " | 20.67 " " " | 20.67 " " " | 20.67 " " " |
| Silver........ | 0.57 " " " | 0.87 " " " | 0.56 " " " | 0.215 " " " |
| Quicksilver... | 0.52 per pound | 1.71 per pound | 1.66 per pound | 0.65 per pound |
| Nickel....... | 0.45 " " | 0.50 " " | 0.35 " " | 0.35 " " |
| Tin.......... | 0.37 " " | 0.85 " " | 0.65 " " | 0.23 " " |
| Aluminum.... | 0.19 " " | 0.37 " " | 0.26 " " | 0.23 " " |
| Copper....... | 0.14 " " | 0.23 " " | 0.128 " " | 0.048 " " |
| Antimony.... | 0.07 " " | 0.14 " " | 0.18 " " | 0.054 " " |
| Zinc......... | 0.05 " " | 0.08 " " | 0.061 " " | 0.031 " " |
| Lead......... | 0.04 " " | 0.07 " " | 0.067 " " | 0.030 " " |
| Pig iron...... | 0.006 " " | 0.013 " " | 0.008 " " | 0.0064" " |

Many interesting data on the total quantities of metals produced in the world and on the largest amounts mined in any one deposit are given by J. H. L. Vogt.[1]

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[1] Beyschlag, Krusch and Vogt, Die Lagerstätten, etc., 1, 1909, pp. 187–200.

## PRODUCTION OF ORE AND METAL

An interesting table reducing metal production in the United States to a uniform basis of short tons was formerly published annually in *Mineral Resources.* From this it is seen that in 1925, for instance, were produced 36,800,000 tons of pig iron, nearly 837,000 tons of copper, 655,000 tons of lead, 556,000 tons of zinc, 2,750 tons of silver, and 100 tons of gold. Iron in long tons; others in short tons.

The ore production of the same year was as follows: Iron ores, 64,000,000; copper ore, 50,000,000; lead ore, 8,600,000; zinc ore, 16,500,000; gold ore and silver ore, 9,000,000, all in round figures. Iron ores in long tons; others short tons.

## WEIGHTS AND MEASURES

Before leaving this part of the subject a few words on weights and measures may be added. The contents of base-metal ores, such as iron, lead, zinc, and copper, are measured by percentage. For lead and copper the figures given often do not mean the exact content by wet analysis, but by the dry assay, which is 1½ per cent or more lower than the exact content. In some cases the lead is determined by wet assay of the button from a crucible assay, which places the percentage obtained still farther below the actual content. The smelter pays for the metals by the "unit," which means 1 per cent, or 20 pounds to the ton, or else by a "basis price" for a given percentage of metal, say 55 per cent for bessemer iron ores or 65 per cent for Joplin zinc concentrates. Tungsten ore is sold per unit of tungsten trioxide for ore carrying 60 per cent or more of this compound. Deductions and allowances based on the presence or absence of certain elements and certain other rules complicate the smelter schedules.[1]

Precious metals in ores are measured in England and its colonies by troy fine ounces and pennyweights, per long or short ton. In the United States decimal fractions are substituted for pennyweights; gold is often reported in dollars and cents, $1 corresponding closely to 1 pennyweight. Silver is measured in fine ounces, the pennyweights always being omitted. The short ton is always used. Practically all other nations measure these metals in grams per metric ton, a far more sensible way. For comparison the following data, computed and arranged by W. J. Sharwood,[2] are given:

**Conversion Tables.**—The gram is taken as 15.4320 grains. The value of a troy ounce of fine gold is assumed as being exactly $20.67,

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[1] C. H. Fulton, The buying and selling of ores and metallurgical products, *Tech. Paper* 83, U. S. Bur. of Mines, 1915.

[2] Conversion tables for assay valuations, *Mines and Minerals,* January, 1909, p. 250.

instead of $20.6718346+, resulting in an error of less than 1 in 10,000. Values in English coin are based on the assumption that an ounce of fine gold is worth 4.25 pounds sterling, or 85 shillings, or 1,020 pence; this is too high by about 1 part in 2,000, the true value being 1,019.45 pence. It is useless to attempt a closer approximation in practical work, for the simple reason that gold bullion assays are rarely reported closer than the nearest half millieme, or to within 1 part in 2,000. At the values adopted one dollar is equivalent to 4.11224 shillings, and one pound sterling to $4.86353.

VOLUME AND WEIGHT OF FINE GOLD

| | One cubic centimeter | One cubic inch | One cubic foot |
|---|---|---|---|
| Fine gold: | | | |
| Weight: grams | 19.3 | 316.269 | 546,513 |
| Weight: troy ounces | .6205 | 10.1680 | 17,570.39 |
| Value: U. S. gold-standard dollar | $\$12.82^{57}$ | $210.17 | $363,180 |
| Value: gold-standard British pound | £2.647 | £43.214 | £74,674 |
| Value: U. S. dollars at $35 per ounce | $21.71 | $355.82 | $614,864 |

The gold standard (étalon d'or; Gold Münzfuss; padron de oro), which defines the currency units as equivalent to a certain approximately uniform weight of gold, was adopted long ago by many countries. In 1931, owing to various causes, Great Britain abandoned the gold standard and several nations followed this movement of depreciation. By presidential proclamation on Jan. 31, 1934 the value of the United States dollar was fixed at 59.06 per cent of the gold dollar, which meant that henceforth gold was valued at $35 per ounce instead of at $20.67+ per ounce. Therefore, figures given in the following tables should be multiplied by 1.693 to obtain the value in the new United States dollar. Approximately the same is true of the values given in pounds sterling (also in Canadian dollars); the exact value varies with the market price of gold, which thus far has been about $35. Opinions differ sharply as to the wisdom or necessity of this depreciation by which many guaranteed values were repudiated. Other changes may conceivably be made. Therefore, the two following tables have not been changed from the gold standard values.

### ASSAY VALUATIONS

| Values | One part in | Per cent | Per metric ton | | | Per long ton of 2,240 pounds | | | | | Per short ton of 2,000 pounds | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | Troy ounces | United States dollars (gold) | Grams | Troy ounces | Oz. | Dwt. | Gr. | United States dollars (gold) | Troy ounces | United States dollars (gold) |
| One per cent | 100 | 1 | 321.50 | $6,645.40^{6}$ | 10,000 | 326.666 | 326 | 13 | 8 | 6,752.20 | 291.666 | 6,028.75 |
| One gram per metric ton = one part per million. | 1,000,000 | 0.0001 | 0.03215 | $0.66^{45}$ | 1 | 0.03266 | | ...... | 15.68 | $0.67^{5}$ | 0.029166 | $0.60^{29}$ |
| One troy ounce per short ton | 29,166.66 | 0.00342857<br>$=\frac{24}{7000}$ | 1.1023 | $22.78^{45}$ | 34.2857 | 1.120 | 1 | 2 | 9.6 | 23.15 | 1 | 20.67 |
| One troy ounce per long ton | 32,666.66 | 0.0030612<br>$=\frac{3}{980}$ | 0.9842 | $20.34^{34}$ | 30.612 | 1 | 1 | ...... | ..... | 20.67 | 0.892857 | $18.45^{9}$ |
| One troy ounce per metric ton | 32,150 | 0.0031104<br>$=\frac{2}{643}$ | 1 | 20.67 | 31.104 | 1.016 | 1 | ...... | 7.7 | 21.00 | 0.90720 | $18.75^{2}$ |
| One dollar gold per short ton | 602,875 | 0.00016587 | .0533285 | 1.1023 | 1.6587 | 0.054185 | .... | 1 | 2 | 1.12 | 0.048379 | 1 |

## CONVERSION TABLES—WEIGHTS

|  | Grains | Penny-weights | Troy ounces | Avoirdupois ounces | Avoirdupois pounds | Grams | Fine gold value |  |
|---|---|---|---|---|---|---|---|---|
|  |  |  |  |  |  |  | United States | British |
| One grain | 1 | 0.041666 | 0.0020833 | 0.00228571 | 0.000142857 | 0.0648 | 4.306 cts. | 2.125 pence |
| One pennyweight | 24 | 1 | 0.0500 | 0.0548571 | 0.00342857 | 1.5552 | $1.0335 | 4.25 shillings |
| One troy ounce | 480 | 20 | 1 | 1.0971428 | 0.0685714 | 31.104 | $20.67 | 85 shillings |
| One troy pound | 5,760 | 240 | 12 | 13.165714 | 0.822857 | 373.248 | $248.04 | £51 |
| One avoirdupois ounce | 437.50 | 18.22917 | 0.911458 | 1 | 0.06250 | 28.35 | $18.84 | 77.474 shillings |
| One avoirdupois pound | 7,000 | 291.666 | 14.58333 | 16 | 1 | 453.60 | $301.4375 | £61.97 |
| One milligram | 0.015432 | 0.000643 | 0.00003215 | 0.000035274 | 0.0000022046 | 0.0010 | 0.06645 ct. | 0.033 penny |
| One gram | 15.432 | 0.643 | 0.03215 | 0.035274 | 0.0022046 | 1 | 66.45 cts. | 2.73275 shillings |
| One kilogram | 15,432 | 643 | 32.15 | 35.274 | 2.2046 | 1,000 | [1]$664.54 | £136.64 |

The price of silver in the United States is stated in cents per troy ounce 1000 fine; in England in pence per troy ounce of sterling silver 925 fine: 1 cent per fine ounce = 0.457 pence per ounce 925 fine; 1 penny per ounce 925 fine = 2.19 cents per fine ounce. With silver at 50 cents per ounce, 3 grams per ton equals about 5 cents United States currency.

Precious stones are measured by the carat One carat (diamond weight) is equal to 3.2 grains troy, or 206 milligrams. The "metric carat," now universally used, is equal to 200 milligrams.

[1] The United States Mint Bureau uses $664.60. The British values are calculated on the basis of the gold-standard pound. The United States values are calculated on the basis of the gold-standard dollar. At present the United States is off the gold standard and export of gold is prohibited except for certain purposes. Thus gold producers must sell their product to the mint at $35 per ounce, United States currency. In the above table the values in present United States currency are: 1 grain, 7.29 cents; 1 pennyweight, $1.75; 1 troy ounce, $35; 1 milligram, 0.1125 cents; 1 gram, $1.125; 1 kilogram, $1125.07.

# CHAPTER II

## THE FORMATION OF MINERALS

### SOLUTION AND PRECIPITATION

**General Features.**—A mineral is an inorganic[1] body of definite chemical composition. Solid minerals[2] are formed by changes in chemical energy in systems which contain one fluid or vapor phase. Their development, therefore, generally indicates a transition from a mobile to a less mobile form of matter. In the great majority of cases the minerals in nature are formed by precipitation from solutions,[3] and this, therefore, becomes a process of the highest importance for mineral genesis.

Supersaturation and consequently precipitation are controlled by the thermodynamic environment of the system. Temperature and pressure are the most important agencies though at times electric energy and light may also be active.

Minerals may form:

1. By supersaturation in a solution.
2. By reactions between liquids or liquid solutions.
3. By reactions between gases or gaseous solutions.
4. By reactions between liquids or liquid solutions, and gases.
5. By reactions between solids and liquids, or liquid solutions or gases.

Few minerals are formed below the freezing point of water. Their upper limit of development is marked by the temperature at which they become unstable or melt.

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[1] This does not mean that minerals cannot be formed by organic processes. If this were so, a large proportion of material in sedimentary rocks (calcite, phosphates, silica) could not be considered as minerals.

[2] The liquid minerals are practically confined to water, mercury and to certain hydrocarbons. The last, as well as certain undercooled liquids like obsidian, are not, however, considered as minerals as they have no definite composition expressible in a chemical formula. To a small extent minerals may form by "unmixing" or "exsolution," when in a cooling solid solution (*e.g.* argentite in galena) a lesser constituent is ejected at a certain temperature and crystallizes as inclusions. Here apparently no fluid phase is involved.

[3] Certain chemical reactions involving the formation of new minerals may apparently take place by the interaction of solids under the influence of heat and pressure, but it is possible that here too a vapor phase or a fluid phase is present in the system.

In minerals subject to polymorphism the arrangement of the molecules may change at certain critical temperatures. In many cases no new names are used for those products, *e.g.* $\alpha$ and $\beta$ quartz, isometric and rhombic chalcocite. In other cases new names are applied, *e.g.* sphalerite and wurtzite for isometric and rhombic ZnS.

After the inception of mineral formation concentration to larger masses or deposits may be effected by the continuation of similar processes or reactions. Quite frequently, however, there are other causes which have contributed: Gravity may cause the sinking of heavy minerals in lighter media as when anhydrite crystals sink to the bottom in sea water, or when magnetite sinks in a residual rock melt. Solution and disintegration may do its share, as when native gold is set free from a quartz matrix, and is separated by gravity into workable gold deposits; or, as when, in a rock composed of calcite and calcium phosphate, the former is dissolved, leaving a phosphate deposit of economic value.

**Influence of Pressure.**—The effect of change of pressure, according to LeChatelier's law, is as follows: When the pressure in a system in equilibrium is increased that reaction takes place which is accompanied by a diminution in volume; and when the pressure is diminished a reaction ensues which is accompanied by an increase in volume. "The direction in which change of concentration will occur with change of pressure can be predicted, if it is known whether solution is accompanied by increase or diminution of the total volume. If diminution of the total volume of the system occurs on solution, as is the usual case, increase of pressure will increase the solubility; in the reverse case increase of pressure will diminish the solubility."[1] In general, a decrease of pressure, which results when solutions ascend in the earth's crust, will be favorable to precipitation. The influence of pressure is, however, in most cases slight. For instance, the solubility of sodium chloride (in grams of salt in 1 gram of solution) at the pressure of one atmosphere is expressed by 0.264 and at 500 atmospheres by 0.270. In systems in which one or more of the components are volatile the effect of pressure may be very great; carbon dioxide, for instance, held in water by pressure, may increase the solubility of calcium carbonate owing to the formation of bicarbonate.

The effects of uniform, or hydrostatic, pressure are much less marked than are the effects of stress, or unequal pressure. Under conditions of stress a given pressure will lower the melting temperature far more rapidly than when the pressure is equal from all sides.[2]

**Influence of Temperature.**—In a solution of various salts in water or in a silicate melt changes in temperature are far more effective in producing precipitation than changes in pressure. Van't Hoff's law states: When the temperature of a system in equilibrium is raised that reaction takes place which is accompanied by absorption of heat; and, conversely, when the temperature is lowered that reaction occurs which

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[1] Alexander Findlay, The phase rule, 1923, p. 87.

[2] John Johnston and Paul Niggli, The general principles underlying metamorphic processes, *Jour. Geol.*, 21, 1913, pp. 481–516; 588–624.

John Johnston, Pressure as a factor in the formation of rocks and minerals, *idem.*, 23, 1915, pp. 730–747.

is accompanied by an evolution of heat.[1] In the great majority of cases increase of temperature promotes the solubility of salts, and decreasing temperature—say by the cooling of ascending thermal waters or of magmas—promotes precipitation. The common rule for salts—to which a number of exceptions may be noted—is that the solubility increases to temperatures of 75° C. or 150° C. beyond which a lessening of the quantity dissolved may often be noted. Breaks in the solubility curve usually indicate the limit of stability for a particular salt, less hydrated forms, for instance, coming in at higher temperatures.

In any hot, complex solution, occurring in nature, decreasing temperature will, in general, cause precipitation of some mineral; with continued cooling a series of other minerals may be precipitated, as the solubility limit of each is reached.

As no compounds are absolutely stable under the varying conditions obtaining in the crust, it follows that minerals once formed may subsequently be brought into solution, transported, redeposited, or indeed wholly decomposed so that their elements may enter into new combinations.

**Precipitation by Evaporation of the Solvent.**—The salts contained in a solution are naturally precipitated when evaporation at the surface so reduces the amount of the solvent that supersaturation ensues. The deposits of gypsum and salt in various formations are familiar results of this process. In some cases carbon dioxide or other gases may be the solvent; the precipitation of calcium carbonate follows, for instance, in springs at their point of issue when the carbon dioxide escapes which holds the salt in solution as a bicarbonate.

**Precipitation by Reaction between Solutions.**—Mingling of different solutions is one of the most common occurrences in nature, as when rivers discharge their waters into the sea or as when ascending hot waters meet surface waters of different composition, or when different solutions meet in the cells of organisms. Precipitation of chemical compounds results when any combination of the various ions in the solution can form to a sufficient extent to be insoluble in the liquid present. Solutions in nature are usually complex and the various reactions are more or less interfered with. In general, according to Nernst's law, the solubility of a given salt is reduced by the presence in the solution of another salt which has a common ion but is increased by the presence of another salt with no common ion. For instance, the solubility of lead chloride is decreased by the presence of the chloride of calcium or magnesium. The presence of alkaline carbonates decreases the solubility of $FeCO_3$; the solubility of NaCl is decreased by $CaCl_2$; while the solubility of $CaSO_4$ (gypsum) is increased in a NaCl solution. If calcite is treated with a saturated

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[1] Alexander Findlay, The phase rule, 1923, p. 46.

solution of $FeCO_3$, $ZnCO_3$, or $MgCO_3$, a part of the calcite will be dissolved while a corresponding part of the other carbonates is precipitated; the solubility of $CaCO_3$ in water is increased by $Na_2SO_4$ or NaCl but decreased by $MgCO_3$.

In mixed solutions precipitation is often delayed, as shown, for instance, by the slow precipitation of barite ($BaSO_4$) due to the presence of sodium and magnesium chlorides in certain mine waters consisting of salt brines. Nernst's law offers an explanation of these anomalies.

Slow precipitation in dilute solutions generally results in large crystals being formed, while rapid precipitation results in colloids or fine powders.

**Precipitation by Reactions between Aqueous Solutions and Solids.**—In nature, solutions act constantly upon solid minerals. At the surface all rocks and mineral deposits are moistened by rain water which also may descend to great depths in porous, fissured, or broken material. Rising hot waters soak into the rocks from the fractures up which they ascend. Minerals are attacked to a greater or lesser degree by these various kinds of waters; they are decomposed and partly or wholly go into solution. From these solutions new minerals are deposited in open spaces. This is a very common mode of mineral formation. But the changes also proceed in the solid rocks themselves and such processes by which new minerals may take the place of old ones are called *metasomatism* or *replacement.* The water penetrates the rocks in capillary openings. By the phenomenon known as adsorption, the film of liquid on the solid contains more than an average amount of material in solution and these films are likely to become supersaturated in advance of the remainder of the solution, so that chemical reactions will be facilitated. In this manner the mineralogical and structural character of rocks may be changed: Chlorite may replace augite, and sericite and quartz may replace feldspars. Metallic ores are often formed by replacement. Limestone may, for instance, be permeated by a solution of zinc sulphate with the result that the calcium carbonate is replaced by zinc carbonate with faithful preservation of the limestone structure, while calcium sulphate is carried off in solution. It is not even necessary that the replacing mineral should have an element in common with the older mineral. Pyrite or galena may replace feldspars or calcite grains. The replacing mineral may even develop as perfect crystals in the older mineral. The phenomena of replacement are of the utmost importance for the genesis of mineral deposits.

**Precipitation by Reactions between Gases or between Gases and Solutions.**—Gases may produce precipitation in solutions. Hydrogen sulphide in some mine waters precipitates cuprous sulphide from cupric sulphate. Less important is the action between gases: Native sulphur may be precipitated in volcanic regions by a mixture of hydrogen sulphide and sulphur dioxide.

**Crystalline Minerals.**—The minerals may be precipitated as crystalloids or as colloids. In mineral deposits formed in depth and at temperatures higher than those prevailing at the surface crystalloids are almost exclusively present, as they also are in igneous and metamorphic rocks. The molecules have been allowed to arrange themselves in the symmetry of one of the six crystal systems and the minerals form homogeneous grains or crystals. Crystalline minerals develop best by slow precipitation in solutions contained in open spaces; under such conditions free crystals may form in silicate melts or magmas and in aqueous solutions. Crusts of minerals may develop where crystals adhere to walls of water-filled fissures; this is a common feature in mineral veins. The first impulse to crystallization may be given by adsorption and supersaturation along the walls. Once started the larger crystals become further enlarged because smaller crystals dissolve more rapidly than large ones and the liquid remains supersaturated with reference to the larger growths.

When crystallization is progressing from a great number of points in the solution, a granular texture is developed by the mutual interference of the crystals. In mineral deposits the resulting textures are usually rather coarse; only very rarely do we find fine-grained aggregates of order of magnitude of the dense groundmass of igneous porphyritic rocks. In mineral deposits a considerable variety of textures result by replacement in solid rock; such textures are in general analogous to those found in metamorphic rocks.

**Colloids.**[1]—Many minerals are deposited as gradually hardening gelatinous or flocculent masses instead of assuming crystalline form. These are called colloidal minerals or colloids. The colloid state consists of electrically charged particles in a state of dispersion ranging from comparatively coarse suspensions down to almost molecular subdivisions. A colloid mixture is a heterogeneous, two-phase system in which one phase divided into small separate volumes is known as the disperse phase, and the other as the dispersion medium. In the cases here considered the disperse phase may be a liquid or a crystalloid; the dispersion medium a liquid, usually water. A disperse system of this kind is often apparently homogeneous and transparent, sometimes opalescent; it is generally called a *sol.* The disperse phase may separate from the mixture in a gelatinous or flocculent form, still retaining some of the medium. Sols are of two kinds:

1. *Suspensoids.*—In these sols the disperse phase is a solid. The viscosity is low; they do not gelatinize, but the addition of an electrolyte causes a flocculent precipitate, which can not readily be returned to the state of a sol. The purple or red gold sols and many other metal sols may serve as examples; also the sulphide sols, such as arsenic sulphide.

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[1] Emil Hatschek, An introduction to the physics and chemistry of colloids, London and Philadelphia, 1922. Especially recommended for brevity and clarity.

R. Zsigmondy, Colloidchemie, Leipzig, 1925, 246 pp.

The latter sol is obtained as a yellow liquid when hydrogen sulphide is added to a solution of arsenious oxide. A small quantity of a gelatinizing colloid prevents the coagulation of colloidal suspensions by electrolytes. Thus gelatine or a silica sol may keep silver chloride indefinitely in a state of colloidal suspension.

2. *Emulsoids.*—In these sols the disperse phase is believed to be a liquid. Upon cooling or evaporation a gelatinous mass—a gel—is obtained. The emulsoids have a high viscosity and are not always coagulated by electrolytes, though some electrolytes produce a sudden setting of the sol to a gelatinous mass, which can again be brought into solution. The sols of gelatine, albumen, silica, and cerium hydroxide may serve as examples of emulsoids. The hydroxide sols, for instance, those of tin, iron, chromium, and aluminum have some properties which seem to connect them with the emulsoid sols.

The sols do not readily pass through membranes (like parchment paper or bladder), probably because the disperse particles are larger than the pores in the membranes. Electrolytes, on the other hand, pass readily through such walls.

In the sols, surface and surface tension are two most important factors. The reactions between sols are quite different from those taking place between electrolytes. In nature, gels and colloid precipitates often harden to rounded, reniform masses referred to as "colloform." The solidified colloids have a tendency to acquire crystallinity and may become transformed into fibrous or cryptocrystalline aggregates. Such crystallized colloids may be called *metacolloids.*[1] The composition of colloid minerals is often variable; this is caused by the peculiar quality of colloids to adsorb foreign substances. Often they contain water.

Colloid minerals form in abundance in open space within the oxidized zone and near the surface, but they may also develop at higher temperatures to a degree formerly not suspected. In solid rocks colloid minerals may replace older minerals, first as gels, later crystallized to metacolloids.

Among the amorphous gel minerals common in nature may be mentioned opal, limonite, halloysite, and garnierite. Examples of crystallized colloids (metacolloids) are numerous. Pyrite, marcasite, wurtzite, malachite, smithsonite, psilomelane, and chrysocolla are often of colloid origin.[2]

The great importance of colloidal solutions for ore deposition lies in the fact that many difficultly soluble substances can easily be dispersed to form relatively highly concentrated sols.

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[1] E. T. Wherry, *Jour.* Washington Acad. Sci., 4, 1914, p. 112.

[2] W Lindgren, The colloidal chemistry of minerals and ore deposits, in "Theory and Application of Colloidal Behavior," R. H. Bogue, editor, New York, 1924.

H. C. Boydell, Role of colloidal solutions in the formation of mineral deposits, *Trans.*, Inst. Min. and Met (London), 34, 1925, pp. 145–337.

H. C. Boydell, Operative causes in ore deposition, *idem,* 37, 1928, pp. 50–177

W. Lindgren, Metasomatism, *Bull.* Geol. Soc. Am., 36, 1925, pp. 247–261.

# CHAPTER III

# ATMOSPHERE, HYDROSPHERE, AND UNDERGROUND WATERS

## INTRODUCTION

As minerals form mainly in the presence of a fluid phase it is pertinent to consider the various forms of liquids and gases involved. They are in brief: (1) *The atmosphere;* (2) *the hydrosphere,* including oceans, lakes, and rivers; (3) *the underground fluids.*

## THE ATMOSPHERE

The volumetric composition of the atmosphere is given in the following table. It contains no other elements except as dust or colloid dispersions. For the origin of mineral deposits it is highly important, however, as the water, carbon dioxide, and oxygen are fluids which promote the oxidation of rocks and the development of ore deposits in the zone of weathering. The fixation of nitrogen by organic life or electricity may be of geologic importance. Halogen salts are carried as dust in the atmosphere from salt plains and oceans or from volcanic eruptions, and form chlorides and bromides in the oxidized zones of ore deposits. A

COMPOSITION OF ATMOSPHERE AND HYDROSPHERE[1]
(In Per Cent)

| Volumetric composition of the atmosphere | | Average composition of the ocean | |
|---|---|---|---|
| Nitrogen | 78.03 | Oxygen | 85.79 |
| Oxygen | 20.99 | Hydrogen | 10.67 |
| Argon | 0.94 | Chlorine | 2.07 |
| Water vapor | Variable | Sodium | 1.14 |
| Carbon dioxide | 0.03 | Magnesium | 0.14 |
| Hydrogen | 0.01 | Calcium | 0.05 |
| Neon | 0.00123 | Potassium | 0.04 |
| Helium | 0.0004 | Sulphur | 0.09 |
| Krypton | 0.00005 | Bromine | 0.008 |
| Xenon | 0.000006 | Carbon* | 0.002 |
| | | * In carbonates and organic matter. | 100.000 |
| | | Average salinity | 3.5 |

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[1] F. W. Clarke and H. S. Washington, The composition of the earth's crust, *Prof. Paper* 127, U. S. Geol. Survey, 1924, p. 33.

little sulphur dioxide or hydrogen sulphide can be introduced by volcanic eruptions or by industrial processes.

## THE HYDROSPHERE

**Ocean.**—In the above table the average elemental composition of the ocean is given according to Clarke and Washington in the quoted publication. The conventional composition of the salts dissolved in sea water is according to the same authority.

COMPOSITION OF THE OCEANIC SALTS

| | |
|---|---|
| $NaCl$ | 77.76 |
| $MgCl_2$ | 10.88 |
| $MgSO_4$ | 4.74 |
| $CaSO_4$ | 3 60 |
| $K_2SO_4$ | 2.46 |
| $MgBr_2$ | 0.22 |
| $CaCO_3$ | 0.34 |
| | 100.00 |

Substances present in small amount in the sea water may be tabulated as follows, mainly from data in Roth's "Allgemeine und Chemische Geologie," vol. 1; in Krummel's "Handbuch der Oceanographie"; and in Clarke's "Geochemistry," where the originals are cited. All figures in milligrams per liter.

| | | | |
|---|---|---|---|
| Aluminum, maximum | few milligrams | Iron, present[2] | few milligrams |
| Ammonium ($NH_3$) | 0.13–1.34 | Lead, present[3] | |
| Arsenic | 0.01–0.08 | Lithium, present | |
| Boron, present, small | | Manganese, present | |
| Barium, present | | Nickel and cobalt[4] | |
| Cesium, present | | Phosphorus, less than 1.0 | |
| Copper,[1] present | | Rubidium | 15.0 |
| Fluorine | 0.3–0.8 | Silica, present, 1.0 in average | |
| Gold, present (see p. 7) | | Silver, present (see p. 7) | |
| Iodine | 0.05–2.38 | Strontium, present | |
| | | Vanadium[5] | |
| | | Zinc | 0.002 |

[1] In molluscs and seaweed; Dieulafait says also in sea water.

[2] Abundant in ashes of organisms.

[3] In ashes of seaweed.

[4] In ashes of seaweed.

[5] In holothurians.

The sea water also contains about 45 milligrams of $CO_2$ per liter and amounts of oxygen, nitrogen, and other inert gases which differ according to the temperature. W. Dittmar[1] found from 14 to 28 cubic centimeters of air per liter of sea water containing, respectively, 35.01 and 33.11 per cent oxygen. The dissolved air is thus much richer in oxygen than the

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[1] *Challenger Repts.*, 1, 1884, p. 203.

air. The amount of free $CO_2$ and $HCO_3$ is variable but there is, in fact, about 27 times more $CO_2$ in sea water than in the air. An unstable equilibrium exists between the free $CO_2$ and that combined as $HCO_3$ in the bicarbonates. The ocean may be considered as the great regulator of the carbon dioxide in the air.

In closed or poorly ventilated basins the sea water may contain hydrogen sulphide; the Black Sea, for instance, contains in its deepest layers 600 or 700 cubic centimeters per liter of this gas. Gaseous hydrocarbons may be locally formed in the presence of much organic matter close to the shores.

In the littoral and neritic (shallow water) environments of the oceans many mineral deposits may form such as mineral salts, iron ores, manganese ores, phosphates, clay, limestones, mainly by manifold chemical and biochemical reactions between sea water and organisms or between sea water and rocks or between sea water and river water, or finally by evaporation.

The relation between the constituents in ocean waters is always as follows: $Cl > SO_4 > CO_3$; $Na > Mg > Ca$.

**Lakes and Rivers.**[1]—The water of rivers and drained lakes is often very pure, as it is derived directly from the rain which falls on the land. Particularly does this apply to the waters of glaciated areas of crystalline rocks. The salinity of such waters may run as low as 14 parts per million. Almost always the relation prevails that $CO_3 > SO_4 > Cl$ and $Ca > Na > Mg$. Silica is variable but often high—as much as 20 per cent of the dissolved material. Iron as $Fe_2O_3$ and alumina as $Al_2O_3$ are each frequently above 1 per cent of the total solids.

In waters of the arid regions and in sedimentary rocks the salinity increases rapidly, often to several hundred parts per million. The $SO_4$ radical also increases, also Cl, in places so as to change the above relation to $SO_4 > CO_3 > Cl$.

In tropical regions of heavy vegetation the organic matter carried may increase to 50 per cent of the total residue. In such cases the silica may also greatly increase and this increase is usually accompanied by considerable amounts of $Fe_2O_3$ and $Al_2O_3$, both probably present as colloids. Some of the rivers of the Amazon region carry up to 13 per cent of $Al_2O_3 + Fe_2O_3$ in terms of the total solids. Manganese is probably universally present in river waters.

Primarily the composition of the river waters depends on the geological formations traversed. They are, therefore, in unstable equilibrium, changing continually by chemical reactions of various kinds. Human contamination by sewage, industrial plants, etc. is apparent in many places.

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[1] F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, Chap. 3, 1924.

Some mineral deposits are formed in lakes and rivers. Among them are clays, iron ores, manganese ores, infusorial earth, and saline residues, but in importance they hardly compare with those originating along the shores of the ocean.

## THE UNDERGROUND FLUIDS

**Fluids of Meteoric Origin.**—The most abundant of these is water of atmospheric (meteoric) origin derived from the rain that falls on the surface, or from water courses, or from lakes, or from the present oceans. It descends into the earth in cavities, fissures, or capillary openings. In part these waters remain stored in the rocks. In part they ascend under hydrostatic conditions and rejoin hydrosphere and atmosphere.[1] The gases, in part dissolved in water, include oxygen, nitrogen, argon, helium, carbon dioxide, and methane. In part they are generated by underground reactions.

**Petroleum.**—Liquid hydrocarbons of varying composition are found in sedimentary rocks, often in large amounts. They are derived by slow distillation of organic material deposited with the sediments and are usually accompanied by gas of variable composition and by salt water of meteoric origin.

**Magmatic or Juvenile Fluids.**—The water of magmatic origin existed in the molten silicate solution which constitutes a magma. It is the most abundant of the volatile constituents and many of the latter are entrained with it or dissolved in it when it is liberated by crystallization or by the irruption of the magma into higher levels of the crust. The magmatic gases include nitrogen, argon, carbon dioxide, hydrogen sulphide, and others. They are in part dissolved in water, in part free. In part they join the atmosphere and the hydrosphere. The dissolved salts are silica, chlorides, sulphates, borates, etc. The quantity of new magmatic water is far smaller than that of meteoric origin.

**Imprisoned Fluids.**—Smaller parts of meteoric or magmatic solutions or gases may permanently or temporarily be withdrawn by being held firmly by capillarity, by forming inclusions in minerals or by entering chemical compounds. Heat, pressure, and chemical action may release part of these imprisoned fluids when rocks sink into warmer zones or are engulfed by rising magmas. While no one may doubt that the magma

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[1] R. A. Daly, Genetic classification of underground volatile agents, *Econ. Geol.*, 12, 1917, pp. 487–504. Daly shows that Posepny's term "vadose" (vadus = shallow) was applied by him to the descending waters above the water level, that is to the *zone of gathering* (p. 34). Authors have used it since with different meanings, in each case including a certain part of the atmospheric waters. It seems as if the science could dispense with the word vadose. Daly following Archibald Geikie has suggested "epigene" to cover the underground activities of both fresh and marine waters. The term phreatic, applied by Daubrée to a somewhat indefinite part of the meteoric waters, may likewise be dispensed with.

contains primary water a certain small part of it may be derived by the melting rocks immersed in magmas.[1] Connate water[2] is sea water of ancient oceans which for geologic periods has been included in the sediments of these seas. There are no physical or chemical criteria by which the origin of a certain water can be determined beyond question. All of these waters are really solutions which range from very dilute to highly concentrated. The rain water is, of course, the purest, followed in order by those of rivers, springs, and oceans. If it should be possible to distinguish between surface waters and magmatic waters, the evidence must be circumstantial and depend on geologic structure, igneous history, rainfall, and drainage basins. Magmatic solutions are naturally likely to be diluted along their upward path by waters of meteoric origin.

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[1] R. A. Daly, *Am. Jour. Sci.*, 4th ser., 26, 1908, p. 48. Igneous rocks and their origin, New York, 1914, p. 249. Daly terms such waters *re-surgent.*

[2] A. C. Lane, *Bull.* Geol. Soc. Am., 19, 1908, p. 502.

## CHAPTER IV

# THE FLOW OF UNDERGROUND WATERS

**General Statement.**[1]—Much of the rain water which descends upon the land runs off in surface drainage, and a smaller part evaporates, but a certain quantity sinks down into the soil and into the porous and fractured rocks. This part of the precipitation adds to the *ground water* and, if there is a sufficiency of rainfall, it saturates the material at a certain varying depth below the surface. This upper limit of the saturated zone indicated by the depth at which water stands in wells or shafts is called the water level, ground-water level, water table, or hydrostatic level. It is, as a matter of fact, a warped surface which feebly reflects the topographic features (Fig. 2). The water may penetrate to considerable depths, particularly along fissures; gravity and heat often establish a circulation of the ground water so that it may again reach the surface after a long underground journey. The ground water performs important geologic work by solution and precipitation which may result in concentrations of useful minerals. The branch of geology which deals with the underground waters is called *Hydrology.*

There is a small class of hot waters, ascending from great depths, which is believed by many geologists to be of magmatic origin, that is, given off by cooling magmas. However important such waters may be in the formation of certain kinds of mineral deposits, they are insignificant in

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[1] C. R. Van Hise, A treatise on metamorphism, *Mon.* 47, U. S. Geol. Survey, 1894, pp. 123–191 and 657–670.

C. S. Slichter, Theoretical investigation of the motion of ground waters, *Nineteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1898, pp. 301–304. Also *Water-supply Paper* 67, U. S. Geol. Survey, 1902, 106 pp. L. C. Graton and H. J. Fraser, Porosity and permeability, *Jour. Geol.*, 1935.

F. H. King, Principles and conditions of the movements of ground-water, *Nineteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1898, pp. 59–294.

J. F. Kemp, Igneous rocks and circulating waters, etc., *Trans.*, Am. Inst. Min. Eng., 33, 1903, pp. 707–711.

J. W. Finch, The circulation of underground aqueous solutions and the deposition of lode ores, *Proc.*, Colo. Sci. Soc., 7, 1904, pp. 193–252.

M. L. Fuller, Total amount of free water in the earth's crust, *Water-supply Paper* 160, U. S. Geol. Survey, 1906, pp. 59–72.

K. Keilhack, Lehrbuch der Grundwasser- und Quellenkunde, Berlin, 1935, 575 pp.

O. E. Meinzer, The occurrence of groundwater in the United States, *Water-supply Paper* 489, U. S. Geol. Survey, 1923, 321 pp. Also *Water-supply Paper* 494, 1923.

Ed. Imbeaux, Essai d'hydrogéologie, Paris, 1930, 704 pp.

quantity compared with the great mass of water of atmospheric origin which is contained in the rocks.

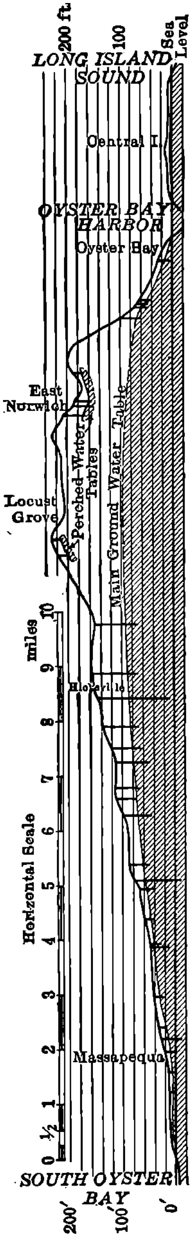

FIG. 2.—Section across Long Island, showing actual position of ground-water table. Upper part of section shows zone of gathering with local ground-water tables due to presence of impermeable strata. Zone of discharge between ground-water table and sea level. Static zone below sea level. (*From Prof. Paper 44, U. S. Geol. Survey.*)

**Pores and Openings in Rocks.**—All rocks contain interstitial openings and are capable of absorbing water. By porosity is understood the percentage of pore space referred to the total volume of the rock. The ratio of absorption is the ratio between the weight of the water absorbed and the weight of the rock tested. When the pores are completely filled the rock is said to be saturated, but a saturated rock after being drained always retains a certain amount of water which adheres to the walls of the pores. The pore space in rocks varies from a fraction of 1 per cent to 40 per cent. In fresh granites and similar compact rocks the porosity is from 0.2 to 0.5, in limestones from 0.53 to 13.36, in sandstones from 5 to 28, and in loose sands and gravels from 26 to 65 per cent.

Under the assumption that a sandstone consists of spherical grains packed in the most compact arrangement possible, the space between the spheres would amount to 25.95 per cent. The porosity is independent of the size of the grains.

The absorbed water may be called "free water" in contrast to that existing in chemical combination in the minerals of the rock. All of the free water is not "available," for instance in wells, because some rocks, like clays, have the peculiarity of holding in their pores great quantities of water which is released only at an extremely slow rate.

On the basis of size, openings in rocks may be divided into (1) openings which are larger than those of capillary size, or super-capillary openings; (2) capillary openings; and (3) subcapillary openings. For water, openings larger than capillary openings may be considered as circular tubes which exceed 0.508 millimeter in diameter, or sheet openings, such as those furnished by faults, joints, etc., whose width exceeds 0.254 millimeter. Capillary tubes or sheet spaces are those smaller than the dimensions indicated but larger than the openings in which the molecular attraction of the solid material extends across the space, and to such openings the laws of capillary flow apply. In sub-capillary openings the attraction of the molecules extends from wall to wall, and this class includes tubes smaller than 0.0002 millimeter in diameter and sheet openings smaller than 0.0001 millimeter in width. According to G. Bakker[1] the spheres of molecular attraction are only six to seven times the molecular diameter and consequently, if this be true, capillary movement can take place in tubes very much smaller than 0.0002 millimeter.

The flowage of water through super-capillary openings nearly follows the ordinary laws of hydrostatics, but is subject to a certain retardation on account of friction. The super-capillary openings include the greater number of faults, joints, partings, and the openings in coarser sediments. In capillary openings the movement is very slow indeed, so that many rocks in which they occur, as shales and clays, are spoken of as impermeable. In sub-capillary openings the water is held firmly as a film glued to the walls by adhesion; there is no free water and the flow is practically nil.

The capillary forces play a considerable part in the movement of underground waters; but, in general, they cannot produce a continuous flow, and they are of secondary importance in comparison with the hydrostatic pressure. At higher temperatures the capillary action decreases and becomes zero at the critical points.

The height to which water can rise by capillarity in porous soils is considered to be from 3 to 8 feet, the maximum being seldom reached.

Openings in rocks do not persist indefinitely in depth, though hard rocks like granite will sustain very heavy loads. The experiments of F. D. Adams[2] indicate that small cavities may exist in granite to a depth of at least 11 miles, or 17,600 meters. Most rocks tend to become plastic at far lesser depths and will then become deformed and flow without fracture. Under such conditions an active circulation of water is difficult. Super-capillary openings will be closed.

Later Bridgman[3] arrived at similar conclusions on basis of extensive experiments. He thinks it extremely probable that minute crevices, at least large enough for the percolation of liquids, exist in the stronger rocks at depths corresponding to pressures of 6,000 to 7,000 kilograms per square centimeter (a depth of about 20+ kilometers).

**Water in Sands and Gravels.**—We may first consider the simpler case of loose material such as sands and gravels, which are so abundant in the uppermost part of the crust, in comparison with which the underlying compact rocks may be regarded as impermeable. Under the

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[1] *Zeitschr. f. phys. Chem.*, 80, 1912, p. 129. See also J. Johnston and L. H. Adams, *Jour. Geol.*, 22, 1914, p. 13.

[2] *Jour. Geol.*, 20, 1912, pp. 97–118.

[3] P. W. Bridgman, *Am. Jour. Sci.*, 4th ser., 45, 1918, pp. 243–280.

influence of gravity the water descends until a depth is reached where the material is saturated, this being the water level. In the valleys the water level will lie close to the ground while it rises slightly under the ridges. The ground water is not stationary but moves slowly from the higher ground toward the valleys; and underneath the valleys a perceptible movement continues down stream until ultimately, with lack of grade, for instance where a river valley opens toward the sea, the movement will become slow and almost imperceptible. That such a movement actually takes place has been proved by the use of fluorescein and other indicators in wells and bore holes.

**Water in Rocks of Uniform Texture.**—In uniform rocks, like granite, water is not only contained in the pores but moves more easily on the ever-present seams and joints. Considerable water may be stored, though the quantity per unit volume of rock will be much smaller than in sands and gravels. The water table is here also a warped surface which follows approximately the topographic relief but is less accentuated. The elevation of the water table or water level may fluctuate considerably, dependent upon seasonal rainfall. A permanently moist fringe may extend for several feet above the water level. This rise is caused by the capillarity of the rock.

In his paper, previously cited, J. W. Finch distinguishes the space above the water level as the *gathering zone* or *zone of percolation*, in which water and air are both present, and in which the water is conducted to the *saturated belt.* Even in very arid regions there is usually a zone of saturation. It would also be possible, however, to have no belt of saturation and the water would then simply percolate feebly downward until the quantity is diminished to zero.

The *zone of discharge* "embraces that part of the belt of saturation which has a means of horizontal escape" by continuous gravitative flow. The movement of the water is usually more rapid in the upper part of this zone than in the lower part, where, in spite of greater pressure, the obstructions and the increasing compactness of the rock retard the flow. The air or gas filling the pore space must also be driven out before the water can enter, and the evaporation of water in the underground atmosphere may, under some conditions, also become an important factor in reducing the supply of water. Still another portion of the water enters into chemical combination by hydration.

The *static zone* is the third and deepest of the divisions proposed by Finch. It extends below the level of the lowest point of discharge and the water in it is stagnant or moves very slowly. It depends upon the zone of discharge for its water, as it is simply the bottom part, with a gradually diminishing quantity of water, of a belt of saturation of which the zone of discharge is the upper and flowing part. The lower limit of the third zone, where the quantity of water becomes exceedingly small, is not entirely a matter of speculation, for definite data are supplied by mining operations and in many places it is not more than 1,500 feet below the surface. Large quantities of water may be stored in the third zone, as, for instance, in the deep artesian basins where impermeable beds prevent escape. A certain amount of "rock moisture" undoubtedly persists to great depths.

**Water in Sedimentary Rocks.**—In a series of sedimentary beds it is common to find impermeable rocks, like clay and shale, alternating with more porous beds, like sandstone and limestone. Under such conditions the distribution of the descending surface waters is likely to be irregular. Near the surface there may be a local water level but below this beds heavily charged with water may alternate with almost dry strata. Each bed may, in a way, be considered as a unit and if it outcrops it has its own zone of gathering, zone of discharge and its static zone. The Cretaceous Dakota sandstone presents an excellent example of a porous stratum in which a large amount of water can be stored. Throughout the Great Plains this is a veritable reservoir of water, which can be tapped by artesian wells as far as 300 miles from its outcrop and at depths of a few hundred to 3,000 feet (Fig. 3). But this stratum at present simply contains a stagnant body of water; and, as in most other artesian basins, the quantity is not inexhaustible. This very case proves how impervious the adjacent sedimentary beds are, for neither upward nor downward is an avenue of escape afforded in spite of the strong pressure. Should profound fissuring take place in the Great Plains a natural avenue of escape would, of course, be opened and a deep circulation established. Kemp and Fuller have both brought out the fact that the deep sedimentary beds are often remarkably dry. The well 4,262 feet deep at Wheeling, West Virginia, was in absolutely dry rocks for the lower 1,500 feet. Wells sunk at Northampton, Massachusetts, and at New Haven, Connecticut, to depths of 3,800 feet have failed to obtain water.

The Goff well, and the Lake well, both in West Virginia, and the Geary well, Pennsylvania, were

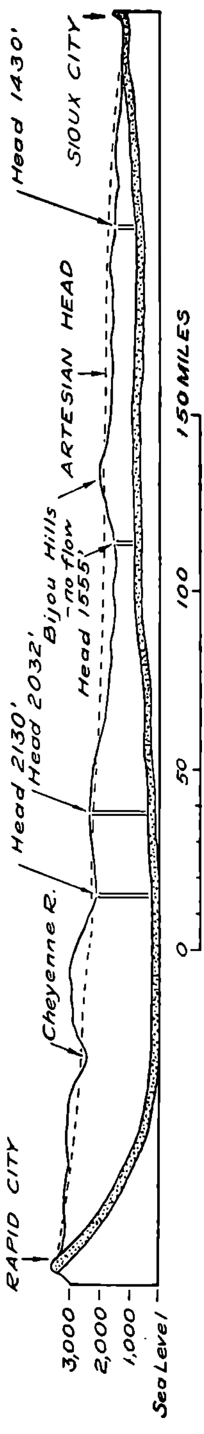

Fig. 3.—Profile showing the gradient in head of artesian water of the Dakota sandstone from intake to outflow. Vertical scale exaggerated. (*After N. H. Darton.*)

drilled to depths of 7,386, 7,579, and 7,248 feet. All three were in sedimentary rocks. In the Goff well no water entered below 2,307 feet, in the Lake well none below 2,118 feet. In the Geary well some beds below 6,045 feet yielded large amounts of salty water under pressure. It is possible, however, that the deep strata in the first two wells, though generally non-porous, may have held more or less rock moisture.[1]

In limestone regions much water is often stored in solution cavities which follow bedding and jointing. Larger open caves are found above the water level and sometimes contain underground water courses.

**Influence of Fractures.**—The simple conditions outlined above are seriously disturbed where extensive fracturing has taken place and paths have been laid out on which the water may move under approximately normal hydrostatic conditions. There may be a comparatively slow descent of the water along devious joints and fractures and a rapid rise under hydrostatic head where the descending water reaches the open paths of important faults and fissures. Friction during the descent undoubtedly seriously diminishes the theoretical head; but the evidence is perfectly clear that in regions of dynamic disturbance, such as the Alps and the Rocky Mountains, strong ascending springs may result from these conditions.

At the point of issue such springs may be warm and their temperature, in regions where no recent igneous action has taken place, may be a good indication of the depth attained by the water. Such springs seldom have a temperature higher than 65° C., and the composition of their salts corresponds to the character of the beds traversed. On the supposition, believed to be well founded, that only a moderate loss in heat takes place during the ascent, a water of the temperature named would be derived from a depth of about 5,500 feet. Large regions of the earth, such as the Scandinavian Peninsula, contain no warm springs; and the eastern part of the American Continent yields very few of them. Fuller says:

> The results of drilling in sedimentary and crystalline rocks, as well as the studies of deep mines, show that in all probability water does not commonly exist in the rocks under great pressure, although such may be exerted in an occasional crevice. It is not believed that hydrostatic waters exist, except possibly rare instances, at depths of over 10,000 feet, and that in reality the estimate of a depth of 6,520 meters, or 20,000 feet, as the limit of the zone of open cavities is closely approximate to the truth . . . If waters were freely circulating at great depths, within the zone of fracture, hot springs would certainly be more common along the numerous faults of the Piedmont, Appalachian, and similar regions.[2]

Van Hise suggests that the decreased density and viscosity of water at higher temperatures may lessen the head necessary for ascending

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[1] O. E. Meinzer, *Water-supply Paper* 489, U. S. Geol. Survey, 1923, p. 44.
E. W. Shaw, *Econ. Geol.*, 12, 1917, p. 610.

[2] *Water-supply Paper* 160, U. S. Geol. Survey, 1906, p. 64.

springs, but it may be doubted whether these factors would ever offset the great friction encountered during the downward passage. Faulting and mountain-building processes develop heat and this disturbance of the conditions of temperature may result in an increased circulation of the water stored in the rocks.

**Influence of Volcanism.**—When magmas are intruded into the zone of fracture the conditions become more complicated. It is thought by some that atmospheric waters are able to descend into the deep regions and become absorbed by the magmas, but this view appears improbable. Before its irruption into the zone of fracture the magma is assuredly far beyond the reach of any waters percolating from the surface. Daubrée's well-known experiment has often been cited, as showing how water may pass through a disc of sandstone against a certain counter pressure of steam. Recent critical examination[1] has shown the fallacies involved in the experiment, and indicates that "the probabilities are all against the notion that appreciable amounts of meteoric waters can ever penetrate into deep-seated and highly heated rock masses." Nevertheless, it is probable that in certain volcanoes where surface waters are abundant, the latter may reach the hot lavas and increase their activity and temperature.[2]

The presence of a heated body in the zone of fracture would undoubtedly quicken the circulation of water by inducing strong convection currents and expelling the stored water from its reservoir. Whether this action is sufficient to account for the remarkable number and volume of hot springs rising in volcanic regions may well be doubted; and it is thought that the magma itself gives up most of its constitutional water, partly when moving up to higher levels, partly when crystallizing to solid rocks.

**Conclusions.**—In conclusion it is believed that water in quantities sufficient to supply an ascending circulation can only exceptionally attain a depth of 10,000 feet and that, except in regions of permeable sedimentary rocks, the active circulation is confined to the uppermost few thousand feet. More commonly the depth of active circulation is measured by the level of surface discharge and the water below that level is practically stagnant; the lower limit of the body of stagnant water then forms an irregular surface descending to greater depths along the fractures and rising higher in the intervening blocks of solid ground.

**Examples of Movement of Water.**—According to Fuller, water supplies in wells in crystalline rocks are usually found within 200 or 300 feet of the surface and it is ordinarily useless to go below a depth of 500 feet

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[1] John Johnston and L. D. Adams, Observations on the Daubrée experiment, etc., *Jour. Geol.*, 22, 1914, pp. 1–15.

[2] A. L. Day and E. T. Allen, The volcanic activity and hot springs of Lassen Peak, *Publ.* 360, Carnegie Inst., Washington, 1925, 190 pp.

(Fig. 4). The occurrence of porous strata which are capable of holding immense quantities of water but in which none whatever is actually found is, according to Fuller, a common experience of drillers in this country, even where the upper strata contain a well-defined water table.

Investigations of joints in the crystalline rocks of Connecticut have shown, according to Fuller, that the water occurs largely in the vertical joints, which have an average spacing of 3 to 7 feet at the surface. In depth these joints diminish rapidly or close up and it is therefore not advisable to go below 250 feet in search of water.

The evidence from many mining regions is of considerable importance. In the Sierra Nevada of California deep canyons are separated by broad-backed ridges capped with Tertiary gravels and andesitic tuffs. For the Sierra as a whole the Great Valley of California forms the ultimate level of discharge. In the whole western part of the range there are no thermal springs and very few strong ascending springs, in spite of the prevalent fissility and jointing in the rocks. Hot springs are encountered only along the eastern slope of the range, a region which in the late Tertiary and Quaternary was the scene of great dislocations and volcanic activity. In the gold-quartz veins contained in the old rocks of the western slope much water is found in fissures to a depth of about 800 or 1,000 feet. Below this little water is met and many stopes and drifts are entirely dry; and this applies both to mines high up on the slope, as at Nevada City and Grass Valley, and to the Mother Lode mines of the foot-hill region.

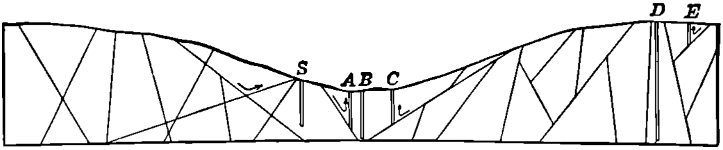

Fig. 4.—Section illustrating flow of water in jointed crystalline rocks. *A*, *C*, flowing wells fed by joints; *B*, intermediate well of greater depth, with no water; *D*, deep well not encountering joints; *E*, pump well adjacent to *D*, obtaining water at shallow depth; *S*, dry hole adjacent to spring. (*After M. L. Fuller, U. S. Geol. Survey.*)

Cripple Creek, Colorado, is another interesting example. Here we have a granitic plateau at an elevation of 9,000 feet above the sea; this plateau contains a volcanic plug about 2 miles in diameter which is largely filled with porous breccias and tuffs. The water fills the volcanic rocks as in a sponge inserted in a cup and the mining operations to a depth of 3,000 feet have tapped heavy flows. But even in this water-logged mass there are solid intrusive bodies, for instance at the Vindicator mine, at a depth of 1,000 feet, which are so dry that water must be sent down for drilling. The data thus far available have led Ransome to the conclusion that even at Cripple Creek the water is slowly diminishing in quantity at increasing depth. The granite which surrounds this water-soaked plug contains very little water and at most places is practically dry, in spite of the great hydrostatic pressure. The ultimate level of possible discharge would be in the valley of the Arkansas, 2,500 feet lower and many miles distant, but it may be gravely doubted whether any water from the Cripple Creek mines ever finds its way through the granite mass to this level.[1]

In the copper mines of Butte, Montana, where the granitic rocks are greatly faulted, much water was encountered, extending in places down to the bottom of the mines. A high range adjoins the mines on the east and conditions seem to be favorable for deep circulation yet the water is probably almost stagnant, and Weed mentions the existence of large bodies of dry rock.[2] Kemp states that the total water pumped in Butte from an area of two square miles to a depth of 2,200 to 2,800 feet is only 3,500 gallons per minute.

At the United Verde Mine, Arizona, only about 50 gallons per minute are pumped from the 3,000-foot level. In the Morning Mine, Coeur d'Alene district, Idaho, only 73 gallons per minute are pumped from the deepest level, 2,200 feet below the main adit.

Leadville, Colorado, is another place where the faulting is extensive and of comparatively recent date. At 1,500 feet, the greatest depth attained, there is still much water, mainly along the faults.

At Rossland, British Columbia, according to Bernard McDonald,[3] the water level is at 40 feet and the quantity increases to a depth of 200 to 350 feet. Below 350 feet a decrease begins, slowly at first but soon more rapid, until at 900 feet there is only a slight seepage and below 1,000 feet the mine is dry. Weed[4] states that in the copper mine at Ely, Vermont, an inclined shaft was carried down for a length of 3,600 feet, attaining a vertical depth of 1,700 feet. There is no water here below a vertical depth of 600 feet. At Przibram, Bohemia, the workings are dry and dusty at a depth of 3,000 feet. In Cornwall, and in Newfoundland mines have been worked underneath the sea, and sometimes close to the sea bottom, without irruptions of salt water.[5] Where fissures exist such irruptions are possible, however. The case of the Alaska-Treadwell mine is well known. In 1917, this property worked to a depth of 2,400 feet below sea level was suddenly flooded by salt water and had to be abandoned.

One of the most convincing examples is that furnished by the deep copper mines of Michigan and fully set forth by A. C. Lane.[6] He shows that

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[1] Lindgren and Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906, pp. 233–251. J. W. Finch, *Proc.*, Colo. Sci. Soc., 7, 1904, p. 204.

[2] M. L. Fuller, *Water-supply Paper* 160, U. S. Geol. Survey, 1906, p. 65.

[3] T. A. Rickard, *Min. and Sci. Press*, June 27, 1908.

[4] M. L. Fuller, *Water-supply Paper* 160, U. S. Geol. Survey, 1906, p. 65.

[5] For other examples see J. F. Kemp, The ground waters, *Trans.*, Am. Inst. Min. Eng., 45, 1914, pp. 3–25.

[6] Mine waters, *Trans.*, Lake Superior Min. Inst., 12, 1908, pp. 154–163.

the surface waters are of the normal, potable type and that they descend in diminishing quantities only to a depth of about 1,000 or 1,500 feet below the surface. Below this depth moisture is scant, but where it appears it consists of drippings of strong calcium chloride brine which cannot in any way be explained as being derived from the surface water. Many levels are absolutely dry and water must be sent down for drilling. This case is particularly convincing, for we have here many features in favor of a strong circulation: Moist climate, inclined position of beds, and great permeability.

The gold mines of St. John del Rey in Brazil and Kolar in India have both passed the 7,000-foot level. In both cases, according to information received from Prof. L. C. Graton, the rocks are dry at that depth. At the Rand in South Africa, there is some difficulty to obtain exact data, for in order to abate the danger of silicosis no attempt is made to keep the surface water down. Water is supplied to all the workings.

**Depth of Water Level.**—In moist climates the water level is usually found within 50 feet of the surface, but in regions with less rainfall there is great diversity in the location of this upper limit of the zone of saturation. In the more arid regions the water is often met 300 or 400 feet below the surface. At the Abe Lincoln gold-quartz mine, New Mexico, a little water began to come in 1,300 feet below the surface. In the rich deposits of Tintic, Utah, the water level in limestone lies 1,700 to 2,000 feet below the surface, but in mines in andesite and porphyry in the same district water may be found at much less depth.

When water is being drained or pumped from a mine the water level is artificially depressed, in the form of a flat funnel. The pump, in this case, does not merely drain the bottom level, but receives water from higher levels farther away from the shaft. It is important to note this, for the water thus obtained from the bottom of a wet mine may not have the same composition as that originally belonging to this level.

**Total Amount of Free Water in Earth's Crust.**—Several estimates have been made of the total amount of uncombined water contained in the upper crust. The older estimates by Delesse, Dana, and Slichter were very high. Van Hise[1] reduced the estimate materially and concluded that it would be equivalent to a sheet of water 226 feet (69 meters) thick over the continental areas.

Fuller[2] estimates, after a careful study of the problem, that the total water would be equivalent to a uniform sheet 96 feet thick over the entire surface of the earth. This estimate is probably more nearly correct than any of the others.

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[1] A treatise on metamorphism, *Mon.* 47, U. S. Geol. Survey, 1904, pp. 128–129, 570–571.

[2] M. L. Fuller, *op. cit.*, p. 72.

# CHAPTER V

# THE COMPOSITION OF UNDERGROUND WATERS

## INTRODUCTION

Water is continually evaporated from sea and land. From the gathered clouds it is precipitated as pure rain water, which, by the aid of absorbed oxygen and carbon dioxide immediately begins the attack on disintegrating rocks. The rivers finally carry suspended particles and dissolved salts to the sea. Considered more closely, the rivers are the combined products of the weak solutions of the immediate run-off and the stronger ground-water solutions from the zone of discharge which have been in longer contact with the rocks and leached them more thoroughly. This inconspicuous process of decomposition of rocks and solution of resulting salts is one of scarcely realized geologic importance. F. W. Clarke[1] states that the Mississippi annually carries to the sea about 108 metric tons of salts from each square mile of territory drained; the Colorado abstracts about 51 tons from the same unit area.

A smaller part of the ground water sinks to enrich the static zone of stagnant waters, and ultimately becomes highly charged with salts. A still smaller part of the water is permanently withdrawn by entering hydrated compounds like kaolin.

To complete the picture we must not overlook the ascending hot solutions which come from great depths and which in part at least are derived from rising magmas.

As most mineral deposits have been formed by aqueous solutions the composition of the waters of rivers, lakes, and seas becomes a study of importance. Even more important is the composition of the underground waters of wells and springs. In considering them from a chemical standpoint it will be best to attempt no artificial distinction between thermal or cold, mineral or non-mineral waters.

The substances dissolved in the ground water depend upon the formations which it traverses. At the immediate surface organic life may influence the composition. In general, each formation yields its characteristic salts to the percolating waters. Each natural water is a chemical system of balanced constituents of more or less dissociated electrolytes, of colloids, and of gases.

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[1] F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 116.
R. B. Dole and H. Stabler, *Water-supply Paper* 234, U. S. Geol. Survey, 1909, p. 78.

## CALCIUM CARBONATE WATERS IN IGNEOUS ROCKS

Igneous rocks, of deep-seated origin, as well as crystalline schists contain only small amounts of soluble salts. The surface waters penetrating them are charged with more or less carbon dioxide, which, at ordinary temperatures, gradually decomposes the silicates, particularly the pyroxene, amphibole, biotite, and the calcium feldspars; the alkali feldspars are more slowly attacked. As a result the springs in such terranes will have a low salinity, rarely above 1,000 parts per million, and will contain principally calcium carbonate, with more or less of the corresponding magnesium salt; a smaller amount of sodium carbonate and much less of potassium carbonate are present. There will be little chlorine and sulphuric acid radical. The silica is relatively high. Such calcium carbonate waters are characteristic not only of superficial springs, but also of the deeper circulation in crystalline terranes; in the latter case the waters may be warm, though usually they are cold. The spring-fed rivers in such terranes have a similar composition.

Where magnesian rocks, like basalt and serpentine, abound, the underground waters are richer in magnesia than usual; and this substance may even equal the calcium. Waters of this calcium carbonate type are common and, when encountered as ascending springs or elsewhere, justify the presumption of surface origin. It often happens that hot springs which are not characterized by an abundance of calcium carbonate are accompanied by numerous other springs of somewhat lower temperature. A comparison of analyses will usually show that in proportion to the lowering of the temperature the quantity of calcium carbonate increases; this indicates a cooling admixture of surface waters bearing calcium.

Waters of the kind described above are generally poor in the rarer metals. A little arsenic is found in some cases and traces of barium, strontium, lithium, boron, and phosphorus are sometimes recorded. Where they traverse mineral deposits, metals contained in the deposits will, of course, be dissolved. The springs may, under favorable conditions, form crusts of calcium carbonate and hydroxide of iron, but as a rule their powers of solution and deposition are weak. Where the rocks contain much pyrite, as often is the case in mining districts, the sulphates, especially calcium sulphate, rapidly increase in the waters.

## CALCIUM CARBONATE WATERS IN SEDIMENTARY ROCKS

Waters of the type described above are not confined to igneous rocks. They are often found in circulation in glacial drift and also in sedimentary rocks—sandstones, limestones, and dolomites. Such waters sometimes contain hydrogen sulphide and carbon dioxide. The derivation of the latter is not always easily explained. In some cases the gas may emanate from a deep-seated magma, but more commonly it is formed by decomposition of carbonates. An example of such water is furnished by the cold Cresson Spring in Pennsylvania, which issues from a shale member between sandstones in a 3,000-foot series of Coal Measures, containing practically no limestone. This water is pure, its salinity being only 442 parts per million, and of this 272 may be calculated as calcium carbonate, 76 as sulphates of sodium, magnesium, and calcium, and 11 as sodium chloride. According to a careful analysis by Genth this water contains traces of nickel, cobalt, iron, manganese, copper, strontium, barium, and fluorine, 0.17 part per million of the last-named element being present. Several analyses of similar well-known waters are quoted below.

COMPOSITION OF SALTS AND TOTAL SALINITY OF CALCIUM CARBONATE WATERS IN SEDIMENTARY AND CRYSTALLINE ROCKS

| | *A* | *B* | *C* | *D* | *E* | *F* |
|---|---|---|---|---|---|---|
| $CO_3$ | 47.14 | 57.80 | 38.46 | 40.02 | 41.47 | 48.64[1] |
| $SO_4$ | 6.67 | 3.10 | 15.35 | 21.73 | 3.93 | 6.30 |
| Cl | 4.18 | 1.30 | 2.81 | 0.64 | 1.27 | 5.40 |
| S | ........ | 0.50 | ........ | ........ | ........ | ........ |
| $PO_4$ | ........ | ........ | ........ | ........ | 0.03 | ........ |
| $NO_3$ | ........ | ........ | ........ | ........ | 0.23 | ........ |
| Ca | 22.67 | 13.90 | 13.24 | 23.35 | 23.54 | 16.56 |
| Mg | 6.17 | 2.30 | 4.33 | 5.82 | 2.56 | 7.64 |
| Mn | ........ | 0.10 | ........ | ........ | 0.17 | ........ |
| Na | 5.32 | 5.50 | 12.86 | 1.81 | 2.38 | 10.36 |
| K | | 0.40 | 3.76 | 2.04 | 0.80 | ........ |
| $(Al\ Fe)_2O_3$ | ........ | 1.70 | ........ | 0.58 | 0.10 | 0.60 |
| $NH_4$ | ........ | ........ | ........ | ........ | 0.03 | ........ |
| $SiO_2$ | 7.85 | 13.40 | 9.19 | 4.01 | 22.85 | 4.50 |
| $BO_2$ | ........ | ........ | ........ | ........ | 0.64 | ........ |
| | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 |
| Salinity, parts per million | 280 | 245 | 282 | 563 | 199 | 222 |

[1] $HCO_3$.

NOTES RELATING TO ABOVE ANALYSES

*A*. Aztec Spring, 4 miles east of Santa Fe, New Mexico. In schist and granite. Cold. Analysis by F. W. Clarke, Data of geochemistry, 1924, p. 68.

*B*. Cold spring in Federal Loan mine, in granite and schist, Nevada City, California. Approximate analysis by W. F. Hillebrand; contains a little manganese and trace of lead. Deposits calcium carbonate, limonite, and some arsenic. Sulphur probably due to reduction from small amount of $H_2S$ after bottling. *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, p. 121.

*C*. Cold water from 500-foot level, Geyser mine, Silver Cliff, Colorado. Analysis by W. F. Hillebrand. In "The mines of Custer County, Colorado," by S. F. Emmons. *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, p. 461. Free and semi-combined $CO_2$, 38.8 parts per million.

*D.* Virginia Hot Springs, Virginia. Analysis by F. W. Clarke, Data of geochemistry, 1924, p. 194. Temperature tepid. Issues from Paleozoic sediments. See also *Bull.* 32, U. S. Survey, 1886, p. 61.

*E.* Hot Springs, Arkansas. Spring No. 16. Temperature 62° C. Issues from sharply compressed folds of Silurian sandstone and shale. $CO_2$ from bicarbonates 28.34 cc. per liter; nitrogen 8.39 cc. per liter; oxygen 2.49 cc. per liter; $H_2S$ none. Arsenic none; trace iodine and bromine. Analysis by J. K. Haywood. The Hot Springs of Arkansas, *Senate Doc.* 282, Fifty-seventh Congress, First Session, 1902, p. 94. Recalculated by F. W. Clarke, Data of geochemistry, 1924, p. 196.

*F.* Cold water from well of Missouri Lead and Zinc Company, Joplin, Missouri. Depth 1,387 feet. In Paleozoic limestone. Analysis by Cleveland and Miller Laboratory. *Water-supply Paper* 195, U. S. Geol. Survey, 1907, p. 137. Recalculated.

These waters frequently form ascending springs, but are generally of meteoric origin.

Carbonate waters are undoubtedly active in solution and deposition in the upper part of the crust, and especially in the formation of concentrations from weathering rocks. They may deposit calcareous sinters and effect concentrations of iron and manganese. Some lead and zinc deposits in limestone may also be genetically connected with them; their power of solution and concentration of rarer metals appears to be weak, unless they contain carbon dioxide and hydrogen sulphide. Such waters in Kansas, Missouri, and Kentucky have been found to contain zinc and probably also lead and copper.

The salts are surely obtained from the rocks traversed.

## CHLORIDE WATERS IN SEDIMENTARY ROCKS

**Infiltration from Present Oceans.**—Wells and springs along the sea coasts usually contain a higher percentage of sodium chloride than those farther inland; this may be caused either by infiltration of sea water into sediments or porous igneous rocks, or by winds carrying finely divided salt from the spray of the waves.

**Solution of Saline Deposits.**—Many past geologic periods included epochs of desiccation and desert climate when salt was precipitated from evaporating waters of closed basins. Surface waters encountering such sedimentary deposits easily dissolve the sodium chloride, and wells and springs rich in this salt are characteristic of many regions. Besides sodium these waters contain calcium and magnesium, and are often rich in calcium chloride. They are poor in silica and potassium and rarely contain much calcium which can be combined with carbon dioxide. The presence of bromine is almost characteristic; traces of iodine and boron are often found. Barium and strontium are almost always present, the former sometimes in considerable amount. Free carbon dioxide and hydrogen sulphide are sometimes found, the latter especially where there is an abundance of calcium sulphate. Waters of this general type are characteristic of certain Paleozoic beds in the eastern United States, as for instance, the Silurian of New York and Michigan and certain parts of the Carboniferous in Michigan. In the western States the "Red Beds," generally of Permian or Triassic age, are sometimes rich in salt and gypsum; and this combination appears in the waters of these terranes.

There are many similar springs and wells in Pennsylvania, and in fact, all through the interior Paleozoic basin, from Arkansas to Canada. The Saratoga Springs of New York, issuing from Silurian limestones, probably belong to this class. Their temperature is about 50° F.; the total solids amount to about 11,000 parts per million, of which the larger part is sodium chloride. Barium is conspicuously present, in some analyses to a maximum of about 34 parts per million, likewise bromine to about 1.20 parts per million. Small amounts of silica, iron, and lithium, and traces of boron, iodine, and fluorine are recorded. The origin of the $CO_2$ so abundant at Saratoga Springs is uncertain. J. F. Kemp believed it to be of magmatic derivation. Examples of such waters are given in the table of analyses on page 46.

Certain of these waters are abnormally rich in calcium chloride, that most easily soluble salt which remains as the last liquid residue in evaporating brines. Several instances of such waters have been interpreted as residual or connate brines, remaining in early isolated Paleozoic basins.[1]

In the lower peninsula of Michigan brines are obtained from deep wells in the Carboniferous and Silurian. One of the springs in this region contains 12,000 parts per million in total solids, with 6,000 calculated as NaCl, 1,600 as $MgCl_2$, and 4,100 as $CaCl_2$. The researches of A. C. Lane have shown that the scanty waters in the deep levels of the copper mines of Michigan have a similar composition, except that here calcium chloride prevails.

In oil-bearing districts salt waters are of very frequent occurrence. They are rich in sodium chloride and often also contain the chlorides of calcium and magnesium, as well as more or less bicarbonates. They are always poor in sulphates, and this is perhaps due to their reduction by the hydrocarbons. Such salt solutions have been variously interpreted as connate waters, as solutions of saline deposits, as of magmatic origin,[2] or as resulting from long-continued base exchange (p. 53).

In the western states many similar waters occur in the Red Beds, but, as stated, they are usually also rich in calcium sulphate. As an example may be cited the tepid Quelites Spring in New Mexico,[3] which ascends

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[1] E. M. Shepard, Underground waters of Missouri, *Water-supply Paper* 195, U. S. Geol. Survey, 1907, p. 81.

[2] C. W. Washburne, Chlorides in oil field waters, *Trans.*, Am. Inst. Min. Eng., 45, 1915, pp. 687–693. G. S. Rogers, Chemical relations of the oil field waters in the San Joaquin Valley, California, *Bull.* 653, U. S. Geol. Survey, 1917.

[3] F. A. Jones, New Mexico mines and minerals, 1904, p. 309.

through "Red Beds" and contains about 2.6 per cent of solids; one-half of which is calculated as sodium chloride and the larger part of the remainder as calcium sulphate. Bromine, boron, and barium are present.

The Triassic strata of the French Alps and the Pyrenees are rich in similar waters, many of which are warm. The mineral combination is

COMPOSITION OF SALTS AND TOTAL SALINITY OF CHLORIDE WATERS

(*Cited from Clarke's Data of Geochemistry*, 1924, pp. 183–188)

| | *A* | *B* | *C* | *D* | *E* | *F* |
|---|---|---|---|---|---|---|
| Cl | 55.83 | 58.79 | 42.00 | 62.31 | 63.55 | 56.58 |
| Br | 0.04 | trace | 1.13 | 0.53 | ........ | 0.04 |
| I | 0.03 | ........ | 0.02 | 0.01 | ........ | trace |
| $SO_4$ | 3.12 | 0.94 | 0.08 | 0.03 | 0.01 | 0.78 |
| $CO_3$ | 2.63 | 0.61 | 18.59 | 0.27 | 0.01 | 3.13 |
| $B_4O_7$ | ........ | ........ | ........ | ........ | ........ | 0.01 |
| Na | 33.09 | 30.38 | 27.62 | 18.35 | 5.63 | 32.60 |
| K | 0.27 | 3.76 | 0.78 | 1.55 | ........ | 1.16 |
| Li | ........ | ........ | 0.08 | 0.04 | ........ | 0.04 |
| $NH_4$ | ........ | ........ | ........ | 0.23 | ........ | 0.07 |
| Ca | 3.72 | 4.90 | 6.03 | 13.86 | 30.78 | 4.05 |
| Ba | ........ | ........ | 0.09 | ........ | ........ | 0.01 |
| Sr | ........ | ........ | ........ | ........ | ........ | 0.12 |
| Mg | 1.13 | 0.40 | 3.41 | 2.53 | 0.01 | 0.61 |
| $Al_2O_3$ | ........ | 0.02 | ........ | 0.02 | ........ | ........ |
| $Fe_2O_3$ | 0.06 | ........ | 0.03 | ........ | ........ | ........ |
| Fe | ........ | ........ | ........ | 0.25 | ........ | 0.04 |
| $SiO_2$ | 0.08 | 0.20 | 0.14 | 0.02 | 0.01 | 0.76 |
| | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 |
| Salinity, parts per million. | 10,589 | 23,309 | 12,022 | 309,175 | 212,300 | 8,241 |

NOTES RELATING TO ABOVE ANALYSES

*A*. Cincinnati artesian well, Cincinnati, Ohio. Analysis by E. S. Wayne, cited by A. C. Peale, *Bull.* 32, U. S. Geol. Survey, 1886, p. 133. This water contains considerable quantities of free $H_2S$ and $CO_2$.

*B*. Utah Hot Springs, 8 miles north of Ogden, Utah. Temperature 55° C. Analysis by F. W. Clarke, *Bull.* 9, U. S. Geol. Survey, 1884, p. 30.

*C*. Congress Spring, Saratoga, New York. Analysis by C. F. Chandler, cited by A. C. Peale, in *Bull.* 32, U. S. Geol. Survey, 1886, pp. 38, 39. Traces of F, P, B, Sr, and Al. Contains much free $CO_2$.

*D*. Brine from well 2,667 feet deep at Conneautville, Pennsylvania. Analysis by A. E. Robinson and C. F. Mabery, *Jour.* Am. Chem. Soc., 18, 1896, p. 915. A little $H_2S$ is present.

*E*. Water from the deep levels of the Quincy mine, Hancock, Michigan. Analysis by George Steiger.

*F*. The Kochbrunnen, Wiesbaden, Germany. Analysis by C. R. Fresenius. This water also contains traces of I, P, and As.

a characteristic mingling of chlorides and sulphates, and undoubtedly all of the constituents are derived from the sedimentary rocks mentioned.

A celebrated group of these chloride springs are found in Germany on both sides of the Rhine. Among them are the waters of Soden, Homburg, Wiesbaden, Kreutznach, Kissingen, and Nauheim. Most of them issue from or ascend through salt-bearing beds of Permian or Triassic age, and their composition is similar. The springs of Kreutznach are especially rich in calcium chloride. Some of the springs cited are hot, others cold; some are rich in carbon dioxide. In regard to Kreutznach and Wiesbaden there is room for doubt, for the former springs stand in intimate relation to eruptive rocks, while the latter issue from a gneiss and are by some authors considered of juvenile origin. The majority of them, at any rate, have certainly derived their salts from sedimentary beds.

The chloride waters, described above, are capable of dissolving and depositing many metallic substances and have strong dehydrating power. Their relation to mineral deposits will be mentioned later.

### CHLORIDE WATERS IN IGNEOUS ROCKS

Waters rich in chlorine are sometimes found as ascending springs in igneous rocks, but almost always close to regions of comparatively recent volcanic activity. Their composition is somewhat different from the brines resulting from the dissolving of salts from sedimentary beds. Bromine is seldom present except in mere traces, while boron appears in considerable amounts. Such tepid salt waters arise, for instance, in the volcanic region around Clifton, Arizona. The Paleozoic rocks of this region are not known to contain either salt or gypsum. Another case is the Glenwood Hot Springs in western Colorado; the springs at this place issue from limestone, but the structural relations show that the basal granite underlies this limestone at slight depths. The temperature is 49.5° C.; the water contains a large amount of sodium chloride and relatively small amounts of carbonates and sulphates. Hydrogen sulphide and free carbon dioxide are present. Still another case is Steamboat Springs, Nevada, which issue from granodiorite near the eastern base of the Sierra Nevada in a region of Tertiary volcanism.

Many of these springs are rich in carbon dioxide and hydrogen sulphide; they often contain many of the rarer elements, as shown in the analyses quoted below, and they usually appear in regions rich in ore deposits. Doubt as to the derivation of the salt may, of course, exist in some cases. Another notable instance of chloride springs of this class is mentioned by Daubrée[1] from the provinces of Antioquia and Cauca in Colombia, where they issue in great abundance from granite, crystalline schist, and late volcanic rocks. Great difficulties arise in attempting to

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[1] Les eaux souterraines, etc., II, 2, Paris, 1887, p. 106.

trace the origin of the sodium chloride in springs of this class to surrounding rocks, even admitting that granite and other crystalline rocks may contain traces of this salt. Sinters of calcium carbonate and silica are often deposited at the orifices of these springs.

COMPOSITION OF SALTS AND TOTAL SALINITY OF SODIUM CHLORIDE AND SILICA WATERS

(*After Clarke's Data of Geochemistry*, 1924, pp. 188 and 197)

| | *A* | *B* | *C* | *D* | *E* |
|---|---|---|---|---|---|
| Cl | 35.00 | 36.61 | 31.64 | 13.52 | 37.52 |
| Br | | | 0.25 | | |
| $SO_4$ | 4.58 | 1.84 | 1.30 | 9.01 | 4.96 |
| S | 0.22 | | | 0.32 | |
| $CO_3$ | 5.08 | 0.15 | 8.78 | 10.16 | |
| $PO_4$ | 0.03 | 0.08 | | | |
| $AsO_4$ | | | 0.24 | | |
| $B_4O_7$ | 8.88 | 2.24 | 1.19 | | |
| Na | 30.35 | 21.44 | 26.42 | 19.71 | 24.22 |
| K | 3.79 | 4.45 | 1.93 | 1.88 | 0.36 |
| Li | 0.27 | 0.22 | 0.40 | | |
| $NH_4$ | | 0.02 | trace | 0.28 | |
| Ca | 0.25 | 0.39 | 0.11 | | 2.59 |
| Mg | 0.01 | 0.08 | 0.04 | 0.08 | 0.19 |
| Fe | | trace | trace | | trace |
| As | 0.10 | | | | |
| Sb | 0.02 | | | | |
| $Al_2O_3$ | 0.01 | 0.76 | 0.12 | | 0.35 |
| $SiO_2$ | 11.41 | 31.72 | 27.58 | 45.04 | 29.81 |
| | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 |
| Salinity, parts per million | 2,850 | 1,830 | 1,388 | 1,131 | 2,735 |

*A*. Steamboat Springs, Nevada. Analysis by W. H. Melville, given by G. F. Becker in *Mon.* 13, U. S. Geol. Survey, 1888, p. 349. Bicarbonate reduced to normal salts. Temperature 85° C. Contains free carbon dioxide and hydrogen sulphide. Traces of iron and quicksilver; deposits cinnabar and stibnite.

*B*. Coral Spring, Norris Basin, Yellowstone National Park. Analysis by F. A. Gooch and J. E. Whitfield, *Bull.* 47, U. S. Geol. Survey, 1888. Temperature 73° C. $H_2S$ none. Free $CO_2$, 42.5 parts per million.

*C*. Old Faithful Geyser. Same locality and analysts. Temperature 84°–88° C. $H_2S$, 0.2 part per million.

*D*. Great Geyser, Iceland. Analysis by F. Sandberger.

*E*. Water of the pink terrace, Roturoa geyser. Analysis by W. Skey.

Closely related to this group are the predominant springs in the great geyser regions of Yellowstone National Park, New Zealand, and

Iceland.[1] They are essentially sodium chloride waters with large amounts of silica, a large quantity of free carbon dioxide, and a little hydrogen sulphide. Large amounts of boron, usually calculated as sodium borate, are often present, and also frequently arsenic. Bromine is rarely recorded in quantities approaching those in the brines from sedimentary formations. The waters are always hot and usually ascend through volcanic rocks, mostly rhyolite; from these the silica is supposed to be derived, but no such explanation seems sufficient to account for the predominating salt, sodium chloride, or for the boron. In the Yellowstone Park a number of the springs issuing near limestone bear evidence of their passage through this rock in increased quantities of calcium and magnesium. Others are rich in sulphate of sodium and other sulphates, but these springs give an acid reaction and the sulphates may be due to the oxidation of hydrogen sulphide.

## SULPHATE WATERS IN SEDIMENTARY ROCKS

The waters which traverse sedimentary rocks are often rich in salts, particularly in sulphates. The gypsum waters have been mentioned and are connected with the sodium chloride waters in a manner corresponding to the association of gypsum and rock salt. By interaction of calcium sulphate and magnesium carbonate, the sulphate of magnesium may be formed, or it may be derived from the decomposition of a pyritic dolomite.

Sodium sulphate waters are almost characteristic of certain formations in the western Cretaceous, for instance; these formations consist mainly of sandstones and carbonaceous shales, the latter often pyritiferous; and the whole series mainly a product of near-shore deposition. The oxidation of the pyrite furnishes solutions containing free sulphuric acid, and by reaction between this and various other substances sulphates of calcium, magnesium, and sodium will be formed. The lowest member of this series, the Dakota sandstone, is particularly noted as a water-carrying formation. The water, which is under artesian pressure, penetrates this formation for several hundred miles underground from its outcrop and in places contains so much sodium sulphate as to be unfit for irrigation purposes when tapped by deep wells. There is no evidence that this water has formed mineral deposits in the sandstone.

In land deposits contained in many series of sedimentary rocks sodium carbonate and sodium sulphate are formed by several well-established reactions, and percolating waters will easily abstract these salts. The interaction of calcium sulphate and sodium carbonate results in sodium sulphate and precipitation of calcium carbonate. Sodium sulphate[2] in

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[1] E. T. Allen and A. L. Day, Hot Springs of the Yellowstone National Park, Carnegie Institute, Washington, D. C., 1935, 525 pp.

[2] F. W. Clarke, Data of geochemistry, 1924, p. 241; E. W. Hilgard, *Am. Jour. Sci.*, 4th ser., 2, 1896, p. 100; Cameron and Bell, *Bull.* 33, Bureau of Soils, 1906; F. K. Cameron, *Bull.* 17, Bureau of Soils, 1901.

the presence of free carbon dioxide will dissolve calcium carbonate, forming sodium bicarbonate and a precipitate of gypsum. Reactions in soils between sodium chloride and calcium sulphate result, according to Cameron, in calcium chloride and sodium sulphate; and similar reactions take place between sodium chloride and calcium carbonate.

In many waters in sedimentary formations chlorides and earthy carbonates appear mixed with sulphates. Waters from artesian wells at Roswell, New Mexico, about 400 feet deep, derived from Permian limestones,[1] have a temperature of 64°–70° F. and contain from 600 to 1,200 parts per million of solid salts, of which 300 to 576 are calcium and magnesium sulphates and the remainder carbonates and chloride of sodium. In regions of dislocations such waters may be hot and then the ordinarily low percentage of silica may increase considerably. The Arrowhead

PERCENTAGE COMPOSITION AND TOTAL SALTS AND ACIDS OF SULPHATE WATERS AND ACID WATERS

| | *A* | *B* | *C* | *D* | *E* | *F* | *G* |
|---|---|---|---|---|---|---|---|
| HCl free | ...... | ...... | ...... | ...... | 0.18 | 5.60 | ...... |
| $H_2SO_4$ free | ...... | ...... | ...... | 9.37 | 1.29 | 59.11 | 31.38 |
| $H_3BO_3$ | ...... | ...... | ...... | ...... | 2.73 | ...... | ...... |
| Cl | 0.48 | 11.10 | ...... | 0.32 | ...... | ...... | 0.81 |
| S | ...... | ...... | ...... | ...... | ...... | ...... | 0.04 |
| $SO_4$ | 66.28 | 59.68 | 76.57 | 68.21 | 67.66 | 20.21 | 46.95 |
| $CO_3$ | 0.60 | 1.67 | ...... | ...... | ...... | ...... | ...... |
| $NO_3$ | ...... | ...... | ...... | ...... | ...... | ...... | 0.02 |
| Na | 30.46 | 13.89 | 1.19 | 0.22 | 0.73 | 8.35 | 1.47 |
| K | 1.08 | 0.49 | ...... | 0.11 | 0.24 | 0.32 | |
| Li | ...... | ...... | ...... | ...... | 0.01 | ...... | ...... |
| $NH_4$ | ...... | ...... | ...... | ...... | 22.85 | ...... | ...... |
| Ca | 0.67 | 2.91 | 5.82 | 0.38 | 1.18 | 0.47 | 1.62 |
| Mg | 0.41 | 10.19 | 3.39 | 1.11 | 0.36 | 0.22 | 2.48 |
| Fe″ | ...... | ...... | 4.28 | 1.19 | trace | ...... | 5.74 |
| Fe‴ | ...... | ...... | ...... | ...... | ...... | 0.33 | 8.22 |
| Al | ...... | ...... | 7.36 | 11.08 | 0.10 | trace | none |
| $SiO_2$ | 0.02 | 0.07 | 1.39 | 7.11 | 2.67 | 5.39 | 1.27 |
| | 100.00 | 100.00 | 100.00 | 99.10 | 100.00 | 100.00 | 100.00 |
| Salinity, parts per million | 74,733 | 15,682 | 3,303 | 464 | 3,365 | 1,862 | 9,760 |

*A.* Abilena Well, Abilene, Kansas, 130 feet deep. Analysis by E. H. S. Bailey, Geol. Survey Kansas, 7, 1902, p. 166. In Permian strata. From Clarke's Data of geochemistry, 1924, p. 189.

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[1] C. A. Fisher, Report on the Roswell artesian area, *Water-supply Paper* 158, U. S. Geol. Survey, 1906.

*B.* King's Mineral Spring near Dallas, Indiana. *Twenty-sixth Ann. Rept.*, Indiana Dept. Geol., 1901, p. 32. Traces of Al, Fe, Ba, Sr, Li, Mn, Ni, Zn, Br, $PO_4$, and $B_4O_7$. Geological horizon Paleozoic shale.

*C.* Alum Well, Versailles, Missouri. Analysis by P. Schweitzer, Geol. Survey Missouri, 3, 1892, p. 131. In Pennsylvanian shale.

*D.* Rockbridge Alum Springs, Virginia. Analysis by M. B. Hardin. Cited in Clarke's Data of geochemistry, 1924, p. 199. From pyritic shale. Contains also 0.69 Mn, 0.01 Li, 0.05 Co, 0.07 Ni, 0.08 Zn, and traces of Cu, $HNO_3$, and $PO_4$.

*E.* Devil's Inkpot, Yellowstone National Park. Analysis by F. A. Gooch and J. E. Whitfield, *Bull.* 47, U. S. Geol. Survey, 1888, p. 80. Contains also 65 parts of free $CO_2$ and 5 parts of $H_2S$ per million. Cited in Clarke's Data of geochemistry, 1924, p. 200.

*F.* Cameron's Bath, Roturoa geyser district, New Zealand. Analysis by W. Skey, *Trans.*, New Zealand Institute, 10, 1877, p. 423. Contains 6 parts per million of $H_2S$. Cited in Clarke's Data of geochemistry, 1924, p. 201.

*G.* Water at Sulphur mine of Cove Creek, Beaver Valley, Utah. Analysis by W. M. Barr, *Water-supply Paper* 217, U. S. Geol. Survey, 1908, p. 20. Contains also much free $H_2S$.

Spring[1] of San Bernardino Valley in southern California, issuing from Tertiary sediments, has a temperature of 184° F. and contains 1,086 parts per million of solids, of which 735 are calculated as sodium sulphate, 69 as potassium sulphate, 23 as calcium sulphate, 3 as magnesium sulphate, 141 as sodium chloride, and 23 as calcium carbonate; 85 are present as silica. In well waters of the same valley the solids range from 191 to 260 parts per million and the relation is $CaCO_3 > MgCO_3 = Na_2SO_4 > Na_2CO_3 > > NaCl$. Silica amounts to 24 to 32 parts per million.

Waters percolating through oxidizing pyritic shales sometimes contain large amounts of the sulphates of aluminum and ferrous iron; evidently this happens only when comparatively large amounts of sulphuric acid, which is capable of attacking aluminous silicates, are set free.

Many such waters are found in the eastern and central states and usually contain small amounts of rarer metals; traces of nickel, zinc, and arsenic are common. The sulphate waters, especially those rich in iron and aluminum, are of great importance in the genesis of deposits in the oxidizing zone, and may form, at their orifices, large quantities of ocherous deposits. Many waters of this kind are known from Virginia, issuing from pyritic shales, and Peale[2] quotes some interesting and reliable analyses. A water from Alleghany Springs in Montgomery County, analyzed by Genth, contained 3,129 parts per million of solids, of which the principal constituents were calculated as 1,955 parts $CaSO_4$, 225 parts $MgSO_4$, and 61 parts $CaCO_3$. Small quantities of strontium, barium, fluorine,

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[1] W. C. Mendenhall, Hydrology of San Bernardino Valley, *Water-supply Paper* 142, U. S. Geol. Survey, 1905.

[2] A. C. Peale, Lists and analyses of the mineral springs of the United States, *Bull.* 32, U. S. Geol. Survey, 1886, pp. 58–65.

and silica and traces of zinc, lead, copper, and cobalt are noted. Some free carbon dioxide and a trace of hydrogen sulphide are present.

The Jordan Alum Springs in Rockbridge County, Virginia, of which several analyses by J. W. Mallet are recorded, contain from 306 to 935 parts per million of solid salts, of which the larger amount consists of aluminum sulphate, 35 to 85 parts of ferric sulphate, and from 8 to 17 parts of manganese sulphate. Small quantities of copper, zinc, cadmium, nickel, and cobalt are determined, also a trace of fluorine. One of the waters contained 102 parts of copper and 9 parts of zinc. Springs of similar composition are found in Pennsylvania and other eastern States. All these acid springs are poor in silica and contain very little chlorine.

### ACID SULPHATE WATERS IN IGNEOUS ROCKS

Sulphate springs in connection with igneous rocks and volcanism are in part at least products of the oxidation of ascending waters of alkaline reaction, containing free hydrogen sulphide. In regions of volcanic activity such oxidation may take place on a large scale. These acid waters effect rock alteration by attacking aluminum silicates and developing alunite. Free hydrochloric acid may be generated in large amounts by decomposition of chlorides by sulphuric acid.

Such waters occur near the orifices of alkaline hot springs, as well as at volcanoes. The development of free acid, of course, displaces the equilibrium and the oxidized water may differ greatly from its parent liquid; thus it happens that a single ascending hot spring may yield a whole series of derivatives of varying temperature and composition by mingling with other waters and by oxidation.

The peculiar water from the Yellowstone National Park known as the Devil's Inkpot contains, besides free acids, a large amount of sulphate of ammonia.

At the geysers of Sonoma County, California,[1] are a great number of springs of varying temperature and composition, all of them heavily charged with hydrogen sulphide. The primary water at a temperature of 110° F. appears to contain chiefly carbonate of magnesium with some of calcium. The total solids amount to about 568 parts per million, most of which consist of the above-mentioned carbonates; there are 92 parts of silica per million. This water is probably of mixed origin; the carbonates are clearly derived from the serpentinoid rocks of the vicinity, but the hydrogen sulphide is most likely of magmatic origin. Oxidation gives rise to a long series of peculiar sulphate waters, most of them rich in dissolved solids and of high temperature. Allen and Day consider these acid springs as the normal emanations from magmas though admixed

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[1] Winslow Anderson, Mineral springs, etc., of California, 1892, pp. 136–154.
G. A. Waring, Springs of California, *Water-supply Paper* 338, U. S. Geol. Survey, 1915, p. 109.

with meteoric waters. If this is true they must originate from near-surface intrusions or from flows, because no deep-seated water could retain its acidity during a long ascent.

Finally, acid water may result directly from the oxidation of deposits of sulphur. A water of the latter type from Beaver County, Utah, is described by W. T. Lee.

### SODIUM CARBONATE WATERS IN SEDIMENTARY ROCKS

Waters containing sodium carbonate in large amounts are not common in sedimentary rocks, but here and there wells or springs of this character are encountered; they are usually cold and often contain some free carbon dioxide and hydrogen sulphide. The alkaline carbonate is probably, as suggested above, derived from a reaction between sodium sulphate and calcium carbonate or between sodium chloride and calcium carbonate. Waters of this kind occur at a few places in the eastern and central states. A good instance is furnished by some Missouri waters in Carboniferous limestone, one of which is quoted under *E* in the following table. Similar waters are those of the wells at La Junta, Denver, and Greeley, Colorado. The artesian wells at Denver, about 1,200 feet deep, are in the Arapahoe, Eocene, while the Greeley well, of the same depth, is sunk in Laramie sandstone. The maximum of total solids is about 1,530, divided between sodium carbonate and sodium chloride. Some free $CO_2$ is present. Many artesian waters in New South Wales are rich in sodium carbonate. Sedimentary beds containing volcanic tuffs often yield sodium carbonate waters. An explanation of this has been given by B. C. Renick.[1] He shows that waters rich in the carbonates and sulphates of calcium and magnesium are converted to sodium carbonate by an exchange of bases, effected by certain hydrous silicates in the sedimentary tuffaceous beds.

### SODIUM CARBONATE WATERS IN IGNEOUS ROCKS

Ascending sodium carbonate waters are mainly found in regions of subsiding or expiring volcanism. During surface eruptions alkaline chlorides and carbonates always appear as sublimates; and waters traversing tuffs, breccias, and lava flows may dissolve these salts together with other volcanic exhalations such as borates. The characteristic sodium carbonate waters are, however, of deep-seated origin and usually break through the older igneous or metamorphic rock underlying the lavas in regions where the active volcanism has ceased; the prevailing opinion is that these waters with their charge are in whole or in part of magmatic origin. They rarely contain much calcium and are poor in silica, but are usually heavily charged with carbon dioxide and sometimes hydrogen sulphide. They almost always contain many rarer substances such as boron, fluorine, iodine, arsenic, and various metals.

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[1] *Water-supply Paper* 520, U. S. Geol. Survey, 1925, pp. 53–72.

PERCENTAGE COMPOSITION OF SALTS AND TOTAL SALINITY OF SODIUM CARBONATE WATERS

(*Cited from Clarke's Data of geochemistry*, 1924, pp. 192–194, and 198)

| | *A* | *B* | *C* | *D* | *E* | *F* |
|---|---|---|---|---|---|---|
| Cl | 6.17 | 8.85 | 11.52 | 13.57 | 6.63 | 4.01 |
| F | ........ | 0.19 | 0.03 | ........ | ........ | ........ |
| $SO_4$ | 3.75 | 5.77 | 31.19 | 0.32 | 6.21 | 4.26 |
| S | ........ | ........ | ........ | ........ | 0.06 | ........ |
| $CO_3$ | 45.57 | 41.91 | 19.15 | 22.38 | 44.76 | 47.45 |
| $PO_4$ | 1.52 | 0.01 | 0.01 | ........ | ........ | ........ |
| $AsO_4$ | 0.04 | ........ | ........ | ........ | ........ | ........ |
| $B_4O_7$ | ........ | 0.16 | ........ | 27.98 | ........ | ........ |
| Na | 35.27 | 38.08 | 32.49 | 33.97 | 41.07 | 40.09 |
| K | 2.88 | 1.20 | 1.35 | 0.48 | ........ | 0.38 |
| $NH_4$ | ........ | ........ | ........ | 0.05 | ........ | ........ |
| Li | ........ | 0.12 | ........ | ........ | ........ | ........ |
| Ca | 2.29 | 0.87 | 2.23 | 0.41 | 0.30 | 0.27 |
| Sr | 0.04 | 0.05 | 0.01 | ........ | ........ | ........ |
| Mg | 1.11 | 0.41 | 0.65 | 0.11 | 0.12 | 0.15 |
| Mn | ........ | ........ | 0.01 | ........ | ........ | ........ |
| Fe | ........ | ........ | 0.02 | ........ | ........ | 0.14 |
| $Fe_2O_3$ | 0.04 | 0.06 | ........ | ........ | ........ | ........ |
| $Al_2O_3$ | ........ | 0.02 | ........ | ........ | ........ | 0.20 |
| $SiO_2$ | 1.32 | 2.30 | 1.34 | 0.73 | 0.85 | 3.05 |
| | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 |
| Salinity, parts per million | 5,249 | 2,614 | 5,431 | 5,096 | 2,069 | 1,668 |

### NOTES RELATING TO ABOVE ANALYSES

*A*. The Grand-Grille spring, Vichy, France. Analysis by J. Bouquet. Small quantity of fluorine present. Temperature 44° C. Issues from Tertiary beds.

*B*. Ojo Caliente spring, near Taos, New Mexico. Analysis by W. F. Hillebrand. Trace of barium and arsenic. In lake beds and gneiss.

*C*. The Sprudel, Carlsbad, Bohemia. Analysis by F. Ragzsky. Contains 0.76 gram free and half-combined $CO_2$ per kilogram. Traces of Br, I, Li, B, Rb, and Cs. Temperature 72° C. In granite.

*D*. Hot water from the Hermann shaft, Sulphur Bank, California. Analysis by W. H. Melville. A little $H_2S$ and a considerable amount of $CO_2$ present. Temperature 80° C. In basalt and sandstone.

*E*. McClelland well, Cass County, Missouri, 45 feet deep, in Carboniferous limestone. Analysis by P. Schweitzer. Contains $H_2S$.

*F*. Artesian water, La Junta, Colorado. Well 386 feet deep. Analysis by W. F. Hillebrand. In Cretaceous beds.

Such waters should attack silicates with great energy; and it is suggested that their strong percentage of sodium may have been leached from walls of the fissures, during the conversion of sodium silicates to potassium silicates as in the process of sericitization, so common in mineral veins. It is certain that these waters are of the utmost importance in the genesis of ore-bearing veins.

An excellent instance of a province of such waters is furnished by the volcanic district of central France.[1] An analysis of the celebrated Vichy Springs is given in the table on page 54. Sodium preponderates as bicarbonate, but smaller quantities of sodium chloride and sulphate are also present. The whole region of the Central Plateau is rich in carbon dioxide, occurring both in springs and as exhalations (for instance, at the Pontgibaud lead-silver mines). The magmatic source of the gas is rarely questioned whatever opinion may be held about the origin of the water.

At the foot of the Erzgebirge, in the Tertiary volcanic region of northern Bohemia, issue a series of hot springs, extending from Teplitz to Carlsbad and Eger. Most of these belong to the class of sodium carbonate waters with free carbon dioxide. They contain an abundance of salts, and in the Teplitz and Bilin springs sodium carbonate predominates. In the Carlsbad (*C* in table on p. 54) and Marienbad springs the sulphuric acid radical is prominent and must largely exist in sodium sulphate. The Carlsbad springs contain fluorine and barium with traces of many rarer metals which are mentioned above.

In the Cordilleran ranges in North America and South America sodium carbonate waters are abundant and always closely connected with areas of Tertiary volcanic activity.

In New Mexico the Ojo Caliente (*B* in analyses on p. 54), Faywood, and Las Vegas springs may be mentioned; in Colorado the Idaho Springs, Middle Park Springs, Poncha Springs, and the water in the Geyser mine at Silver Cliff; in Idaho the Boise Hot Springs. In California sodium carbonate waters are especially abundant and characteristic; they follow the Coast Range from San Diego to Mendocino County and appear to stand in some causal connection with the late Tertiary or Quaternary eruptions of basalt. Some of the waters are clearly admixed with magnesium from the serpentinoid rocks which they have traversed, but in general the type is perfectly distinct. The data on p. 56 are taken from the U. S. Geological Survey, *Bulletin* 32, by A. C. Peale.

An interesting type of these waters is represented by the hot spring of Sulphur Bank (*D* in analyses on p. 54), which contains boron and is depositing cinnabar. On the whole these waters are rich in unusual constituents and have great solvent powers.

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[1] Jacquot and Willm, Les eaux minérales de la France, Paris, 1894.

| Source of water | Salinity, parts per million | Composition and quantity of principal salts |
|---|---|---|
| San Juan Capistrano (T. 50° C.). | 290 | $HNaCO_3 >$ NaCl $> SiO_2$<br>111 105 70 |
| Skaggs Springs (T. 54° C.) ..... | 2,556 | $HNaCO_3 > B_4O_7 > SiO_2$<br>2,083 176 151 |
| Paso Robles Springs (T. 42° C.). | 1,581 | $HNaCO_3 >$ NaCl $> Na_2SO_4$<br>850 469 136 |
| New Almaden Vichy (T. 17° C.). | 7,361 | $HNaCO_3 > CaSO_4 > CaCO_3$<br>3,400 680 544 |
| Napa Soda (T. 17° C.).......... | 1,156 | $HNaCO_3 > MgCO_3 >$ NaCl<br>561 187 85 |
| Pacific Congress (T. 10° C.)..... | 5,678 | $HNaCO_3 >$ NaCl $> CaCO_3$<br>2,091 1,923 289 |
| Ukiah Vichy (T. 34° C.)........ | 4,624 | $HNaCO_3 >$ NaCl $> MgCO_3$<br>3,369 459 374 |

## SODIUM SULPHIDE WATERS

It is believed that in some of the springs already referred to—for example, Steamboat Springs, Nevada—sodium sulphide is present. In the Pyrenees of France and Spain is found a group of springs in which sodium sulphide is constantly present. These springs have a high temperature and a low salinity, containing from 250 to 350 parts per million of salts; they usually issue in crystalline schists or on the contact of the schists with Paleozoic strata. A characteristic spring mentioned among others by Jacquot and Willm[1] contains total $CO_2$ 52, S (in sulphides) 31, Na 97, $CO_3$ 26, Cl 55, and $SiO_2$ 93 parts per million. Some organic matter is present and strong traces of boron, arsenic, copper, etc., are mentioned. There appears to be considerable difficulty in the explanation of this combination on the hypothesis of leaching from the surrounding country rock. Where contaminated by surface water or where locally issuing through Triassic strata they become calcic. By oxidation they acquire hyposulphites.

## SUMMARY

In sedimentary formations, beyond the influences of igneous activity, the waters are of many differing types dependent upon the composition of the rocks traversed. Some contain mainly calcium carbonate; others are of the chloride type, with sodium or calcium as the prevalent base; still others, a very abundant class, are rich in calcium or sodium sulphates; a rarer type is that of the sodium carbonate waters. Naturally many

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[1] Les eaux minérales de la France, Paris, 1894.

waters show a mingling of these types. Most of these waters are cold; many are tepid; few of them are hot. Whether warm or cold, both hydrogen sulphide and carbon dioxide may be present.

In older igneous rocks where the effects of volcanism have subsided the types vary less widely. The ordinary surface waters are always—unless some disturbing influence interferes—of the calcium carbonate type, often with sodium chloride, ferrous and magnesium carbonate, and considerable silica, but low salinity. These waters sometimes, but not often, appear as tepid ascending springs. If the rocks contain iron disulphide the waters may locally contain free sulphuric acid and the sulphates of calcium, aluminum, and iron.

The remaining classes of water in igneous rocks are ascending and confined to regions of recent or Tertiary volcanic activity. They are tepid to hot, though cold waters are also known. They easily fall into two classes: (1) the sodium chloride waters, of which the siliceous "geyser waters" form a sub-class; (2) the sodium carbonate waters, which are generally rich in free carbon dioxide. Transitions between the two classes are plentiful, and such waters may also contain notable amounts of sodium sulphate; of this class the Carlsbad Springs form a prominent example. Acid waters may result from the oxidation of rocks containing sulphides or sulphur; or they may form from the oxidation of ascending hot waters containing $H_2S$; or, near volcanoes, they may develop from admixture of hot acidic emanations with surface waters. It is clear that no acid waters can travel far without becoming alkaline or neutral.

## INTERPRETATION OF WATER ANALYSES

Analyses of waters are usually stated in parts per million of radicals and metals. From this form a calculation will be necessary to ascertain whether the water is alkaline, neutral, or acid.

Stabler[1] has suggested that for this purpose the quantities determined may be multiplied by the reciprocals of the equivalents. The products are called the reacting values. If the water is neutral the reacting values of acids and basic radicals should balance. Palmer,[2] in his method of geological interpretation of water analyses, finds it convenient to express the reacting values in percentages, thus eliminating the factor of concentration. Palmer's classification emphasizes the fact that a solution in which strong acids are exactly balanced with strong bases is relatively inert, whereas one in which either group exceeds the other is relatively

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[1] Herman Stabler, The mineral analysis of water for industrial purposes and its interpretation by the engineer, *Eng. News*, 60, 1908, p. 356. Also, chapter on the industrial application of water analyses in *Water-supply Paper* 274, U. S. Geol. Survey, 1911, pp. 165–181.

[2] Chase Palmer, Geochemical interpretation of water analyses, *Bull.* 479, U. S. Geol. Survey, 1911.

active. It is of special use in showing the relationship and the nature of chemical action of different waters.

*Alkalinity* and *salinity* are the fundamental properties. Salinity is measured by the strong acid radicals ($SO_4$, Cl). If the basic radicals are partly or wholly alkaline metals their proportion of the salinity is said to be *primary*. The remaining salinity due to radicals Ca, Mg, Fe is called *secondary*. If the acid radicals are in excess, *tertiary* salinity or acidity results. The measure of *primary alkalinity* is the excess of alkaline metal radicals over the strong acids; the weak-acid radicals $CO_3$ and $HCO_3$ which balance any excess of the alkaline earth metals over the stronger acids produce *secondary alkalinity*.[1]

In spite of an objectionable terminology Palmer's method furnishes a convenient basis for comparative study but as a classification of natural waters it is unwieldy and uncertain. It is not always a safe guide to the geological history of the water.

The constants used in converting grains per gallon to parts per million and vice versa are as follows:

1 grain per U. S. gallon = 17.138 parts per million
1 grain per Imperial gallon = 14.285 parts per million
1 part per million = 0.0588 grain per U. S. gallon
1 part per million = 0.07 grain per Imperial gallon

## MINE WATERS[2]

### Chloride Waters

Mine waters consist as a rule of the normal ground waters of the region modified by the salts resulting from the decomposition of the ores. In districts free from sulphides the mine waters of the upper levels are ordinarily of the potable calcium carbonate type, but as depth is gained they acquire much more chlorides. Examples of such waters are found in the iron and copper mines of Michigan. In the following table analyses 1 and 2 represent average waters of the upper circulation; 3 and 4 give the composition of the deep waters of the same region, which differ very materially from the shallow waters and contain an abundance of calcium chloride. The deep waters in the copper region contain some copper, zinc, nickel, and traces of boron.

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[1] *Cf.* F. W. Clarke, Data of geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 63.

G. S. Rogers, The interpretation of water analyses by the geologist, *Econ. Geol.*, 12, 1917, pp. 56–88.

[2] W. H. Emmons and G. L. Harrington, A comparison of the waters of mines and of hot springs, *Econ. Geol.*, 8, 1913, pp. 653–659.

E. T. Hodge, The composition of waters of sulphide ores, *Econ. Geol.* 10, 1915, pp. 123–139.

W. H. Emmons, The enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917, pp. 83–106.

ANALYSES OF MINE WATERS FROM UPPER AND LOWER LEVELS OF IRON AND COPPER MINES OF MICHIGAN
(Parts per Million)

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Cl | 3.5 | 6.00 | 25,360 | 176,027 |
| Br | ... | ... | ... | 2,200 |
| $CO_3$ | 24.2 | 41.60 | Not det. | ... |
| $SO_4$ | 18.8 | 12.10 | 1,045 | 110 |
| Ca | 12.9 | 15.20 | 7,902 | 86,478 |
| Mg | 2.0 | 9.60 | 566 | 20 |
| Na | 11.3 | 4.00 | 7,290 | 15,188 |
| K | ... | 1.50 | | 411 |
| $Al_2O_3$ | 4.0 | 0.60 | 700 | 10 |
| $Fe_2O_3$ | | 1.23 | | |
| Mn | ... | ... | ... | 4 |
| Cu | ... | ... | ... | 16 |
| Ni | ... | ... | ... | 6 |
| $SiO_2$ | 14.5 | 8.43 | ... | 20 |
| Loss | 2.8 | ... | ... | ... |
| Total | 94.0 | 100.26 | 42,863 | 280,490 |

1. Mass copper mine, Michigan. Water from upper levels. Analyst, Dearborn Chemical Works, Chicago. A. C. Lane, Mine waters. Thirteenth Annual Meeting, Lake Superior Mining Institute, June, 1908, p. 31.
2. Newport mine, Gogebic district, Michigan. Water from upper levels. Analyst, R. D. Hall, Free $CO_2$ 18.0. Residue dried at 100° C. 108.30. Van Hise and Leith, *Mon.* 52, U. S. Geol. Survey, 1911, p. 543.
3. Republic iron mine. Seventeenth level. Temperature 57° F. Analyst, G. Fernekes. Calculates to NaCl 18,510 and $CaCl_2$ 21,800. A. C. Lane, *op. cit.*, p. 10.
4. Quincy copper mine. Drippings on fifty-fifth level. Similar waters from the Calumet and Hecla mine also contain some zinc. Analyst, G. Fernekes. Chiefly calcium and sodium chlorides and sodium bromide. *Op. cit.*, p. 48. Trace boron and strontium. No barium or carbon dioxide.

Salt waters, containing mainly sodium chloride to the amount of several per cent, are reported from the Kalgoorlie mines in Western Australia, where they began to come in at water level, a few hundred feet below the surface. At the Great Boulder Proprietary the water contained 9 per cent of sodium chloride and also much magnesium chloride.[1] Similar waters came into the deep levels of the Bendigo mines.[2] Sulphides are not abundant in these mines. The analysis of a sample taken 4,280

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[1] T. A. Rickard, Formation of bonanzas, *Trans.*, Am. Inst. Min. Eng., 31, 1901, pp. 198–220.

[2] W. J. Rickard, Deep mining at Bendigo, *Mining Mag.*, 1910, p. 281.

feet below the surface, in the Victoria Reef quartz, where the temperature of the water is 114° F., is as follows, contained in parts per million:

$NaCl$, 1,308.45; $Na_2SO_4$, 75.79; $Na_2CO_3$, 37.18; $CaCO_3$, 124.41; $MgCO_3$, 45.76; $SiO_2$, 21.45; $(Al,Fe)_2O_3$, 2.86; total, 1,615.90.

According to T. A. Rickard[1] the mine water at Mammoth, Pinal County, Arizona, contains 86 parts per million of sodium chloride, and that from Stratton's Independence mine, at Cripple Creek, Colorado, 51 parts of the same salt.

## Carbonate Waters

The mine waters from the Wardner lead mines, in the Coeur d'Alene district, Idaho, are rich in ferrous carbonate (from siderite in the ore) and deposit abundant limonite. A sample from the Reed level, Bunker Hill and Sullivan mine, showed 70 parts per million of total solids, chiefly bicarbonate and sulphate of calcium.[2]

A number of analyses of waters from the lead mines of southeastern Missouri are given by E. R. Buckley.[3] The waters come from the La Motte sandstone and Bonneterre dolomite, generally at depths of a few hundred feet. The total solids are at most 500 parts per million, of which 200 parts or more are calculated as calcium-magnesium carbonates. The sulphates, calculated as the magnesium salt, are at most 200 parts per million, while sodium chloride averages only 50 parts. Silica is low. All contain a little lead, at most 1 part per million, calculated as lead sulphate, and generally a trace of zinc. They are weak waters mainly on account of the small amount of pyrite in the deposit.

## Sulphate Waters

**Oxidation of Pyrite.**—Where pyrite is present in notable quantities its oxidation materially changes the composition of the waters. The sulphuric acid radical increases rapidly and displaces the equilibrium so that the normal calcium carbonate waters are changed into those containing mainly calcium sulphate. When the free sulphuric acid increases still further the water becomes rich in the sulphates of aluminum (by the decomposition of sericite and other silicates) and iron, the latter present as both ferrous and ferric sulphate. Free hydrochloric acid is sometimes present. In waters above or at the water level these sulphates may be present in large quantities. Below the water level free acid is rarely found and the sulphate of aluminum is absent. The iron is present as ferrous sulphate and diminishes in quantity with increasing depth. The characteristic calcium sulphate waters persist for wide spaces around

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[1] *Trans.*, Am. Inst. Min. Eng., 31, 1901, pp. 198–220.

[2] F. L. Ransome, *Prof. Paper* 62, U. S. Geol. Survey, 1908.

[3] Missouri Bur. Geol. and Mines, 9, pt. 1, 1909, p. 249.

pyritic deposits and also reach considerable depths. Besides the sulphates mentioned, the mine waters of the oxidized zone contain almost all the metals occurring in the deposit. Zinc sulphate is especially abundant; copper sulphate is usually present, lead much more rarely; arsenic is common and antimony rare.

The waters of coal mines show plainly the result of the oxidation of the pyrite and marcasite occurring in the beds. Such waters are often rich in the sulphates of ammonium, calcium, iron, and aluminum, and even in free sulphuric acid. In the drainage from the mines the iron appears as ferrous sulphate, from which, by oxidation, ferric sulphate is formed. Coarsely crushed coal washed with distilled water is said to yield free sulphuric acid in the filtrate. Mine waters from coal mines occasionally contain zinc, copper, cobalt, and nickel. A water from the coal mine of the Dravo-Doyle Company, in Pennsylvania, showed, as analyzed by the Pittsburgh testing laboratory of the Bureau of Mines:

| | |
|---|---|
| Free $H_2SO_4$ | 117 parts per million |
| $Fe_2(SO_4)_3$ | 4,970 parts per million |
| $Al_2(SO_4)_3$ | 140 parts per million |
| $FeSO_4$ | 54 parts per million |

More or less of the sulphates of calcium and magnesium are also usually present.

ANALYSIS OF MINE WATERS FROM THE COMSTOCK LODE
(Parts per Million)

| | 1 | 2 | 3 |
|---|---|---|---|
| Cl | 1.27 | 19.00 | 127.60 |
| $SO_4$ | 380.38 | 474.00 | 209,100.00 |
| $CO_3$ | 115.03 | 20.45 | ........ |
| K | 8.39 | 53.40 | ........ |
| Na | 57.13 | 132.00 | 535.00 |
| Ca | 148.10 | 100.10 | 1,286.00 |
| Mg | 154.03 | 5.88 | 6,590.00 |
| Al | ........ | 1.37 | 9,670.00 |
| Mn | ........ | ........ | 885.10 |
| Cu | ........ | ........ | 147.50 |
| $SiO_2$ | 30.50 | 133.40 | 616.00 |
| Fe | ........ | 6.33 | 5,025.02 |
| H | ........ | ........ | 2,575.00 |
| Total salinity | 894.83 | 945.93 | 236,557.22 |

1. Water from the 600-foot level of the Savage mine. G. F. Becker, *Mon.* 3, U. S. Geol. Survey, 1882, p. 152.

2. Waters from the C. and C. shaft at the 2,250-foot level. John A. Reid, *Bull.* California Univ. Dept. Geology, 4, 1905, pp. 177–199. Contains 2.92 milligrams of silver and 0.298 milligram of gold per ton of solution.

3. Surface water from the Central tunnel. *Idem.*

**Examples.**—Analyses (page 61) of the Comstock waters, Nevada, by J. A. Reid well illustrate the occurrence of sulphate waters. No. 3 is a concentrated sulphate water from the oxidized zone; Nos. 1 and 2 are deeper hot waters, resulting from the reaction between an ascending sodium-carbonate water and sulphuric acid from the upper zones. The ores contain mainly gold and silver and are not rich in pyrite.

ANALYSES OF MINE WATERS

| | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| Cl | 2.7 | 12.4 | 6.8 | 13.0 | 0.8 |
| $HCO_3$ | .......... | .......... | 13.5 | .......... | 210.0 |
| $SO_4$ | 6,153.2 | 124.8 | 406.5 | 2,672.0 | 1,088.0 |
| $SiO_2$ | 107.6 | 18.0 | 23.0 | 47.7 | 64.0 |
| $AsO_4$ | .......... | .......... | Trace. | .......... | .......... |
| Ca | 345.3 | 46.4 | 151.2 | 132.5 | 564.7 |
| Mg | 25.2 | 14.5 | 28.2 | 61.6 | 22.4 |
| Na | 49.9 | .......... | 16.2 | 39.6 | 51.2 |
| K | 0.5 | .......... | 7.1 | 13.1 | 7.1 |
| $Fe''$ | 474.6 | 6.6 | 1.8 | .......... | .......... |
| $Fe'''$ | | | .......... | 159.8 | .......... |
| Mn | 1.7 | .......... | 0.5 | 12.0 | .......... |
| Zn | 2,412.0 | 8.9 | 0.3 | 852.0 | .......... |
| Cd | 9.0 | .......... | .......... | 41.1 | .......... |
| Al | 142.1 | .......... | .......... | 83.5 | .......... |
| Cu | 3.7 | .......... | Trace. | 59.1 | .......... |
| Co + Ni | .......... | .......... | .......... | 0.5 | .......... |
| Sn | .......... | .......... | .......... | 17.0 | .......... |
| | 9,727.5 | 231.6 | 655.1 | 4,204.5 | 2,008.2 |

1. Water from Alabama Coon mine, Joplin, Missouri. H. N. Stokes, analyst.
2. Water from the Rothschönberger Stolln, Freiberg, Saxony, at point of discharge. Analysis by Frenzel. Recalculated by F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 647.
3. Water from 2,200-foot level, Green Mountain mine, Butte, Montana, remote from veins. W. F. Hillebrand, analyst.
4. Water from 1,200-foot level, crosscut St. Lawrence, Butte, Montana. Tin possibly accidentally introduced? Faintly acid. W. F. Hillebrand, analyst. $Fe''$ probably changed to $Fe'''$ during exposure to air.
5. Cripple Creek, Colorado, water from El Paso Tunnel draining the lowest workings. R. C. Wells, analyst. Fe + Al, etc., 0.6. Free $CO_2$ trace.

Some more recent analyses of Comstock waters are discussed by E. S. Bastin.[1] He also presents analyses of mine waters from the San Juan districts, Colorado.

Some of the mine waters of the Joplin zinc region, where the deposits contain, besides sphalerite and galena, some pyrite or marcasite, are extremely rich in zinc sulphate and contain also the sulphates of iron and aluminum. (See analysis No. 1 in the table, p. 62.)

The water of the Rothschönberger tunnel, draining the mines at Freiberg, Saxony, is a good example of a dilute mine water which has traversed the old workings of veins carrying pyrite, galena, and sphalerite. (See analysis No. 2, p. 62.)

ANALYSES OF MINE WATERS FROM DUCKTOWN, TENNESSEE, AND CANANEA, MEXICO

G. W. Hawley, Analyst (Cananea); R. C. Wells (Ducktown)

(Parts per Million)

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| SO$_4$ | 5,064 | 4,457 | 415.8 | 476.8 |
| Cl | Not determined | 22 | 0.7 | 0.4 |
| SiO$_2$ | 76 | 56 | 47.0 | 49.9 |
| Al | Not determined | 22 | 14.5 | 19.1 |
| Fe'' | 305 | 524 | 71.4 | 89.2 |
| Fe''' | ... | ... | 20.3 | 55.9 |
| Ca | 436 | 753 | 19.7 | 30.4 |
| Mg | 61 | 86 | 5.2 | 6.2 |
| Mn | 236 | 153 | 0.2 | 0.1 |
| Zn | Not determined | 252 | 2.4 | 2.9 |
| Cu | 1,659 | 60 | 28.1 | 11.0 |
| K | Not determined | 198 | 2.7 | 2.2 |
| Na | | | 5.2 | 5.5 |
| Acidity as (H$_2$SO$_4$) | 970 | Nil | 210.2 | 97.5 |

1. Cananea 300-foot level.
2. Cananea 900-foot level.
3. Calloway shaft, Ducktown, Tennessee, standing water, at water level, 90 feet below surface.
4. Calloway shaft, Ducktown, Tennessee, standing water 37 feet below water level.

The same principle is illustrated by the analyses of two waters from the mines at Butte, Montana. No. 3 is from a deep level, but rather far from the principal vein system; No. 4 is from the 1,200-foot level in one of the principal mines; it has acquired the habit of a water of the upper oxidized zone because the water level has been artificially lowered and the oxidation of the pyrite is progressing rapidly. No. 5 is a deep water from Cripple Creek (p. 62).

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[1] *Bull.* 735, U. S. Geol. Survey, 1923, pp. 57, 67, and 107.

A. C. Lawson[1] describes the mine water from the Ruth mine 335 feet below the surface, in the chalcocite blanket in the porphyry of Ely, Nevada. The temperature was 16° C., decidedly higher than the average annual temperature of the region. The total solids were 1,094 parts per million, of which 359 parts were calculated as calcium sulphate, 130 as magnesium sulphate, 93 as alkaline chlorides, 160 as ferrous sulphate, and 7 as ferric sulphate.

Four analyses of the mine waters at the copper mines of Cananea, Mexico,[2] and Ducktown, Tenn.,[3] illustrate the change as depth is gained. The Cananea waters come from an upper and a deeper level and have percolated through a sericitized rock with a considerable amount of chalcocite and pyrite. The Ducktown waters are also acid and are taken from the upper part of the groundwater (p. 63).

The deeper waters contain much more calcium sulphate as well as ferrous sulphate, but much less copper. Chalcocite probably reduces the ferric sulphate to ferrous.

In mines containing molybdenite the waters from the oxidized zone may contain this metal usually as the blue oxide. Such a water from a tunnel in the Gilpin County gold mines contained nearly 8 grams per liter of a molybdenum oxide. The water had a deep greenish-blue color.

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[1] *Bull.* California Univ. Dept. Geol., 4, 1906, pp. 287–357.

[2] W. H. Emmons, Enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917, p. 87.

[3] W. H. Emmons and F. B. Laney, *Prof. Paper* 139, *idem*, 1926, p. 69.

## CHAPTER VI

# THE SPRING DEPOSITS AT THE SURFACE

Dissolving and precipitating processes are in continual progress in the underground waters. As will be seen later the conclusion seems justified that a majority of mineral deposits have resulted from reactions in the underground solutions. Only at the surface or in mines, however, is it possible to study the actual progress of these chemical changes, and great interest, therefore, attaches to the deposits formed by the natural waters where they issue as springs from their underground path. The precipitation taking place in rivers, lakes and seas will be described in later chapters.

On the whole the composition of the material deposited by springs is simple. Three main divisions are recognized: deposits of *limonite* (iron hydroxide), *calcium carbonate,* and *silica.* Mixtures of two or all of these substances are frequently observed. The deposits are known as (1) ochers, (2) tufas, travertines, or calcareous sinters, (3) sinters or siliceous sinters.

The precipitation is in part due to oxidation, cooling, or escape of carbon dioxide but algæ and micro-organisms frequently aid by secreting silica jelly, calcium carbonate, colloidal ferric hydroxides, or manganese dioxide.[1]

**Deposits of Limonite and Calcium Carbonate.**—Limonite is frequently deposited by superficial meteoric waters which contain ferrous carbonate and ferrous sulphate. Many such ochers contain a little manganese and traces of copper, arsenic, nickel, and cobalt. Other waters also deposit some limonite so that many sinters and tufas are stained by this compound. Analyses of such ochers are quoted by F. W. Clarke.[2]

Calcium carbonate is probably the most common spring deposit, though the ordinary dilute surface waters rarely are able to form important precipitates. Hot carbonated waters issuing from limestone often deposit large masses of such tufa, covering many acres with thick terraced beds. The Mammoth Hot Springs in the Yellowstone Park offer a beautiful example of such tufa. The precipitates are almost pure calcium carbonate with a little magnesium carbonate. In many of these springs

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[1] W. H. Weed, *Ninth Ann. Rept.,* U. S. Geol. Survey, 1889, pp. 613–676.
H. Molisch, Die Eisenbakterien, Jena, 1910.
*Jour.* Am. Chem. Soc., 92, pt. 2, 1907, p. 888, abstract.

[2] Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 206.

the calcium carbonate is the least soluble constituent which remains after the others have been carried away. Thus, the sodium chloride springs of Glenwood, Colorado, yield a considerable deposit, and the sodium carbonate springs of Ojo Caliente, New Mexico, which are very poor in calcium, deposited at their former point of issue a porous tufa containing over 90 per cent of calcium carbonate. This carbonate is no doubt deposited in crystalline form, though it is usually fine grained. Such deposits are not always calcite, for the presence of aragonite has been proved in many spring deposits, for instance those of Hammam Meskoutine, in Algeria, and of Carlsbad, in Bohemia.[1]

**Deposits of Silica.**—At hot springs containing much silica, this substance is abundantly precipitated because of evaporation, through mixture with other waters, or, according to W. H. Weed, by the action of certain hot-water algæ. The material is deposited as a colloid gel which subsequently crystallizes to opaline or chalcedonic silica. Such sinters are formed by the hot springs of the Yellowstone Park and may contain up to 95 per cent of silica. Sodium is often present as chloride or carbonate. The Steamboat Springs of Nevada[2] deposit a sinter of pure silica or mixtures of calcium carbonate and silica, the latter being present as chalcedony, or small crystals of quartz (see Fig. 5). This sinter contains weighable quantities of sulphides of mercury, lead, copper, arsenic, and antimony; the presence of gold and silver was also determined, and traces of manganese, zinc, cobalt, and nickel were found. Antimony sulphide ($Sb_2S_3$) is deposited as the amorphous "metastibnite" in quantities large enough to color the sinter red in places. In a shaft sunk into the gravel immediately adjoining the granite hill from which the springs issue, Lindgren[3] discovered delicate crystals of stibnite covering the pebbles and associated with thin crusts of black opal and grains of pyrite or marcasite.

The opaline sinter (Fig. 7) of the Yellowstone Park often contains arsenic, especially in the form of scorodite ($FeAsO_4.2H_2O$), and near one of the springs which was impregnated with pyrite Weed noted rhyolite that contained traces of gold and silver. On the whole, however, the Yellowstone spring deposits are poor in the rarer metals. The same author, associated with Pirsson,[4] reports the occurrence of orpiment and realgar with native sulphur in a siliceous sinter from the Norris geyser basin. De Launay mentions a deposit containing orpiment at St. Nectaire, Puy-de-Dôme, France.

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[1] H. Vater, *Zeitschr. Kryst. u. Min.*, 35, 1902, p. 149.

[2] G. F. Becker, The quicksilver deposits of the Pacific coast, *Mon.* 13, U. S. Geol. Survey, 1888, p. 341.

[3] W. Lindgren, *Trans.*, Am. Inst.Min. Eng., 36, 1906, pp. 27–36.

[4] Occurrence of sulphur, orpiment, and realgar in the Yellowstone National Park, *Am. Jour. Sci.*, 3d ser., 42, 1891, pp. 401–405.

FIG. 5.—Section of chalcedonic spring deposits from Steamboat Springs, Nevada, showing colloform structure. White areas microcrystalline quartz. Magnified 25 diameters.

FIG. 6.—Section of chalcedonic spring deposits, from DeLamar, Idaho, showing vegetable remains. Magnified 35 diameters. Ordinary light.

A calcareous sinter deposited by an ascending sodium carbonate spring in the Geyser mine, Silver Cliff, Colorado, on the 2,000-foot level, yielded traces of lead, copper, zinc, nickel, and cobalt. At Hammam Meskoutine, in Algeria, a similar spring, according to Daubrée, deposits tufas and pisolitic sinters in which, in the concretions, shells of calcium carbonate alternate with shells of pyrite; strontianite is deposited by the same spring.

Quicksilver, gold, and silver have been recognized in the spring deposits of the geyser districts in New Zealand. From the Whakarewarewa hot springs at Roturoa (sodium chloride-silica type) sinters[1] have been analyzed which yielded nearly 5 ounces of silver and about $1 in gold per ton.

FIG. 7.—Section of opaline silica showing colloform structure. Yellowstone National Park. Magnified 32 diameters. (*After H. C. Boydell, "Economic Geology."*)

At De Lamar, Idaho, Lindgren found spring deposits of flinty chalcedony in rhyolite, which included casts of vegetable remains and yielded traces of gold and silver[2] (Fig. 6).

It has been shown that springs, hot or cold, may deposit limonite in abundance, with arsenic, manganese, and traces of other metals; and it is likewise proved that the carbonate and silica sinters of hot springs, particularly those of the NaCl or $Na_2CO_3$ type, contain small quantities of the rarer metals, including gold, silver, copper, lead, zinc, antimony, arsenic, tin, and quicksilver. In very few instances has commercial ore been

[1] J. M. Bell, *First Ann. Rept.*, N. Z. Geol. Surv., 1907, p. 100.

[2] The gold and silver veins of Silver City, etc., *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, p. 187.

obtained from spring deposits at the surface. Quicksilver ores have been mined in New Zealand and ores of iron and manganese have been utilized in rare instances. The evidence that such waters have formed workable ore deposits is therefore strong but hardly conclusive; the remarkable poverty in metals of the deposits of the springs in the Yellowstone National Park, for instance, will to many seem an argument against the hydrothermal theory of the genesis of ore deposits.

**Deposits of Other Gangue Minerals.**[1]—Calcite, quartz, chalcedony, and opal are common products of deposition at the surface, but besides these the mineral deposits often contain such minerals as barite, ankerite and siderite, fluorite, and more rarely gypsum, strontianite, celestite, and zeolites.

Fluorine is present in traces in many waters, both superficial and deep, but appears in larger quantities in waters of the sodium carbonate type. Few authenticated instances of actual deposition of fluorite by springs are recorded; the substance rarely occurs in crystallized form and the chemists have probably often neglected to test the sinters for fluorine. The Carlsbad Springs deposit a pisolitic sinter of aragonite and calcite. According to Berzelius[2] and later chemists this contains a notable quantity of calcium fluoride. A limonitic variety of the spring deposit yielded 0.272 per cent arsenic.[3] The following analyses also demonstrate that various phosphates may be precipitated as well as the carbonates of iron and strontium.

COMPOSITION OF DEPOSITS OF CARLSBAD SPRINGS
Berzelius, Analyst

| | *A* | *B* |
|---|---|---|
| Ferrous carbonate | 12.13 | ........ |
| Ferric oxide | 19.35 | 0.43 |
| Manganese oxide | ........ | trace |
| Calcium carbonate | 53.20 | 96.47 |
| Strontium carbonate | ........ | 0.30 |
| Basic ferric phosphate | 1.77 | ........ |
| Aluminum phosphate | 0.60 | 0.10 |
| Calcium phosphate | ........ | 0.06 |
| Calcium fluoride | ........ | 0.99 |
| Silica | 3.95 | ........ |
| Water | 9.00 | 1.59 |

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[1] The best résumé of the older data regarding spring deposits are found in Roth, Allgemeine und Chemische Geologie, 1, 1879, pp. 564–596.

[2] *Pogg. Ann.*, 74, 1823, p. 149.

[3] Blum and Leddin, *Am. Chem. Pharm.*, 73, 1850, p. 217.

At Plombières, in the Vosges, springs with a temperature of 70° C. issue from granite. They have a low salinity (360 parts per million) and contain mainly sodium sulphate and silica, also traces of arsenic and fluorine. The derivation of these salts is doubtful and the springs are apparently not directly related to volcanic rocks. They issue from well-defined fissure veins containing quartz and fluorite, and Daubrée[1] found that the waters had actually deposited calcite, aragonite, and fluorite. The bricks and cements used by the Romans 2,000 years ago in the construction of the baths at Plombières were found to contain zeolites, chiefly apophyllite (containing fluorine) and chabazite, with opal and chalcedony. Chabazite is also reported by Daubrée as deposited from springs at Luxeuil and at Bourbonne-les-Bains, which have a temperature of 46° and 68° C. respectively.

Crystals of gypsum occur commonly near springs charged with calcium sulphate. Weed[2] has described how the Hunter Hot Springs, near Livingston, Montana, deposit this mineral in fractures in Tertiary sandstone; stilbite, a zeolite, is forming with the gypsum. The springs have a temperature of 64° C. and are weak mineral waters. The presence of stilbite has also been noted by Weed[3] in the vein-like deposits, containing gold and silver, believed to be made by the present Hot Springs at Boulder, Montana; the stilbite occurs in the predominating quartz, chalcedony and calcite. Lindgren noted the presence of a little adularia in the material.

Hewett[4] describes a mildly alkaline spring in basalt, which appears to deposit calcite, chabazite, mesolite, analcite, stilbite, and opal.

According to Lindgren[5] a spring deposit in New Mexico contains about 89.60 per cent of calcium carbonate and 0.9 per cent of calcium fluoride. There are no springs now at this place, but it is probable that the sodium carbonate water of Ojo Caliente, a short distance lower down in the valley, formerly issued here. As shown by an analysis on page 54, the water contains a notable amount of fluorine. A vein of white crystalline fluorite is opened by a shaft close by the calcareous tufa and is believed to have been formed by the same waters. Both tufa and vein contain traces of gold and silver, and a few crystals of barite were observed in the vein material. W. H. Emmons and E. S. Larsen[6] have described

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[1] Les eaux souterraines, 3, Paris, 1887, p. 31.

[2] Economic value of hot springs and hot springs deposits, *Bull.* 260, U. S. Geol. Survey, 1904, pp. 298–604.

[3] Mineral vein formation at Boulder Hot Springs, Montana, *Twenty-first Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1899–1900, pp. 233–255.

[4] D. F. Hewett, E. V. Shannon, and F. A. Gonyer, Zeolites from Ritter Hot Springs, Grant County, Oregon, *Proc.*, U. S. Nat. Mus., 73 (16), 1928, p. 18.

[5] *Econ Geol.*, 5, 1910, pp. 22–27.

[6] *Econ. Geol.*, 8, 1913, pp. 235–246.

a similar case from Wagon Wheel Gap, Colorado. The travertine contains in per cent 0.22 F, 0.007 ZnO, and 0.045 BaO. Barite, fluorite, pyrite, opal, and jaspery silica are deposited. Small amounts of gold and silver are contained in the replaced rock close by.

Veins and replacements of fluorite in quartz porphyry and Cretaceous sandstone near the sodium carbonate springs of Teplitz, Bohemia, have been described by J. E. Hibsch[1] and the evidence is convincing that fluorite was deposited by these thermal waters.

Barite is deposited far more abundantly than fluorite. As shown above, many carbonate and even sulphate waters contain a little barium. It has been proved that alkaline bicarbonates with an excess of carbon dioxide can hold barium in solution, notwithstanding the presence of sulphates; sodium chloride and other salts also seem to retard the formation of barium sulphate. Haidinger observed that barite was deposited by the hot waters at Carlsbad[2], and Becke noted the same at Teplitz.[3] At Idaho Springs, in Colorado, a hot sodium carbonate water issues from granite rocks, and barite crystals are found in abundance in a cellular and decomposed dike rock at the mouth of the spring. Spurr[4] shows that the barium is contained in this dike rock and believes that the barite resulted from the reaction of the water on the rock.

Barium is, however, far more commonly contained in sodium chloride waters, particularly in the brines of sedimentary strata. Many writers record the deposition of barite from such waters, and it is probable that wherever this mineral appears in large quantities in mineral deposits waters of this type have been active.

An excellent example is reported from a mine near Clausthal, Germany,[5] where a spring of strong brine, which undoubtedly derived its salts from sedimentary strata, was encountered at a depth of 1,200 feet; this brine contained in grams per liter 67.555 sodium chloride, 10.509 calcium chloride, 4.360 magnesium chloride, 0.350 potassium chloride, 0.314 barium chloride, and 0.854 strontium chloride. Where it mingled with the ordinary mine waters, which carried sulphates, abundant precipitation of barium and strontium sulphate took place in the pumps and elsewhere, so that within a few years the capacity of the pipes became much reduced by this deposit. This compact material contained 92.44 per cent barium sulphate and 4.32 per cent strontium sulphate. The

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[1] *Tschermak's min u. petr. Mitt.*, 25, 1906, pp. 482–488.

[2] *Jahrb.* K. k. Reichsanstalt, 5, 1854, p. 142.

See also R. Delkeskamp, Entstehung und Wegführung des Baryts, *Notizblatt d. Ver. f. Erdkunde* (Darmstadt) (4), 21, 1900, pp. 55–83.

[3] *Tschermak's min u. petr. Mitt.*, 5, 1883, p. 115.

[4] J. E. Spurr, *Prof. Paper* 63, U. S. Geol. Survey, 1908, p. 165.

[5] *Lattermann*, Die Lautenthaler Soolquelle und ihre Absätze. *Jahrb.* Preuss. geol. Landesanstalt, 1888, pp. 259–283. Ref. in Stelzner and Bergeat, Die Erzlagerstätten, II, 1905–1906, p. 1218.

reaction is believed to be retarded by the presence of sodium and magnesium chlorides.

According to P. Krusch,[1] barite is precipitated in the pumps at some Westphalian coal mines by a similar reaction between strong salt brine from the Triassic sandstones and potable water with sulphates, each ascending on separate faults and each deriving its contents of salts from sedimentary strata. Veins of barite with small amounts of galena, pyrite, etc., appear in the Carboniferous and in the Devonian. At a lower horizon quartz veins contain galena and sphalerite, both kinds of deposits having been made, according to Krusch, by these saline waters. Similar deposits of barite in the pipes of the pumping apparatus have been described from English coal mines.[2]

An account of Headden[3] describes an interesting group of springs on the North Fork of the Gunnison River, Delta County, Colorado. They are cold, but contain free carbon dioxide and a little hydrogen sulphide and are essentially sodium chloride waters. At least one spring contains barium and all of them yield a little strontium. The Drinking Spring has a total salinity of about 1,656 parts per million. Small quantities of calcium, potassium, magnesium, barium (0.0132 gram per liter), strontium (0.0066 gram per liter), lithium, manganese, ammonia, iron, aluminum, also a trace of zinc, are present in the order stated; also sulphuric acid radical (0.6254 gram per liter), silica, boron, and bromine, the latter three in very small amounts. The spring deposits a calcium carbonate sinter, which was found to contain 5.42 per cent barite, but no gypsum or sulphur.

Ferrous carbonate, or siderite, is sometimes observed, as in the Carlsbad "Sprudelstein" and in deposits of limonite formed in bogs and peat. Deposits of magnesium minerals are rare. H. Leitmeier[4] describes a deposit of hydrous carbonate of magnesia from the springs of Lohitsch in Styria; many springs, especially those whose waters have traversed sedimentary beds, contain organic matter and are probably competent to deposit hydrocarbons.

Siebenthal[5] found that at the Sulphur Springs of Arkansas, a sodium chloride water with some $CO_2$ and $H_2S$, deposited small quartz crystals, calcareous concretions, and crystallized pyrite. The dried sediment contained 0.4 per cent Zn with much less Pb and Ca; also 2 per cent Fe, mostly as sulphide. He considers that the zinc-lead deposits of the region were formed by such waters.

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[1] *Monatsberichte* Deutsch. geol. Gesell., 1904, p. 36; Ref. in *Zeitschr. prakt. Geol.*, 12, 1904, p. 252.

[2] J. T. Dunn, *Chem. News*, 35, 1877, p. 140.

[3] The Doughty Springs, etc., *Proc.*, Colo. Sci. Soc., 8, 1905, pp. 1–30.

[4] *Zeitschr. Krystall. u. Min.* 47, 1909, p. 104.

[5] C. E. Siebenthal, Spring deposits at Sulphur Springs, Arkansas, *Econ. Geol.*, 9, 1914, pp. 758–767.

The more common gangue minerals in certain classes of veins are thus deposited by spring waters, particularly by the warm sodium chloride and sodium carbonate springs. There are, of course, a great number of gangue minerals like tourmaline, garnet, feldspars, and similar silicates which cannot be expected to develop in water under the conditions of temperature and pressure prevailing at the surface.

**Summary.**—The deposits of ascending springs of undoubted meteoric origin contain opal, chalcedony, calcium carbonate, limonite, hydroxide of manganese, barite, siderite, and pyrite. They often deposit sulphur by the oxidation of hydrogen sulphide. The ochery deposits very frequently yield small quantities of arsenic, copper, lead, zinc, nickel, and cobalt.

The springs of the sodium carbonate and sodium chloride-silica type in volcanic regions yield abundant deposits of opal, chalcedony, quartz, calcium carbonate, limonite, barite, siderite, sometimes also pyrite. They also deposit fluorite which is rarely if ever found in the sinters of meteoric waters and yield smaller quantities of quicksilver, antimony, arsenic, lead, copper, zinc, tin, silver, and gold. The rarer metals are thus more prominent and the waters are particularly characterized by a relative abundance of borates, arsenates, and fluorides.

The list of recognizable minerals deposited by springs at the surface is as follows: Sulphur, quartz, opal, chalcedony, limonite, wad, psilomelane[1] calcite, aragonite, siderite, strontianite, barite,[2] gypsum, celestite, fluorite, scorodite, pyrite, marcasite, realgar, orpiment, cinnabar, stibnite, chabazite, apophyllite, and stilbite, probably also adularia.[3]

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[1] Sometimes containing tungsten. See W. Lindgren, *Econ. Geol.*, 17, 1922, pp. 201–206. In Cuba, Hewett found a hydrous calcium-manganese silicate called orientite which he considers to have been deposited by hot springs. *Am. Jour. Sci.*, 5th ser., 1, 1921, p. 491.

[2] Plumbiferous barite (Hokutolite), an isomorphous mixture of barite and anglesite, has been described as a deposit from hot springs in Japan by R. Ohashi, *Mineralog. Mag.*, 19, 1920, pp. 73–76. Also from hot springs, in Formosa, by I. Suganuma, Chem. Soc. Japan, *Bull.* 3, 1928.

[3] A. L. Day describes (Carnegie Inst. Wash. *Yearbook*, 28, 1929, pp. 76–79) a 400 feet deep drill hole near the Old Faithful Geyser basin in Yellowstone Park which penetrated hot spring deposits and reached rhyolite and obsidian. The material brought up contained secondary quartz and calcite, also some chlorite and specularite as well as much (probably secondary) orthoclase (adularia). The temperature was 180° C., the hydrostatic head sufficing to hold down the steam pressure.

## CHAPTER VII

## RELATIONS OF MINERAL DEPOSITS TO MINERAL SPRINGS

The deposition of many valuable minerals can be directly observed in nature; limonite, for instance, from the evaporation of water containing iron, or from precipitation in bogs and lakes; sulphur by the decomposition of hydrogen sulphide dissolved in water; residual deposits of limonite, nickel silicates, and pyrolusite by the decomposition of rocks by meteoric waters; common salt and borax by the evaporation of lake waters. A large class of deposits, such as the deep-seated veins containing metals and ores developing near intrusive contacts, we can never hope to observe in nature in the process of formation.

Ascending mineral springs are not uncommon in mineral deposits, particularly in those which follow fissures, but caution must be used in attributing a genetic rôle to these waters. If we find such a spring in a contact-metamorphic deposit or in a vein of deep-seated origin, as a cassiterite vein, it would be unlikely indeed that this spring had anything to do with the formation of the deposit, for it could scarcely be assumed that the circulation of underground waters could be maintained in the same path during the many vicissitudes of deep erosion, involving the laying bare of rocks once many thousand feet below the surface. The formation of ore deposits usually occupies comparatively short epochs, and the agencies to which they owe their origin are evanescent phenomena.

In a rather large class of veins, however, of which we know that they were formed near the surface and in recent geological times, we may look with more confidence for a maintenance of the originating solutions, but even here it is well to investigate carefully; the spring may simply be a water of the upper circulation which selected the fissure as a convenient path.

The case of Plombières has already been mentioned (p. 70) and there seems to be little reason to doubt that the quartz-fluorite veins at this place have been deposited by the same hot waters which still issue from the fissures. Daubrée[1] cites the frequent occurrence, in the Triassic beds of the Central Plateau and the Vosges, of veins and extensive silicification similar to that at Plombiéres. Barite, fluorite, and sometimes galena accompany the quartz.

The sodium chloride springs of Bourbon-l'Archambault,[2] in the same region, issue from a fracture in Triassic strata, which contains quartz

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[1] Les eaux souterraines aux époques anciennes, Paris, 1887, p. 151.

[2] Jacquot and Willm, Les eaux minérales de la France, Paris, 1894, p. 107.

with galena, barite, and fluorite. The fissures follow dikes of micaceous porphyry (minette?). The waters have a temperature of 53° C. and the total solids aggregate 3,186 parts per million, of which 1,770 are sodium chloride. Bromine, iodine, fluorine, arsenic, and copper are present, and the saline constituents are attributed to the Triassic and Permian strata. The spring deposits contain earthy carbonates and 0.07 per cent copper oxide.

The springs of Lamalou, near Montpelier, southern France, have a temperature of 34° to 47° C. and 1,500 parts per million of total solids; the alkaline carbonates prevail, but they also contain calcium and magnesium carbonates, suggesting an admixture of meteoric waters. Traces of barium, arsenic, copper, lead, nickel, and cobalt were determined.[1] These springs are believed to be genetically connected with the eruption of a neighboring basalt area and stand in close relationship to veins containing pyrite, arsenopyrite, and chalcopyrite in a gangue of quartz and barite. Barite is believed to be deposited by the present waters.

The sodium carbonate springs at Ems,[2] according to Delkeskamp, issue from a fissure which forms the extension of a quartz vein and contains chalcopyrite. Basalt occurs in the same vicinity.

Sandberger and Delkeskamp state that the hot sodium chloride springs of Wiesbaden are closely connected with a quartz vein containing tetrahedrite; veins of barite and calcite are common, as are impregnations of barite; the latter are attributed to earlier (Tertiary) spring waters.

Close connection with ore-bearing veins is also, according to Delkeskamp, indicated by the sodium chloride springs of Kreutznach, which issue close to a number of veins containing calcite, barite, and fluorite with ores of copper and quicksilver. Here, also, Tertiary strata higher than the springs are impregnated with barite, suggesting a considerable age and a formerly higher point of issue of the springs. The saline constituents of the water are believed to be derived from sedimentary rocks. The widespread occurrence of barite, in close association with strong sodium chloride springs, is assuredly suggestive.

Mineral springs with a maximum temperature of 26° C. have been opened at several places in the mines of Freiberg, Saxony, and are described in some detail by Stelzner and Bergeat,[3] but there is little reason to believe that they are genetically connected with the deposits. Warm springs have been encountered in the mines of Cornwall; and one of them in a tin vein near Redruth is said to have contained much lithium, which is not surprising considering the general distribution of lithium-bearing muscovite in that region. It seems difficult to believe that these springs are the remains of the waters which deposited the veins, for the

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[1] L. de Launay, Recherche, etc., des sources thermominérales, 1892.

[2] *Verhandl.* Gesell. deutscher Nat. u. Aerzte, 2, pt. 1, 1903.

[3] Die Erzlagerstätten, 2, 1905–1906, p. 1227.

veins were formed at a great depth and under high pressure and temperature at a remote geological time.

In the Cordilleran region of the United States examples of mineral springs in mineral veins are not so common as might be expected from the coexistence of a late mineralization and present abundance of thermal waters. One reason for this lack lies probably in the great physiographic changes which in most parts of this region have taken place in relatively late times and which would tend to lower or divert the discharges of the springs. At Silver Cliff, S. F. Emmons[1] found issuing from the 2,000-foot level of the Geyser mine a strong spring of sodium carbonate water with free carbon dioxide, yielding small quantities of copper, lead, and zinc; the temperature was 26.5° C. The shaft was sunk to a depth of 1,850 feet in rhyolite tuff; at this depth, at the contact between the tuff and pre-Cambrian gneiss, a vein was found containing galena, sphalerite, tetrahedrite, argentite, etc., in a gangue of calcite, barite, and quartz. The water deposited a calcium carbonate sinter with traces of lead, zinc, copper, nickel, and cobalt. In this instance it is possible that the ascending water may have had a genetic connection with the deposit.

At the Comstock Lode[2] hot waters were encountered at an early date and have made exploitation difficult. It can scarcely be doubted that these waters stand in causal relation to the vein and Reid thinks that they now dissolve and precipitate gold and silver, as well as pyrite. The heat of the lode has been shown by Becker to be clearly due to the ascending waters. Reid[3] has examined the evaporated residue from water collected on the 2,250-foot level of the C. and C. shaft. He found 2.92 milligrams of silver and 0.298 milligram of gold per ton of solution. This water, which has a temperature of 46° to 81° C., contains 965 parts per million of solids, mostly sulphates of calcium and sodium but including 133 parts of silica. This water is assuredly not one of the pure types of ascending waters; its composition is in the main the same as that of the ordinary mine waters and it may be a mixture of meteoric mine waters with a very hot ascending water. Hot waters of a similar calcium sulphate type with some silica and sodium carbonate have been analyzed from the Tonopah, Nevada, mines, but they contain no sulphides and Bastin holds the evidence is by no means conclusive that they have caused the ore deposition. Particularly convincing of the competency of the ascend-

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[1] The mines of Custer County, Colorado, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, p. 461.

[2] G. F. Becker, Geology of the Comstock lode, *Mon.* 3, U. S. Geol. Survey, 1882, p. 230.

[3] John A. Reid, The structure and genesis of the Comstock lode, *Bull.* 4, California Univ. Dept. Geology, 1905, pp. 177–191.

E. S. Bastin, Bonanza ores of the Comstock lode, *Bull.* 735, U. S. Geol. Survey, 1923, pp. 57–63; Genesis of the ores at Tonopah, Nev., *Prof. Paper* 104, U. S. Geol. Survey, 1918, pp. 26–30.

ing "volcanic" springs to deposit gold and silver-bearing veins are the data given on page 54 in relation to the Ojo Caliente springs of New Mexico and those of Wagon Wheel Gap in Colorado. To this is added the evidence of the gold and silver-bearing sinters of New Zealand (p. 68) and Steamboat Springs, Nevada (p. 66).

The widely cited occurrence at Sulphur Bank, in Lake County, California,[1] is considered to furnish good proof of deposition of cinnabar by hot sodium chloride waters, also heavily charged with boron (analysis on page 54). The springs issue through Quaternary basalt in which cinnabar was deposited with opal as crusts along crevices, sometimes as delicate crystals loosely attached to the walls, or as impregnations of the porous basalt; the pyrite or marcasite was mostly disseminated in the rock, but occurred also as crusts alternating with cinnabar and opal. At the surface no cinnabar was observed, but sulphur, derived from the oxidation of $H_2S$, was present. A few feet below the surface the cinnabar appeared and continued down to about 300 feet, into the sandstones on which the basalt rested. No quicksilver was found in the water, but no one who has studied the occurrence has doubted that cinnabar, pyrite, and opal have been precipitated from the water which still ascends in these channels. The gases dissolved in the water consist mostly of carbon dioxide, with hydrogen sulphide, hydrocarbons, nitrogen, and some ammonia. The evidence gains in importance when it is realized that the mineral combination and general mode of occurrence cited are characteristic of the quicksilver deposits of the Coast Ranges. A number of other instances of deposition of cinnabar by ascending waters are given in Chapter XXIV.

**Summary.**—There is then, convincing testimony that deposits of quicksilver, antimony, arsenic, gold, and silver, may be formed close to the surface by hot ascending waters of the kind related to volcanic phenomena. It is probable, indeed, that the majority of fissure veins which contain notable amounts of gold and silver together with sulphides of the baser metals have been formed by these solutions. Of this, more conclusive evidence is yielded by the many veins which so frequently, like a metallic aureole, surround areas of igneous intrusive rocks.

On the other hand it is certain that warm and even cold waters of the meteoric circulation in non-volcanic regions are likewise competent to form mineral deposits of the baser metals containing oxides and carbonates of iron and manganese, and sulphides of copper, lead, and zinc

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[1] G. F. Becker, Geology of the quicksilver deposits of the Pacific slope, *Mon.* 13, U. S. Geol. Survey, 1888, pp. 251–268.

Joseph Le Conte and W. B. Rising, The phenomena of metalliferous vein formation now in progress at Sulphur Bank, *Am. Jour. Sci.*, 3d ser., 24, 1882, pp. 23–33.

F. Posepny, The genesis of ore deposits, 2d ed., pub. by the Am. Inst. Min. Eng., 1902, pp. 32–36.

with very small quantities of gold and silver. There is also evidence that such waters may develop deposits of minerals of vanadium, and uranium with radium (Chap. XXII, p. 409). The waters most competent to perform this work appear to be the calcium carbonate solutions and the chloride brines which at the same time contain carbon dioxide and hydrogen sulphide.[1]

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[1] C. E. Siebenthal, Zinc and lead deposits of the Joplin region, *Bull.* 606, U. S. Geol. Survey, 1915, p. 154.

# CHAPTER VIII

# ORIGIN OF UNDERGROUND WATERS

## THE UNDERGROUND FLUIDS OF METEORIC ORIGIN

A preliminary division of the underground fluids and solutions has been given in Chapter III (p. 29). It is certain that the bulk of the water in the rocks is of meteoric origin, *i.e.* that it consists of infiltrated rain water or water from present rivers, lakes, and seas. The sediments contain more or less of such water imprisoned for many eons, filling their pores or contained as inclusions or in newly formed hydrous minerals. The igneous rocks likewise contain such imprisoned water, usually in small quantities. In regions of dynamic metamorphism much water from one source or another is combined as hydrous silicates, particularly in sericite or chlorite. In mineral deposits, such substances as mica and chlorite hold much imprisoned water.

The ordinary waters in non-volcanic regions from the zone of circulation or from the static zone range from weak to medium and strong solutions of various salts dissolved from the rocks. The minerals of the rocks, in part, are dissolved directly; in larger part, they are decomposed by the waters, this decomposition often resulting in an insoluble residue and various soluble salts.

There are many substances in meteoric waters which may again be precipitated and form mineral deposits. Those in sea water and lakes and rivers have found mention on pages 27 and 28. In the underground waters such elements are iron, manganese, sulphur, copper, nickel, cobalt, zinc, cadmium, lead, barium, strontium, arsenic, bromine, phosphorus, boron, iodine, and fluorine are more rarely encountered and only in traces. During the oxidation of ore deposits, many other rare substances may be found in the surrounding ground waters. It is certain that deposits may be formed from waters containing a very small quantity of metals. For example, the Steamboat Springs in Nevada contain only 0.5 part per million of $Sb_2O_3$, but antimony sulphide is deposited throughout the sinter and in fairly considerable amounts below the surface.[1] This is, of course, another type of water, but the same reasoning holds. Many undoubtedly meteoric deep waters of the Mississippi valley contain zinc in about the same amount. Many springs depositing iron and manganese hydroxide certainly contain only a very small amount of these metals,

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[1] W. Lindgren, *Trans.*, Am. Inst. Min. Eng., 36, 1906, pp. 27–36.

say one or two parts per million. Therefore provided that the conditions favor the deposition of an insoluble compound, even very small quantities will suffice to produce a considerable deposit.

Meteoric waters often contain carbon dioxide and hydrogen sulphide, also hydrocarbons. These can be easily accounted for, considering that many sedimentary rocks contain pyrite, various forms of carbon and organic matter, and that by several kinds of reactions these gases may be set free.

Other underground fluids unrelated to volcanism are gases containing nitrogen, oxygen, argon, helium, and hydrocarbons. The last are important fuels. From such gases helium has been recovered in considerable amounts for aerial navigation. In gases from Kansas and Texas fields helium occurs up to 1.84 per cent, and 7 per cent is contained in a gas well southwest of Pueblo, Colorado.[1] Nitrogen may be derived from air from which the oxygen may have been removed by chemical processes. Helium may conceivably be derived from radioactive substances, such as carnotite, which is known to be present in certain formations. The hydrocarbons are distilled from organic material in the sediments.

Another possible source of carbon dioxide is the reaction of acid waters, such as are found in many mines, with adjacent limestone. It is also known that many granular rocks, like granite, contain minute inclusions of carbon dioxide. Laspeyres has calculated that a cubic kilometer of granite contains enough carbon dioxide to furnish the springs of Nauheim, in Germany, with this gas for 273,000 years, but such computations carry little conviction to those who realize the difficulty involved in the absorption of any but a minimal quantity of this gas from the quartz grains by percolating waters.

Gases may also, perhaps, be derived from a slow distillation from rock masses exposed to high temperatures in the lower part of the crust. This assumes, of course, that openings are available for such escaping fluids and this is, of course, more or less questionable. The interesting experiments of A. Gautier and others on the gases included or occluded (absorbed) in the minerals of a rock and set free on heating have been summarized by several writers, including F. W. Clarke[2] and F. L. Ransome.[3] A great number of exact analyses of these gases were made by

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[1] G. S. Rogers, Helium-bearing natural gas, *Prof. Paper* 121, U. S. Geol. Survey, 1921, pp. 113.

P. Ruedemann and L. M. Oles, Helium—Its probable origin and concentration in the Amarillo fold, *Bull.* Am. Assoc. Pet. Geol., 13, 1929, pp. 799–810.

S. C. Lind, Origin of terrestrial helium, *Proc.*, Nat. Acad. Sci., 11, 1925, pp. 772–779.

R. B. Moore, Helium, a national asset, *Trans.*, Am. Inst. Min. Met. Eng., 69, 1923, pp. 110–122.

[2] Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 276–288.

[3] *Econ. Geol.*, 1, 1906, p. 688.

R. T. Chamberlin,[1] who found in general that the various pulverized rocks yielded a total amount of gases (at 0° C. and 760 millimeters pressure) equal to from a fraction up to as much as thirty times the unit volume of rock. The carbon dioxide is believed to be derived from the decomposition of small quantities of secondary carbonates, while in part it may also be included or occluded. The hydrogen and the carbon compounds are probably due to reactions of water vapor and carbon dioxide with some of the substances contained in the rock, notably ferrous compounds. There is indeed a possibility that some of the carbon dioxide in deep waters may have been derived in this way. But it is, perhaps, scarcely recognized that there is a great difference between heating a small quantity of pulverized rock in the air and obtaining the same amount of gases from a solid mass at great depth. It seems probable that pressure would prevent the escape of these gases, and if the mass of rock were heated to the melting-point it would undoubtedly acquire a capacity for absorption of far greater amounts of gases than those expelled by heating the powder to redness.

## MAGMATIC OR JUVENILE WATERS

Volcanic phenomena are almost always accompanied by the emission of large quantities of steam and other volatile substances, and geologists generally have agreed that part of this water is a contribution to the atmosphere and hydrosphere from the magmas.

Day and Shepherd[2] subjected the gases of the Kilauea crater on the island of Hawaii to a very careful study, and ascertained that when free from contamination of air they consist of nitrogen, water gas, carbon dioxide, sulphur dioxide, and hydrogen. They concluded that the water released from the liquid lava as it reaches the surface is entitled to be considered an original component of the lava with as much right as the sulphur or the carbon. It follows logically that some of this water from cooling lavas, with associated gases, must mingle with the waters of meteoric origin.

Regarding plutonic rocks the direct evidence is lacking but indirect testimony is supplied by the inclusions of aqueous solutions so commonly found in granular rocks and by the presence of minerals like mica and amphibole which contain the hydroxyl molecule.

The best general evidence of the existence of juvenile waters is furnished, not by observation of the present springs, but by the study of old intrusive regions. Here the granites merge into pegmatite dikes, the latter change into pegmatite quartz, and this into veins carrying quartz

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[1] R. T. Chamberlin, The gases in rocks, Carnegie Inst. Washington, *Publ.* 106, 1908, p. 80.

[2] Arthur L. Day and E. S. Shepherd, Water and volcanic activity, *Bull.* Geol. Soc. Am., 24, 1913, pp. 573–606.

and metallic ores, such as cassiterite and wolframite. Here we have evidence difficult to controvert that dikes consolidated from magmas gradually turn into deposits the structure and minerals of which testify to aqueous deposition; this admitted, it is difficult to see what would prevent such waters from reaching the surface in the form of ascending springs.

Elie de Beaumont[1] was the first to give full expression to this view. He believed that there were two classes of hot springs: The first (the more common) is intimately related to volcanism and derives its waters and dissolved solids from this source; the second, and more exceptional, derives its water from simple infiltration. This view was accepted by de Lapparent, but Daubrée arrived at the contrary conclusion, that all thermal springs result from the infiltration of water from the surface; similar views were held by Fouqué and have been adopted by de Launay who, however, admits the presence of magmatic exhalations in certain ascending springs.[2] The views of Daubrée found general acceptance in other countries; in the United States they were accepted by Le Conte, Van Hise, and others. All waters appearing at the surface were considered of atmospheric origin and their salts were dissolved from the rocks percolated. About the year 1900 the importance of magmatic exhalations for the formation of mineral deposits began to be reasserted by various mining geologists—among them Vogt in Norway, and Spurr, Kemp, Weed, and Lindgren in the United States. In 1902 Suess,[3] the eminent Austrian geologist, announced his belief that many of the springs in volcanic regions were of "juvenile" origin—that is, that they now reach the surface for the first time and yield a permanent addition of water and salts, carried up from magmas cooling at great depth. As an excellent example of this the Carlsbad Springs were cited.

The question now arises whether it be possible to establish criteria by which the magmatic waters may be distinguished from those of meteoric origin. Delkeskamp in Germany has attempted the solution of this problem in a series of suggestive papers.[4] He rightly considers temperature of little value as a criterion and points out that many springs of meteoric origin are hot, while some, strongly suspected to be of juvenile

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[1] *Bull.* Soc. Géol. de France, série 2, 1847, tome 4, p. 1272.

[2] L. de Launay, Recherche, captage et aménagement des sources thermominérales, Paris, 1892.

[3] *Verhandl.* Gesell. deutscher Nat. u. Aerzte, Karlsbad, 1902, pp. 133–150.

[4] R. Delkeskamp, Juvenile und vadose Quellen, *Balneolog. Zeitg.*, 16, No. 5, Feb. 20, 1905, p. 15.

R. Delkeskamp, Die Genesis der Thermalquellen von Ems, Wiesbaden, und Kreutznach und deren Beziehungen zu den Erz und Mineralgängen des Taunus und der Pfalz. *Verhandl.* Gesell. deutscher Nat. u. Aerzte, 2, pt. 1, 1903.

A. Gautier, *Compt. Rend.*, 150, 1910, p. 436.

See also reference in *Econ. Geol.*, 1, 1906, pp. 602–612.

origin, are cold. The constant admixture with meteoric waters forms another difficulty, but accounts well for the many derivatives of varying characteristics which accompany every spring of deep-seated origin. Seasonal variations of temperature, salinity, and quantity of water constitute excellent proofs of superficial origin. A practical constancy of salinity, temperature, and quantity is said to be the best proof of a juvenile origin. Among the juvenile springs are those of Carlsbad in Bohemia, Ems and Wiesbaden in Germany

These criteria are all doubtful. Much more work must be done before we shall be able to establish the magmatic origin of any given spring.[1]

**Examples of Springs in Volcanic Regions.**—As pointed out on page 57 there are two types of ascending hot waters which may be of juvenile origin. They are the sodium carbonate and the sodium chloride-silica types, both common in regions of expiring volcanism. The former appear, for instance, in central Germany, in central France, in California, and at various places in our Western States. The latter characterize the great geyser regions of Yellowstone Park, Iceland, and New Zealand. The two classes break up through volcanic rocks and through the underlying plutonic rocks or crystalline schists. Whether these waters are wholly or partly of magmatic origin is a doubtful question. Arnold Hague,[2] who spent many years in the study of the Yellowstone Park has expressed the decided opinion that the present hot springs at this locality are of meteoric origin and this opinion is apparently shared by members of the Carnegie Geophysical Laboratory, who are now investigating this area. Such an origin is probably more difficult to establish for the geyser district of New Zealand. Almost all geologists are of the opinion that some of the dissolved salts and gases at all of these places are of magmatic or juvenile origin.

**Salts of Volcanic Springs.**—Some of the hot ascending springs in volcanic regions carry much sodium carbonate (Fig. 8). The long-continued action of the hot water saturated with carbon dioxide on the feldspars of the surrounding rock undoubtedly yields this salt in large quantities, and the scarcity of calcium and magnesium carbonates is explained by their precipitation with increasing percentage of alkaline carbonates.

Considerable quantities of sodium chloride are, however, always associated with the sodium carbonate and sometimes indeed predominate; to find an adequate explanation of this is more difficult. Igneous rocks average, according to Clarke's calculation, only 0.07 per cent of chlorine, and while there are some exceptional rocks containing sodalite, the sodium chloride waters are by no means particularly associated with this mineral. Considering that the water could extract only a small part of this chlorine,

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[1] Recently doubts have been expressed as to the magmatic origin of the water in the Carlsbad Springs.

[2] Origin of the thermal waters in the Yellowstone National Park, *Bull.* Geol. Soc. Am., 22, 1911, pp. 101–122.

it is not easy to estimate the amount of rock which must be percolated to obtain a sustained flow of chloride waters of the concentration often found in hot springs. It might be imagined that surface waters moving downward could have become charged with sodium chloride while traversing saline sedimentary rocks, but such an explanation seems somewhat forced in the case of springs which issue from granite in a region where no such sedimentary beds are known to occur. Boron is a common constituent of many of these springs, for instance, the Steamboat Springs, Nevada, and Ojo Caliente, New Mexico, both of which issue from granitic rocks. It is still more difficult to find a reasonable explanation for the presence of this substance on any hypothesis of leaching. Tourmaline and datolite are of course present in some rocks, but the springs carrying boron exhibit no marked relation to areas where such boron minerals occur. It is true that boron occurs in saline sedimentary beds and that traces of it are often found in waters traversing them, but the quantities do not compare with those determined in many waters of volcanic associations. Similar statements can be applied to fluorine.

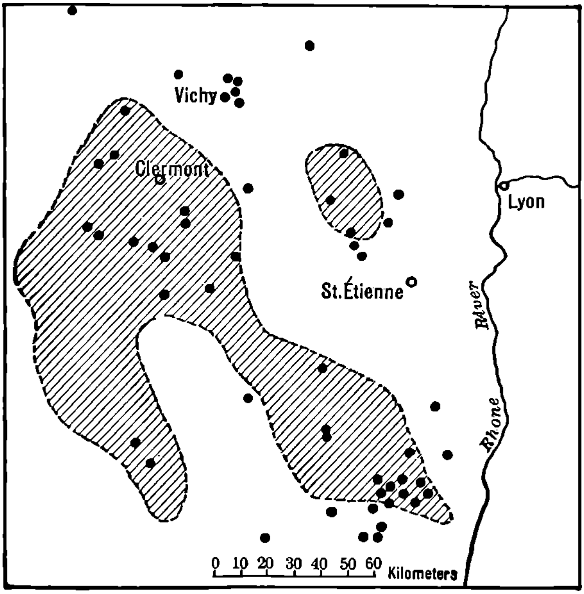

Fig. 8.—Carbon dioxide and sodium carbonate springs of central France. Black dots are springs. Shaded area shows extent of basaltic eruptions.

The geyser springs of Iceland, the Yellowstone Park, and New Zealand are rich in silica, and as most of them ascend through easily decomposed rhyolitic rocks, that substance may well be derived from leaching of the country rock. And yet when we note how veins rich in quartz are at places closely connected with pegmatite dikes, and how strong the evidence is against their deposition by leaching from surrounding rocks, we may well wonder whether this silica in the thermal waters is necessarily derived by solution of rock comparatively near the surface. And again, when we observe that chlorides form part of magmas, as indicated by the presence of sodalite and scapolite, and remember that sodium chloride occurs as small crystals in the fluid inclusions of quartz phenocrysts, and finally note the abundance of chlorides at volcanic eruptions, would it not then be easier to account for this salt in the springs of volcanic regions by an easily effected concentration of volatile substances while the magma was still fluid, than by a laborious search for traces of chlorides in the congealed igneous rocks?

**The Gases in Volcanic Springs.**—Many springs in volcanic regions contain large amounts of carbon dioxide, often also hydrogen sulphide. It is very difficult to account for these gases by assuming that they have been derived from reactions with the uppermost part of the crust, especially when the springs ascend through granite or other crystalline rocks below which sediments can hardly exist. The opinion of most geologists is that they represent emanations from deeply buried magmas brought to the surface by meteoric waters, or by magmatic waters or by a mixture of both.

The hot springs of Arkansas are a peculiarly interesting case. Their composition certainly indicates meteoric origin, but they have an extremely high temperature and also contain boron (p. 43). It has been suggested that in this case meteoric waters have descended to unusual depths and that they have been heated by gaseous emanations from a deeply buried magma.

The presence of liquid carbon dioxide in cavities in minerals of igneous rocks is proof of its occurrence in the molten magma consolidated in depth. Every eruption brings new evidence of exhalations from magmas congealing near the surface; and almost every volcanic district of recently closed igneous activity testifies to the persistence of this gas in escaping from the cooling lavas below. The Cripple Creek district, where gold-tellurium veins cut through the core of an old volcano, presents an excellent illustration of this condition. Imperceptible at the surface, exhalations of carbon dioxide become more marked in depth and their temperature, higher than that of the surrounding rocks, indicates that they came from below. In the extinct volcanoes of the Auvergne in France and of the Eifel on the Rhine, waters highly charged with carbon dioxide and exhalations of the same gas are extremely abundant. It seems difficult to escape the conclusion that the enormous quantities of this gas contained in the ascending waters of volcanic regions are of igneous origin.

These considerations apply equally well to the hydrogen sulphide with which some of these springs are so abundantly supplied. The decomposition of sulphates by organic matter or other reducing agents may be appealed to in places; but in igneous rocks, like granite, it does not appear to be quantitatively sufficient, and as we know that this gas plays a prominent part in volcanic eruptions we may well feel justified in believing that the waters ascending in regions of such eruptions may absorb this gas or alkaline sulphides and carry them to the surface.

When intrusive magmas break through limestone or dolomite the carbonates are often replaced by silicates and carbon dioxide is set free. There is good reason to believe that this process is going on at present. The quantity of carbon dioxide is large, and it is likely that many thermal springs are fed from such sources, but this explanation does not suffice for all cases.

**Rarer Elements in Volcanic Springs.**—In hot ascending sodium chloride springs which issue in volcanic regions rarer elements have often been determined; such waters are often rich in boron. Steamboat Springs, Nevada, contain notable amounts of arsenic and antimony with traces of quicksilver and many other metals. The springs of the Yellowstone Park carry boron and arsenic, but are poor in other rarer constituents.

The ascending sodium carbonate springs in volcanic districts also frequently contain boron and fluorine in notable amounts. Arsenic and copper have been found in the spring of Ems, and the same metals with lead also at Vichy. The Carlsbad Sprudel contains, according to Göttl,[1] traces of bromine, iodine, fluorine, selenium, phosphorus, boron, barium, strontium, lithium, titanium, tin, arsenic, antimony, copper, chromium, zinc, cobalt, nickel, and gold.

The presence of silver has apparently not been recorded, and that of gold only from the Carlsbad Springs. Gold, however, is found in the spring deposits at the surface, for instance in New Zealand and in the Steamboat Springs of Nevada. Quicksilver and large quantities of antimony seem to occur only in sodium chloride or sodium carbonate waters of the volcanic type, of which also higher amounts of boron and fluorine are characteristic. Arsenic is probably the most common of the rarer metals and has been found in all kinds of water. Copper, zinc, nickel, and cobalt are not uncommon, both in waters of sedimentary and in those of igneous origin. Lead is of rare occurrence. Iron is present in meteoric waters but occurs only in minute quantities in the hot springs. Probably it has been precipitated at lower levels.

**The Igneous Emanations.**—At several places above, the igneous emanations have been mentioned and the inference drawn that the waters in the crust of whatever origin must in places have absorbed such volatile substances. It may be well to describe briefly the character of these emanations.

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[1] J. Roth, Allgemeine und chemische Geologie, Berlin, 1, 1879, p. 570.

The active volcanoes constantly emit volatile matter from lava flows, craters, and fumaroles. Some of the less volatile materials crystallize as sublimates near the gas vent; other parts escape into the atmosphere.

In conclusion, considering the great variety of volatile substances given off as emanations from volcanoes (p. 112) and the many rare substances of similar kinds dissolved in the hot springs of volcanic affiliations, it seems certain that such ascending springs must in many places carry a considerable load of material emanating from the magmas.

Some hot springs which are found on the flanks of active or dying volcanoes doubtless yield meteoric waters which have received their heat and an admixture of fumarolic exhalations from masses of lavas, comparatively near the surface.[1] Such springs carry little or no silica, and make small deposits at the surface. The abundant sulphates are considered to be derived from oxidation of $H_2S$ in the fumaroles.

There are, however, many springs in regions of extinct volcanism, which are alkaline and carry much sodium carbonate, sodium chloride, and silica, and which have been proved to deposit cinnabar, stibnite, and realgar. They also are probably in most cases mixtures of meteoric and magmatic waters though some are possibly wholly of magmatic origin. Geophysicists have lately stated that the magmatic portion of such springs must have been contributed as gaseous exhalations from deep-seated magmas, and express the opinion that when the gases condense after leaving the magma they have no ascending power and must be carried up, if at all, by meteoric waters.

Geologists, on the other hand, believe that the magmatic liquids have ascending power because (1) they are propelled upward by continuous steam pressure from below; and (2) they carry large loads of nonvolatile material, principally silica, which is abundantly deposited all the way up. Some of this silica is indeed carried to the surface.

Much work has lately been done on the volcanic gases, fumarolic products, and on the gases retained by lavas and other igneous rocks.[2]

The gases obtained directly from volcanic vents are essentially identical in their components with those extracted from consolidated fresh lavas, though the ratios show wide variation. All agree in having water as the main constituent, to the amount of 80 to 90 per cent of the total gas content. The total amount of volatiles per gram, at 1200° C. and

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[1] A. L. Day and E. T. Allen, The volcanic activity and hot springs of Lassen Peak, *Publ.* 360, Carnegie Inst. Washington, 1925, 190 pp. See also symposium on hot springs, etc., *Jour. Geol.*, 32, 1924, Nos. 3, 4, 5, and 6.

[2] E. S. Shepherd, *Jour. Geol.*, 33, 1925, pp. 289–370; Carnegie Inst. Washington, *Yearbook*, 30, 1931, pp. 78–82.

E. S. Shepherd and H. E. Merwin, *Jour. Geol.*, 35, 1927, pp. 97–116.

E. G. Zies, The fumarolic incrustations in the Valley of Ten Thousand Smokes, *Contrib. Papers*, Nat. Geog. Soc., 1, No. 4, 1929, 79 pp.

760 millimeters is about 5 to 10 cubic centimeters, corresponding in weight percentage of rock to from 0.1 to 0.5. Fresh, coarsely crystalline rocks free from decomposition products yield about 30 cubic centimeters. If for any reason the volatiles were unable to escape during crystallization, the amounts may be much higher. The minor constituents of the gases are $CO_2$, CO, $H_2$, N, S, Cl, and F. In volume percentage at 1200° they vary from less than 1 to 20. In volcanic vents S, Cl, and F generally escape as hydrogen compounds. Certain elements usually considered non-volatile, such as copper, zinc, molybdenum, and silica, may also be carried along. The fumarolic deposits and gases may in fact contain many other metals, as well as boron, selenium, and tellurium.

The emanations of gases and liquids in an igneous body do not cease entirely even after the rock has congealed. Such changes shown, for instance, by the replacement of the primary minerals by albite, microcline, and magnetite have been designated by J. J. Sederholm as *deuteric* and should not be confused with replacements caused by later solutions of various kinds.[1]

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[1] See *Econ. Geol.* 24, 1929, pp. 100, 335, 869.

# CHAPTER IX

# METAMORPHIC ZONES AND MINERAL DEPOSITS

## THE CHEMICAL WORK OF UNDERGROUND WATER

**Stability of Minerals and Rocks.**—The underground water plays a very important part in the changes which take place in rocks; and the majority of mineral deposits are formed by the aid of it. Near the surface it may completely saturate the rocks or move in large volumes on fractures. At greater depths where there is no active circulation it may be sparingly present as rock moisture. The great mass of underground water is of atmospheric origin but as all magmas contain water which is given off upon solidification some waters in the rocks may be of magmatic origin. Solution and precipitation go on continuously; one or the other may predominate at any given place. The reactions which take place in the underground solutions extend over a wide range as to temperature, pressure, substances, concentration, and time; and they differ markedly under the varying conditions. The study of these reactions was first seriously undertaken by G. Bischof and Justus Roth[1] and these pioneers have been followed by many eminent geologists who have devoted themselves to the study of chemical geology.

One of the most fruitful conceptions developed in recent years is that of the limits of stability of minerals and rocks Conforming to increasing heat and pressure, zones exist in the earth's crust, gradually merging into one another but each characterized by certain groups of minerals that are stable only under the conditions prevailing in that particular zone. No mineral is absolutely stable. If subjected to certain conditions of temperature or in contact with certain solutions it will melt, decompose, dissociate, or dissolve. At the surface under the influence of atmospheric

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[1] G. Bischof, Lehrbuch der chemischen und physikalischen Geologie, 1863–1866.
Justus Roth, Allgemeine und chemische Geologie, 1, Berlin, 1879.
C. R. Van Hise, A treatise on metamorphism, *Mon.* 47, U. S. Geol. Survey, 1904.
U. Grubenmann and P. Niggli, Die Gesteinsmetamorphose, Berlin, 1924.
F. Becke, Ueber Mineralbestand und Struktur der krystallinen Schiefer, Ninth Session Internat. Geol. Congress, Vienna, 1903; also *Sitz.-ber.*, k. k. Akad., Vienna, 1903.
John Johnston and Paul Niggli, The general principles underlying metamorphic processes, *Jour. Geol.*, 21, 1913, pp. 481–516; 588–624.
C. K. Leith and W. J. Mead, Metamorphic Geology, New York, 1915.
E. B. Knopf, Retrogressive metamorphism, *Am. Jour. Sci.*, 5th ser., 21, 1931, pp. 1–28.

waters with oxygen and carbon dioxide practically no minerals are stable except a few oxides, hydroxides, and native elements.

In consequence of the reversible nature of chemical processes under changing conditions each mineral has its stability field or "critical level" which it can not leave without undergoing decomposition. The mineral aggregates, that is, the rocks, also follow this law and as the rock minerals have usually been formed in closely analogous ways most of the component minerals will become unstable more or less simultaneously.

Certain minerals, few in number, are less sensitive than others to such changes and recur under the most different conditions. They are designated "*persistent minerals*" and are in general of simple composition and do not contain the hydroxyl molecule; among them are quartz, hematite, pyrite, chalcopyrite, fluorite, calcite, and native gold. Orthoclase, all plagioclases, biotite, augite, olivine, the spinels, cordierite, and garnets develop and are fully stable only at high temperatures. Minerals rich in water, like chlorite, serpentine, and talc, are characteristic of lower temperatures. Other minerals, like muscovite, zoisite, epidote, hornblende, and albite, develop readily under strong pressure.

The varied composition of the crust, the unequal distribution of the underground water, the changing pressure, and the great differences in temperatures even at the same horizon make it difficult to establish strict rules and well-defined zones. One merges into another. Besides, stability is a relative term. Some rocks, like granite, are really stable only shortly after their complete consolidation. Under the influence of percolating deep waters the minerals of the granite are unstable, as they are in the zone of weathering. But the changes take place so slowly that at many places they can scarcely be perceived. Other rocks, like calcareous shales, are stable at moderate depths, but easily subject to recrystallization under pressure and rising temperature. The results of the reactions differ widely according to the composition of the waters. The minerals that develop in a rock charged with a slight amount of moisture are not the same as those that appear when the rock is penetrated by rapidly moving solutions, charged with salts and gases of foreign origin.

**Metamorphism.**[1]—The term *metamorphism* meaning strictly "a change in form," was proposed by Lyell in 1833 to express the changes of sedimentary beds to slates, quartzite, crystalline limestone, etc. Later it was extended to the development of schists and slates from igneous rocks by pressure and recrystallization. Still later, for instance, by C. R. Van Hise it has been employed in a wide sense so as to cover any change in the composition and structure of any rock, through whatever agency and with or without gain or loss of substance. This would include

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[1] For a thorough discussion of the various uses of this term, see: R. A. Daly, Metamorphism and its phases, *Bull.* Geol. Soc. Am., 28, 1917, pp. 375–418.

weathering and the development of any kind of epigenetic deposit, such as mineral veins, in a rock. Geologists have not generally accepted this wide definition. Metamorphism is here reserved for the processes which result in a partial or complete crystallization or recrystallization of solid masses of rocks, as in gneiss from granite or mica schist from clay shale. Though the mechanical effects of pressure may be conspicuous, metamorphism is always characterized by chemical changes in the component minerals.[1] The composition of the rock as a whole, while remaining fairly constant undoubtedly has a tendency to change.

For practical purposes we may distinguish between *static*, *dynamic*, *igneous*, and *hydrothermal* metamorphism. Static metamorphism proceeds without stress, at slight depths and under influence of a slight amount of water. At great depths and high temperatures a static recrystallization under great load may effect changes.[2] Dynamic metamorphism is effected under stress at higher or lower temperatures. These two are regional and proceed without strongly marked changes in composition.

Igneous metamorphism (pyrometamorphism) includes the effects of magmas on adjacent rocks and is a high temperature process. It is about equivalent to contact metamorphism but includes also the effect of igneous injection and pegmatitization.

Hydrothermal metamorphism includes the changes effected in rocks by hot ascending waters. Igneous metamorphism may be local or regional and in part involves changes of composition. Hydrothermal metamorphism is local and almost always involves changes of composition.

**Metasomatism or Replacement.**—The geological importance of metasomatism or replacement has already been pointed out on page 23. The word metasomatism, meaning a change of body, first used by C. Naumann to designate some kinds of pseudomorphism, is now applied to the *process of practically simultaneous capillary solution and deposition by which a new mineral of partly or wholly differing chemical composition may grow in the body of an old mineral or mineral aggregate.* The secondary minerals of any metamorphic rock result from metasomatic action. Rocks are termed metasomatic if their composition has been materially changed by replacement of the original minerals. Pseudomorphs and petrifications often furnish direct and incontrovertible evidence of processes of replacement.

Metasomatism is met everywhere and at all depths in sedimentary and igneous rocks and shows that the rock minerals have been subjected to conditions under which they were unstable. The development of

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[1] In *mylonitization* the rocks along faults and overthrusts are completely crushed under conditions where it cannot lose coherence, but even in this case there will be a variable amount of newly formed minerals.

[2] R. A. Daly, *Bull.* Geol. Soc. Am., 28, 1917, p. 400.

chlorite in augite, sericite or kaolin or calcite in feldspars, or galena in limestone is due to metasomatism. The typical metasomatic processes, traced with the highest magnifying power, show no space between the parent mineral and the metasome, as the newly developed mineral may be designated. The fibers and blades of sericite project into quartz without the slightest break in the contact. Rhombohedrons of siderite develop in quartzite, their crystal faces cutting across the grains without any interstices. Perfect prisms of tourmaline develop in feldspar grains, and sharp cubes of pyrite in primary feldspar or quartz.

Metasomatic rocks, that is rocks which have suffered a change in composition, are very common in mineral deposits and are often produced by strong and rapidly moving solutions (usually aqueous, sometimes gaseous) which penetrate the material through fissures and pores. There are many cases of complete or almost complete metasomatism, for instance of limestone by sulphides and quartz in which the chemical composition has been absolutely changed. In contrast to this the ordinary metamorphic processes in rocks are carried on by the scant rock moisture and while there is metasomatism in detail, the composition as a whole is but little changed. For description of metasomatic processes in ores and for criteria of metasomatism see Chapter XIV.

*Alteration* and *decomposition* are synonymous with certain phases of partial metasomatism of rocks.

*Dissemination* refers to grains or crystals distributed in a rock and is without definite genetic significance.

*Impregnation* is a genetic term and means that the mineral introduced is later than the rock; it may have developed by metasomatic processes or by filling of pore spaces or other cavities.

*Mineralization* (mineralize: to impregnate or supply a rock with minerals) is now mainly a mining term, conveying the meaning that a rock contains an ore mineral or a mineral indicative of ores, *e.g.*, pyrite near a copper deposit.

*Metallization* is a recent term to designate the introduction of useful metals in certain districts, *e.g.*, the tin metallization of Cornwall.

*Cementation* is used to indicate the filling of interstices in porous or shattered rocks.

**The Law of Equal Volume.**—It is necessary to distinguish between (1) metasomatic changes proceeding in free crystals or grains, or in loose or plastic aggregates under light load, where the force of crystallization can easily overcome the restraining pressure; and (2) metasomatic changes proceeding in rigid rocks where the new mineral is forced to make room for itself by solution of the host mineral.

In the first case the volume changes proceed according to the chemical formula. In the second case, *the replacing mineral occupies exactly the space formerly filled by the primary mineral;* the force of crystallization is of little or no direct influence, but as the pressure differs in intensity according to the crystallographic directions and as solution proceeds most actively at points of greatest pressure the development of crystal faces is thereby explained.[1]

The chemical formulas by which some kinds of replacement are usually expressed do not represent the actual change for these formulas are based on equal weights and will indicate definite changes in volume. The conversion of orthoclase to sericite is usually considered to take place according to the following reaction, which involves a decrease in volume

$$\underset{\text{Orthoclase}}{3KAlSi_3O_8} + H_2O + CO_2 = \underset{\text{Sericite}}{KH_2Al_3Si_3O_{12}} + K_2CO_3 + 6SiO_2$$

of 15.5 per cent even if the $SiO_2$ is assumed to have recrystallized as quartz. If, however, one volume of orthoclase has been replaced by an equal volume of sericite this equation is not correct, and by a calculation of the quantities of silica, alumina, etc., contained in one cubic centimeter of orthoclase and sericite, respectively, it will be found that a considerable addition of alumina is necessary. The actual formula is probably very complicated and could be established only if all the reactions taking place in the solution during the conversion of one mineral to the other were known. Many kinds of metasomatism, for instance, galena or barite replacing calcite (Figs. 67 and 68) can not be expressed by chemical formulas. One crystal, for instance, of pyrite may simultaneously replace parts of adjacent grains of different minerals, or may replace an aggregate of minerals in a fine grained rock. These well known facts will at once show that replacement is not the expression of one definite chemical reaction.

The law of equal volumes has been repeatedly verified by many independent observers; and there is little doubt that it holds for most metasomatic processes, on both a large and a small scale, both in general metamorphism and in mineral deposits. The most fundamental changes in rocks take place with practical constancy of volume.[2] A great deal has been written on changes of volume and energy liberated or absorbed, that is absolutely valueless as a measure of the processes that have been going on in rocks. The time will soon come when these relations are more clearly recognized.

**General Definition of the Metamorphic Zones.**—The conceptions of Albert Heim developed by C. R. Van Hise led to a division of the litho-

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[1] W. Lindgren, The nature of replacement, *Econ. Geol.*, 7, 1912, pp. 521–535. Volume changes in metamorphism, *Jour. Geol.*, 26, 1918, pp. 542–555. Metasomatism, *Bull.* Geol. Soc. Am., 36, 1925, pp. 247–262.

H. C. Boydell, Metasomatism and linear "force of growing crystals," *Econ. Geol.*, 21, 1926, pp. 1–55.

[2] When a rock is permeated by strong, rapidly moving liquids, solution may overtake deposition and a drusy structure may result. But even in this case the bulk volume is likely to remain about constant, except under very heavy pressure.

sphere into an upper zone of fracture and a lower zone of rock flowage, in which only sub-capillary openings exist and deformation is effected by granulation and recrystallization. Between them intervenes a middle zone of combined fracture and flowage. The limits of these zones are very indefinite owing to the greatly differing plasticity of rocks, *e.g.*, a granite and a calcareous shale. The experimental proof given by F. D. Adams[1] that in supported rocks in depth openings in granite can persist to depths of at least 11 miles, or about 58,000 feet, at a uniform pressure of 70,000 pounds per square inch and at temperatures supposedly corresponding, that is, 550° C. shows that the zones overlap widely and have value only as relative conceptions.

Van Hise divided the zone of fracture into an upper zone of weathering and a lower zone of cementation. The zone of flowage corresponds to the deep metamorphic zones in which minerals form by replacement only and in which the temperature is high and the pressure largely stress.[2] Later experiments by Bridgman[3] indicate that stresses twenty times greater than the crushing strength may be necessary to close cavities in rocks by hydrostatic pressure.

Any rock may be deformed under stress.[4] The thrust required to develop deformation in marble at a pressure corresponding to 4.2 miles would be 66,400 pounds per square inch; in case of granite, 138,500 pounds per square inch. At greater depths the required stress increases markedly. The pressure necessary for plastic deformation is very much greater than the crushing strength of the rock at the surface.

**Zone of Weathering.**—The best defined zone is that of weathering, the depth of which is determined by the level of the ground water, or by the depth to which free oxygen can penetrate in large quantities. In the zone of weathering the water percolates downward more freely than in the underlying zone, there is a tendency to the destruction of the rocks as units, and active transportation and concentration are characteristics.

Chemical work progresses by means of water solutions and gases, also extensively through the medium of organic life; mechanical disintegration is also important. The chemical reactions are oxidation, carbonatization, desilication, and hydration, the two first named mainly through decompo-

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[1] F. D. Adams, *Jour. Geol.*, 20, 1912, pp. 97–118.

[2] Van Hise called the upper two zones the realm of katamorphism and the lower that of anamorphism. In the zone of katamorphism (*kata*, down) complex silicates break down and simpler, less dense minerals form. In the zone of anamorphism (*ana*, up) silicates are supposed to be built up with forming of denser minerals and compact texture. Since Leith and Mead have changed these conceptions (Metamorphic Geology, 1915) and now confine katamorphism to processes of weathering, a confusion has been introduced that is best cured by the dropping of both terms.

[3] P. W. Bridgman, The failure of cavities in crystals and rocks under pressure, *Am. Jour. Sci.*, 4th ser., 45, 1918, pp. 243–268.

[4] F. D. Adams and J. A. Bancroft, *Jour. Geol.*, 25, 1917, pp. 597–637.

sition of silicates by water containing carbon dioxide. As a consequence of these reactions the volume should increase, but so much is carried away by solution that a great reduction of volume ensues.

Disintegration works hand in hand with decomposition and in advance of it; calcium, magnesium, sodium, and potassium are leached; the final products are a small number of minerals, largely hydrated compounds with low specific gravity and, for the most part, comparatively simple molecules. Almost all rock-forming minerals are unstable, as are the sulphides. These processes give rise to many mineral deposits of oxidized ores, which will be described in a later chapter.

The great extent of weathering and the intensity of the changes are especially emphasized in regions of soluble rocks like limestone. Weathering is caused by long-continued action by extremely dilute solutions. This is shown by the relative purity of the surface waters, which contain calcium and magnesium carbonates with lesser amounts of alkaline salts. The soluble products in part escape into the rivers through the zone of discharge, which ordinarily lies below the zone of weathering, and finally into the oceans. Colloid processes and products are common features of weathering.

**The Epi-metamorphic Zone.**—The rocks immediately below the zone of weathering are often saturated with water which diminishes in quantity with increasing depth. The small pressure permits fracturing and brecciation; and the openings created by these processes, as well as those resulting from porosity, are filled with minerals deposited by circulating solutions. To a small extent these minerals result from material abstracted from the zone of weathering, but as that zone is shallow the salts available from the weathering are, to a large extent, carried away by the surface drainage. The larger part of the minerals deposited have been derived from the rocks themselves; to a considerable extent they may be derived from deep-seated sources, as, for instance, in the cementation by quartz veins and veinlets near igneous intrusions. Hydration and carbonatization are the principal processes. Minerals like chlorite, serpentine, talc, sericite, epidote, and calcite develop, largely by metasomatic processes. Replacement and filling work together.

Where stress is present it is mainly in one direction and shearing and schistosity may develop; in metamorphic schists some of the minerals formed are muscovite, chlorite, talc, hornblende, zoisite, epidote, and albite; also quartz, pyrite, and calcite, probably magnetite and specularite. The clay slates with muscovite and albite, the chloritic schists, and the talc schists belong to this zone. The term "epi-metamorphic" was introduced by Grubenmann.

**The Meso-metamorphic and Deeper Zones.**—In the deeper belts (included by Van Hise under the name of the anamorphic zone) the pressure and temperature are high; the latter in general above 200° C. Little water is present. Minerals are formed mainly by replacement. Temperature and pressure work in the direction of diminished molecular volume. The pressure is largely stress—that is, acting in one direction—but hydrostatic pressure (transmitted in all directions) is becoming of importance. The important reactions are dehydration, the development of silicates, and deoxidation. Often there is a mixture of the epi-metamorphic and the hypo-metamorphic minerals. The minerals produced are numerous, stable, heavy, and complex. The rocks formed are compact and strong. However, the temperature is not sufficiently high to break up the molecules in which hydroxyl is contained.

Among the minerals of this zone are muscovite, microcline, albite, microperthite, oligoclase, biotite, zoisite, epidote, hornblende, staurolite, garnet, cyanite, titanite, magnetite, and ilmenite. Most of the micaceous and hornblendic gneisses containing garnet, staurolite, etc., belong to this zone; also the mica schists, amphibolites, and glaucophane rocks.

Where there is no stress in this zone many igneous rocks, like granite, basalt, and rhyolite, are stable.

Under the influence of stress in the metamorphic zones a parallel separation takes place in rocks. This may be caused by shear, by plastic deformation, or by new crystal growth, or by all of these combined. When shearing planes are closely spaced, the name schistosity is used. Recrystallization takes place according to the law of Riecke, solution prevailing at places of maximum pressure and deposition at those of minimum pressure. The crystallized products may assume lamellar structure, extending perpendicularly to the pressure; this results in a "schistosity by crystallization."

On the basis of Becker's[1] fundamental work, Bruno Sander[2] has studied the textural fabric of stressed rocks, not only as to the form of the minerals but also in regard to their optical orientation. Important information is thus obtained of direction and character of the forces to which the rock has been exposed. Practical applications result in the case of structural materials, and a similar study of stressed mineral deposits should produce valuable results.

**The Hypo-metamorphic Zone.**[3]—In the lowest zone, according to the classification of Grubenmann, the temperature is high and the tendency is toward an increase of volume. The hydrostatic pressure is enormous and stress almost non-existent, but high temperature is the dominant feature. There are no minerals containing the hydroxyl molecule except biotite, and the characteristics are, therefore, the prevalence of anhydrous minerals of small molecular volume. Characteristic minerals of this zone

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[1] G. F. Becker, Finite homogeneous strain flow and rupture of rocks, *Bull.* Geol. Soc. Am., 4, 1893, pp. 13–90.

[2] Bruno Sander, Gefügekunde der Gesteine, Wien, 1930, 352 pp.

[3] Grubenmann calls this the kata-metamorphic zone.

in the crystalline schists are orthoclase, all plagioclases, biotite, augite, olivine, garnet, cordierite, sillimanite, spinel, magnetite, and ilmenite. Many of the minerals of this zone also appear in the massive igneous rocks and in the contact metamorphic rocks. The rocks are mostly gneisses, gradually approaching granites; also granulites, eclogites, and augite gneisses. Most of the igneous rocks are stable in this zone.

The orthoclase or microcline in the crystalline schists of the deepest zone tends to microperthite in the middle depths and to sericite in the upper zone. Plagioclases of the deep zone may be transformed into albite and anorthite and finally to albite and zoisite or sericite. The augites change to hornblende and finally to chlorite. Olivine of the deep zone is transformed to hornblende or (with feldspar) to garnet and becomes serpentine in the upper zone.

Carbon dioxide and water doubtless escape from the deep zones upward wherever calcareous rocks containing free water or hydrated compounds become submerged in it.

Exceptional supplies of heat contributed by igneous intrusions may carry the reactions of the lower zones close to the surface.

**Relation of Mineral Deposits to the Metamorphic Zones.**—Though certain kinds of mineral deposits have originated at the surface or in the zone of weathering, the largest number have undoubtedly been formed in the zone of fracture, where circulation of solutions is comparatively easy. It is safe to assert that the great majority of ore deposits have been formed within 15,000 feet of the surface.

Ore deposits do not, as a rule, form in the zone of flowage where the passage of solutions is difficult. An exception to this is where hot emanations from intrusive bodies penetrate and impregnate certain rocks like limestone without the necessity of ducts and cavities.

Ore deposits may also form in the hottest zone where the solutions consist of magmas in which the free rearrangement of molecules is possible.

During the ordinary metamorphic processes under static or dynamic conditions extensive changes in mineral composition and structure may be effected with small changes in the chemical composition of the rocks, so that it is possible to trace the origin of highly metamorphosed rocks by the aid of analyses. Metamorphism can be, and usually is, effected with the aid of minute quantities of rock moisture and during the process there is little opportunity for extensive concentration of rarer constituents. Mineral deposits due to simple hydration or chemical rearrangement within the mass may result. Examples: soapstone by hydration of magnesian minerals; magnesite from carbonatization of serpentine; sulphur from reduction of gypsum by organic compounds; garnets developed in crystalline schists; concentration of hematite from lean primary ores; and many similar instances.

A comparison of the mineral records of ore deposits, formed at various levels in the earth's crust, with the results obtained by a study of general metamorphism soon brings out the fact that the same laws do not apply to both cases, although there are points of similarity. Attention was called to this important feature in a paper on the metasomatic processes in fissure veins, and increasing knowledge emphasizes the distinction.[1] Neither the rules of Van Hise nor the three zones of Grubenmann will fit closely the case of the ore deposits. The reason for this is not difficult to find. In metamorphism one deals with small quantities of solutions, free from large amounts of carbon dioxide and hydrogen sulphide. The majority of ore deposits, on the other hand, were formed by large quantities of waters rich in these gases and heavily charged with alkaline salts. A large number of silicates and other minerals, fairly stable under the influence of ordinary deep ground water, are incapable of existence in many vein-forming solutions. Biotite, amphibole, soda-lime feldspars, often also chlorite, serpentine, and magnetite are included among these.

## UNDERGROUND TEMPERATURES[2]

The increment in temperature in the upper part of the earth's crust is generally assumed to average 1° C. for 30 meters or nearly 100 feet. Beginning with a surface temperature of 11° C. at a depth of 100 feet, corresponding to the mean annual temperature of a place in the temperate zone, we would have at a depth of 1,000 feet, 20° C.; at 9,000 feet, 100° C.; at 20,000 feet, 210° C.; and at 35,000 feet, 360° C., which is near the critical temperature of water (374° C.). As a matter of fact but little is known about the increment at great depths. Actual measurements within the accessible zone or to depths of about 7,000 feet show considerable divergences from the average figures given above.

The best summary of the results obtained in widely separated parts of the world has been given by Koenigsberger,[3] who has also given important data regarding the influences which increase or diminish the geothermal gradient.

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[1] W. Lindgren, *Trans.*, Am. Inst. Min. Eng., 30, 1900, p. 601.

W. Lindgren, The relation of ore deposition to physical conditions, *Econ. Geol.*, 2, 1907, pp. 105–127. Also Compte Rendu de la X$^{ème}$ session du Congrès Géologique International, Mexico, 2, 1906, pp. 701–724.

[2] J. D. Everett, Evidence before the Royal Commission on coal supplies, London, 1904. Also in Reports of the British Association, 1882–1904. An increment of 1° C. in 100 feet equals 1° F. in 55 feet.

[3] J. Koenigsberger and M. Mühlberg, Ueber Messungen der geothermischen Tiefenstufe, *Neues Jahrbuch*, Beil. Bd. 31, 1911, pp. 107–157. (Contains also list of literature and technique of measuring temperatures.)

For a general treatment see C. E. Van Orstrand, On the nature of isogeothermal surfaces, *Am. Jour. Sci.*, 5th ser., 15, 1928, pp. 494–519.

The following data are abstracted from the tables of Koenigsberger:

GEOTHERMAL GRADIENTS IN UNALTERED ROCKS (NOT RECENT ERUPTIVES) AND REGIONS OF FLAT RELIEF. BORE HOLES

| Locality | Gradient in meters per 1° C. | Gradient in feet per 1° C. | Depth in meters | Depth in feet | Author |
|---|---|---|---|---|---|
| Martincourt,[1] France.... | 31.0 | 101 | 1,200 | 3,937 | |
| Sperenberg, Berlin....... | 32.5 | 107 | 1,268 | 4,160 | Dunker. |
| Sennewitz, Halle........ | 36.6 | 120 | 1,048 | 3,438 | |
| Schladebach, Merseburg. | 35.7 | 117 | 1,236 | 4,055 | Dunker. |
| Czuchow,[1] Silesia........ | 29.6 | 97 | 2,239 | 7,346 | Michael and Quitzew. |
| Bay City, Michigan..... | 36.8 | 121 | 1,050 | 3,445 | Lane. |
| Marietta, West Virginia.. | 37.9 | 124 | 1,360 | 4,462 | Hallock. |

[1] At Martincourt and Czuchow some coal beds are present.

One of the deepest bore holes in the world is that of the Goff Farm near Clarksburg, West Virginia. In 1918, this had reached 7,350 feet and the temperature gradient was 1° F. in 92 feet. The boiling point of water should be reached at about 10,000 feet.[1]

Bore hole 96 Olinda, Fullerton field, California, has attained 8,046 feet at which depth a temperature of about 220° F. is reported.[2]

In 1931, a well in the Tuxpam field, Vera Cruz, Mexico, reached a depth of 10,585 feet. From 2,700 feet to the bottom it was in the Lower Cretaceous limestone.[3]

Underneath high ridges and mountains the increase is slow:

| Locality | Gradient in meters per 1° C. | Gradient in feet per 1° C. | Depth in meters | Depth in feet | Author |
|---|---|---|---|---|---|
| Mont Cenis (summit).... | 50 | 164 | ......... | ....... | Giordano. |
| Gotthard (summit)...... | 44 | 144 | ......... | ....... | Stapff. |
| Simplon (summit)....... | 43.5 | 143 | ......... | ....... | Schardt. |

In or near recent eruptive rocks the increase is often rapid. This rapid increase is even noticeable in Tertiary eruptions. The following data are from borings:

| Locality | Gradient in meters per 1° C. | Gradient in feet per 1° C. | Depth in meters | Depth in feet | Author |
|---|---|---|---|---|---|
| Sulz (Wurttemberg)..... | 24.1 | 79 | 710 | 2,329 | Braun and Waitz. |
| Macholles, France....... | 14.2 | 46 | 1,005 | 3,329 | Michel-Lévy. |
| Buda-Pest.............. | 15.0 | 49 | 903 | 2,963 | Szabó. |

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[1] *Press Bull.* 357, U. S. Geol. Survey, 1918.

[2] *Oil and Gas Jour.*, Sept. 30, 1926.

[3] *Oil Weekly*, Oct. 2, 1931.

In the vicinity of hot waters, or where chemical processes of decomposition are active, the increase is especially rapid.

| Locality | Gradient in meters per 1° C. | Gradient in feet per 1° C. | Depth in meters | Depth in feet | Author |
|---|---|---|---|---|---|
| Idria, Austria | 10 | 33 | 320 | 1,079 | Scheinpflug and Holler. |
| Comstock, Nevada | 17.1 | 56 | 457<br>672 | 1,499<br>2,205 | G. F. Becker. |

In coal mines and in borings in coal-bearing strata the increase is more rapid than the normal, owing to the chemical processes in the coal beds.

| Locality | Gradient in meters per 1° C. | Gradient in feet per 1° C. | Depth in meters | Depth in feet | Author |
|---|---|---|---|---|---|
| Charmoy, Creusot (bore hole). | 26 | 85 | 1,168 | 3,832 | Michel-Lévy. |
| Paruschowitz, Silesia (bore hole): | | | | | |
| Above coal | 26 | 85 | 1,122 | 3,681 | Hendrick. |
| Below coal | 35 | 115 | 1,959 | 6,427 | Hendrick. |
| Gelsenkirchen, Germany. | 23.5 | 77 | 705 | 2,313 | |
| Rosebridge, England (cooled by ventilation). | 29.5 | 97 | ......... | ....... | |

Similar relations appear to exist in regions producing petroleum.[1]

| Locality | Gradient in meters per 1° C. | Gradient in feet per 1° C. | Depth in meters | Depth in feet | Author |
|---|---|---|---|---|---|
| Santa Maria, California | 23.0 | 76 | 1,097 | 3,599 | Arnold and Anderson. |
| Berekei, Caucasus | 23.0 | 76 | 1,000 | 3,281 | Kelickij. |
| Apsheron, Russia | 28.4 | 93 | 300–700 | 984–2,297 | Solubjatnikow. |

The following table embodies some late measurements. It emphasizes the steep gradients in oil wells in sedimentary rocks. In some metal mines like those of Freiberg (Saxony) and Cornwall, a similar steep gradient seems to exist. In others, principally gold-quartz mines in old and stable regions of considerable geological age, the gradients and the

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[1] For later information on this subject see H. Hoefer, Temperature in oil regions, *Econ. Geol.*, 7, 1912, 536–541 and E. L. De Golyer, The significance of certain Mexican oil temperatures, *Econ. Geol.*, 13, 1918, pp. 275–301. Also N. H. Darton, Geothermal data of the United States, *Bull.* 701, U. S. Geol. Survey, 1920.

increase of temperature are very gradual so that some of these mines can be worked to depths of over 7,000 feet. Owing to differences in observations and calculations the figures in some cases vary considerably. In many cases the depth-temperature curve is practically a straight line; but more commonly there is a slight flattening so that the gradient becomes somewhat smaller in depth; in other words, the increase in temperature is somewhat slower as depth is attained.

| Locality | Gradient in meters per 1° C. | Gradient in feet per 1° C. | Greatest depth in meters | Greatest depth in feet | Temperature at maximum depth | | Author |
|---|---|---|---|---|---|---|---|
| | | | | | ° C. | ° F. | |
| North Star Mine, California... | 104.1 | 341.7 | 1,128 | 3,700 | 22.4 | 72.3 | W. D. Johnston, Jr.[1] |
| Mother Lode, Amador County, California................ | 82.3 | 270.0 | 1,282 | 4,200 | 30.0 | 86.0 | A. Knopf.[2] |
| Calumet & Hecla, Michigan... | 60.7 | 194.4 | 1,731 | 5,679 | 35.0 | 95.0 | Ingersoll.[3] |
| St. John Del Rey Mine, Brazil | 68.4 | 224.7 | 1,871 | 6,140 | 46.5 | 115.7 | Van Orstrand. |
| Bendigo, Victoria............ | 42.7 | 140.0 | 1,110 | 3,645 | (44.5 | 112.0)* | Jenkins. |
| Johannesburg, South Africa.... | 110.9 | 363.9 | 2,143 | 7,032 | 36.1 | 97.0 | Marriott, Tillard and Ranson. |
| Freiberg, Saxony............ | 31.0 | 102.0 | 457 | 1,500 | ...... | ........ | D'Aubuisson. |
| Dolcoath, Cornwall........... | 33.0 | 108.0 | 610 | 2,001 | ...... | ........ | Haldane and Thomas. |
| Longmont, Colorado (oil)...... | 22.8 | 75.0 | 1,981 | 6,500 | 99.2 | 210.4 | Van Orstrand. |
| Ligonia, Pennsylvania (oil).... | 35.0 | 115.0 | 2,077 | 6,815 | 68.7 | 155.7 | Van Orstrand. |
| Fairmont, West Virginia (oil).. | 36.0 | 118.4 | 2,286 | 7,500 | 70.7 | 159.3 | Van Orstrand. |
| Longbeach, California (oil).... | 28.2 | 92.5 | 1,067 | 3,500 | 58.2 | 138.4 | Van Orstrand. |

* At 4,600 feet.

[1] *Jour.* Washington Acad. Sci., 22, May, 1932, pp. 267–271.

[2] *Prof. Paper* 157, U. S. Geol. Survey, 1929; also *Jour.* Washington Acad. Sci., 22, August, 1932, pp. 389–393.

[3] *Tech. Publ.* 481, Am. Inst. Min. Met. Engs., 1932.

At Bendigo,[1] where gold-bearing quartz veins occur in Ordovician sandstone, the rock temperature at the greatest depth attained, 4,600 feet, is 112° F. (44.5° C.). At this depth the water is salty and has a temperature of 114° F. At St. John del Rey,[2] a gold mine in the schists of southern Brazil, at 4,000 feet below the adit tunnel, the temperature is 95° F., or 35° C. At a vertical depth of 6,500 feet the rock temperature is 118° F. with an air temperature of 110° F. This is reduced to about 98° F. by a powerful system of fans and air cooling, made difficult by the high percentage of moisture in the air. At 7,100 feet the rock temperature is said to be 127° F. On the Witwatersrand, in the Transvaal,[3] a temperature of 65° F. prevails at 500 feet; the increase down to

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[1] W. J. Rickard, Deep mining at Bendigo, *Mining Mag.*, London, 1910, pp. 281–282.

[2] T. T. Read, *Eng. and Min. Jour.-Press,* Oct. 10, 1925. See also *Eng. and Min. Jour.*, May 25, 1931, p. 447.

[3] H. F. Marriott, An investigation of earth temperatures, etc., *Trans.*, Inst. Min. and Met., London, 15, 1906.

3,900 feet is regular at the rate of 1° C. per 360 feet; at a vertical depth of 4,500 feet the rock temperature in the City Deep Mine is 84° F. According to Marriott the natural ventilation reduces the temperature near the workings 5° to 6° C. The plotted curve is almost a straight line. By extrapolation a temperature of 109.5° should be reached at 10,000 feet based on plotting as a straight line.

At the Kirkland Lake mine, Ontario, a rock temperature of 69° F. is reported at 4,300 feet of depth, making a gentle gradient of 150 feet.[1]

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[1] *Eng. and Min. Jour.*, May 25, 1931, p. 447.

## CHAPTER X

# THE MAGMA IN ITS RELATION TO MINERAL DEPOSITS

### DEFINITION AND CONSTITUTION OF THE MAGMA

We know by experience that the lavas, that is, the effusive igneous rocks, solidify from a hot, fluid silicate melt. By study and inference we know also that all the rocks which, on the basis of certain characteristics, we designate as igneous rocks have congealed from such melts. Rapid cooling produces glasses or fine-grained rocks; slow cooling may produce rocks with coarsely granular texture; igneous rocks which solidified at great depth, beyond our observation, usually have such a granular texture.[1] Such silicate melts are called magmas and may be defined as liquids of high temperature, which consist of a mutual solution of silicates and silica with admixed oxides, and always contain a certain amount of dissolved volatile substances.[2]

The oxides present in igneous rocks are chiefly those of silicon and iron; they separate out as quartz and magnetite. The small but important quantity of volatile substances consists of predominating water with some carbon dioxide, sulphur, chlorine, fluorine, etc., in all rarely exceeding a few per cent. Observations on lavas plainly show their presence; and we infer, from many lines of deduction, that they existed in the magmas which solidified as intrusive rocks. Fenner[3] believes that under intracrustal conditions these substances form homogeneous solutions with those of less volatile matter and that they are in a state of molecular dispersal. Critical temperatures are so modified that they have little relation to those of the pure substances; gaseous properties are obscured and all the substances should be regarded as parts of a mutual solution. The crystalline compounds and the gases that escape later may be far from representing the state of combination in the solution.

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[1] Most coarse-grained rocks like granites and diorites have undoubtedly congealed from a hot fluid melt. Some petrologists hold that similar rocks may be produced by a high-temperature replacement of quite different rocks, like quartzite, by the action of emanations from contiguous magmas. In the nature of things such products must be of subordinate importance. Again certain basic igneous rocks may have formed indirectly from a melt by the crystallization and settling of heavy silicates like olivine.

[2] Magmas, rich in silica, may perhaps be considered as a solution of all other constituents in silica.

[3] C. N. Fenner, The Katmai magmatic province, *Jour. Geol.*, 34, 1926, p. 697.

That a magma is a solution is inferred from the lowering of the freezing point as shown by the order of crystallization, and from the fact that some of the last residues of crystallization approximate eutectic mixtures. Dissociation takes place to some extent and the magmas are electrolytes.

Though some basic rocks can be reproduced by dry fusion, the presence of volatile substances appears to be necessary for the crystallization of acidic rocks like granite and monzonite.

With lowering pressure and congealing of the non-volatile material the volatile compounds tend to escape. In the magma their rôle is to decrease viscosity and to lower the melting point. When magmas consolidate, these constituents are expelled except as they may enter the composition of rock minerals or become imprisoned in them as gaseous or liquid inclusions. When expelled, they (particularly the halogens) carry with them many heavy metals like iron, copper, zinc, lead, or tin because volatile compounds are formed, such as $SiH_4$, $SiF_4$, $SiCl_4$, $TiCl_4$, $AsCl_4$, $AlCl_3$ $WF_6$, $BCl_3$, $BF_3$, etc.

Barus obtained liquid solutions of various glasses with water at high temperature and pressure, which congealed at comparatively low temperatures as undercooled liquids or "rigid solutions." Upon heating, the water is expelled and a pumice-like, difficultly fusible mass remains. Many pitch stones and obsidians which contain much water behave in the same way when heated.

Similar experiments more closely paralleling the conditions of pegmatite magmas have recently been undertaken by Morey.

The average composition of igneous rocks is thus not exactly that of the average magma for the volatile components in the latter are not accounted for in the analyses; on the other hand the analyses include some unavoidable products of later decomposition. There are many kinds of magmas yielding rocks ranging from very acidic to very basic composition, such as rhyolite, granite, andesite, granodiorite, monzonite, diorite, basalt, diabase, gabbro, and peridotite. The common rock-forming minerals are relatively few in number and include feldspars, feldspathoids, pyroxenes, amphiboles, olivine, micas, quartz, iron ores, and apatite. Various silicate molecules enter into the constitution of the magma, and they are miscible in all proportions. According to J. W. Greig, the oxides of Ca, Mg, and Fe show limited miscibility with silicates but only at high temperature and acidity. Sulphides are miscible with the silicates to only a limited extent, the amount increasing with temperature and basicity.

## Crystallization of Magmas

As in an aqueous solution, the successive crystallization of given minerals is dependent upon their solubility in the rest of the magma and does not follow their temperature of fusion. When a salt dissolves in water the temperature of solidification is changed. Water freezes at 0° C., but an addition of sodium chloride to it depresses its melting or solidifying point many degrees. Alloys show the same behavior—for example, those with extraordinary low temperature of fusion, sometimes below 100° C. In the same way an igneous rock may become fluid at a temperature far below the average melting point of its constituent minerals, or even lower than the lowest of these.

On the other hand, no mineral can separate if the temperature, for a given pressure, is higher than the point of fusion of this mineral. Below this point crystallization takes place whenever the point of saturation of the solution for this mineral is exceeded. Some of the components of the solution will form isomorphous mixtures, but a part of it may remain in eutectic proportions.

According to the empirical rule of Rosenbusch the separation of crystals in a silicate magma follows an order of decreasing basicity, so that at every stage the residual magma is more acidic than the aggregate of the crystals already separated out. This rule is subject to important exceptions, especially in basic magmas, but in the granitic and dioritic rocks the basic and difficultly fusible minerals, such as zircon, magnetite, apatite, ilmenite, and rutile, crystallize first. Then follow biotite, hornblende, and augite, or in general the magnesium and iron silicates, then the soda-lime feldspars, later orthoclase, and finally the residual quartz, which probably separates at about 800° C. The "mother liquor" of a granite thus becomes successively richer in silica. The "mineralizers," or the volatile substances, like boron, fluorine, and tin, follow the acidic rather than the basic constituents. The residual magma, in granitic rocks, is a solution rich in alkalies and silica, probably with water, which under certain circumstances may be eutectic and may be pressed out of the partly consolidated magma, as from a sponge, and crystallize as pegmatite in fissures held open by the hydrostatic pressure of the fluids. The eutectic is said by Vogt to have the composition of 73 per cent orthoclase and 27 per cent quartz. Petrographers, however, are not unanimous in regard to the existence of a true eutectic in rocks.[1] Another fact to be considered is that the rest-magma is certainly not always highly acidic but that in basic rocks it may retain much iron to a very late stage.

Newton and Teall[2] say of certain basalts from Franz Josef Land that progressive crystallization leads to the formation of a mother liquor, poor in silica and rich in iron. This agrees with conclusions derived from the study of some high-temperature magnetite deposits.

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[1] C. N. Fenner, *op. cit.*, p. 750.

J. J. Sederholm, On synantetic minerals and related phenomena, *Bull.* 48, Comm. géol. de Finlande, 1916.

E. S. Bastin, Origin of the pegmatites of Maine, *Jour. Geol.*, 18, 1910, p. 312.

[2] *Quart. Jour.* Geol. Soc., 53, 1897, p. 477.

Near the surface there often appear two generations of crystals caused by interrupted crystallization, and sometimes an older generation, of hornblende, for instance, may be resorbed and almost obliterated. In rock-forming minerals the volume of the crystallized substances is smaller than that of the corresponding fluid substance; their fusibility and also their solubility diminish somewhat with increasing pressure. A sudden release of pressure may then act as an increase of temperature and newly formed crystals may be remelted.

The mode of crystallization in a binary magma is illustrated in Fig. 9, presenting the most simple conditions. In this temperature-composition diagram the saturation- or crystallization-curves are indicated; they show that with decreasing temperature separation of one component takes place until with a certain composition the whole remaining liquid will crystallize as a eutectic. Above the curves the magma is fluid; at a certain point say with 10 per cent of the second substance the first substance will begin to crystallize; this changes the composition of the magma and as the temperature is still further lowered crystallization continues. At *E*, the eutectic point, the remaining mixture of *A* and *B* solidifies as an intimately intergrown or fine granular mass.[1] Between the curves and a horizontal line through *E* the magma consists of fluid and crystals. Below *E* there are crystals only. Similar diagrams have been constructed for many substances. The crystallization of a three-component magma can be represented by a more complicated triangular diagram.[2]

It has been assumed that the constituents of a binary magma are of unchanging composition. But many rock-forming minerals consist of two or more components which are isomorphous and miscible in all proportions, forming homogeneous crystals. The plagioclases, the pyroxenes, and the olivines are examples of this. The process of crystallization of such compounds may be represented by a temperature-composition diagram (Fig. 10).

In mix crystals, for instance, of albite and anorthite, there is only one freezing-point curve, instead of two meeting at a eutectic point. As in the case of non-mixing components it is called the *liquidus* (the upper curve in the diagram). The melting-point curve, or *solidus*, does not coincide with the freezing-point curve. A mix crystal does not melt at a

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[1] C. N. Fenner, The significance of the word "eutectic," *Jour. Geol.* 38, 1930, pp. 159–165. "Eutectic" is applicable with strict propriety only to a point on the phase diagram. The lines and surfaces related to the eutectic point are called "boundary lines" and "boundary faces." At this point the temperature is fixed until the crystallization is complete and the total composition of the solid phases is exactly that of the liquid from which they crystallize. The system is invariant. In a magma impregnated with volatiles, the final crystals are deposited along a boundary curve of decreasing temperature and varying composition until the last of the volatiles escape or until a more or less dilute aqueous residue is attained.

[2] G. W. Tyrrell, Principles of petrology, 1926, p. 67.

definite temperature, but melting is spread over an interval of which the lower limit is fixed by the solidus, and the upper limit by the liquidus. A mixture of composition $Ab_1An_1$ begins to crystallize at 1450°, the first crystals having the composition $Ab_1An_5$. With further cooling both liquid and crystals change their composition. When the temperature has fallen to 1285° the crystals have the composition $Ab_1An_1$, and the liquid has just disappeared. There is no eutectic point. It is evident that if cooling proceeds very slowly the crystals will gradually change their composition. The complete readjustment of the composition of

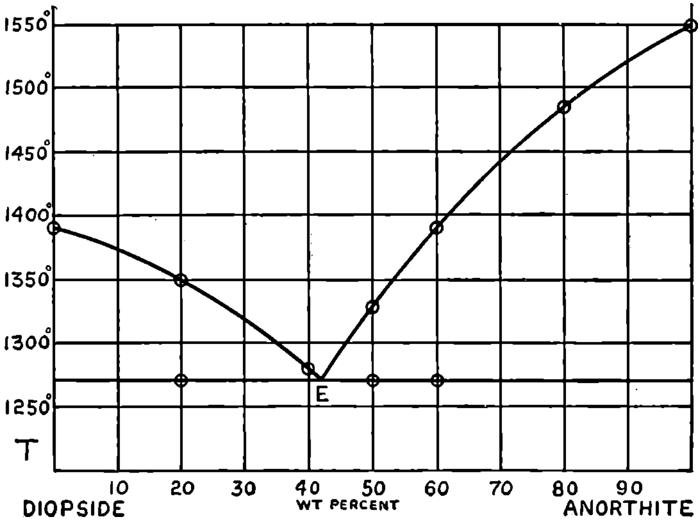

FIG. 9.—Equilibrium diagram of the system, anorthite-diopside. (*After N. L. Bowen.*)

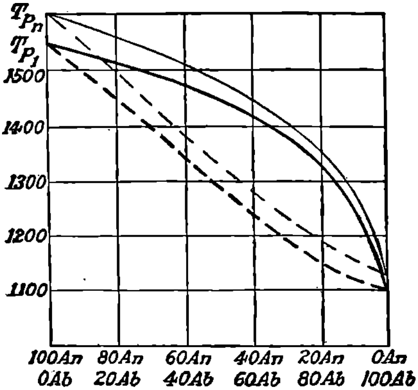

FIG. 10.—Temperature-composition diagram, illustrating crystallization of mixtures of albite and anorthite. Upper curve liquidus; lower curve solidus. Also showing slight difference between the curves at low pressure, $T_{p_1}$, and at high pressure, $T_{p_n}$. (*After J. H. L. Vogt.*)

the crystals is only theoretically possible. Actually, zones are formed around a kernel, each of the composition appropriate to the temperature of formation and to the liquid with which it was in equilibrium. Thus arises the well-known zonal structure of plagioclase crystals.[1] The resultant product is made up of crystals whose average composition is $Ab_1An_1$, but whose inner cores are more calcic and whose outer rims more sodic. The diagram also shows the relatively slight displacement which is effected in these curves by a strong increase in pressure.

## CAUSES OF VOLCANISM

It is not the purpose to discuss in detail the causes of volcanism, a subject which as yet is in a highly speculative stage. It is generally admitted that the sources of the magmas are very deep, perhaps from 20 to 60 miles and that by various causes, in various places, the rigid crust is subject to melting and the development of local magma basins. Structural lines undoubtedly determine the position of these. The ultimate forces which bring the magma up to the surface or within short distances of it may be contraction with tangential pressure, folding, and thrust, or may be isostatic adjustment. Here opinions differ widely. Among the direct causes of volcanic action the following are mentioned by Daly: minute deformations of the magma chamber; effervescence of lavas; assimilation of rocks in depth resulting in increase of volume; increase in volume through heating or by chemical processes. To this may be added the possibility of explosion and eruption by surface waters or oceanic waters, near the surface.

## DIFFERENTIATION IN MAGMAS[2]

Differentiation, according to Iddings, means the separation of a homogeneous rock magma into chemically unlike portions. For each region, in each separate "magma basin," there is probably one essentially homogeneous magma from which by some process of differentiation the various rock types have been derived. In general it is thought that the primary magma was of intermediate or basic composition and has been separated into basic and acidic forms, like basalts, latites, and rhyolites.

[1] G. W. Tyrrell, *op. cit.*, p. 66.

[2] For fuller treatment see chapters on Magmas and Differentiation in: Natural history of igneous rocks, by Alfred Harker, 1909; Igneous rocks, by J. P. Iddings, 1909; Igneous rocks and the depths of the earth, by R. A. Daly, 1933; Data of geochemistry, by F. W. Clarke, 1924; The principles of petrology, by G. W. Tyrrell, 1926; J. H. L. Vogt, The physical chemistry of the differentiation of igneous rocks, *Jour. Geol.*, 29, Nos. 4, 5, 6, 7, 1921; N. L. Bowen, Later stages of the evolution of igneous rocks, *idem*, 23, 1915, Supplement; Crystallization-differentiation in igneous magmas, *idem*, 27, 1919, pp. 393–430; The reaction principle in petrogenesis, *idem*, 30, 1922, pp. 177–198; N. L. Bowen, The evolution of igneous rocks, Princeton, 1928, 338 pp.

Symposium on the physical chemistry of igneous rock formation. A general discussion, *Trans.*, Faraday Soc., 20, pt. 3, 1925, with many important papers.

That differentiation has actually occurred is proved by the consanguinity of different rocks in a given district; by successive effusion of unlike lavas from the same vent; by the different products found in one and the same intrusive body; and by the study of individual flows. The explanation of these processes is not always satisfactory.

A magma may yield gases, or aqueous solutions rich in silica; it may also yield melts of the oxides of iron, titanium, or chromium; or finally it may yield sulphide melts or sulphide solutions. These separate parts should not be termed magmas but *magmatic products.*[1]

The earlier attempts to account for differentiation may still prove to be of some value. They include the application of "Soret's principle," which states that when two parts of a solution are at different temperatures there is a tendency toward selective concentrations of certain ingredients in the cooler portion.[2] G. F. Becker showed, however, that molecular diffusion in a viscous magma would require almost unlimited time and H. Bäckström pointed out that the action could not alter the relative proportions among the dissolved substances.

"Gravitative adjustment" was thought by J. Morozewicz and R. A. Daly to play a considerable part in differentiation. A mass of magma, like a high column of a salt solution would separate into a denser substratum and a lighter upper part.

G. F. Becker first indicated the possible importance of fractional crystallization, thus regarding differentiation as a consequence of the general cooling process. This view has been very fully developed by N. L. Bowen in a notable series of papers. Bowen's theory of differentiation by crystallization is now accepted by many petrographers (p. 110).

**Liquid Immiscibility.**—A separation may occur in the magma if with lowering temperature certain elements are no longer fully miscible with the silicate melt. It is conceivable that in such a way dissolved sulphides may separate out in globules which would grow and eventually, if the magma continues fluid, would settle to a molten sulphide layer. This theory has much to commend it, and it is thought that some high-temperature sulphide deposits may have been concentrated in such a manner.

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[1] Naturally, the question may arise where the line should be drawn, for many of such oxide or sulphide melts contain admixed silicates and transitions exist. Many geologists speak of quartz magmas or magnetite magmas or sulphide magmas. An exact definition is perhaps of no great importance but it would be advisable to retain the term "magma" for melts which contained any considerable amount of silicates, say about 25 per cent. The use of "magma" or "ore magma" for the volatile components and aqueous solutions does not appear justified. My attitude in relation to the so-called "carbonate magmas" except as local products of assimilation, or as products of carbonate replacement, is frankly one of doubt.

[2] A. E. Lagorio, *Tschermak's min. u. petr. Mitt.*, 8, 1887, pp. 421–529.

**Differentiation by Crystallization.**—Bowen's theory emphasizes the fact that the early crystals are usually of heavy minerals like magnetite, ilmenite, and olivine, and laboratory experiments have indeed shown that olivine crystals actually sink in the melt.[1] It is easily seen that the settling of these early crystals may lead to a concentration of iron ores and, in general, to a differentiation of the magma. Many cases have been investigated where, in flows and intrusives, such a settling of olivine crystals has taken place.

**The Reaction Series.**—Tyrrell[2] concisely explains Bowen's reaction series as follows:

> Once produced, a mineral formed as part of a eutectic system in magmas is no longer concerned in the equilibrium. Its composition can not be further changed; it is magmatically "dead." On the other hand, a mineral of a mix-crystal series is in continual reaction with the liquid from which it has crystallized, and its composition is continually being modified. The plagioclase series forms an ideal mix-crystal series. In their magmatic relations, Bowen proposes to call such series *continuous reaction series.*
>
> Another type of reaction relation also occurs in magmas in which an early crystallized mineral reacts at a certain temperature with the liquid in such a way as to form a mineral of different composition. In the system $MgO$-$SiO_2$, for example, mixtures of appropriate composition produce olivine as the earliest mineral, but with falling temperature the olivine reacts with the liquid to form a pyroxene (clino-enstatite). The two minerals thus related by reaction are called a *reaction pair.* A reaction relation of the same character may exist between three or more minerals which, when arranged in the proper order of succession, constitute a *discontinuous reaction series.* Minerals which are connected by this relation often exhibit *incongruent melting, i.e.*, they have no definite melting point but on heating break up into some other mineral plus liquid.

Both kinds of reaction series may co-exist in the same magma. A common discontinuous reaction is: olivine, pyroxene, amphibole, biotite, muscovite, and quartz. A common continuous reaction series is exhibited by the feldspars as follows: Calcic plagioclase, lime alkali plagioclase, alkali plagioclase, potash feldspar, muscovite, and quartz. The two branches converge and merge into a single discontinuous series of which quartz is the final product.

Without doubt, crystallization, with removal of crystals as formed, is capable of effecting large changes of composition in a magma. Gravity is an ever-present force, tending to effect the required separation. In this manner the splitting up of basaltic or intermediate magmas into more acid and more basic portions can be well accounted for. Large intrusive sheets like that of Sudbury, Ontario, are often gradually separated into a lower norite and an upper quartzose rock. However, the

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[1] *Am. Jour. Sci.*, 4th ser., 39, 1915, p. 175.

[2] The principles of petrology, London, 1926, pp. 76–77.

recent controversy regarding the petrology of Sudbury has shown that opinions about this supposed differentiation are far apart. Differentiation by crystallization is surely highly important. But probably it is not the whole story as has been recently emphasized by C. N. Fenner.[1]

It has been pointed out above (p. 105) that in many cases, at least, the residual magma is not rich in silica. Certainly more information is required before a complete solution of the problems of differentiation can be attained.

Another difficult question in regard to crystallization-differentiation is the very apparent lack of differentiation in batholiths of monzonitic or granodioritic type. Some of these are most remarkably uniform though deep sections have been exposed by erosion. In many cases the only evidence of differentiation consists in minor accumulations of more basic rocks at their margins.

If the crystallization and settling take place in the magma chamber, as indeed must be the case, the solids must be melted again in order to ascend as separate irruptions say of iron ore or peridotites. This point has been emphasized by Vogt. Late investigations[2] seem to indicate that most important changes of state may be produced in a system (in a magma, for instance) by pressures of 5,000 to 10,000 atmospheres.

**Differentiation by Gas Transfer.**—The importance of the volatile substances in effecting differentiation has been emphasized by several authors[3] and minimized by others. In a large magma basin no complete equilibrium can exist; both pressure and temperature vary from place to place. Probably there is a stream of gases through the magma reservoir, some of it escaping upwards through volcanic conduits and some penetrating the rock walls.

> In the ascent of a gas bubble reactions among its constituents will ensue and reactions with the surrounding magma will likewise take place . . . . The effectiveness of magmatic gases as agents of differentiation, therefore, lies both in their ability to carry away material when they escape and in their ability to make selective transfer of material from lower to upper levels.[4]

It is also likely, as suggested by W. H. Goodchild, that the gas bubbles will attach themselves to minute crystals and to droplets of sulphides and may thus effectively serve as an agent of flotation.

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[1] The Katmai magmatic province, *Jour. Geol.*, 34, October, November, 1926.

[2] L. H. Adams, *Yearbook*, Carnegie Inst. Washington, 29, 1930, pp. 69–72.

[3] C. H. Smyth, Jr., The chemical composition of the alkaline rocks, etc., *Am. Jour. Sci.*, 4th ser., 36, 1913, pp. 33–46; *Proc.*, Am. Philos. Soc., 66, 1927, pp. 535–580.

A. C. Lane, Wet and dry differentiation of igneous rocks, Tufts College Studies 3, No. 1, 1910, pp. 39–53.

R. A. Daly, The genesis of alkaline rocks, *Jour. Geol.*, 26, 1918, pp. 97–134.

[4] C. N. Fenner, *op. cit.*, p. 744.

The alkaline rocks appear to be derived from the sub-alkaline clan by differentiation; this is pretty clearly proved by their mode of appearance in the San Francisco volcano, Arizona,[1] and in the Leadville-Boulder[2] belt of intrusives, to pick out two of many examples. It is well possible, as suggested by Daly, that assimilation of limestone may locally produce alkaline rocks.

Besides their richness in alkalies these rocks carry much titanium, zirconium, chlorine, fluorine, phosphorus, barium, strontium, and other unusual constituents. It is held that these alkaline rocks were probably differentiated near the top of the magma chamber from the liquid magma, largely by gaseous transfer.

### RELATION OF VOLCANOES AND LAVA FLOWS TO MINERAL DEPOSITS

At volcanoes and in the lava flows we may study the relations of the magma at close range. It seems clear that the magmas which are poured out on the surface are derived from very considerable depths, from magma basins perhaps 20 to 60 miles below the surface. The well-known volcanic phenomena indicate violent explosions from the release of pressure of the volatile components, probably also from the sudden expansion of meteoric liquids reached and assimilated by the magmas near the surface. At other places basic magmas much less rich in volatile matter pour out quietly from fissure eruptions without much explosive activity.

That gases are present in the lavas is clearly shown by the vesicules so common in flows and is plainly seen in the bubbles that rise from the lavas at the surface. In more concentrated form the volatile substances escape from the fumaroles or steam jets which are so common in active volcanoes. Detailed observations have been made at the Italian volcanoes, at Santorini, in Iceland, in Hawaii, at the Katmai volcano in Alaska, at Lassen Peak, California, and at many other places.[3]

Bergeat[4] lists the following elements and compounds occurring as sublimates from Vulcano, in the Eolian Islands: Sulphur (seleniferous), tellurium, realgar, boric acid, chlorides of sodium, ammonium, and iron, glauberite ($Na_2SO_4.CaSO_4$), glauber salt ($Na_2SO_4.10H_2O$), lithium sulphate, alum (with thallium, rubidium, and cesium), hieratite (potassium silico-fluoride), cobalt, zinc, tin, bismuth, lead, copper, iodine, phosphorus.

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[1] H. H. Robinson, The San Franciscan volcanic field, Arizona, *Prof. Paper* 76, U. S. Geol. Survey, 1913.

[2] S. H. Ball, *Prof. Paper* 63, U. S. Geol. Survey, 1908, pp. 67–70.

F. L. Ransome, *Prof. Paper* 75, *idem*, 1911, pp. 60–62.

[3] For summary on this subject see F. C. Lincoln, *Econ. Geol.*, 2, 1907, pp. 258–274.

[4] A. Bergeat, Die Eolischen Inseln, *Abhandl.* k. bayer. Akad. d. Wiss., 2 Cl., 20 Bd., 1 Abth., 1899, p. 193.

For data on Vesuvius see F. Zambonini, Mineralogia Vesuviana, Naples, 1910. The results of Zambonini's late work is abstracted in a paper by Albert Pelloux on The minerals of Vesuvius, *Am. Mineral.*, 12, 1927, pp. 14–21.

Among the fumarolic products of Vesuvius are present most of the minerals just enumerated. Also selenium, magnetite, hematite, hausmannite, pyrite, galena, chalcopyrite, pyrrhotite, covellite, millerite, cuprite, tenorite, and an extensive list of chlorides (among them atacamite and cotunnite) and sulphates.

Yellow incrustations of a vanadate of lead and copper were found in the lava flow of 1631.

Various Ca-Mg-Fe silicates, with sulphides, magnetite, and hematite, occur in ejected limestone blocks which form an exact equivalent to the pyrometasomatic (contact-metamorphic) deposits.

At Katmai, in the fumaroles of the Valley of Ten Thousand Smokes, crusts of specularite, magnetite, pyrite, galena, sphalerite, and covellite in recognizable crystals were found and investigated by E. G. Zies.[1] There are also small amounts of tin and molybdenum. The highest temperature in any of the many fumaroles was about 650° C. In the one depositing magnetite the gas temperature was 230° C. The principal constituent of the fumaroles is usually superheated steam, the heat being derived from the gases escaping from the crystallizing magma comparatively near the surface; but except in the fumaroles, which issue directly from the hot lava, there is likely to be more or less water of meteoric origin mixed with the pure magmatic emanations. Reaction with adjacent rocks or with organic matter underneath the flows may contribute foreign material. The other constituents, which usually make less than 1 per cent, are carbon dioxide, nitrogen, hydrogen sulphide, also hydrocarbons and various acid molecules as HCl, HF, $SO_2$. The gases are not in equilibrium; constant changes take place. Allen concludes that the heat is not primarily caused, nor is any volcanic activity initiated by the oxidation of the gases, though these chemical reactions may contribute to keep up the temperature of the lavas near the surface.[2]

Very small quantities of Zn, Pb, Cu, Cd, and Mo were found in the fresh pumice of the valley floor, and the question is discussed whether or not the metals in the incrustations were derived from this source. Probably this could not apply to the specularite, which is supposed to form as follows: $2FeCl_3 + 3H_2O \rightleftarrows 6HCl + Fe_2O_3$.

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[1] The fumarolic incrustations in the Valley of Ten Thousand Smokes, Nat. Geog. Soc., Contributed Tech. Papers, Katmai ser., 1, No. 3, 1924.

[2] E. T. Allen, Chemical aspects of volcanism, *Jour.* Franklin Inst., 193, 1922, pp. 29–80.

For many analyses of volcanic gases see F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 262–271. An interesting example of fumaroles are the "soffioni" of Tuscany. From these vents steam of a temperature of 190° C. and a pressure of 2 to 4 atmospheres is emitted. They contain much boric acid, doubtless of magmatic origin.

Exhalations of carbon dioxide and nitrogen are observed in many mines located in volcanic districts, thus, for instance, at Cripple Creek, Colorado; Creede, Colorado; Tonopah, Nevada; and the Thames district, New Zealand.

In active volcanoes few mineral deposits are formed. This is probably because the tumultuous eruptions offer less opportunity for a concentration of the metal-bearing emanations. They are present but have no chance for gradual and persistent escape. On the other hand, in the enormous flows of basic lavas which are poured out quietly in some regions, the bulk of the emanations have already been separated during differentiation in the magma chamber and there is little left but the siderophile elements like Ni, Cr, Pt, and elements like copper, which have entered into the composition of the silicates.

Where the meteoric water is abundant and the gas emanations more feeble, hot ascending springs may develop, particularly in volcanoes in process of extinction. Such springs, exemplified by those at Lassen Peak, are usually neutral sulphate solutions, or they may carry acid waters. They produce no important ore deposits.

The mineral deposits of the volcanoes and the lava flows originated close to the surface or say within a few hundred feet of the surface. They should be sharply distinguished from the ore deposits formed by the alkaline hot springs which ascended from far deeper sources though they may be found in the effusive rocks near the surface.

The "volcanic" deposits are few in number and of little economic importance. They comprise native sulphur (with tellurium and selenium) found in the craters of many volcanoes (p. 380); exhalations of boron (sassolite and boric anhydride); local developments of alunite (p. 341) produced by the action of sulphuric acid solutions on potassic rocks; and little else. But in this class should perhaps also be included certain exudation products which evidently originated from the flows themselves, such as topaz and other minerals in rhyolite, and the peculiar local high-temperature tin deposits found also in certain rhyolites of Bolivia, Nevada, and New Mexico, which show the odd combination of colloidal wood-tin, chalcedony, and specularite (p. 659). They have not proved of economic importance.

## THE RELATIONS OF INTRUSIVE MAGMAS TO MINERAL DEPOSITS[1]

A certain pressure, for instance a couple of hundred atmospheres, corresponding to a depth of about 2 kilometers, suffices to force the

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[1] P. Niggli, Versuch einer natürlichen Klassification, etc., *Abhandl. prakt. Geol.*, Bd. 1, 1925, Halle, pp. 1–69. Translated by H. C. Boydell under title of Ore deposits of magmatic origin, London, 1929, 93 pp.

C. S. Ross, Physico-chemical factors controlling magmatic differentiation and vein formation, *Econ. Geol.*, 23, 1928, pp. 864–886. An excellent paper.

volatile constituents to form solutions with the other units of the magma. The full sequence of events in the cooling of such a system can, therefore, be developed only in intrusive bodies.

What happens is represented in the temperature-composition (left) and temperature-pressure (right) diagram (Fig. 11). It shows in greatly simplified form a binary system, *A* being the volatile; *B* the non-volatile component. *A* may represent water; *B* a rock-forming silicate. All complications such as isomorphic crystals or combinations between *A* and *B* are left out of consideration; *B* always separates as the same mineral. As a starting point assume the concentration $x_1$ in which there is very little of the volatile constituent. The left-hand curve represents the freezing points of any mixture of *A* and *B*. Owing to the presence of *A* the freezing point of *B* is lowered to $t_1$ for the concentration $x_1$. *B* crystallizes and the solution gradually changes its composition, following the curve. At first but little of *B* separates out, but a further lowering of *T* causes a large separation of *B*. This represents the flat part of the curve in which the composition of the rest-magma is greatly altered by a slight cooling. When almost all of *B* has crystallized there remains a small amount of fluid in which *A* is present in large excess. The vicinity of $t_2$ represents the close of the magmatic period. The principal part of the melt is crystallized and there remains a solution of very different character. The change is, however, gradual and continuous. Here, near $t_2$, begins the formation of pegmatite solutions. We know that transitions exist from normal igneous rocks to pegmatites, and thence to hydrothermal phenomena.

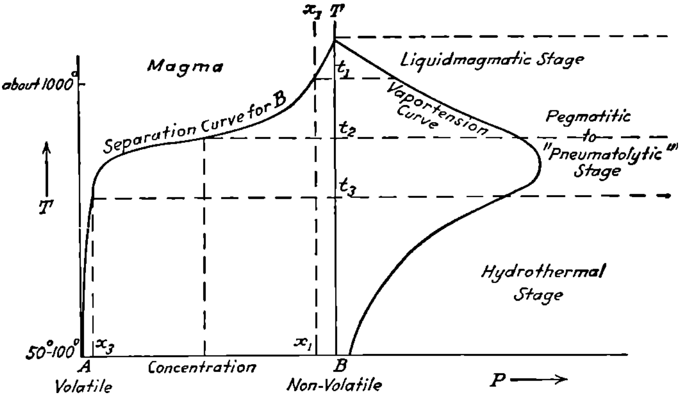

FIG. 11.—Temperature-composition diagram (left) and temperature-pressure diagram (right) for a system of volatile component (*A*), *e.g.*, water and non-volatile component (*B*), *e.g.*, silicate. (*After P. Niggli.*)

The pegmatites stand in closer relation, however, to the igneous rocks than to the hydrothermal products like ore-bearing veins. The pegmatite stage or, as it may also be called, the stage of the salic extracts (Si, Al), develops following the curve. The main part of silica and alkali-aluminum silicate separates out rapidly as the temperature is lowered slightly. When only small parts of more aqueous rest-magma remain, the crystallization begins to follow the steep part of the curve, that is, it develops slowly and the solubility changes but slightly with sinking $T$. This corresponds to the later phase of pegmatites, often characterized by a drusy development. Minerals like albite, tourmaline, beryl, phosphates, cassiterite, and columbite, separate. The beginning of the period of the pegmatite druses would correspond to the turn in the curve between $t_2$ and $t_3$. Still later follows the stage of the ore minerals, zeolites, and carbonates; that is, the beginning of the hydrothermal phase, with slow crystallization from a residue, now rich in the principal volatile, *i.e.*, water. These emanations are products of magmatic differentiation, but they are not magmas.

It should be understood that while many new minerals are deposited directly by the escaping magmatic products, a host of other minerals are formed by reactions between these products and already deposited minerals, or minerals contained in the surrounding rocks. Such replacements, of infinite variety, may be observed in the igneous rocks themselves, in the pegmatites, and in all other mineral deposits of magmatic affiliations.

No sudden separation of the gaseous phase will take place; in a closed system, under heavy pressure this would not be expected; it could occur only if the vapor tension of the solutions is greater than the external pressure.

Now, let us turn to the right-hand side of Fig. 11 which represents the vapor tensions belonging to a saturated solution at given temperatures. During the beginning of the crystallization the vapor pressure will be low; there are much liquid and little of the volatile constituents; as the temperature falls much of the non-volatile matter is crystallized and the water becomes concentrated in the residual liquid. But to increase the water in the fused rest of the silicate means an enormous increase in the pressure needed to hold the water in the liquid. In the part of the curve indicated as the "pneumatolytic" stage the pressure attains a maximum. The pressure does not increase indefinitely, for the decrease in equilibrium temperature resulting from the increased lowering of the freezing point tends to decrease the vapor pressure.

The maximum pressure, according to Niggli, indicates an interval during which the residual solution is above its critical temperature; this he calls the pneumatolytic stage.[1] It immediately follows the

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[1] There exists a tendency to extend unduly the "pneumatolytic" stage; that is, the stage formed by gaseous fluids. For instance, many authors, without justification, refer any deposit containing tourmaline and cassiterite to this group.

pegmatitic stage; during this period the solutions push outward from the magma at all places where the pressure is reduced. The intrusive power of the hydrothermal solutions is weaker; they will advance only along fractures.

Morey looks at this question somewhat differently. According to him:

> The pegmatites correspond to the maximum pressure or about 575° C. On further cooling crystallization will continue but is of a different type. The earlier crystallization is mostly of silicate minerals; following this we get the crystallization of the accessory components: tourmaline, topaz, fluorite, apatite, cassiterite, and other high-temperature vein minerals. This range may lie between 500° and 300° C. The liquid now contains far more water than at the maximum pressure but the vapor pressure will be much reduced and at the critical temperature of water will have a pressure much less than that of pure water. (Critical point for $H_2O$: T 374°; P 218 at.)
>
> The addition of any soluble substance will raise the critical temperature. As the cooling magma eliminates one component after the other, the residual magma will at all times contain enough material of less volatile character (borates, phosphates, chlorides) to hold the compounds of greater volatility in solution. The critical phenomena are never shown by the saturated solutions formed by the cooling magma.[1]

Day and Allen,[2] as well as Morey, say that any pressure which would condense the steam would drive the water back into the magma. The liquid solutions have no power to rise to the surface.

While all this may be true it seems, however, that the solutions do rise to the surface and carry with them large amounts of silica and other substances. The driving power would be furnished by the continuous escape of gases pushing the column of condensed solution ahead of it.

The figures compiled by Vogt[3] show that there are many substances—$SiH_4$, $SiF_4$, $SiCl_4$, $SnCl_4$, $SO_3$, $TiCl_4$, $AsCl_3$, $AlCl_3$—which have a critical temperature approximately similar to that of water. Some five and six valence chlorides and fluorides of W, Mo, U, Ni, Ta have extremely low boiling points; $WF_6$ boils at 19° C.

Many of the common soluble compounds, chlorides of the alkalies, for instance, have, however, extremely high critical temperatures.

## GENERAL REVIEW OF ORE DEPOSITS WITH MAGMATIC AFFILIATIONS

Before examining the probable nature of the fluids which form ore deposits of the magmatic cortege, necessarily a more or less speculative

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[1] G. W. Morey, *Jour. Geol.*, 32, 1924, pp. 291–295.

[2] Symposium on the temperatures of hot springs, etc., *Jour. Geol.*, 32, 1924, Nos. 3, 4, 5, and 6.

[3] *Econ. Geol.*, 21, 1926, pp. 207 *et. seq.*

discussion, we may present, more in detail, some facts regarding the various types.

The types of deposits are in brief: (1) liquid-magmatic, (2) pyrometasomatic, (3) pegmatitic, (4) hypothermal, (5) mesothermal, and (6) epithermal.

The liquid-magmatic deposits contain few and simple minerals, mainly oxides and sulphides of the siderophile elements: Fe, Ni, Cr, Ti, Pt, Cu; among the "mineralizers" phosphorus is most prominent; there is little quartz, but otherwise the gangue minerals are the common rock-forming minerals. Naturally replacement is a subordinate process; the ores crystallized from melts.

The pyrometasomatic deposits formed by direct action of magmatic emanations (gaseous or liquid) on limestone contain, likewise, simple minerals of the oxide and sulphide types; magnetite and specularite are abundant but the simple sulphides, like pyrite, pyrrhotite, sphalerite, arsenopyrite, chalcopyrite, and bornite, are also plentiful. Owing to their occurrence in carbonate rocks and to additions of magmatic Fe and Si, the gangue minerals are mainly silicates; quartz is subordinate. Replacement is the chief process. Much material is carried away, as well as added.

The pegmatites are regarded as intrusive residual magmas pressed out from the almost congealed melt. They carry quartz, feldspars, micas, and pyroxenes, and the quartz is followed by a characteristic association of lithium, tin, tungsten, columbium, tantalum, cerium metals, thorium, beryllium. Oxides and complex silicates, likewise phosphates and tantalates are present. There are a few sulphides and arsenides but rarely pyrite. Large amounts of fluorine, boron, and sometimes chlorine enter into the minerals. The pegmatites form dikes and there is some replacement in the adjoining rock. In the pegmatites themselves there is a succession of minerals of later introduction, such as albite, tourmaline, etc. The pegmatites are a well-defined type but transitions to igneous rocks and to veins are found.

In the hydrothermal deposits (4, 5 and 6) Cr, Ti, V, Zr, U, Ce, Ta, and Pt are absent. There are some exceptions to this rule.

In the hypothermal deposits the metals are of the chalcophile group, such as Fe, Sn, W, Au, Ag, Cu, Zn, Pb. Cerium minerals occur in some rare types. The oxides are less prominent. There are more complex minerals and more evidence of a series of deposition corresponding to slowly decreasing temperature. Sulphur, arsenic, phosphorus, tellurium, chlorine, and boron are present. If the deposit occupies cavities the material is mainly quartz. The silica is not inclined to enter vigorously in replacement. Replacement is, however, extremely important; the rocks are penetrated by fluids which introduce large quantities of substance and which also extract large amounts. Gold and silver appear in notable quantities.

In the mesothermal deposits the diversity becomes greater. The metals are Fe, Cu, Zn, Pb, Au, Ag, Ni; Sn and W in places; the oxides are subordinate. Arsenic, antimony, fluorine, sulphur, and tellurium are present, much less commonly chlorine, boron, and phosphorus. Quartz is the dominant gangue mineral and, as before, prefers the open spaces. Replacement is very prominent and, as before, large quantities of material are added and removed. Replacement processes go on both in rock and in filling. The series of minerals is still further extended by complex sulpharsenides and sulphantimonides, which usually appear toward the close of the mineralization.

The epithermal deposits contain Au, Ag, Fe, Cu, Pb, Zn, Hg, rarely Sn, W, and Ni. Sulphur, fluorine, antimony, arsenic, selenium, and tellurium are present; chlorine, boron, and phosphorus usually absent. Quartz is the prominent gangue mineral; as before, it favors the open spaces. Filling is increasingly prominent but replacement is also important. Large quantities of material are abstracted and introduced. Replacement processes go on both in rock and in filling.

It is held that the hypothermal and the mesothermal deposits are formed by emanations from intrusive rocks, generally batholiths. It is also thought that the epithermal deposits are formed by emanations from very deep magma chambers in which the differentiation of the lavas is proceeding.

## THE ZONAL THEORY

The zonal theory seeks to derive some laws from the arrangement of metals in ore deposits with reference to the parent igneous body. This may find expression in a lateral or vertical direction or both. Foreshadowed by old observations in the mining districts of Europe, it seems to have been brought out independently by L. deLaunay,[1] J. H. Collins,[2] George Waller,[3] and J. E. Spurr.[4] Of these men Spurr has discussed the subject most extensively. W. H. Emmons[5] has also con-

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[1] Les variations de filons métallifères en profondeur, *Rev. gén. sci.*, 11, 1900, pp. 575–580.

[2] Notes on the principal lead-bearing lodes of the West of England, *Trans.*, Royal Geol. Soc. Cornwall, XII, pt. 8, 1902, p. 713 (with map).

[3] Report on the Zeehan silver-lead mining field, *Bull.* Tasmania Geol. Survey, 1904, p. 24.

[4] The ore magmas, 2, 1923, p. 611. A theory of ore deposition. *Econ. Geol.*, 2, 1907, pp. 781–785.

[5] Primary downward changes in ore deposits, *Trans.*, Am. Inst. Min. Met. Eng., 70, 1924, pp. 964–992; Relations of metalliferous lode systems to igneous intrusions, *idem*, 74, 1926, pp. 29–70. Contains a full account of mining districts in which a zonal arrangement is shown.

tributed much valuable material to this subject, and many other authors have participated.

The following table copied from W. H. Emmons gives a schematic expression to the zonal theory on the assumption (which is not proved) that all veins originate from batholithic intrusions. It is understood the whole succession would not be shown in any vein, though it is believed that some vein systems may have a vertical range of 10,000 feet or more; it is also understood that in some veins several of the zones may be lacking.

A RECONSTRUCTED VEIN SYSTEM FROM SURFACE TO NEAR BATHOLITH ROOF

(*After W. H. Emmons*)

Surface

Barren, 1.—Barren zone, chalcedony, quartz, barite, fluorite, etc. Some veins carry a little mercury, antimony, or arsenic.

Mercury, 2.—Quicksilver veins, commonly with chalcedony, marcasite, etc. Barite-fluorite veins.

Antimony, 3.—Antimony ores—stibnite often passing downward into lead, with antimonides. Many carry gold.

Gold and Silver, 4.—Bonanza ores of precious metals. Argentite, antimony and arsenic minerals common. Silver minerals, some copper, lead and zinc sulfides, quartz, calcite, rhodochrosite, adularia, alunite, etc.

Barren, 5.—Most nearly consistent barren zone, represents the bottoms of many Tertiary precious metals veins. Quartz, carbonates, etc. with pyrite and small amounts of other sulfides.

Silver, 6.—Argentite veins, complex antimony silver sulfides, stibnite, etc. Galena veins with silver. Commonly silver decreases with depth. Quartz gangue, siderite common, often increasing with depth.

Lead, 7.—Galena veins, commonly with some silver. Sphalerite generally present, increasing with depth. Chalcopyrite common. Gangue is quartz and often carbonates (Fe, Mn, Ca).

Zinc, 8.—Sphalerite veins with some lead and chalcopyrite, quartz gangue.

Copper, 9.—Tetrahedrite veins, commonly argentiferous, chalcopyrite present. Some pass downward into chalcopyrite. Enargite veins generally with tetrahedrite and tennantite.

Copper, 10.—Chalcopyrite veins, generally with pyrite, often with pyrrhotite. The gangue is quartz and in some places carbonates. Some pass downward into pyrite and pyrrhotite with a little chalcopyrite. Generally carry silver or gold.

Gold, 11.—Gold veins with quartz, pyrite, and commonly arsenopyrite and chalcopyrite. At places zones 10 and 11 are reversed.

Bismuth, 12.—Bismuthinite and native bismuth with quartz and pyrite, etc.

Arsenic, 13.—Arsenopyrite with chalcopyrite and often tungsten ores.

Tungsten, 14.—Tungsten veins with quartz, pyrite, chalcopyrite, pyrrhotite, etc. Arsenopyrite is commonly present.

Tin, 15.—Cassiterite veins with quartz, tourmaline, topaz, etc.

Barren, 16.—Quartz with small amounts of other minerals.

In many veins only one of the upper portions (2, 3, or 4) can be observed. Few of these veins have been developed beyond the barren zone, 5. Nearly all of the zones below 5 have been observed at many places grading one into another.[1]

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[1] W. H. Emmons, *op. cit.*, 70, p. 983.

The precipitation of a given metal in a complex solution is a problem dependent on no single factor. There are involved: Temperature, pressure, solvent, concentration, relative abundance, reactions in the solution as precipitation progresses, character of country rock, and probably several other factors. High-temperature minerals like tourmaline, topaz, magnetite, and cassiterite with some sulphides would generally be deposited near the magma. Many examples substantiate this. As the temperature drops or as the distance from the magma increases, zinc, copper, gold, lead, and silver would be successively deposited. The veins far away from the intrusion would be mainly filled by barren gangue. The deposits close to the magma would be called *perimagmatic;* those in the surrounding areas, *apomagmatic;* and those far away from the intrusive centers, *telemagmatic* (Bergeat).

The deposits of Cornwall, England, furnish probably the best illustration of the zonal theory, with tin in depth, copper in the cooler slates, and lead and zinc deposits still farther away. At Bingham, Utah, the copper deposits center in or near the monzonitic intrusion while lead deposits occur farther away. In the Leadville vein at Bingham a low-temperature lead-zinc deposit changes at a depth of about 2,000 feet into tennantite and chalcopyrite ore. Many lead veins change at slight depth into poor sphalerite-siderite-pyrite ores. In the epithermal veins, the base metals are quite subordinate and the gold and silver seem to have been precipitated within a couple of thousand feet of the surface. These are what J. E. Spurr[1] calls "telescoped" veins, which are said to be produced by gradually increasing temperature and with indefinite or superimposed succession of minerals.

Upon close examination many disconcerting facts appear, which suggest that the zonal theory in its present form is a rather weak structure. Tin is supposed to be confined to the high-temperature deposits, but in places (Potosi and Chocaya, Bolivia) it appears far up in the mesothermal and epithermal zones. Tungsten should act more or less like tin but this metal also forms important deposits (Nederland, Colorado) in a zone pretty close to the epithermal veins, and it reaches the surface in large amounts in hot spring deposits (Uncia, Bolivia). Lead is said to be a metal deposited at low temperatures but three of the largest deposits of the world are of the hypothermal class: Broken Hill, New South Wales, Sullivan mine, British Columbia, and Alta mines Montana. Some lead-bearing veins are impoverished at slight depth; others continue rich over a vertical interval of over 4,000 feet. If we assume, with Spurr, that the Mississippi Valley deposits are of magmatic affiliations, lead reaches the surface in abundance, thousands of feet above the supposed magma basin and where, according to Spurr, we should expect only "barren gangue." A similar argument, on similar

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[1] The ore magmas, 1, 1923, pp. 292–308.

assumption, applies to zinc. The problem becomes still more complicated when the influence of country rock is considered. The solutions may deposit galena in limestone but at the same time entirely similar solutions may form pyritic ores in rocks of different composition. Finally, in the epithermal veins there is no zonal arrangement, except a common impoverishment in depth; they are too far from their starting point. They can be correlated with no exposed igneous parent masses.

## SUCCESSION OF MINERALS IN ORES[1]

In any one ore the minerals have usually separated out in a certain well-defined succession, the later minerals in part replacing the earlier. As there are usually also overlapping and repetition the problem may be studied best in ores which show both crustification and replacement.

There is reason to believe that in all ore deposits of magmatic affiliations somewhat similar laws apply.

In magmatic and pyrometasomatic ores the succession is simple: Gangue minerals are early; ore minerals are late. The succession will be silicates, magnetite, specularite, pyrite, pyrrhotite, sphalerite, chalcopyrite. Processes of "unmixing" are common, finding expression in inclusions of chalcopyrite and pyrrhotite in sphalerite and of pentlandite in pyrrhotite.

In hypothermal and mesothermal ores the succession is generally as follows: Quartz (usually long continued and overlapping), pyrite, arsenopyrite, cobalt and nickel arsenides, cassiterite, wolframite, molybdenite, bismuthinite, pyrrhotite, sphalerite, enargite, tennantite, chalcopyrite (overlapping), bornite, galena, chalcocite, argentite, silver sulpharsenides and antimonides, lead sulphantimonides. Overlapping and repetition may change the succession in places. Naturally all these minerals are hardly ever present in one deposit. The later minerals may replace any of the earlier minerals. Where gold appears it is usually one of the latest minerals. "Unmixing" is of considerable importance especially in galena, pyrrhotite, sphalerite, and stannite.

In epithermal veins the succession is usually quartz, pyrite, sphalerite, chalcopyrite, galena, tetrahedrite, finally argentite, gold, electrum, and complex sulphantimonides and sulpharsenides of silver and lead. Unmixing is of less importance; reversal of the order sphalerite-chalcopyrite is rather frequent;[2] reversal of the order sphalerite-galena is rare. Later generations of different gangue and ore minerals may appear.

Generally the order of deposition is: silicates, oxides, and sulphides, the groups falling in the order of the decreasing electrode potentials of the elements. In any group of oxides and sulphides the elements seem

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[1] W. Lindgren, Magmas, dikes, and veins, *Trans.* Am. Inst. Min. Met. Eng., 74, 1926, p. 88.

[2] Horatio Brown, The mineral zones of the White Cross district, etc. (San Juan), Colorado School of Mines, *Mag.* 15, March, 1926.

to be deposited in the decreasing order of their potentials. This is apparent from the table presented by Butler and Burbank.[1] Elements with electrode potentials above +1 occur only in oxygen combinations. Elements with negative potentials (metallogenic elements) occur only as sulphides or native metals. There is also an intermediate group. Thus, molybdenum, zinc, and iron sulphides would form early; silver and mercury late, and lead and copper between these extremes. It is assumed that the deposition takes place during gradually decreasing temperature.

Several other serial properties have been suggested, such as increasing atomic weight, increase in metal content, decrease in sulphur, decrease in hardness, and decrease in heat of formation.[2]

Newhouse[3] points out that any sulphide being replaced by another must of necessity be more soluble, or replacement would not take place.

Thus it would appear that solubility is the most important factor. This was confirmed by Wells[4] in an important paper. It was shown that the order of precipitation of sulphides was practically the same as determined in 1888 by E. Schürmann and very similar to the solubility table of Weigel for sulphides in water. The order ascertained is Hg, Ag, Cu, Bi, Cd, Pb, Zn, Ni, Co, Fe (ferrous), and Mn. The solution of a salt of any of these metals will be decomposed by the sulphide of any succeeding metal, and the first metal precipitated as a sulphide. A mixture of two metallic salts yields, by fractional precipitation, an initial precipitate containing the sulphides of both metals; but, as a rule, if the mixture is heated or allowed to stand, one sulphide largely or wholly dissolves. Note analogy with replacement.

It seems, therefore, that *solubility* is the main factor in the succession of sulphides in mineral deposits, although other factors may at times interfere with this law. In complex sulphides the explanation is naturally more difficult.

## CHARACTER OF SOLUTIONS FORMING MINERAL DEPOSITS

The quicksilver and stibnite deposits are satisfactorily proved to have been formed by ascending hot alkaline springs, as at Sulphur

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[1] B. S. Butler and W. S. Burbank, Relation of electrode potentials to formation of hypogene mineral deposits, *Trans.*, Am. Inst. Min. Met. Eng., *Yearbook*, 1929, pp. 341–353.

[2] I. de Magnée, Essai d'explication physicochimique du phénomène de la répartition zonaire des minérals d'origine magmatique, *Ann.* 55, *Mém.*, Soc. géol. de Belgique, 1932, pp. M17–M35.

G. Gilbert, The relation of hardness to the sequence of ore minerals, *Econ. Geol.*, 19, 1924, pp. 668–673.

[3] W. H. Newhouse, The time sequence of hypogene ore mineral deposition, *Econ. Geol.*, 23, 1928, pp. 647–659.

[4] R. C. Wells, Fractional precipitation of some ore-forming compounds at moderate temperatures, *Bull.* 609, U. S. Geol. Survey, 1915, 46 pp.

Banks, California, and at Steamboat Springs, Nevada. The deposition proceeds almost under our eyes; small amounts of gold, silver, copper, lead, and zinc are also found in the sinter. Becker proved that the mercury and the antimony were contained in the waters as double alkaline sulphides; and he suggested that most of the other metals, perhaps excepting silver, were in similar combination.

Regarding other deposits inference and speculation must be used. The epithermal deposits are so closely connected with hot-spring action and with deposition of colloidal silica that it is not possible to doubt that they were formed by similar ascending springs.

For the more deep-seated deposits it is convenient to begin with those of the liquid-magmatic class. They were probably formed by consolidation from a hot melt, itself a product of magmatic differentiation; there were other substances present, however, like phosphorus in apatite, which must have materially reduced the melting point of the ores. Regarding sulphide melts similar considerations apply. The presence of certain mineralizers is indicated pretty clearly in the occasional occurrence of minerals such as chlorite, epidote, and garnet foreign to the normal succession of crystallization of igneous rocks.

The pegmatitic residual magmas probably contained a large amount of water and many other mineralizers; they were evidently in aqueo-igneous fusion, intermediate between a magma and an aqueous solution.

The solutions from which the hydrothermal ore deposits were formed are more nearly related to the pyrometasomatic extracts than to pegmatite magmas. Undoubtedly they were often quite concentrated. From the fact that in most kinds of replacements large amounts of material, such as difficultly volatile compounds of potassium, sodium, calcium, magnesium, and other substances were consistently carried away we may conclude with confidence that the agents which effected this were mainly liquids and not gases. This applies to the pyrometasomatic as well as to the tin deposits and other hypothermal and following classes. Gases could not effect this enormous transfer.

It is not believed, therefore, that "pneumatolytic" processes played an important part in the formation of such deposits. From the work of Day, Allen, Sheperd, and others, it would seem probable that many of the volcanic gases are acid when given off, but are in unstable equilibrium, and could not have preserved this acid character after having been long in contact with the surrounding rocks. It is believed that they were in small part oxidized to sulphates, for instance;[1] but in larger part they were composed of hydrogen sulphide, alkaline sulphides, and various halogen compounds. Doubtless, however, free carbon

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[1] B. S. Butler, Some relations between oxygen minerals and sulphur minerals in ore deposits, *Econ. Geol.*, 22, 1927, pp. 233–245.

dioxide was present Becker's view of the double alkaline sulphides as important constituents of the solutions probably holds good.

The metallic sulphides are very slightly soluble in water or aqueous solutions, but some of them, particularly As, Sb, and Hg, can be dissolved in alkaline fluids as shown by the experiments of Becker, Lenher, Ditte, Grout, Ravicz, and Horatio Brown.

The sulphides have rather high melting points: ZnS 1650°, FeS 1174°, PbS 1114°, $Ag_2S$ 837°, $Cu_2S$ 1130°, HgS 1450° (at 120 atmospheres), $Sb_2S_3$ 550°, $As_2S_3$ about 307°, $Na_2S$ 970° C.

A replacement ore ($FeS_2$, PbS, ZnS) from the Sullivan Mine, British Columbia, melted at 1380° and remained liquid to 1200° C.[1]

Fusing these sulphides with $Na_2S$ reduced the melting point to from 560° to 660° C. This held good for Pb, Zn, Fe, Ni, Co, Mn, Mo, As, Sb, Ag, Au, and Cd.

Water breaks down the double sulphides and develops colloidal sulphide solutions. Silica gel was formed on fusing quartz with $Na_2S$ and dissolving in water. Gold nuggets were fused with $Na_2S$; the melt dissolved in water and colloidal gold was obtained. As is well known, sulphides can be dispersed by the action of $H_2S$ and in many cases the action is instantaneous and striking (chalcocite, arsenic sulphide).[2] By hydrolysis of ferric chloride, iron can be easily carried as ferric hydroxide sol in high concentrations.

All these experiments suggest strongly that the metals may have been largely transported as colloidal solutions, or sols.[3] Colloidal silica, probably present, may have exerted a protective action in keeping the metals in a dispersed state.

J. E. A. Kania[4] conducted experiments using sodium sulphide and various metallic sulphides at temperatures up to 90° C. On fragments of limestone, pyrite, marcasite, and chalcopyrite were obtained in crystallized form. It was shown that a mineral may be of colloidal origin although it has no colloform structure.

It is probable, indeed, that the silica has been carried in colloidal solution, and has been precipitated as a gel which almost immediately would crystallize to quartz. The well-known association of gold and quartz is most easily explained by colloidal processes. Silicification is not the most prominent metasomatic process and this is probably because the

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[1] Horace Freeman, *Eng. Min. Jour.-Press*, Dec. 19, 1925.

[2] Tolman and Clark, *Econ. Geol.*, 9, 1914, pp. 559–592.
Clark and Menaul, *idem*, 11, 1916, pp. 37–41.
Young and Moore, *idem*, 11, 1916, pp. 349–365.

[3] H. C. Boydell, The rôle of colloidal solutions, etc., *Trans.*, Inst. Min. Met., London, 1924–1925, p. 184.

[4] *Bull.* Geol. Soc. Am., 43, 1932, p. 187.
Doctor's thesis, Massachusetts Inst. Tech., 1930.

colloidal silica found difficulty in entering and replacing the country rock. By an osmotic process the solutions separated into a colloidal and an electrolytic part.

We often think of colloidal processes as confined to low temperatures but probably our opinions in this matter should be revised. Epithermal vein quartz often shows colloform deposition and recrystallization to quartz; but some mesothermal deposits like those of Tintic show similar phenomena. Takeo Kato (p. 658) describes colloform deposition of quartz and cassiterite in a mesothermal or hypothermal copper-tin vein. W. V. Smitheringale[1] investigated a quartz vein in British Columbia which carries magnetite, specularite, pyrite, chalcopyrite, and other sulphides. The quartz and the pyrite in this deposit of probable hypothermal affiliation show unmistakable colloform characteristics. These examples serve to illustrate the possible range of colloidal phenomena.

The vein-forming solutions were probably liquid from the start, and often highly concentrated. They contained water as the principal constituent and were at all times in a mobile non-viscous condition. In their constitution colloidal dispersions entered prominently.

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[1] Mineral association at the George gold-copper mine, Stewart, B.C., *Econ. Geol.* 23, 1928, pp. 193–208.

# CHAPTER XI

## FOLDING AND FAULTING[1]

### FOLDS

Sedimentary beds and ore deposits contained in them are often bent, corrugated, and folded in a more or less complex manner. Extensive folding is usually effected by horizontal or "tangential" thrust, but minor bends and monoclines (Fig. 12) may originate by thrust in any direction. In extreme cases any fold or bend may pass over into a break or fault. In folding on a large scale it is necessary that the sedimentary complex have beds of sufficient strength (competent beds) to transmit the thrust and support the structures; if the complex is plastic it will be deformed by flowage and no regular folds will result.

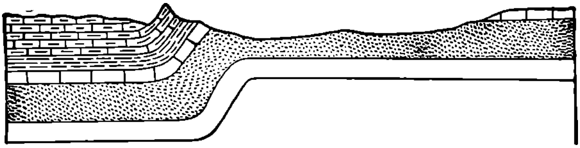

FIG. 12.—Monocline near Gallup, New Mexico. (*After E. Howell.*)

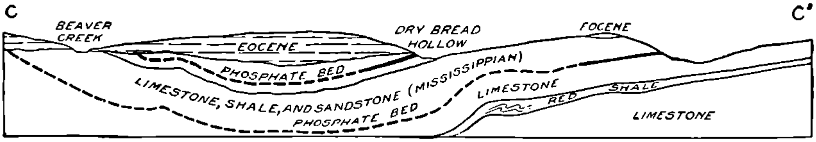

FIG. 13.—Open syncline showing carboniferous phosphate beds unconformably covered by Eocene beds. Beaver Creek, Utah. (*After E. Blackwelder, U. S. Geol. Survey.*)

Folds are synclinal (Figs. 13 and 14), trough-like; or anticlinal (Fig. 15), shaped like a saddle. A plane which bisects the average angle between the limbs is called the axial plane of the fold. By complex movements the axial plane may become a curved surface. If this

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[1] E. de Margerie and A. Heim, Les dislocations de l'écorce terrestre, Zürich, 1888, pp. 49–63.

Bailey Willis, The mechanics of Appalachian structure, *Thirteenth Ann. Rept.*, U. S. Geol. Survey, 1894, pp. 211–281.

C. R. Van Hise, Principles of North American pre-Cambrian geology, *Sixteenth Ann. Report*, U. S. Geol. Survey, pt. 1, 1896, pp. 589–633.

F. H. Lahee, Field geology, 3d ed., New York, 1931.

axial plane is vertical the limbs dip at like angles; if the axial plane is inclined the limbs have unequal dips and the fold is asymmetrical. In close folding the limbs dip steeply (Fig. 16). When the axial plane of folds inclines strongly in one direction we speak of overturned folds; and these by further compression may easily pass over into overthrust faults (Figs. 16–17), causing a part of the folded series to slide over the other. In flat overthrust faults the horizontal movement may amount to many miles.

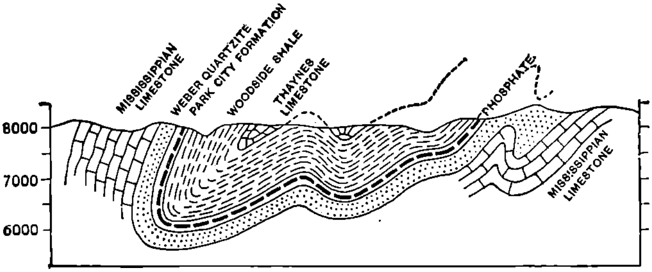

FIG. 14.—Eroded syncline, Georgetown Canyon, Idaho, showing phosphate bed. (*After H. S. Gale, U. S. Geol. Survey.*)

Synclines and anticlines extend naturally about perpendicularly to the direction of compressive stress. Their direction is indicated by a line passing through all the highest or lowest points of a given stratum.

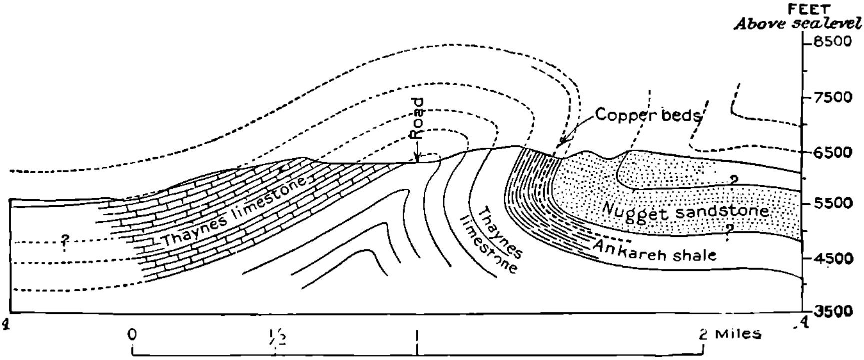

FIG. 15.—Eroded anticline, Montpelier, Idaho, showing bending of copper-bearing beds of Triassic age. (*After H. S. Gale, U. S. Geol. Survey.*)

These crest lines or trough lines have usually a distinct dip; the angle of this line with the horizontal is called the *pitch* of the fold. Minor plications on the limbs often indicate the pitch of the fold. Thrusts in two directions result in cross-folding with the development of bending in forms known as canoes, domes, and basins.

When beds are lifted in dome shape so that they dip away from a central point they form a *quaquaversal.*

In sharp folding of a sedimentary complex, the strata become thicker by compression at the points of greatest bending (Fig. 18). The harder strata of sandstone or limestone will yield to tension by breaking or tearing; the softer, shaly strata do not break, but yield to deformation.

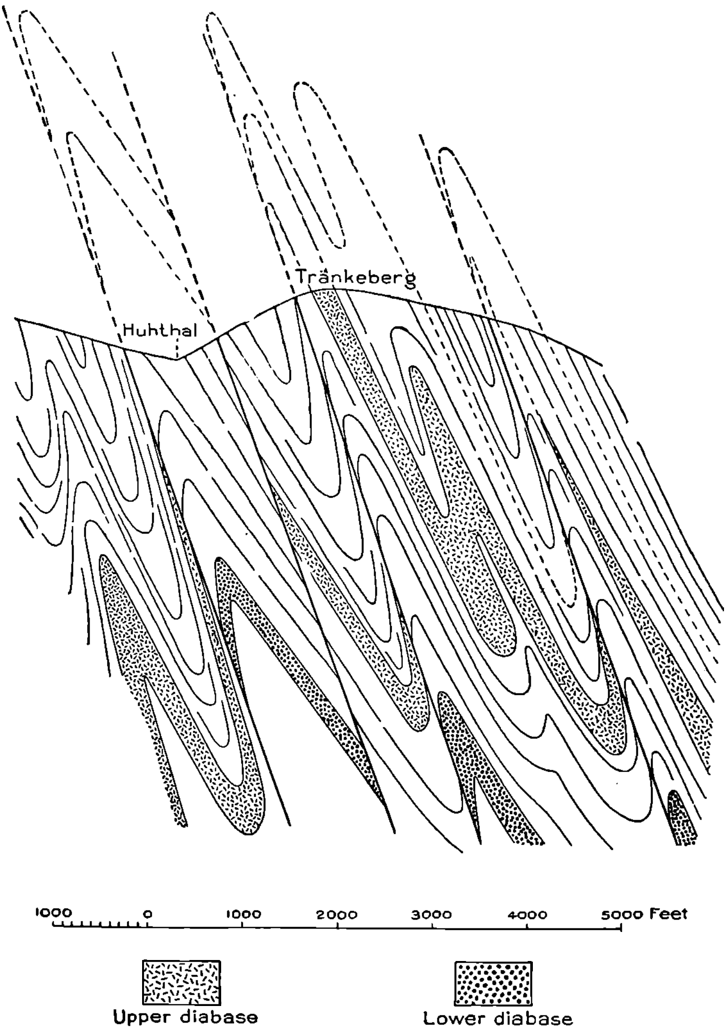

Fig. 16.—Close folding with overthrusts and thickening of strata by duplication. (*After M. Koch.*)

Strata of differing hardness may slide over one another at such points; and openings may be produced which, for instance, may later be filled with quartz. In slates and crystalline schists which have been deformed at great depths by flowage the harder or "competent" layers, like quartz

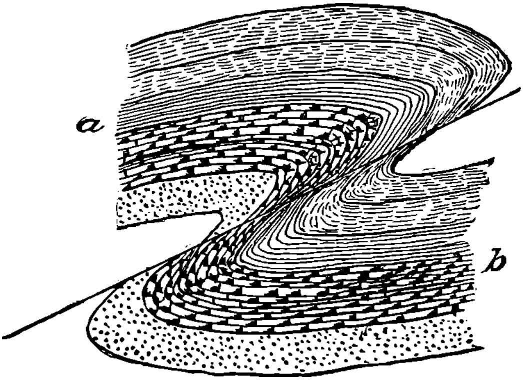

FIG. 17.—Diagram showing development of an overthrust fault from a fold. (*After C. R. Van Hise, U. S. Geol. Survey.*)

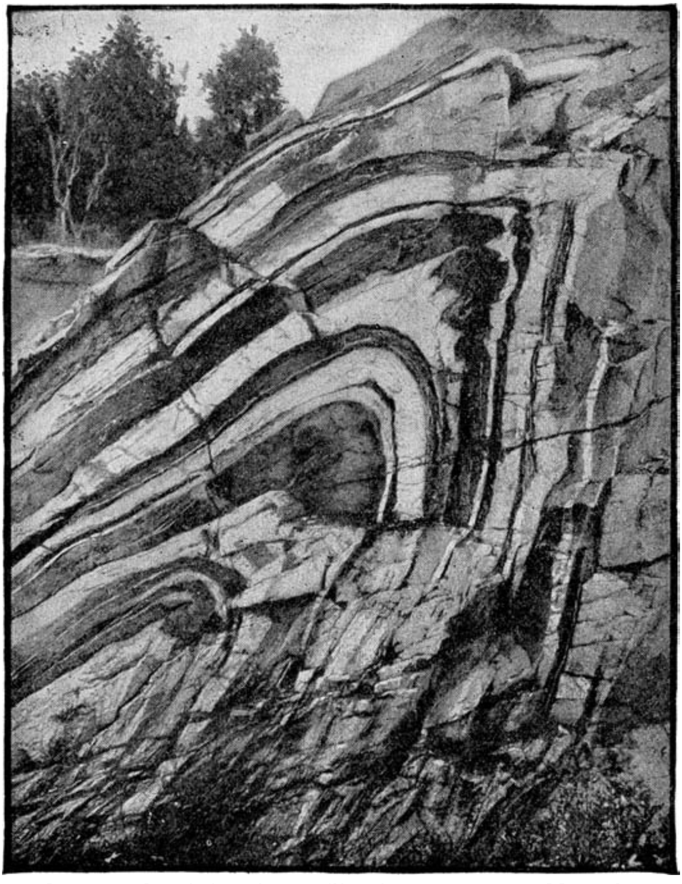

FIG. 18.—Overturned anticline of crystalline limestone, Lenox, Massachusetts, showing thickening and breaking of strata at points of bending. (*After T. Nelson Dale, U. S. Geol. Survey.*)

veins, may be corrugated in an extraordinarily complicated manner. Some quartz veins of Nova Scotia, called "barrel quartz," are believed to owe their form to such conditions (Fig. 19).

Drag folds are minor corrugations in an incompetent bed between two harder competent beds.

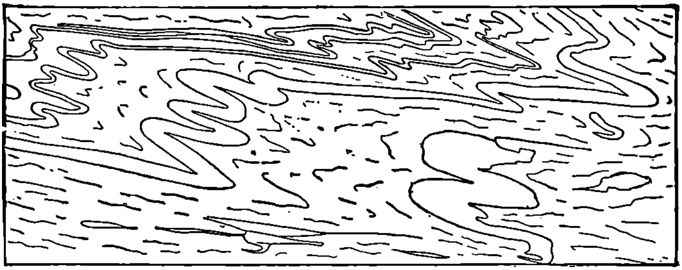

Fig. 19.—Gneiss with corrugated veinlets of quartz. (*After C. R. Van Hise, U. S. Geol. Survey.*)

## FAULTS[1]

Sedimentary beds and deposits, as well as deposits of later origin which persistently follow a certain horizon in a sedimentary series, are sometimes abruptly cut off by structural planes.

When such an occurrence is found it is safe to conclude that the interruption is due to a fault—that is, to a fracture along which movement has taken place—and that the continuation of the deposit exists somewhere beyond this break (Fig. 20).

In the case of epigenetic deposits not closely following the original lines of structure in the rocks, such a conclusion is often, but not always justified. The interruption of the ore-body may be due to an actual post-mineral dislocation, or it may be caused by a cessation of mineralization on account of structures existing before the mineralization began. The replacement of limestone by galena or shale by pyrite (Fig. 21) may stop suddenly at a clay-coated seam, which offered a barrier to the solutions. A vein-filled fissure may terminate abruptly or split up within a

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[1] J. E. Spurr, Measurements of faults, *Jour. Geol.*, 5, 1897, p. 723.

J. E. Spurr, Geology applied to mining, New York, 2nd ed., 1926.

F. L. Ransome, The direction of movement and the nomenclature of faults, *Econ. Geol.*, 1, 1906, p. 777.

C. F. Tolman, Jr., *Econ. Geol.*, 2, 1907, pp. 506–511.

C. F. Tolman, Jr., Graphic solution of fault problems, *Min. and Sci. Press*, June 17, 1911, *et seq.* Reprinted, San Francisco and London, 1911.

H. F. Reid, Geometry of faults, *Bull.* Geol. Soc. Am., 20, 1909, pp. 171–196.

H. F. Reid, W. M. Davis, A. C. Lawson, and F. L. Ransome, Report of the committee on the nomenclature of faults, *Bull.* Geol. Soc. Am., 24, 1913, pp. 163–186.

C. K. Leith, Structural geology, New York, 1923.

F. H. Lahee, Field geology, 3d ed., New York, 1931, Chaps. VII and VIII.

few feet upon encountering softer and more plastic rocks, such as clay shales, thick gouge seams, or soft tuffs. A vein traversing formations of varying hardness often suffers abrupt deflection at rock contacts. It may also be deflected by encountering older dikes or fissures, either barren or filled with vein material (Fig. 22).

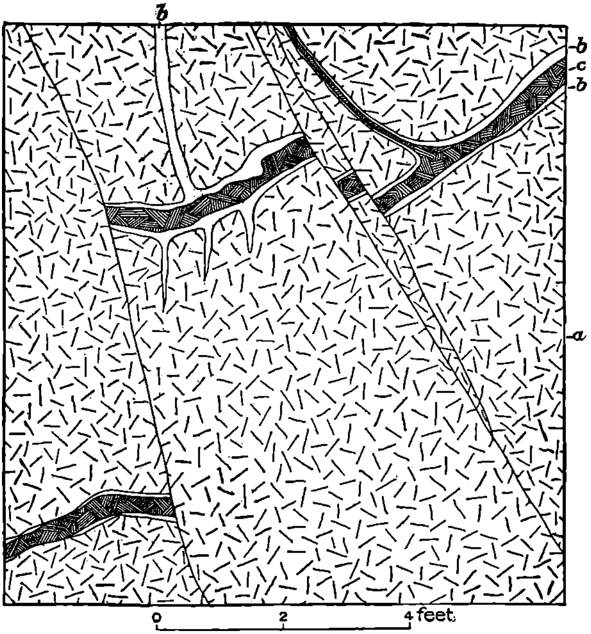

FIG. 20.—Faulting of Mendota vein, Silver Plume, Colorado. *a*, granite; *b*, quartz; *c*. galena and sphalerite. (*After J. E. Spurr, U. S. Geol. Survey.*)

FIG. 21.—Sketch showing replacement of shale by pyrite. Natural size. The small fissures are older than the pyrite and are crossed by its banded structures. (*After F. L. Ransome, U. S. Geol. Survey.*)

The distinction between faults and deviations is most important. The appearance of detached fragments of the ore—usually termed "drag" —on the faulting plane (Fig. 23), the direction of the striations, and the interrelations of dip and strike of faulting, fissure, and ore-body are all valuable data which must be interpreted in each case.

There are many geometrical rules for the finding of a faulted ore-body, but they are of little value unless the character of the interruption is known. Each case must be considered and judged by itself.

Too often faulting is considered in only two dimensions—that is, as either normal or reverse movement in a vertical plane. The fact is that most faulting movements have horizontal as well as vertical components; every mining engineer knows the frequent occurrence of inclined or horizontal striation on fault planes.

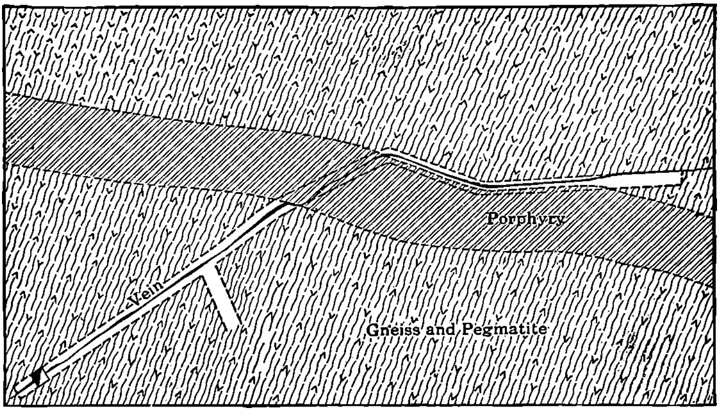

Fig. 22.—Plan of a vein in the Homer Mine, Idaho Springs, Colorado, showing deflection of the vein upon meeting a dike. (*After J. E. Spurr, U. S. Geol. Survey.*)

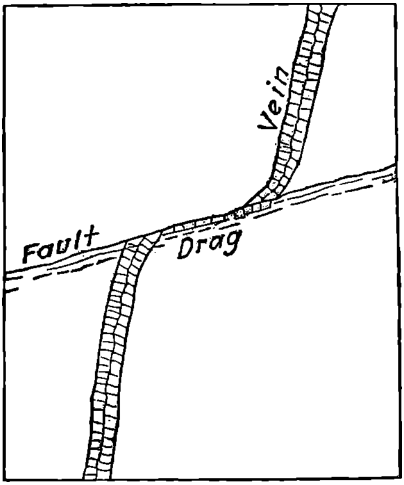

Fig. 23.—Sketch of faulted vein showing "drag."

Several proposals have been made looking to a uniform nomenclature of the various elements involved in faulting movements; the best of these are advocated by J. E. Spurr, C. F. Tolman, Jr., and H. F. Reid. A committee of the Geological Society of America was instructed to examine this question in more detail, and their conclusions, in large part based on the work of Reid, will probably be adopted by American geologists. An abstract of their report[1] will be found in the following pages.

Measurements of fault movement are made in the fault plane itself, in a plane normal to the trace of the faulted body on the fault plane, in any normal plane, and in a horizontal plane.

### General Terms

A fault is a fracture in the rock of the earth's crust, accompanied by a displacement of one side with respect to the other in a direction parallel with the fracture.

As we pass from one part of a fault to another, we find that certain characteristics vary. Definitions descriptive of characteristics must therefore be considered as referring to the parts of the fault to which they are applied and not necessarily to the fault as a whole.

A *closed fault* is one in which the two walls of a fault are in contact.

An *open fault* is one in which the two walls of a fault are separated. The same fault may be closed in one part and open in another. The open part is called the *fault space.*

A *fault surface* is the surface along which dislocation has taken place; if without notable curvature it is called a *fault plane.*

A *fault line* is the intersection of a fault surface with the earth's surface or with any artificial surface of reference, such as a level of a mine.

When a fault is made up of a number of slips on closely spaced surfaces it is called a *shear zone* or a *sheeted zone.*

The *fault breccia* is the assembly of broken fragments which is frequently found along faults.

*Gouge* is a clayey attrition product, usually a mixture of fine-grained minerals, often occurring along the walls of a fault.

A *horse* is a mass of rock broken from one wall and caught between the walls of the fault.

The *fault strike* is the direction of the intersection of the fault surface, or the shear zone, with a horizontal plane. It is measured from the astronomic or from the magnetic meridian.

The *fault dip* is the vertical inclination of the fault surface, or shear zone, measured from a horizontal plane.

The *hade* is the inclination of the fault surface, or shear zone, measured from the vertical; it is the complement of the dip.

The *hanging wall* is the upper wall of the fault.

The *foot wall* is the lower wall of the fault.

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[1] Reid, Davis, Lawson, and Ransome, *op. cit.*

## GENERAL CLASSIFICATION OF FAULTS

Faults of *parallel displacement* are those in which all straight lines on opposite sides of the fault and outside of the dislocated zone, which were parallel before the displacement, are parallel afterward.

*Rotatory faults* are those in which some straight lines on opposite sides of the fault and outside of the dislocated zone, parallel before the displacements, are no longer parallel afterward—that is, where one side has suffered a rotation relative to the other.

Determinations of throw are almost always relative, and hence we can rarely tell which side of the fault has moved; therefore the terms "upthrow" and "downthrow" are objectionable. They are, however, in general use and should be retained; but it should be definitely understood that they refer merely to a relative displacement.

**Faults of Parallel Displacement.**—No faults of any magnitude consist of simple parallel displacements over their whole length. Faults die out at their limits, and the displacement is not uniform along their courses, so that there is necessarily some slight rotation, varying in amount in the different parts of the fault's course. The variations in rotation and displacement are permitted by slight plastic deformation. If, however, we confine our attention to a small length of the fault, we may describe the displacement there as if a parallel displacement had occurred.

The word "displacement" should receive no technical meaning, but is reserved for general use; it may be applied to a relative movement of the two sides of the fault, measured in any direction, when that direction is specified; for instance, the displacement of a stratum along a drift in a mine would be the distance between the two sections of the stratum measured along the drift. The word "dislocation" will also be most useful in a general sense.

There are two methods of defining the displacement due to a fault; we may define the apparent relative displacement of a bed by naming the distance between its two disrupted sections measured in any chosen direction, such as the vertical distance between the sections, measured in a shaft, or the perpendicular distance between the lines of intersection of the two sections with the fault plane; or we may define the actual relative displacement and its components in important directions. The apparent displacements are those usually measured; the actual displacement must be worked out for a complete understanding of the fault.

Only four important technical words are used to denote the various displacements caused by faulting, qualifying words being added to indicate the component of the displacement referred to. These words are:

*Slip,* which indicates the relative displacement of formerly adjacent points on opposite sides of the fault, measured in the fault surface. The qualifying words relate to the strike and dip of the fault surface.

*Shift*, which indicates the relative displacement of regions on opposite sides of the fault and outside of the dislocated zone. The qualifying words relate to the strike and dip of the fault surface, except in the expression "vertical shift," which is self-explanatory.

*Throw*, which indicates a displacement not related to the strike or dip of the fault plane.

*Offset*, which indicates the shortest horizontal distance between the outcrops of a dislocated bed.

By keeping in mind the general meaning of these four words, all confusion in the uses of the proposed nomenclature can be avoided.

There is no generally accepted word in present use to denote the slip. Willis and Tolman use "displacement"; Spurr uses "throw." We have reserved "displacement" for general use, and the word "throw" is here used in quite a different sense. The word "shift" also suggests the meaning attached to it; there is no distinctive word now in use to describe the shift.

In mines, where the fault surface itself is visible, the slip will generally be determined; it is of paramount importance in mining. In field surveys, where the fault is studied by the dislocation of the outcrop of strata, or dikes, often at a considerable distance from the fault, the shift is determined. In the larger problems of geology the shift is of greater importance than the slip. The distinction, not heretofore recognized, between the slip and the shift is important. The perpendicular throw is of the greatest importance. It is frequently the only displacement determined, and in strike faults all the displacements in a vertical plane at right angles to the fault strike—that is, all the displacements which have heretofore received the most attention—can be expressed in terms of perpendicular throw. The offset is often the most important surface measurement made.

**Faults in Stratified Rocks.**—Among stratified rocks the character of the displacement of the strata due to a fault is so much influenced by the relation of the strike of the fault to the strike of the strata that subclasses are generally recognized.

A *strike fault* is one whose strike is parallel to the strike of the strata.

A *dip fault* is one whose strike is approximately at right angles to the strike of the strata.

An *oblique fault* is one whose strike is oblique to the strike of the strata.

These terms are, of course, not directly applicable in regions of unstratified rocks; but they might be used in such regions with respect to the strike of a system of parallel dikes or veins.

Similarly with regard to the general structure of the region:

A *longitudinal fault* is one whose strike is parallel with the general structure.

A *transverse fault* is one whose strike is transverse to the general structure.

**Slip.**—The word "slip," as stated above, indicates the displacement as measured on the fault surface.

The *slip* or *net slip* is the maximum relative displacement of the walls of the fault, measured on the fault surface, along the line of the movement; it is given by *ab* in Figs. 24 and 25.[1]

The *strike-slip* is the component of the slip parallel with the fault strike or the projection of the net slip on a horizontal line in the fault surface; *ac* in Figs. 24 and 25.[2]

The *dip-slip* is the component of the slip parallel with the fault dip, or the projection of the slip on a line on the fault surface perpendicular to the fault strike; *bc* in Figs. 24 and 25.[3]

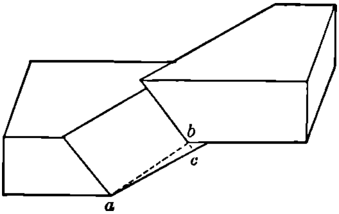

Fig. 24.—The slip.

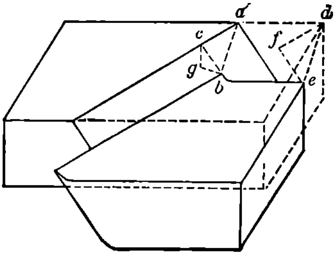

Fig. 25.—The shift.

**Shift.**—It frequently happens that a fault has not a single surface of shear, but consists of a series of small slips on closely spaced surfaces; and in some faults the strata in the neighborhood of the fault surface are bent, so that the relative displacements of the rock masses on opposite sides of the fault may be quite different from the slip and not even parallel with it. The word "shift" is used to denote the relative displacements of the rock masses situated outside of the zone of dislocation; the qualifying words relate to the strike and dip of the fault, with one exception, in which the meaning is clear.

The *shift*, or *net shift*, is the maximum relative displacement of points on opposite sides of the fault and far enough from it to be outside of the dislocated zone; *de* in Figs. 25 and 26, where *d* is the position of a selected point before and *e* after the faulting.

The *strike-shift* is the component of the shift parallel with the fault strike; *df* in Figs. 25 and 26.

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[1] Spurr and Tolman call this the "total displacement."

[2] Tolman calls this the "horizontal displacement."

[3] Tolman calls it the "normal displacement."

The *dip-shift* is the component of the shift parallel with the fault dip; *e* in Figs. 25 and 26. (The triangle *def* is parallel with the fault surface in Fig. 25.[1])

The bending of the strata near the fault may be so great that the direction of the shift is no longer even nearly parallel with the fault surface; it is better then to use the three following terms for the components of the shift:

The *strike-shift* is the horizontal component of the shift parallel with the fault strike, as already defined.

The *normal shift* is the horizontal component of the shift at right angles to the fault strike. It equals the horizontal shortening or lengthening of the earth's surface at right angles to the fault strike, due to the fault.

The *vertical shift* is the vertical component of the shift. These terms may evidently be used equally well when the shift is parallel with the fault plane.

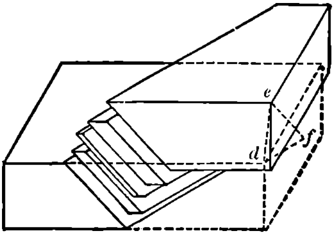

Fig. 26.—The shift.

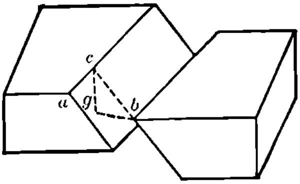

Fig. 27.—The throw.

**Throw.**—The word "throw" will apply to components of the displacement having no immediate bearing on the strike or dip of the fault plane.

The *slip throw* is the vertical component of the slip; *cg* in Figs. 25 and 27, *de* in Figs. 28 and 29. The word throw is almost universally used in this sense, but A. Geikie uses it to designate the vertical distance between the two parts of a dislocated bed, projected if necessary—a very different thing. Geikie's "throw" would be represented by *df* in Figs. 28 and 29. Spurr uses "throw" to designate the distance between the two parts of a dislocated bed measured on the fault plane.

The *heave*[2] is the horizontal component of the slip, measured at right angles to the strike of the fault; *bg* in Figs. 25 and 27, *eg* in Figs. 28 and 29. The word "heave" has been used in many senses; J. Geikie, Willis, Scott, and Fairchild use it as defined above; A. Geikie and Spurr use it to designate what we have called the "offset" of a bed;[2] Jukes-Brown apparently

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[1] The *dip-shift* and *strike-shift* are not accurately shown in Fig. 25, because the net shift, *de*, is not parallel with the fault plane, and the lines *de*, *df*, and *fe* would not lie in one plane. But the definitions are clear and the figure illustrates them fairly well.

[2] Sometimes called the "horizontal throw."

used it for the strike-slip; so did Ransome; and Scott also uses it in this sense when he refers to "heave faults."

The *separation* of a bed, dike, vein, or of any recognizable surface, is the distance between the two parts of the disrupted bed, etc., measured perpendicularly to the bedding plane or to the plane of the surface in question. It is measured, therefore, in a vertical plane at right angles to the strike of the disrupted surface.[1] The importance of the separation of the strata is so great that it is convenient to have special terms for it; these are given below.

The *stratigraphic separation* is the distance between the two parts of a disrupted stratum measured at right angles to the plane of the stratum; *ab* in Figs. 28 and 29. The stratigraphic separation is in general the simplest throw to determine; it can be found from the distance between the outcrops of the two parts of the same stratum, the dip of the stratum, and the slope of the ground.

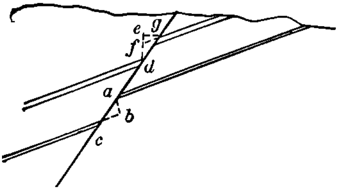

FIG. 28.—Section of a normal fault.

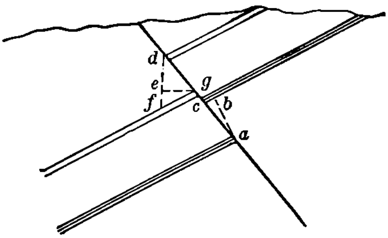

FIG. 29.—Section of a reverse fault.

The *dip throw* is the component of the slip measured parallel with the dip of the strata; *cb* in Figs. 28 and 29.

**Offset.**—The *offset* of a stratum is the distance between the two parts of the disrupted stratum measured at right angles to the strike of the stratum and in a horizontal plane.[2]

If Figs. 30 and 31 represent the ground plans of oblique faults on a level surface, *ab*, and not *ac*, would be the offset of the stratum; *ac* would be the horizontal displacement of the stratum parallel with the fault strike.

**Faults Classified According to the Direction of the Movement.**—Faults may be classified, according to the direction of the movement on the fault plane, into three groups, as follows:

*Dip-slip faults*, where the net slip is practically in the line of the fault dip.

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[1] Spurr calls it the "perpendicular separation." Tolman's "perpendicular throw" would under certain conditions correspond in meaning with our expression.

[2] A. Geikie and Spurr use the term "heave" for this offset.

*Strike-slip faults,* where the net slip is practically in the direction of the fault strike. A vertical fault is one with a dip of 90 degrees (see below); and, by analogy, a horizontal fault should be one with a zero dip and the term should not be applied to strike-slip faults.

*Oblique-slip faults* where the net slip lies between these directions.

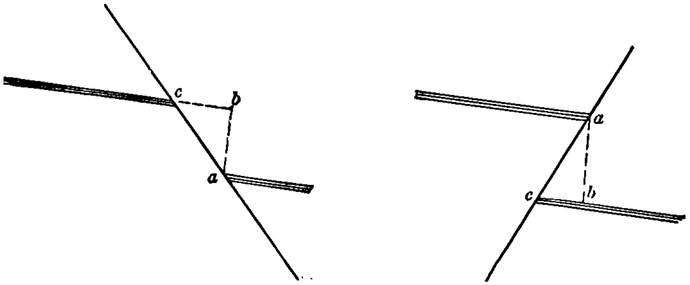
For purposes of simplification, faults are often represented as dip-slip faults and are then classified into:

*Normal faults,* where the hanging wall has been depressed relatively to the footwall.

*Reverse faults,* where the hanging wall has been raised relatively to the footwall.

*Vertical faults,* where the dip is 90 degrees.

The relative displacement has usually been determined by means of a dislocated bed. Although exception may well be taken to these terms, their retention is recommended, because they are in general use and are well understood. The word "reverse" is preferable to "reversed."



FIGS. 30 and 31.—Plan of an oblique slip.

The horizontal distance between two points on opposite sides of a fault, measured on a line at right angles to the fault strike, is always shortened by a reverse strike fault, lengthened by a normal strike fault, and unchanged in length by a vertical fault.

The expressions "normal" and "reverse" may be used in connection with oblique and dip faults, even when these are strike-slip or oblique-slip faults, provided they are applied to designate the apparent relative displacement of the two parts of a dislocated stratum, or other recognized surface, in a vertical plane at right angles to the fault strike. It does not follow, in the case of oblique-slip faults, that a horizontal line at right angles to the fault strike would be lengthened by a normal or shortened by a reverse fault. This has been pointed out by Ransome[1] and can be illustrated by Figs. 32 and 33. In Fig. 32 a reverse fault, as determined by the displacement of the stratum or vein, has caused an extension at right angles to the fault strike. It is evident that if the hanging wall had

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[1] *Econ. Geol.*, 1, 1906, pp. 783–787.

moved, as in Fig. 33, with the stratum or vein dipping as there represented, we should have had a normal fault and a contraction at right angles to the fault strike. The relations of the two parts of the disrupted stratum in Fig. 32 are exactly the same as if we had had a simple reverse dip-slip fault, and in Fig. 33 as if we had had a simple normal dip-slip fault; and if there are no disrupted dikes or other means of determining the amount of the strike-slip, the movements described could not be distinguished from simple dip-slip faults. It very frequently happens that nothing more than the apparent displacement of the strata can be determined, and the terms "normal" and "reverse" faults as defined should be used purely for purposes of description and not for the purpose of indicating extension or contraction, tension or compression, vertical or horizontal forces.

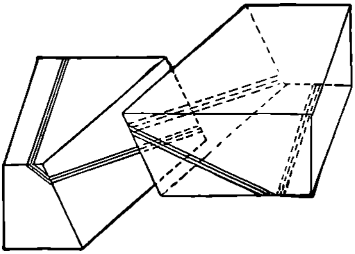

FIG. 32.—A reverse fault due to an oblique slip.

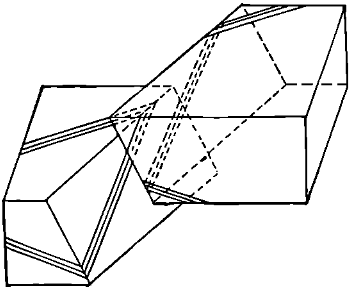

FIG. 33.—A normal fault due to an oblique slip.

**Overthrusts.**—These are reverse faults with low dip or large hade. In some cases the dip-slip has been enormous, amounting to tens of kilometers. The word "overthrust" has been generally used for this kind of fault and is very descriptive.

**Mineralization of Faults.**—Any fault may become a *fissure vein* by filling and replacement along its course. However, it is rather unusual to find large structural faults, normal or overthrusts, which have been extensively mineralized. Shear and sheeted zones often result in veins or lodes.

**Complexity of Faulting.**—During mining operations excellent and detailed instances of the complexity of faulting are often found. Normal and reverse faults may occur in close proximity. A fault consists more frequently of a series of closely spaced breaks than of a single fracture. Displacement occurs usually along each of these breaks, the result being a distortion or "shift" of the deposit within the faulted zone.

Figure 34 shows a case of complicated normal faulting from the Berlin vein, Nevada.[1] Besides the faults indicated there are a great number of

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[1] Ellsworth Daggett, The extraordinary faulting at the Berlin mine, *Eng. and Min. Jour.*, Mar. 30, 1907.

other dislocations with horizontal displacement. The deposit is a filled quartz vein, 2 to 3 feet wide, carrying 2 per cent of sulphides with silver and gold.

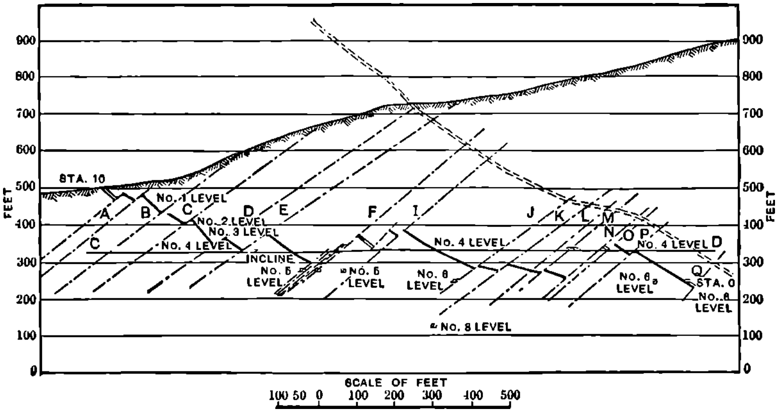

FIG. 34.—Vertical section of a faulted vein, Berlin mine, Nevada, showing also its probable original position. (*After Ellsworth Daggett.*)

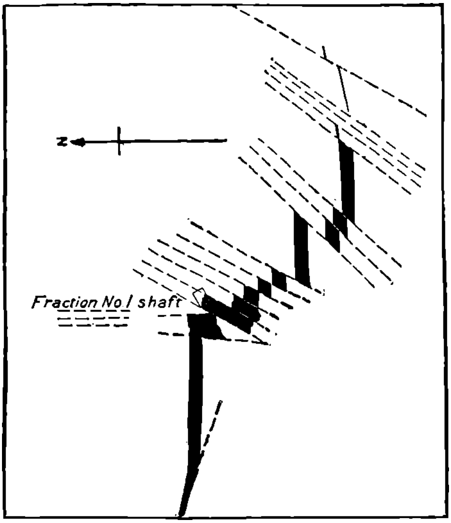

FIG. 35.—Horizontal plan showing faulted vein, Tonopah, Nevada. Scale: 50 feet = 1 inch. (*After J. E. Spurr, U. S. Geol. Survey.*)

The great complications ensuing where faulting takes place along two intersecting fault systems have been described by Spurr[1] in his report on the Tonopah district, Nevada. The result of such structures is likely to be a zigzag distribution of the fragments of the faulted vein with an

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[1] *Prof. Paper* 42, U. S. Geol. Survey, 1905.

average movement determined by the two components. Repeated small dislocations practically result in a deflection of the vein (Fig. 35).

Overthrusts of great magnitude, such as are found in the Alps, may have had most important results as to the continuation in depth of ore deposits. As these dislocations may be measured in miles, it follows that whole groups of deposits contained in the overthrust portion of the strata may have been cut off entirely from their continuation in depth.[1]

**Joints.**—Rocks are commonly traversed by non-persistent fractures along which no displacement or only an exceedingly small displacement has taken place. These are called *joints* and are produced by various stresses as explained in the following chapter. Frequently there are two or more parallel sets of joints which thus tend to break the rocks into more or less regular blocks. Joints near faults may be produced by the same stresses which originated the faulted fractures and are followed by the same mineralizing solutions which filled the fault fissures. Therefore, the study of the joint systems is often of great importance in determining the structure and origin of the mineralization and may guide to the discovery of new ore-bodies.

Closely spaced parallel joints may form transitions to shear zones or sheeted zones and even to slaty cleavage.

[1] B. Granigg, Ueber die Erzfuehrung der Ostalpen, Leoben, 1913.

# CHAPTER XII

# OPENINGS IN ROCKS

Chemical processes and alteration in general may go on in a rock without cavities other than pore space and capillary or sub-capillary openings. Such processes are, however, metamorphic rather than metasomatic; they simply effect a mineralogical rearrangement without much chemical change; the composition of the rock remains constant. The formation of epigenetic mineral deposits usually implies a considerable addition of foreign material by solutions and these solutions must be guided to the place of deposition by open spaces, such as fissures, joints, or cracks. As a matter of fact the great majority of mineral deposits were formed where the path of the solution was prescribed by openings in the rocks other than those of ordinary pore space. After the solutions have gained access to the rock they may, of course, enter the pores and capillary openings and effect metasomatic changes.

The discussion which follows relates mainly to openings of super-capillary size. Such openings are chiefly found in the zone of fracture (p. 94). Few of our mineral deposits have been formed at depths much greater than 15,000 feet. Small openings may, however, exist in hard rocks at a distance below the surface much greater than the figure just indicated (p. 33). The possibility is, therefore, shown that solutions from great depths may gain access to the upper zone of fracture.

## ORIGIN OF OPENINGS

Rock cavities may originate in various ways:

1. **By the Original Mode of Formation of the Rocks.**—Many volcanic flows contain abundant gas pores, or blow holes produced by the expansive forces of gases escaping from the magma. Zeolites and calcite, sometimes with native copper, often accumulate in these pores; and such rocks are usually termed "amygdaloids" and the filled cavities "amygdules" (Fig. 36). Some sandstones and conglomerates contain much pore space in which solutions may deposit ores or other substances.

2. **By Solution.**—Solution cavities are found mainly in easily soluble rocks, such as limestone, dolomite, gypsum, and salt. Posepny justly maintains that the solvent power of water suffices to produce long galleries or passages in rock salt and mentions several examples.[1] Joints in lime-

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[1] F. Posepny and Others, Genesis of ore deposits, 2d ed., 1902, p. 20. Publ. by Am. Inst. Min. Eng., New York.

stone are often irregularly enlarged by solution and when subsequently filled with ores such cavities are known as *gash veins* or *pipe veins*. Caves in limestone are likewise made by atmospheric water of the upper circulation, containing dissolved carbon dioxide. Such caves are generally formed above the ground-water level in the zone of oxidation, though cases are known which suggest that the process can go on also below this level. Caves occur in all limestone regions and are sometimes of enormous extent; the Mammoth Cave of Kentucky has passages more than 40 miles in length and has been formed by the removal of millions of cubic yards of rock. The extent of caves is generally determined by faults and dislocations, and rock openings on a smaller scale are usually determined by the prevailing joint systems. The breaking in of caves near the surface produces the "sink-holes" so characteristic of certain limestone plateaus. Both caves and sink-holes have a certain importance in the origin of the class of zinc-lead deposits common to many limestone areas, and caves of dissolution in the oxidized part of ore deposits in limestone are sometimes the receptacles for a great variety of secondary minerals. The floors of caves are usually covered with red "cave earth," a residual deposit of silica, kaolin, limonite, etc., derived from the less soluble constituents of the limestone. Deposits of bat guano and nitrates are sometimes found in caves. Small solution cavities are often found in more resistant rocks that have been exposed to hot solutions of great solvent power.

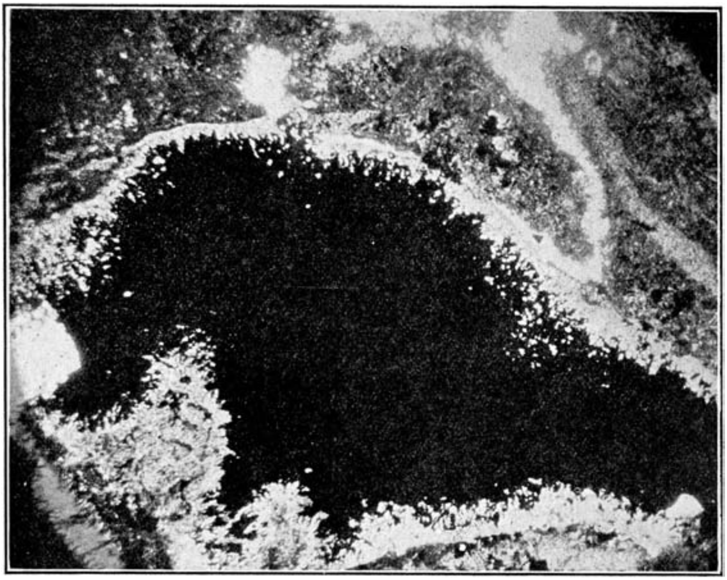

FIG. 36.—Photomicrograph of basalt showing blow hole filled with epidote, pumpellyite, calcite, and native copper (black), Lake Superior. Magnified 30 diameters.

3. **By Fractures of Various Modes of Origin.** (*a*) *Contraction Joints Produced by Tensile Stress in Igneous Rocks.*—When magmas congeal to igneous rocks tensile stresses which result in fissures and joints are developed. This is best exemplified in effusive rocks, which often show regular columnar structure and which are always full of irregular joints and cracks. No doubt these open spaces may guide metal-bearing solutions. In the literature many authors attribute fissure veins in effusive rocks to contraction, but usually without sufficient reason. The tensile stresses cannot produce long fissures with regular strike and dip.

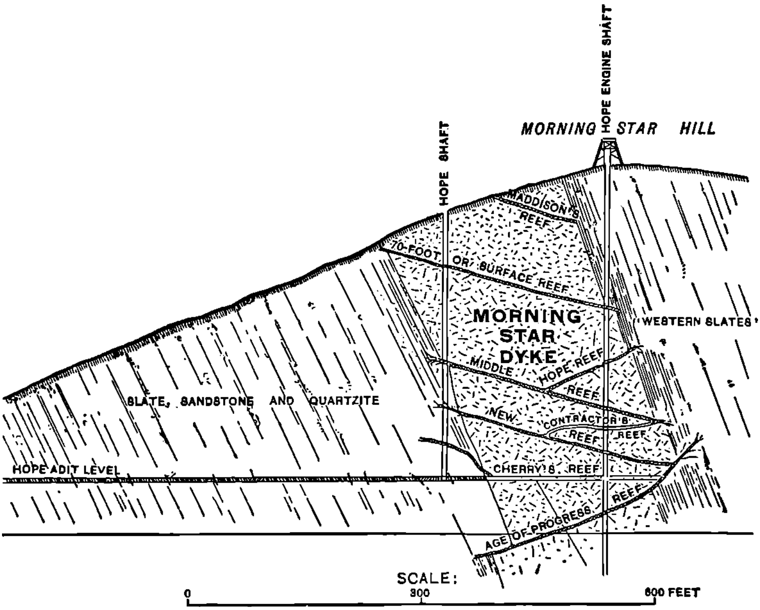

FIG. 37.—Section of Morning Star dyke, Woods Point, Victoria, showing ladder veins. (*After O. A. L. Whitelaw.*)

According to the views of many geologists, smaller irregular veins in dikes or other intrusive rock masses fill contraction fissures. This explanation has been advanced for the horizontal tin-bearing joints in the Zinnwald granite, Saxony, and for other similar "stockwork"; also for the so-called "ladder veins," which are short transverse fissures in dikes, usually extending only from wall to wall. Well-known examples of this kind in Telemarken, Norway,[1] have been described by Vogt; in Victoria, Australia,[2] by Whitelaw; and at Beresowsk, in the Ural Mountains,[3] by Rose, Helmhacker, Karpinsky, Posepny, and Purington. In

[1] Zur Klassification der Erzvorkommen, *Zeitschr. prakt. Geol.*, 1895, p. 149.

[2] *Mem.*, Geol. Survey, Victoria, 3, 1905, p. 11.

[3] *Guide*, Seventh Int. Geol. Congress, 1897, p. 42.

F. Posepny, *Archiv für prakt. Geol.*, 2, 1895, p. 499.

C. W. Purington, *Eng. and Min. Jour.*, 75, 1903, pp. 854–855, 893–894, 929–931.

places, however, the transverse fissures may extend over the contact into the wall rock or correspond to the general joint systems of the vicinity, a fact which throws some doubt on the correctness of the explanation given (Fig. 37).

(*b*) *Contraction Joints by Shrinking of Limestone When Changed to Dolomite.*—Dolomite is not uncommonly formed near certain metal deposits and it is possible that this process when carried on by rapidly moving solutions and in comparatively free space may result in openings suitable as receptacles for ore minerals.

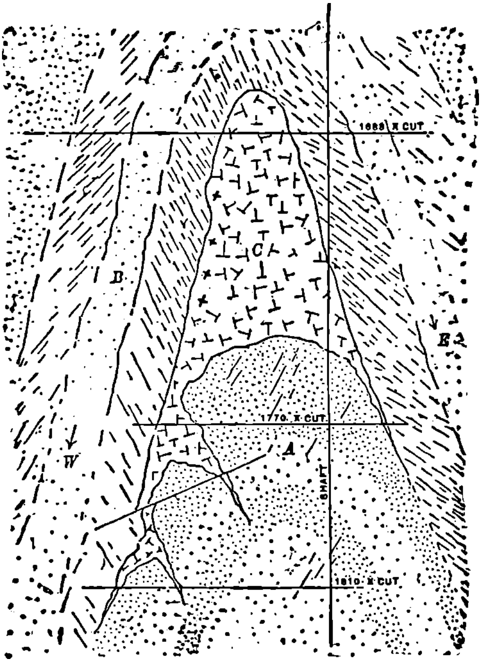

FIG. 38.—Section through a saddle reef, Bendigo, Victoria. *A*, sandstone; *B*, shaly sandstone; *C*, gold-bearing quartz. (*After T. A. Rickard.*)

(*c*) *Expansion Joints Produced by Increase of Rock Volume.*—Peridotite upon change to serpentine near the surface and near fissures is believed to increase its volume greatly and such serpentine often breaks into smooth fragments. Extreme irregularity is a characteristic of all expansion joints and they are of little importance in ore deposition.

(*d*) *Fissures Produced by Torsional Stress.*—The celebrated experiment by Daubrée[1] carried out by twisting a thick glass plate has shown that torsional stress may result in several systems of long and radiating fissures.

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[1] Études synthétiques de géologie expérimentale, Paris, 1879, p. 316.

This experiment has frequently been cited by geologists to explain divergent vein systems; but G. F. Becker[1] has pointed out that such fissures do not follow approximate planes, like fissure veins, but are decidedly curved and warped. C. K. Leith and W. J. Mead have shown that the torsion fissures do not extend through any considerable thickness of material.

(*e*) *Openings Produced by Folding of Sedimentary Rocks.*—The bedding planes of sediments are primary structures which often serve as ducts for metal-bearing solutions. Better passageways for such solutions are provided when a series of sediments of unequal resistance is folded. A sandstone, for instance, will accommodate itself to bending with difficulty and will easily break at anticlines or synclines, whereas softer shales will bend without breaking; the same process may cause a slipping between the various members. Such tensional stresses may then easily produce open cavities. The quartz-filled "saddle reefs" of the gold mines of Bendigo and other places in Victoria are believed to have been formed in this manner by tensile stresses, but they are also accompanied by irregular masses or "makes" of quartz which fill spaces of discission across the beds (Figs. 18 and 38), and which apparently can not be accounted for by replacement.

(*f*) *Openings Produced by Shearing Stress under the Influence of Gravity.* In many disturbed regions the rocks are broken by normal faults along which the various blocks have settled down under the influence of gravity. Such normal faulting is especially characteristic of regions which do not bear evidence of strong compressive stress. Step-faulting is common and friction breccias and crushed zones frequently follow the faults; the open spaces, more or less continuous, offer good paths for the circulation of water; the fault planes are often long and regular. But in spite of all this, mineral deposits are not common along such faults. At Clifton, Arizona, for instance, faults are abundant, but the copper deposits do not ordinarily appear in them excepting in the one case of the Coronado fault. At Creede, Colorado, and Mogollon, New Mexico, the silver veins occupy important fault fissures, and similar cases are known from the silver-lead veins of the Harz Mountains in Germany.

In volcanic regions, such as Silverton and Cripple Creek (Fig. 39), in Colorado, systems of nearly vertical fissure veins contain rich deposits. They have, obviously, little connection with the main structural features of the country; the dislocations are usually small, and the veins were formed shortly after the close of volcanic activity. F. L. Ransome[2] believes that these fissure systems were generated by stresses resulting from slight vertical movements, or settling, following an enormous transfer of volcanic material from an intratelluric to a superficial position.

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[1] The torsional theory of joints, *Trans.*, Am. Inst. Min. Eng., 24, 1894, pp. 130–138.

[2] F. L. Ransome, *Bull.* 182, U. S. Geol. Survey, 1901, p. 66.

Vertical upthrusts of underlying magmas may have caused faulting accompanying or following vein formation.[1]

Locke[2] has called attention to the importance of brecciation due to removal of support by the dissolving action of solutions during the early stages of mineralization, followed by ore deposition in the brecciated masses at a later stage. He believes that many so-called "pipes" have been formed in this manner and cites Pilares copper mine, Sonora, as a good illustration (Fig. 40). In this case the collapse, "crackling," and brecciation can be proved by a sagging andesite-latite contact in the mine, but he does not appear to have proved that this collapse has been produced by the dissolving action of the solutions; it may have been of pre-mineral origin. This deposit has an oval horizontal section, 1,000 by 2,000 feet and has been developed to a depth of 1,800 feet. In a

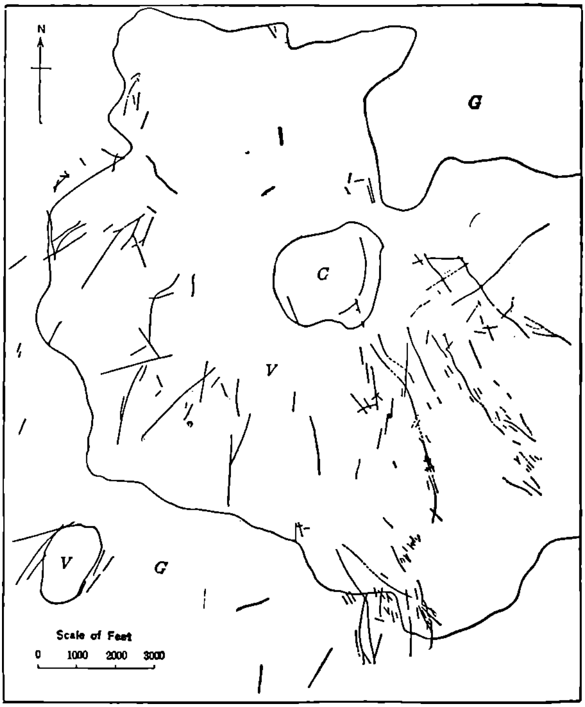

FIG. 39.—Plan of the principal veins of the Cripple Creek district, Colorado, showing a roughly radial distribution. *G*, granite and gneiss; *V*, Tertiary volcanic rocks. (*After Lindgren and Ransome, U. S. Geol. Survey.*)

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[1] J. E. Spurr, Relation of ore deposition to faulting, *Econ. Geol.*, 11, 1916, pp. 601–622.

[2] A. Locke, The formation of certain ore-bodies by mineralization stoping, *Econ. Geol.*, 21, 1926, pp. 431–453.

subsequent paper Wisser[1] described the doming, cracking, and marginal fissures produced by the oxidation and caving of an ore-body. Such caving when extending to the surface may give valuable information as to the location of ore-bodies.

(*g*) *Openings Produced by Compressive Stress.*—In contrast to recently congealed lavas, the rocks which have formerly been far below the surface of the earth but which have been exposed by erosion are usually traversed by more or less regular joint systems, persistent over large areas. While some of these joint systems may be caused by the inherent texture of the rock, they are in most cases the effect of compressive stress. Closely spaced joint systems form transitions into slaty cleavage, and recrystallization of minerals takes place by preference along these planes. In extreme cases fissility or cleavage in very thin laminæ develops. Joints and cleavage present narrow paths for mineralizing solutions and ore deposits are often determined by their direction.

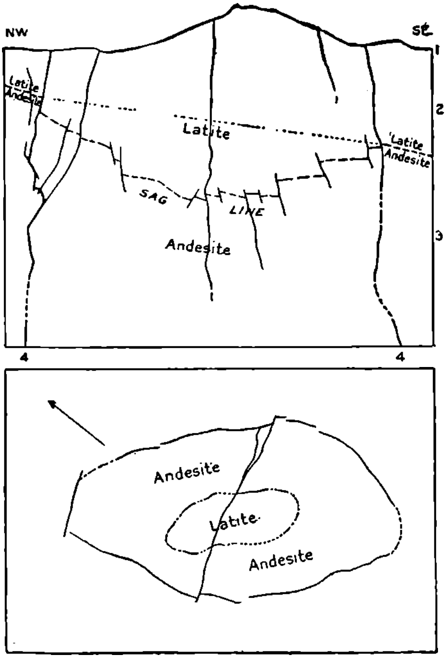

FIG. 40.—Plan and section of the Pilares pipe, Sonora, Mexico. 1, surface; 2, latite; 3, andesite; 4, walls of pipe. (*After A. Locke, adapted from Bjorge, "Economic Geology."*)

There are all stages of transition from joints to fissures along which perceptible movement has occurred. In many districts the fissures which have received the ores are identical in strike and dip with the joint systems

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[1] Edward Wisser, Oxidation subsidence at Bisbee, Arizona, *Econ. Geol.*, 22, 1927, pp. 761–790.

of the country rock. A common condition is that two sets of veins and joints occur which have the same strike, but dip in opposite directions (Fig. 41). Such vein systems are termed conjugated fractures. The explanation of such joints, fissures, and occasionally accompanying schistose structure is furnished by certain experiments by Daubrée[1] and by the mathematical deductions of G. F. Becker.[2] These show that compression may develop several systems of fractures, but that the two principal systems lie along the planes of maximum shear; these shearing planes are inclined to the direction of maximum stress. The accompanying dislocations will largely be reverse faults in which the hanging wall has relatively moved upward. In Daubrée's experiment on a mass of beeswax and resin two conjugated systems of joints and fissures were formed, making an angle of 45 degrees with the line of pressure; similar results have been obtained by testing cubes of building stones. If the stress is not exerted horizontally the dip of the veins will be correspondingly affected. At Grass Valley, California, and in many other districts there are two such conjugated systems of fissures which have been filled with ore. W. D. Johnston, Jr., and E. Cloos[3] believe the Grass Valley fissures originated during an early phase of the consolidation of the rock by stresses imposed by the intrusive itself.

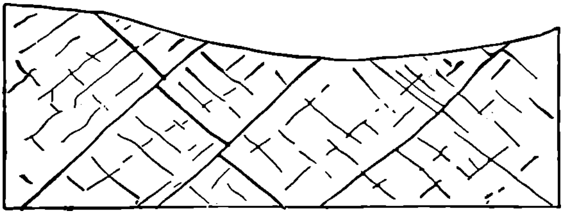

FIG. 41.—Vertical section of a conjugated system of fractures.

Leith and Mead have shown that the three fault systems occupied by the copper veins at Butte were caused by stresses applied successively within one period of mineralization.

In the locality just mentioned the majority of the dislocations are small, but tangential stresses sometimes produce great dislocations. The Mother Lode of California, a vein system nearly 100 miles in length, is believed to represent a reverse fault or system of faults with considerable throw and dips to the east.

When rocks are recrystallized in the deeper zones of the earth's crust they may become so plastic that deformation by rupture cannot

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[1] Études synthétiques de géologie expérimentale, Paris, 1879, p. 316.

[2] Finite homogeneous strain, etc., *Bull.*, Geol. Soc. Am., 4, 1893, p. 13.
C. K. Leith, Structural geology, New York, 1923, pp. 13–62.
W. J. Mead, *Jour. Geol.*, 28, 1920, pp. 505–523.

[3] Paper presented at Boston meeting of Geol. Soc. Am., 1932.

take place. The growth of crystals then probably takes place predominantly in a plane perpendicular to the stress and a close schistose structure like that in many gneisses may develop which offers scarcely any interstitial space available for the circulation of solutions.

If the fissures were perfect planes it would be difficult to conceive of open spaces along them except by tensional stresses pulling the walls apart; but as they are not, movement along them tends to produce a series of openings, alternating with numerous touching points. As a matter of fact, the mode of mineral deposition shows that open spaces existed and that they sometimes were large, in exceptional cases even 20 feet or more in width. In mine workings in hard rock old stopes frequently remain open for an indefinite length of time, and it is probable that such large open spaces may exist down to a depth of at least several thousand feet. Moreover, it is to be remembered that at the time of deposition the fissures were filled by water under a pressure at least equal to that of the hydrostatic column. The depositing solutions emanating from magmas under conditions of far stronger pressure may even have made way for themselves in the manner of an igneous dike or pegmatite vein, actually forcing the rocks apart. Some of the phenomena of deep-seated veins are difficult to explain on any other assumption.

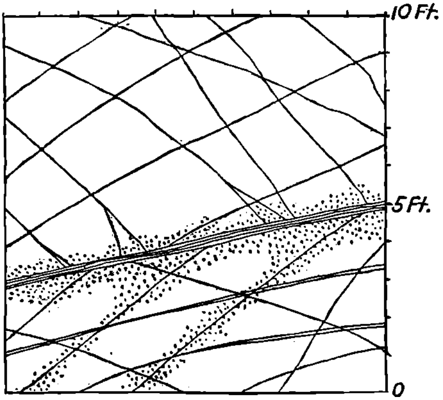

Fig. 42.—Joints varying from 6 inches to 2 feet apart. Mineralization spreading from joints. This ground is "blocky" where slips are in three planes. (*After G. T. Harley.*)

Gaping fissures are not, however, necessary for the circulation of solutions. Water may ascend along a number of closely spaced fissures —usually called a sheeted zone—in which very little open space exists. But in this case mineral deposition is usually effected by replacement. The solutions are forced into the adjoining rock and transform its minerals into ore.

The stresses set up in a mass consisting of various rocks are extremely complex and it may only be possible to ascertain the dominant mode of fracturing. To differentiate strictly between joints and fissures produced by tension and those produced by rotational or nonrotational compression is very difficult. The common condition of fractured ground is shown in Figs. 42 and 43.

**Force of Crystallization.**—Minerals crystallizing from solutions exert a certain pressure on the walls which confine them.[1] Many geologists have held that this force is sufficient to enlarge cavities along fractures and thus make room for mineral deposits. There is strong evidence in the structure and texture of veins which is unfavorable to such a view, except where conditions of light load prevail, as near the surface or near open spaces. Boydell points out the limitations to the exertion of the force of crystallization imposed by the consideration of influence of pressure on solubility and holds that this force is not an important factor in vein formation. Some curious phenomena in regard to inclusions of rocks in veins may find their explanation by the action of this force, for in fissures filled with solutions a comparatively slight force might suffice to detach fragments from the walls.

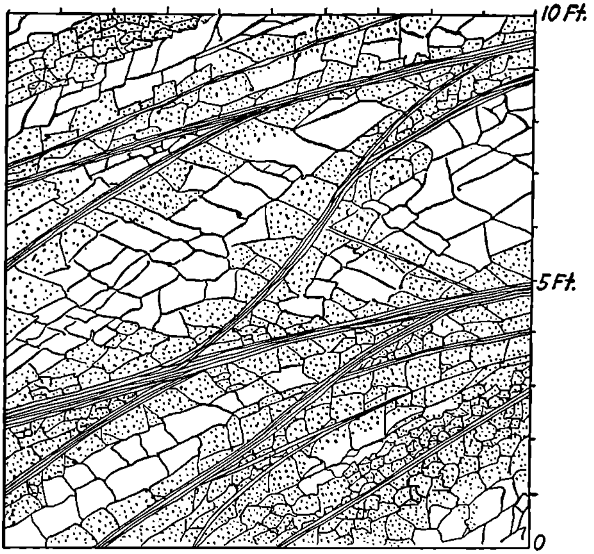

FIG. 43.—Thoroughly jointed and crushed ground illustrating condition in fissure veins. Mineralization is spreading from fissures. (*After G. T. Harley.*)

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[1] G. F. Becker and A. L. Day, The linear force of growing crystals, *Proc.*, Washington Acad. Sci., 7, 1905, pp. 282–288.

S. Taber, Pressure phenomena accompanying the growth of crystals, *Proc.*, Nat. Acad. Sci., 3, 1917, pp. 297–302. *Am. Jour. Sci.*, 4th ser., 41, 1916, p. 535. The mechanics of vein formation, *Trans.*, Am. Inst. Min. Met. Eng., 61, 1920, pp. 1–36. The origin of veinlets in the Silurian and Devonian strata of Central New York, *Jour. Geol.*, 26, 1918, pp. 56–73.

H. C. Boydell, A discussion on metasomatism and the linear force of growing crystals, *Econ. Geol.*, 21, 1926, pp. 1–55. Discussion by S. Taber, *idem*, pp. 717–727.

# CHAPTER XIII

## THE FORM AND STRUCTURE OF MINERAL DEPOSITS

The form of ore deposits is always important, for the mining methods used for a body of irregular outline must, for instance, be very different from those for a tabular vein. In the great majority of deposits the form is rudely tabular, for they usually follow the planes of dislocations, or tabular dikes, or the bedding of sedimentary rocks. The form is particularly important in replacement deposits, and no effort should be spared in ascertaining their internal structure and their relation to the structure of the country rock, that is, to contacts, beds, folds, joints, and fissures. Often this is even more important than the mineralogical composition of the ore.

A convenient and fundamental, though not strictly genetic, classification divides mineral deposits into *syngenetic*, or those formed by processes similar to those which have formed the enclosing rock and in general simultaneously with it; and *epigenetic*, or those introduced into a pre-existing rock.

**Syngenetic Deposits.**—The syngenetic deposits include the magmatic segregations or accumulations of useful minerals formed by processes of differentiation in magmas, generally at a considerable depth below the surface. Their form may be wholly irregular or roughly spherical, but more often they are rudely tabular or lenticular, and they are usually connected by transitions with the surrounding rocks. They are either wholly enclosed in the igneous mass, or lie along its margins, or, in some cases, form dikes or offshoots from a deep-seated reservoir. The last class of ores may be called epigenetic with reference to the rocks incasing the dikes. The width and thickness of these deposits may range from a few inches to several hundred feet, and in rare cases, their length may exceed one mile. Masses of chromite in peridotite or titanic iron ore in anorthosite, furnish examples of this type.

The syngenetic deposits also include sedimentary beds; they have, as a rule, a tabular, sheet-like, or flat lenticular form; they are horizontal if not disturbed, but are frequently folded and faulted. Parallel to their bedding their extent may be counted by miles, as in the case of the Clinton hematite ores of the Appalachian region, or the Jurassic iron ores of France; nevertheless, each bed usually thins out in wedge-shaped form and may be replaced by others at a slightly different horizon. In deposits of metallic ores the thickness is rarely more than 20 feet and this may include intercalated beds of barren material. Coal beds, especially those of lignite, or brown coal, may attain a thickness of 100 feet or more. Beds of rock salt, anhydrite, and gypsum are in some cases several hundred feet thick. In all sedimentary deposits displacements and folding may locally produce an appearance of great thickness. In plastic material like rock salt such deformation is especially effective.

**Epigenetic Deposits.**—The epigenetic deposits have various forms, but among those which follow fissures, the so-called fissure veins, the tabular or sheet-like form is most common. Deposits concentrated in the zone of weathering are often extremely irregular and of limited extent, and several of them are usually found in close proximity. Some hematite ores, like those of the Mayari district in Cuba, which are developed by the weathering of serpentine, may form superficial sheets of great extent.

Replacement deposits in limestone are extremely irregular, although their form as a whole is often dependent upon the bedding, the fissuring, or the contact with other rocks. They are seldom large, but in a few cases, like the galena deposits in southeastern Missouri or the sphalerite deposits in the Joplin region of the same state, they may be followed at a general horizon for several miles.

The ore deposits in metamorphic rocks which have undergone strong mechanical deformation and chemical changes usually assume lenticular form, and the occurrence of successively overlapping lenses is particularly characteristic. In these deposits a steep dip is a common feature, but the main trend of the ore-body in the plane of its strike is usually not in the direction of the dip.

**Definitions.**—The *strike* of a tabular or lenticular deposit is the direction of a horizontal line in the plane of the deposit, measured with reference to a meridian.

The *dip* is measured by the vertical angle between a horizontal plane and the plane of the deposit. Complementary to the dip is the *hade* or *underlie*, which is measured by the angle between the vertical and the plane of the deposit.

The *plunge*[1] (Fig. 44) of an ore-body is the vertical angle between a horizontal plane and the line of maximum elongation of the body. In lenticular ore-bodies in metamorphic rocks which have undergone strong mechanical deformation, the plunge is an important factor, and often it is determined by the direction of the cleavage or schistosity. In fissure veins the *pitch* of the ore-shoot is usually defined as the angle between its axis and the strike of the vein, and it is measured on the plane of the vein[2] (p. 192).

**Spacial Relations of Veins.**—Veins are tabular or sheet-like masses of minerals occupying or following a fracture or a set of fractures in the

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[1] Called "pitch" or "rake" by many authors.

[2] See discussion in *Trans.*, Am. Inst. Min. Eng., 39, 1908, pp. 898–916.

enclosing rock; they have been formed later than the country rock and the fractures, either by filling of the open spaces or by partial or complete replacement of the adjoining rock, or most commonly by both of these processes combined.

Such alteration or replacement does not ordinarily extend far from the fissure. In regions where the vein-forming solutions have acted with unusual intensity a partial alteration may extend from the deposit over considerable areas.

No sharp distinction can be drawn between the filled veins and replacement veins. If open spaces are available the metalliferous solutions which formed the veins in most cases found it easier to deposit their load in these spaces than to replace the country rock. Quartz is more likely to be deposited in the open paths, and likewise most of the heavy metals, unless the country rock is one particularly adapted for replacement, such as limestone. Gases like carbon dioxide and hydrogen sulphide penetrate the wall rocks with ease.

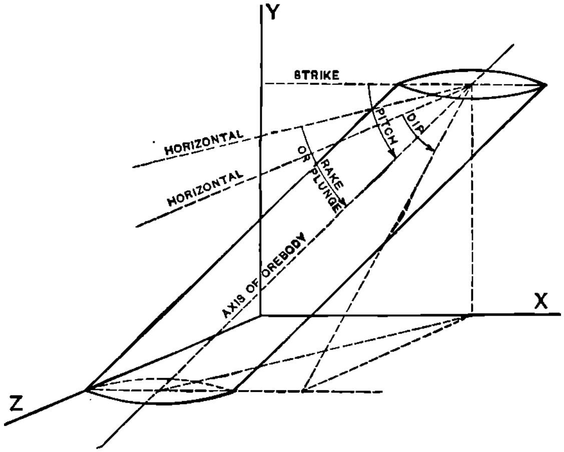

FIG. 44.—Stereogram illustrating strike, dip, pitch, and plunge of an ore-body.

Many veins correspond closely to the old definition of a "true fissure vein," in which the ore occupies the once open spaces along the fracture, with some alteration spreading into the wall rocks. Of such character are the majority of the gold-quartz veins of California and many other occurrences. When the fissures are very small they are referred to as veinlets or seams, and all transitions to a slight mineralization of joint planes are found. The walls may be smooth and separated from the vein material by a *clay gouge* or the filling may closely adhere to the country rock. In the latter case the vein is said to be *frozen to the walls.*

Instead of a single break we may have a fracture consisting of a number of approximately parallel fissures, irregularly connected and

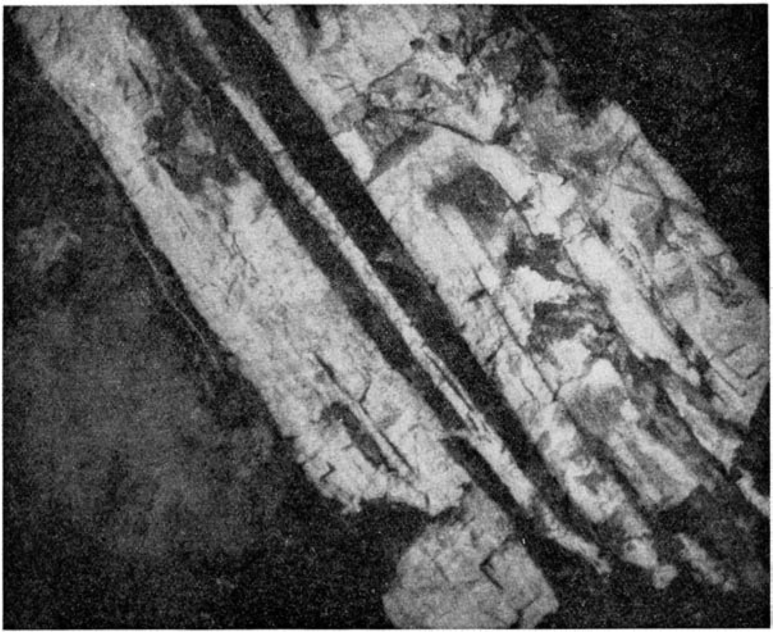

FIG. 45.—Southern Republic vein 3 feet wide. Shows banding in the quartz vein and included stringers of country rock (propylitic andesite). (*After Howland Bancroft, U. S. Geol. Survey.*)

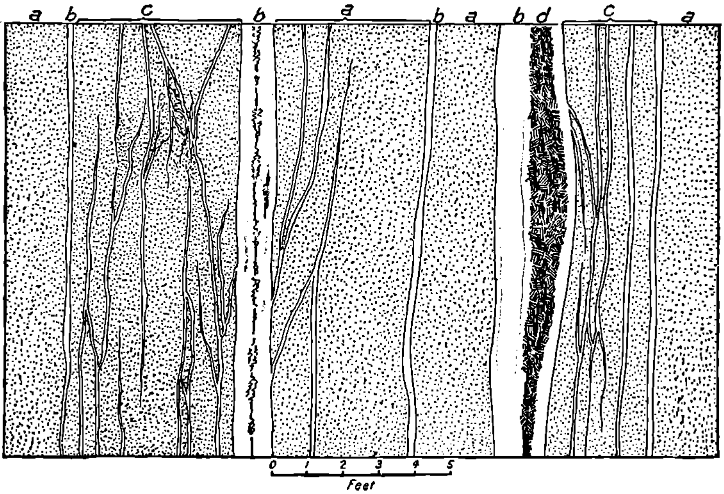

FIG. 46.—Section of Silver Crown lode, Silverton, Colorado, showing lode structure. *a*, andesite; *b*, quartz; *c*, andesite and quartz stringers; *d*, ore. (*After F. L. Ransome, U. S. Geol. Survey.*)

spaced over a considerable width, which may attain 100 feet or even several hundred feet. These large fracture zones, when filled with ore and partially replaced country rock, are called *composite veins* or *lodes* (Fig. 46). The Comstock lode in Nevada illustrates this occurrence; its width in places amounts to several hundred feet.

Lodes often contain two systems of fractures, intersecting at an acute angle, as shown roughly on Fig. 46. This is sometimes referred to as *hammock structure.* A number of adjacent parallel veins are called a vein system. If connected by diagonal veins the term *linked veins* (Fig. 47) is used.

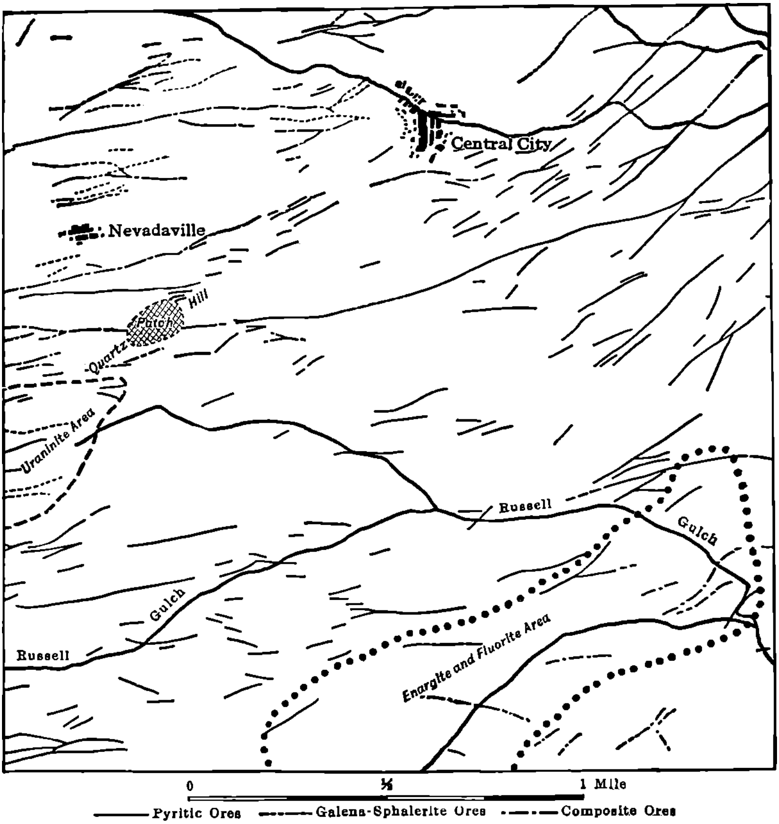

FIG. 47.—Map showing veins of Central City, Colorado, and vicinity. (*After E. S. Bastin, U. S. Geol. Survey.*)

When the fractures are closely spaced and parallel we speak of a *sheeted zone* or a *shear zone* (Fig. 48). Many of the Cripple Creek veins form good illustrations of this mode of occurrence. The width of a sheeted and mineralized zone is rarely over 50 feet and ordinarily much less.

A mass of rock irregularly fractured in various directions by short fissures along which mineralization has spread is called a *stockwork.* Gold-quartz deposits sometimes assume this form; each seam in the several joint systems intersecting the rock may contain a thin but often strongly auriferous sheet of quartz; the mass may be mined as a whole, furnishing low-grade ore. In deeply weathered regions the upper parts of such deposits may be sufficiently disintegrated to be washed by the hydraulic method. In California such mines are called "seam diggings."

A shattered zone cemented by a network of small non-persistent veins is called a *stringer lead* or *stringer lode.*

Sometimes ore deposits are wholly irregular brecciated masses, the ores filling the interstices between the fragments. Again, the breccia may be localized at the intersection of two fractures and a *pipe-like* deposit will be formed, the ore cementing the fragments. Or again, ore deposition may have proceeded in a volcanic vent filled with fragments of rocks due to explosive action. Of such character was the celebrated Bassick deposit in Custer County, Colorado.

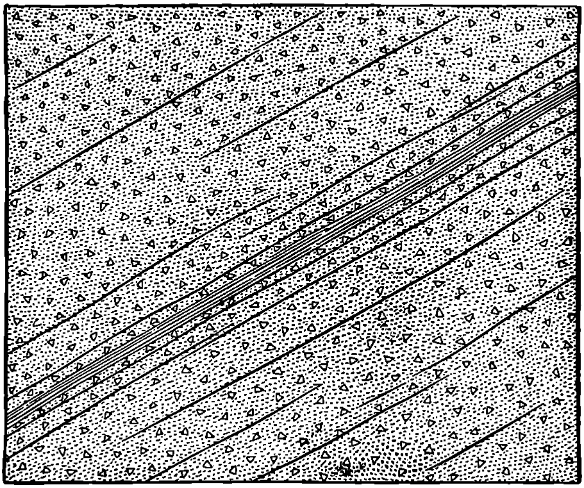

FIG. 48.—Section of the Howard vein, Cripple Creek, Colorado, showing a sheeted zone. Ore follows the close sheeting in the center. Scale: 1 inch = 13 feet. (*After Lindgren and Ransome, U. S. Geol. Survey.*)

Brecciation, shattering, and mineralization often follow lines of weakness along dikes; in such cases, illustrated by the Douglas Island mines in southern Alaska, where a dike of diorite intrudes metamorphic clay slates, the mineralized dike is often referred to as a lode.

*Ladder veins* are deposits filling short transverse fissures sometimes occurring in dikes of intrusive rocks (see Fig. 37).

*Lenticular veins* (Fig. 49) are confined mainly to metamorphic schists and their form is sometimes caused by deformation of an older deposit;

or again the lenticular shape may be due to stresses causing bulging of the schistose layers. It is common to find a number of short lenses of gold-bearing quartz, for instance, scattered along a certain line or zone. Their ends sometimes overlap.

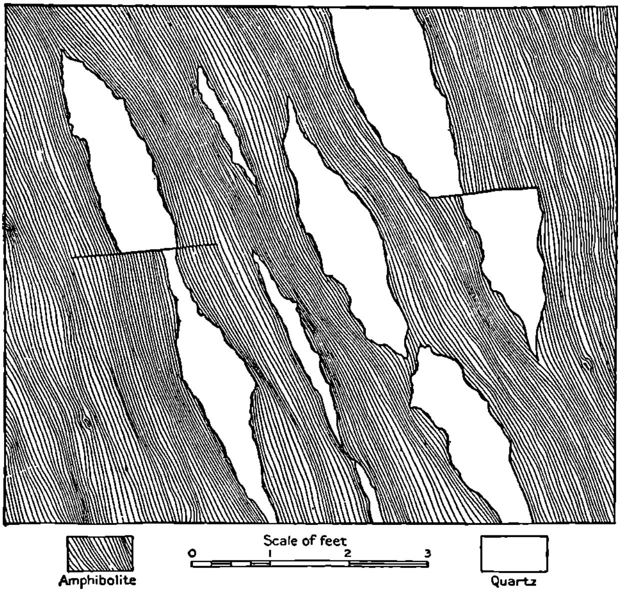

Fig. 49.—Vertical section of Schlegelmilch quartz vein, South Carolina, showing lenticular vein structure in schist with offsets along joint planes. (*After L. C. Graton, U. S. Geol. Survey.*)

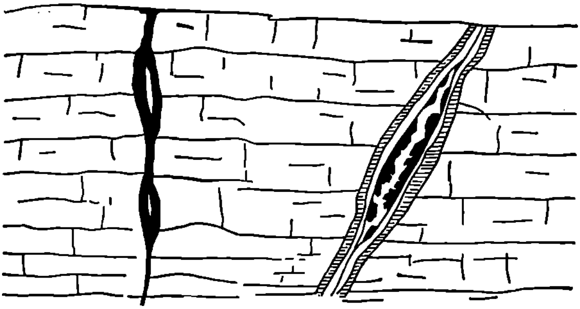

Fig. 50.—Vertical section of gash veins filled with galena (black), or with pyrite, sphalerite, and galena in order of deposition. Drusy cavities in center. Lead mines of Wisconsin. (*After T. C. Chamberlin.*)

*Bed veins* follow the bedding planes in sedimentary rocks.

*Gash veins* are deposits filling non-persistent openings that are of fair width but soon cease when followed along strike or dip; they are particularly characteristic of deposits of galena and sphalerite in limestone and are believed to have been opened by tensional stress, often aided by solution (Fig. 50).

Fig. 51.—Horizontal section of part of Butte district, Montana, on level about 1,500 feet below surface. (*After Reno Sales.*)

Where soft sedimentary beds have been folded and crushed, irregular open spaces are more likely to result than well-defined straight fissures.

In such rocks ores may be found in the spaces opened along anticlines and synclines or in irregular fractures breaking across such folds.

Veins and lodes rarely occur single but, on the contrary, have a tendency to cluster in *vein systems* such as illustrated in Figs. 39, 47, and 51. In some places may be found several intersecting vein systems of great complexity and differing ages and mineralizations as, for instance, in the case of the great copper district of Butte, Montana, illustrated in plan and section in Figs. 51 and 52.[1] The peculiar divergent fractures at the Leonard mine form what is sometimes called a horsetail structure. The veins at Butte are, moreover, in many places disrupted by later faults.

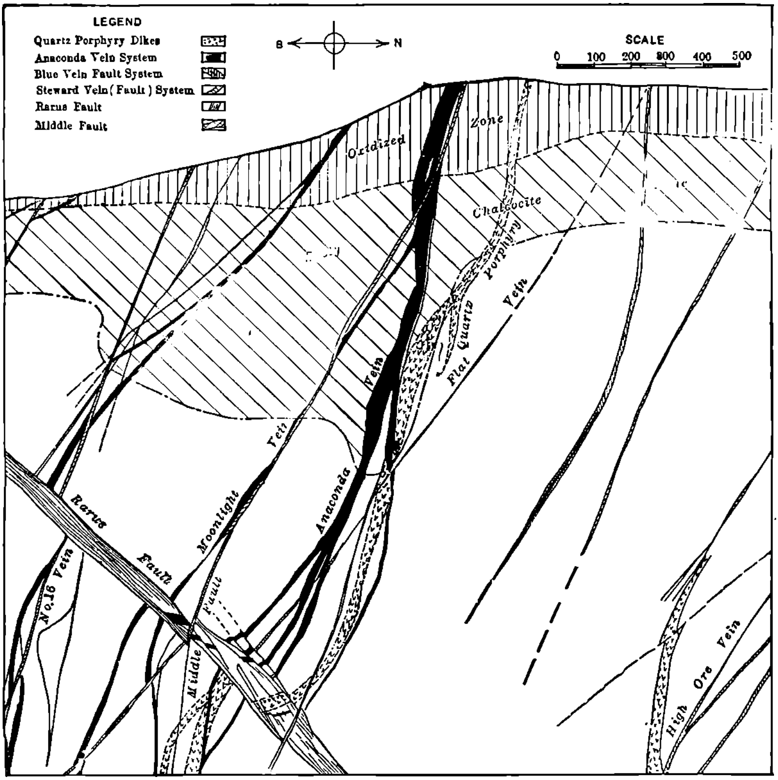

FIG. 52.—North-south section across Butte district, showing structure and ore zones. (*After Reno Sales.*)

**Veins in Relation to the Country Rock.**—Veins crossing the bedding in stratified rocks are referred to as cross veins; those parallel to the

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[1] Reno Sales, Ore deposits of Butte, Montana, *Trans.*, Am. Inst. Min. Eng., 46, 1913, pp. 1–109.

stratification or schistosity are often called bedded veins or bed veins. Differences in the texture and hardness of the rocks traversed influence the form of the vein markedly. In hard dikes crossed by the vein the deposit often splits up into stringers, resuming its typical form beyond this barrier. In fractures formed under light load near the surface there is a great tendency to irregularity and brecciation, especially in the hanging wall. Following G. F. Becker's proposal such may be called *chambered veins.* In a vein of strong dip there will also be a tendency for the hanging wall to settle accompanied by the development of minor vertical fissures. Such conditions were found, for instance, in the Comstock lode, Nevada (Fig. 185) and in the El Oro mines of Mexico (Fig. 179); at both places the vertical hanging wall veins were exceptionally rich, the richness being possibly caused by the impeded movement of the depositing waters. Large masses of country rock included in the vein material are called *horses.* Frequently the vein follows a fissure along the walls of a dike; the lamprophyric dikes, which are the last phases of batholithic intrusions, are especially favored places for ore deposition.

Clayey and soft rocks are most resistant to the development of regular fissures; a fracture in hard rock will suddenly die out when encountering such material; many veins pinch immediately upon entering clay shales or masses of clayey gouge. One of the best examples of this is furnished by the veins of Rico, Colorado, which do not extend through the whole sedimentary series in that district, but suddenly cease at a certain stratum of yielding, plastic rocks, termed *the blanket,* under which almost all the ore-bodies occur. A consequence of this peculiarity of fissuring is that in some regions rich ores are often found just below certain horizons of shale (Fig. 85). In southern New Mexico a persistent Devonian shale plays this part of "indicator." The vein solutions were arrested at this horizon and there deposited their load.

When a vein follows the contact between two formations, say between granite and andesite, we speak of it as a *contact vein.* The contact is usually caused by faulting movements in the plane of the fissure, and such veins are in no wise different from ordinary fissure fillings. They should not be confused with contact-metamorphic deposits, which belong to a separate class.

**Vein Walls.**—In a simple filled fissure vein we have well-defined foot and hanging walls, Fig. 45, which often are smooth surfaces and represent a single fissure opened by a small or large movement along its slightly curved plane. In a replacement vein the fissures are comparatively tight and in most cases appear to have been formed under strong compressive stress that reduced the open spaces to a minimum. The vein-forming solutions were forced into the country rock, and the ores formed by replacement gradually merge into unaltered rock. In such cases we may find a single fissure plane with ore on both sides and not limited by any well-defined walls. The exact limits of commercial ore can be found only by assay and are often spoken of as "assay walls."

In a composite vein or lode or in a sheeted zone there may be several smooth walls and if no cross-cutting is undertaken there is danger that parallel ore-bodies separated by sheets of country rock may be overlooked.

**Outcrops.**—The character of the outcrop of a vein, or in fact of any deposit, is determined by the predominant minerals and by the prevailing surface conditions. In regions of long-continued rock decomposition and inactive erosion, as, for instance in some of the Southern Appalachian States, even the most resistant outcrops may be reduced by weathering; and nothing but fragments scattered over a wide area may be visible at the surface.

Under conditions of fairly active erosion veins with predominant quartz stand out prominently and can be easily traced.

On the other hand, veins with carbonate gangue are likely to weather more rapidly than the surrounding rock; and the deposits may be indicated by little depressions or by notches in the ridges. Where sulphides are abundant, their oxidation is conspicuously reflected in the outcrops. Deposits of mingled quartz and sulphides then form prominent outcrops of limonite and residual quartz; this is the *gossan* of the Cornish, the *ironstone* of the Australian, the *eiserner Hut* and the *chapeau de fer* of the German and French, and the *colorados* of the Spanish terminology. (See Chapters XXI and XXXII.)

**Length and Depth of Veins.**—Where veins follow great dislocations their length may be considerable. One of the more recent veins of Freiberg, Saxony, called the Halsbrücker Spat, has been followed for almost 5 miles. Some of the lead-bearing veins in the Harz Mountains, Germany, are traceable for 12 miles. Exceptionally long single ore-bearing fissures are found in the Silverton quadrangle, San Juan region, Colorado; some of them are 5 miles long. Some of the Mother Lode veins in California can be traced for many miles. The longest single quartz vein known appears to be that known as the Pfal, in the Bavarian Forest, which is said to be traceable in a straight line practically without interruption for 140 kilometers through the pre-Cambrian rocks.[1] The quartz is said to be barren of metals.

The great majority of single ore-bearing veins are short and their outcrops can rarely be traced for more than one mile; they do not, as a rule, occupy great dislocations, but rather subordinate fissures. The great dislocations are formed during mountain building by tangential stresses, whereas the ore-bearing veins are, as a rule, formed after epochs of igneous activity. In the Coeur d'Alene district, Idaho, for instance,

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[1] E. Suess, Das Antlitz der Erde, Leipzig, 1883, 1, pp. 270–272.
W. von Gümbel, Geologie von Bayern, Cassel, 1894, 2, pp. 461–464.

the rich galena veins show little connection with the principal structural faults of the region and were probably not formed at the same time.

Veins do not necessarily continue to great depths. There are all kinds of fissures, some disappearing within a short distance below the surface, others continuing down to the greatest depths attained, or about 7,000 feet (Morro Velho, Brazil). Deep tunnels have been run to intersect veins of favorable appearance on the surface and have failed to disclose their continuation in depth. There is no definite relationship between depth and length of a fissure, though it is true that fissures showing strong movement and shattering are likely to continue to great depths. Again the ore-body may be limited in depth, while the barren fissure continues below it as strong as ever.

Bends and curves in strike and dip are common in veins, but as a rule a vein retains its general angle of dip with remarkable persistence. The dip may be at any angle, but veins dipping from 50° to 80° are most common. The North Star vein at Grass Valley, California, is one of the best instances of a low-dipping vein of great length; with a dip of 20 degrees it has been followed for 9,000 feet. Still flatter veins are called blanket veins and are seldom very persistent or uniform.

# CHAPTER XIV

## TEXTURE OF MINERAL DEPOSITS

### FILLING AND REPLACEMENT

**Introduction.**—The ore minerals and gangue which make up an ore deposit present various types of texture. The texture of an ore is dependent upon many factors. Space available for deposition, concentration and composition of the generating solutions, time, temperature, and pressure—all are of importance in determining the primary texture. Many changes take place in a deposit once formed; the secondary textures, so far as they are caused by solution and redeposition, are influenced by the same factors, and, in addition, deformation by pressure plays an important rôle.

**Texture of Deposits of Igneous Origin.**—The ores consolidated from magmas have in general the texture of igneous holocrystalline rock. The principal minerals comprise chalcopyrite, pyrite, pyrrhotite, magnetite, chromite, and ilmenite. The texture is ordinarily coarse granular, hypidiomorphic; the chalcopyrite and pyrrhotite are rarely crystallized, but may contain phenocrysts of pyrite and magnetite, both of which are frequently developed with crystalline outlines. The ores may, of course, contain phenocrysts and anhedrons of other rock-forming minerals. particularly soda-lime feldspars, olivine, and pyroxene. Eutectic texture results if the magma was a eutectic mixture from which two minerals crystallized simultaneously after the manner of graphic granite. Approximation, at least, to such texture is shown by some intergrowths of magnetite and apatite.

If the ores have been subjected to dynamic metamorphism, granulation and metasomatic development of hornblende, garnet, biotite, sericite, and epidote in coarse or fine aggregates follow, and the ore may acquire schistose structure.

**Texture of Pegmatite Dikes.**—The pegmatite dikes are believed to have been deposited by magmatic solutions of great fluidity and low temperature (about 600° C.). In some cases the pegmatites form transitions between igneous rocks and veins deposited by hot solutions. Their texture is coarsely crystalline, often drusy, and the minerals have a strong tendency to idiomorphic development. Large crystals are the rule, and sometimes they attain enormous dimensions; crystals of spodumene at the Etta mine, South Dakota, are 40 feet or more in length. Quartz crystals several feet long have been observed in these deposits. A rough tendency to crustification is often present, and the walls of the dikes are then lined with crystals of feldspar or mica. In the druses are found the minerals of later generations, like albite and tourmaline.

**Texture of Sedimentary Deposits.**—Ores and minerals of sedimentary deposits are usually fine-grained, and in many cases have been deposited as colloids in which subsequent fine-grained crystallization has developed. Coarsely crystalline, allotriomorphic structure may develop in deposits consisting of calcite, salt, or gypsum.

In many cases the structure is clastic with development of new-formed minerals between the grains. Newly formed quartz, if present, nearly always assumes a microcrystalline or cryptocrystalline texture. Subsequent metamorphism is likely to enlarge the crystalline grains and result in coarser-grained ores.

**Concretions.**[1]—Concretions are rounded bodies of some mineral aggregate which are often found in shale and sandstone. Calcite, silica, siderite, pyrolusite, barite, pyrite, marcasite, and limonite are among the minerals which most commonly form concretions. The structure is often concentric or radial. In some cases the stratification planes pass through the concretions, while in other cases they may bend around them. These structures are of some economic importance as regards ores of iron and manganese. They often have a center of a clastic grain or a fragment of a fossil shell or leaf. Concretions generally result from processes of solution and precipitation in soft or semi-consolidated sediments, less frequently in hardened rocks. The precipitates are usually gels, which later become crystalline (metacolloids). As a rule, removal by solution of the rock material (*e.g.*, shale) is necessary and its place is taken by the precipitated gel. Most concretions are essentially gel replacements. For instance, precipitation of colloidal iron sulphide may start around decomposing organic matter evolving $H_2S$ and the process continues while iron solutions and hydrogen sulphide are available. Concretions generally derive their substance from the surrounding rock. They frequently show cracks and cavities caused by contraction of the original gel material. When small, uniform and abundant they are called oölites.[2]

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[1] J. E. Todd, Concretions and their geological effect, *Bull.* Geol. Soc. Am., 14, 1903, pp. 353–368.

See also, W. A. Tarr, *Bull.* Geol. Soc. Am., 32, 1921, pp. 373–384, and W. A. Richardson, *Geol. Mag.*, 1921, pp. 114–124.

W. Lindgren, Metasomatism, *Bull.* Geol. Soc. Am., 36, 1925, pp. 247–261.

W. H. Newhouse, Some forms of iron sulphide, etc., *Jour. Geol.*, 35, 1927, pp. 73–83.

W. H. Twenhofel, Treatise on sedimentation, 2d ed., Baltimore, 1932, pp. 696–716, 757–769.

[2] For literature regarding the origin of oölites see *Fortschritte Mineral. Krist. Petr.*, Jena, 3, 1913, p. 43.

W. H. Twenhofel, *op. cit.*

W. H. Bucher, On oölites and spherulites, *Jour. Geol.*, 26, 1918, pp. 593–609.

The oölitic texture is characteristic of many deposits of calcite, siderite, bauxite, calcium phosphate, barite,[1] limonite, and psilomelane; pyrite or marcasite rarely assumes this form. The oölites are often affected by later alteration and recrystallization.

**Texture of Residual and Oxidized Deposits.**—In the residual deposits of the zone of oxidation, the ore-bodies are usually very irregular in structure and texture. In large part they were deposited as colloids, which subsequently in part have developed fine-grained crystalline texture.

Earthy, clayey concretionary, mammillary, stalactitic, or pisolitic textures are common, the last being defined as a coarser development of the oölitic form. Coarser crystalline form is assumed by some minerals like calcite, barite, smithsonite, calamine, and cerussite. Crustification or drusy structure is common in places. Quartz, where developed, is usually fine-grained or cryptocrystalline.

## THE TEXTURE OF EPIGENETIC DEPOSITS

**Primary Texture of Filled Deposits.**[2]—The epigenetic deposits are of manifold form and origin, but the majority of them result from aqueous solutions either by filling of open cavities or by replacement of surrounding rocks. Precipitation from complex solutions in open spaces takes place in a certain orderly succession, and the deposits, therefore, readily assume a banded texture; crystallization is facilitated by the open spaces, but the older crystals interrupt the development of the products of later crystallization. Hence a hypidiomorphic to panidiomorphic texture is most common.

Banding by deposition is called *crustification,* a term introduced by Posepny. In many classes of veins, whether banded or not, a drusy texture is common.

*In deep-seated veins* formed at a temperature but slightly lower than that of the pegmatites the texture is usually coarsely crystalline and massive; sometimes even drusy cavities are lacking. Delicate and repeated banding is absent, but a coarsely banded or comb structure recalling that of the pegmatite veins is sometimes encountered. It is usually expressed by quartz crystals developing from the sides or by metallic minerals like wolframite or cassiterite attached to the walls of the fissure.

*In veins formed at intermediate temperatures* a coarsely crystalline massive texture is most common; combs and rough banding by deposition are by no means unknown, especially where the deposit contains calcite or barite. In quartz veins the filling appears to have taken place rapidly

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[1] E. S. Moore, Oölitic and pisolitic barite, *Bull.* Geol. Soc. Am., 25, 1914, pp. 77–79.

[2] S. F. Adams, A microscopic study of vein quartz, *Econ. Geol.,* 15, 1920, pp. 623–664.

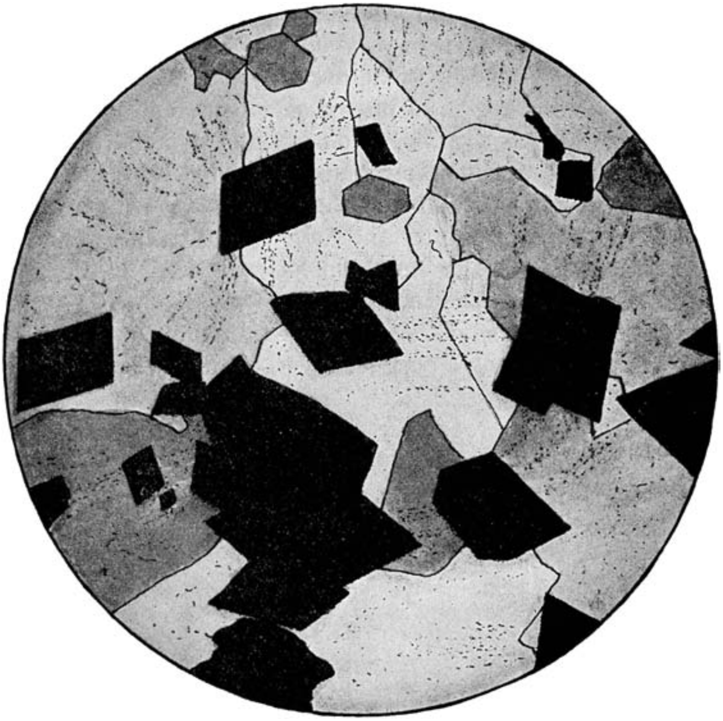

FIG. 53.—Thin section showing normal texture of quartz filling. Black, arsenopyrite; remainder, hypidiomorphic quartz with fluid inclusions. Magnified 32 diameters. Gold-quartz vein, Grass Valley, California.

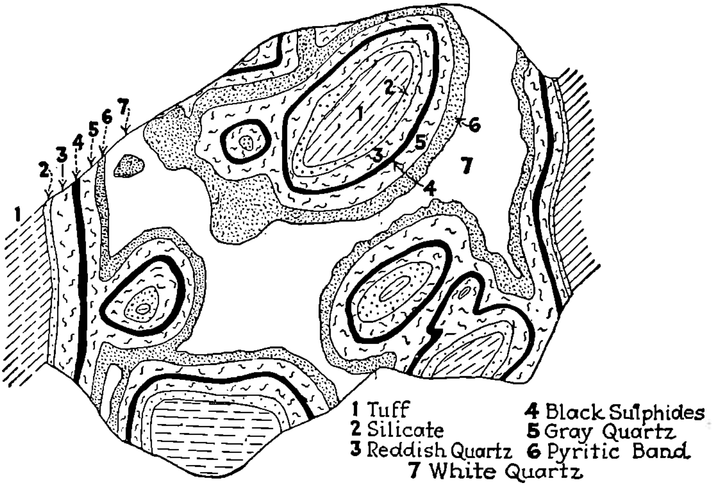

FIG. 54.—Cockade structure in Alacran vein, Mexico. The country rock is a volcanic tuff. (*After J. E. Spurr, "Economic Geology."*)

and completely, so that the resulting ore consists of an irregular massive mixture of quartz and sulphides. That here too the deposition began from the walls is indicated by some occurrences of partially filled veins which form a loose aggregate of prisms to which small crystals of early sulphides may be attached. Any thin section of such quartz will usually show long crystals of earlier growth around which the later quartz has been deposited in large individuals (Fig. 53). Lines of inclusions often penetrate from one grain into another. These inclusions consist of aqueous solutions, often with small cubes or grains of transparent salts suspended in the liquid. Inclusions of carbon dioxide have been reported, but are extremely scarce. The optical continuity of the crystals or grains is often disturbed by a peculiar divergent "flamboyant" structure caused by slight differences in orientation, and not by internal strains.

Inclusions of country rock may be surrounded by concentric rings of different sulphides, and a primary brecciated vein structure may result.[1]

The sulphides are coarsely crystalline and sometimes roughly banded, parallel to the walls. Pyrite and arsenopyrite, both among the earliest minerals, have a strong tendency to crystal development, while galena, sphalerite, chalcopyrite, and tetrahedrite are much less commonly found with crystal faces.

A banded structure sometimes results from the filling of several closely spaced fissures. In quartz veins in fissile rocks a peculiar *book structure* may result from numerous parallel sheets of slate, alternating with quartz. It has been thought that this and other features difficult to explain by the assumption of open cavities are due to the opening of spaces by the *force of crystallization.* Such views have been expressed by E. Suess, W. O. Crosby, E. J. Dunn, S. Taber, and others. It is improbable that crystallization could have opened the cavities. More likely they were supported by the strong pressure of magmatic waters. But within such spaces a slight force exerted by crystallization could readily detach slices of shale from the walls.

Stalactites are unknown in deposits formed at high or intermediate temperature.

*In veins formed at lower temperatures and comparatively shallow depths* crustified and drusy forms and fine granular texture predominate. The quartz filling is usually fine-grained, ranging to cryptocrystalline and chalcedonic near the surface.

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[1] Such structures are called "cockade ore" and fine instances are described in the old literature from Freiberg and Clausthal. For other illustrations see Pl. VII in W. Lindgren's report on Grass Valley and Nevada City, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, 1896, pt. II. Also paper by Spurr, *Econ. Geol.*, 21, 1926, pp. 519–537. Cockade ore with repeated deposition of cassiterite and pyrite has been described by M. J. Buerger and J. L. Maury, *Econ. Geol.*, 22, 1927, pp. 1–13. In many cases such concentric deposits are metacolloids, as suggested by H. C. Boydell and others.

The sulphides are found in small crystals or small anhedrons; large crystals of pyrite, so common elsewhere, are rare in these veins. On the other hand, where calcite, dolomitic carbonates, rhodochrosite, fluorite, or barite are gangue minerals the crystals may be much larger than those found in other deposits. An example is furnished by the magnificent crystals of calcite at Joplin, Missouri, and here galena also appears in unusually large, well-developed individuals. Symmetrical and delicate crustification is often present (Figs. 180, 193, and 204). Brecciated structure of primary origin is common.

Inclusions of country rock are found in veins of all kinds. Sometimes they form a rubble, between which the ores have been deposited. In other cases they appear unsupported in the vein material; generally, they are more or less altered. Such inclusions may have dropped from the walls while the vein filling was partly completed; or may have been detached from the walls by the force of crystallization of the vein matter. Most fissures are probably originally occupied by comminuted rock material and fragments of all sizes. During mineralization the fine-grained matter is easily removed and replaced while the larger fragments remain and are only partly replaced.[1] Some "inclusions" are simply remnants which have escaped the general replacement of the rock.[2]

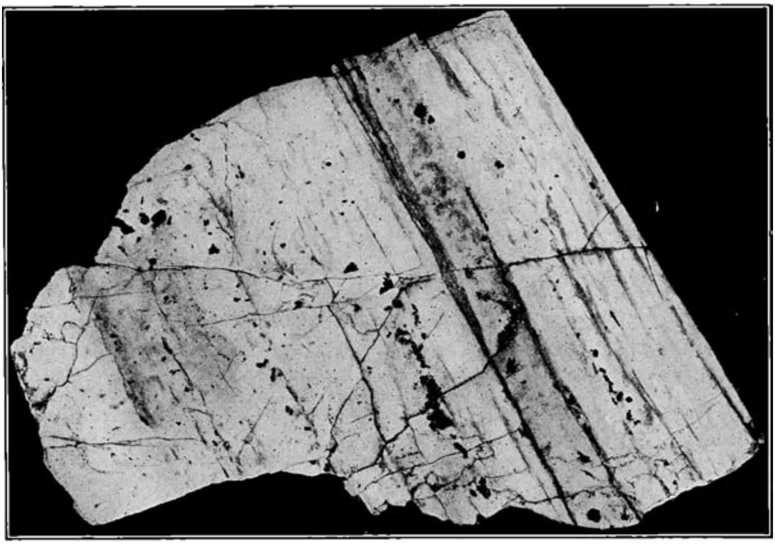

FIG. 55.—Specimen of quartz from Nevada City, California, showing ribbon structure by sheeting. Two-thirds natural size.

**Secondary Textures and Structures of Filled Deposits.**—Crushing and brecciation of the early minerals are extremely common; indeed, few

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[1] J. C. Schagen van Soelen, *Eng. Min. Jour.* 122, 1926, pp. 777–778.

[2] A. M. Bateman, *Econ. Geol.*, 19, 1924, pp. 504–520.

FIG. 56.—Thin section of vein quartz from Nevada City, California, showing crushing and incipient ribbon structure. Magnified 12 diameters. Crossed nicols.

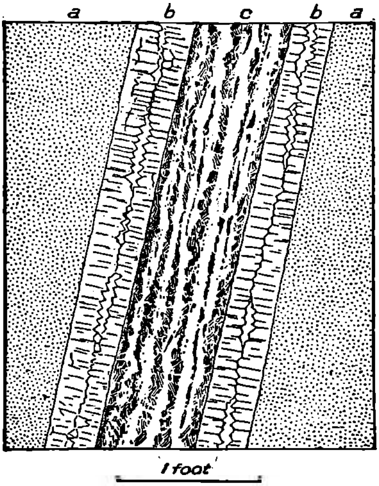

FIG. 57.—Cross section of Japan vein, Silverton, Colorado, showing structure produced by repeated opening of original fissure. *a*, country rock; *b*, quartz; *c*, ore. (*After F. L. Ransome, U. S. Geol. Survey.*)

veins are entirely free from it. Repeated opening of fissures (Fig. 57) and the deposition of new generations of vein material often take place and the cementing ore may be enriched at the expense of the older generations.

A banded or sheeted structure often results from the development of shear planes in the old filling; examples of this are seen in many gold-quartz veins of California (Figs. 55 and 56). Along these shear planes the quartz is deformed and granulated; and gold may be deposited along them by processes which may be called secondary, though, as a rule, they take place shortly after the vein formation. The shearing stress exerted either before or after the filling may affect the walls of the vein and render them close-jointed or even distinctly schistose.

In some deposits, especially those containing zeolites, calcite, or barite, secondary replacement processes play an extensive part. A vein filled by calcite may be replaced by quartz, which then plainly shows its secondary nature by its hackly or lamellar texture, casts of cleavage, pieces of calcite, or imprints of cleavage lines. Such pseudomorphic textures are sometimes accompanied by a marked enrichment of the metallic content of the deposit.

**Metasomatism in Mineral Deposits.**—The nature of metasomatism or replacement has already been described on pages 23 and 91. Many deposits have been formed by solutions containing various salts and gases and capable of attacking certain kinds of rocks. Guided by fissures or other open ducts, the solutions deposit part of their load in the open super-capillary spaces whenever supersaturation takes place; thus is produced the filling of fissures. As almost all rocks are porous and as the solutions are frequently under heavy pressure they will be forced into the rocks and will produce chemical and mineralogical changes in them. At the same time the porous rock acts undoubtedly as a semi-permeable membrane through which various substances will diffuse at differing rates—electrolytes and gases most easily, colloids and difficultly ionized compounds very slowly. Thus any vein will usually be accompanied by a strip of altered country rock in which the solutions have effected certain metasomatic changes. The minerals in the open fissures will ordinarily differ from those formed in the metasomatic zone. We may find, for instance, a quartz filling with various sulphides and gold, while the minerals developed in the country rock consist of pyrite, sericite, and calcite with little if any gold. In some cases no perceptible alteration may be observed in the country rock. The only difference between a filled vein accompanied by metasomatism and a so-called replacement deposit is that in the latter the filling of the narrow open spaces is negligible and the bulk of the ore has been formed by metasomatic processes.

**Metasomatic Processes.**—In a solid rock replacement may be caused by many kinds of solutions the only requirements being that some or all of the rock minerals must be unstable in the penetrating fluids. The usual substances, most active in aqueous solutions, are oxygen, carbon dioxide, sulphuric acid, ferric sulphate, hydrogen sulphide, alkaline sulphides, and alkaline carbonates. Replacement may occur at all temperatures above the freezing point of the solution and below the melting point of the rock; it is naturally most effective in hot solutions. Replacement may proceed at any pressure. It may be effected by the ordinary surface waters, by sea water, by hot ascending waters, and by magmatic emanations whether gaseous, fluid, or above the critical temperature. According to the most important products we speak of silicification, carbonatization, chloritization, sericitization, tourmalinization, pyritization, etc.

There is no rock that is proof against replacing natural solutions of some kind. Limestone and dolomite are most easily replaced even at ordinary temperatures, for instance, by iron carbonate (siderite) or by zinc carbonate (smithsonite). Granite, diorite, and, in fact, all igneous rocks are also subject to replacement. Even quartzite, slate, and aluminous shale may be replaced by other minerals though they are more resistant than others. Replacement by sulphides may to some extent take place at ordinary temperatures (for instance chalcocite replacing pyrite) but large deposits of sulphide ore are usually formed by hot solutions.

**Mode of Replacement.**—As pointed out in previous chapters replacement is effected by concentrated solutions filling capillary openings of extremely small size (sheet openings larger than 0.0001 mm., p. 33) which are just above or below the limit of microscopic visibility. Cases have been noted where replacement begins from a crack doubtless filled with a film of solution and connecting a series of just visible fluid inclusions. Solution and precipitation go on practically simultaneously dependent upon the constantly changing equilibrium, the supply of solvent and the facility of escape for the dissolved material. Two or several minerals may be dissolved at the same time to make room for the new as in the replacement of shale by a pyrite crystal. The volume of the rock remains constant, held by pressure. The moment a place is available some mineral will separate out from the concentrated solution. This law fails to apply in free crystals or when rock pressure can be overcome by the force of crystallization, or when a solid is replaced by a gel, or when the solutions circulate so rapidly that there is a strong balance in favor of solution. As crystal grains develop they will exert a different amount of pressure in various directions thus facilitating solution in the direction of greatest pressure. The development of crystals in the host mineral is a result of this action.

The power of crystallization of the different minerals varies greatly, for some are found only as anhedrons in metasomatic rocks, while others always assume their crystal form. The following list gives the relative power of crystallization in solid rocks of some minerals, as beginning with those of strongly emphasized individuality: Rutile, tourmaline, staurolite, arsenopyrite, pyrite, magnetite, barite, fluorite, epidote, pyroxene, amphibole, siderite, dolomite, albite, mica, galena, sphalerite, calcite, quartz, orthoclase.

When a crystal has ceased to grow solution may still continue parallel to its faces. As no more material for the crystal is at hand the voids are immediately filled by the next precipitate available. Thus are explained the thin films of quartz or calcite which so frequently surround metasomatic pyrite crystals.

FIG. 58.—Replacement veinlets of galena (white) in cryptocrystalline quartz (dark gray) with vugs (black). Tintic, Utah. Magnified 11 diameters.

When taking place under the law of constant volume replacement cannot ordinarily be expressed by the simple chemical formulas[1] usually given. The reactions are likely to be more complicated.[2]

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[1] W. Lindgren, Volume changes in metamorphism, *Jour. Geol.*, 26, 1918, pp. 542–554.

[2] Smithsonite often replaces calcite with preservation of structures indicating constant volume. The reaction is supposed to follow the formula

$$CaCO_3 + ZnSO_4 = CaSO_4 + ZnCO_3,$$

both $ZnSO_4$ and $CaSO_4$ being water soluble salts. One cubic centimeter of calcite contains 1192 milligrams $CO_2$ and 1518 milligrams CaO, while one cubic centimeter of the resulting smithsonite contains 1514 milligrams $CO_2$ and 2787 milligrams ZnO. It is clear then that the principle of equal volumes requires more $CO_2$ than is available in the calcite. If the process follows the formula, shrinkage of volume will necessarily result.

In metasomatic processes gangue minerals like sericite, calcite, siderite, barite, and fluorite replace all silicates. Ferromagnesian silicates will be attacked first, then the soda-lime feldspars, lastly orthoclase and albite. The degree of attack on quartz depends probably on the amount of alkaline carbonates in the solution. All sulphides replace all silicates as well as quartz (Fig. 58). Sulphides and sulphosalts readily replace other sulphides. A succession common in many ores is (1) pyrite (oldest), (2) sphalerite, (3) enargite, (4) bornite, (5) chalcopyrite, and (6) galena. Any of the later minerals may replace any of the earlier products (Fig 59). Our knowledge of these manifold replacements has been greatly increased through the study of polished sections in reflected light. Gangue minerals like sericite, chlorite, calcite, quartz, fluorite, and barite very rarely replace sulphides.

FIG. 59.—Feathery lead sulphantimonide (white) replacing galena. Tintic, Utah. Magnified 24 diameters.

**Diffusion.**—By diffusion is meant the movement of the molecules in a solution from a place of higher concentration to a place of lower concentration. This process is most important for the phenomena of replacement and ore deposition in general. The diffusion power varies with the concentration and differs markedly for different substances. The general formula for diffusion without chemical reaction is $A = k\sqrt{z}$, in which $A$ is the amount of substance diffusing through unit area within a given time $z$, and $k$ is a constant for each concentration at a fixed temperature.

The material has the tendency to distribute itself uniformly through the available space. But the process, as will be seen from the formula, is exceedingly slow, does not act over large distances and ceases after a certain time. Diffusion is incompetent to originate mineral deposits; it must be supplemented by the movement of the solution on fractures or other open spaces. The great importance of diffusion is probably in the mechanism of replacement. In the ultimate small spaces available for metasomatism, there is constant change of concentration; and diffusion attends to the moving up of the new molecules and the removal of the by-products of replacement. Diffusion acts easily in a gel. Diffusing substances may meet and produce uniform or rhythmical replacements.[1] Diffusion also operates in other media such as in sand or other fine-grained material, also in water. It may likewise operate in permeable rocks, such as limestone[2] or sandstone, and may there produce uniform replacement or rhythmical banding, as, for example, in the Leadville limestone where sphalerite and pyrite sometimes alternate or as in the banded magnetite in iron ores from Sweden and from Cuba. Such banding is coarse and lacks the colloform characteristics.

**Texture of Metasomatic Rocks.**—In metasomatism new minerals develop at countless points in the old rock, some growing with crystal form (metacrystic or crystalloblastic series, p. 174) while others grow into irregular grains. Each new grain may be called a metasome, each new crystal a metacryst (pseudophenocryst).[3] The resulting textures will be holocrystalline; the new minerals frequently contain inclusions of the old (sieve texture); and if the replacement is incomplete, as often is the case, enough of the old texture may be preserved to indicate the original rock (relict texture). It is characteristic of some replacements that even if the process has been carried to completion the original texture may be preserved, as in silicified oölitic limestone and in silicified dolomites (Fig. 61). In many cases, however, the original texture is wholly destroyed.

The structure of a rock may be faithfully preserved even when metasomatic action has destroyed its texture. Such preserved structures are, for instance, stratification, joints, breccias, folds and vesicules in lavas. Preservations of texture of limestones which have been completely

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[1] R. E. Liesegang, Geologische Diffusionen, 1913.

M. Watanabé, Some problems of diffusion in the special reference to the study of ore deposits, *Sci. Repts.*, Tohoku Imp. Univ., sec. III, vol. 2, Sendai, 1924. Both of these publications are highly recommended.

[2] A. R. Whitman, Diffusion in ore genesis, *Econ. Geol.*, 23, 1928, pp. 473–488.

[3] This term was first introduced by A. C. Lane, *Bull.*, Geol. Soc. Am., 14, 1903, p 369.

Grubenmann and Becke use the terms xenoblast and idioblast. N. Grubenmann, Die kristallinen Schiefer, Berlin, 1910, p. 91.

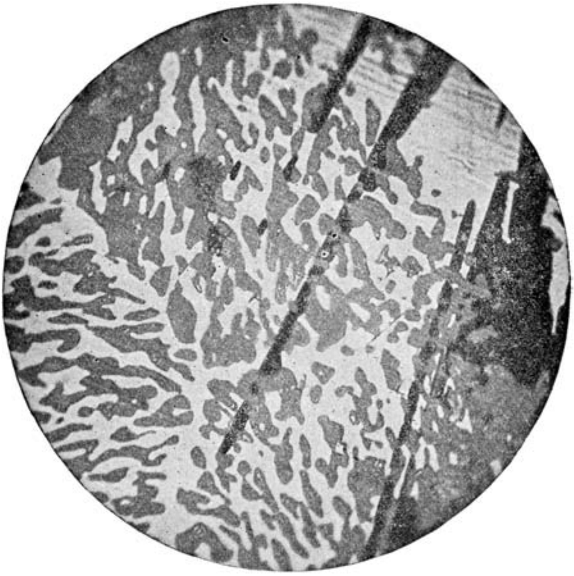

FIG. 60.—Galena (light gray) replaced by tennantite and pearceite ($9Ag_2S.As_2S_3$). Earlier barite plates are not replaced by galena but by later minerals. Tintic, Utah. Magnified 690 diameters.

FIG. 61.—Thin section of dolomite completely silicified, but retaining texture and crystal form. (*After J. D. Irving, U. S. Geol. Survey.*)

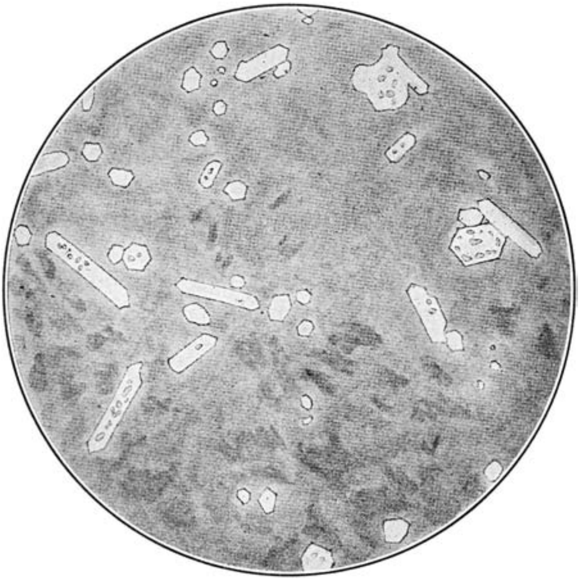

FIG. 62.—Incipient silicification of limestone. Aspen, Colorado. White areas represent quartz crystals with small inclusions of limestone. Magnified 30 diameters.

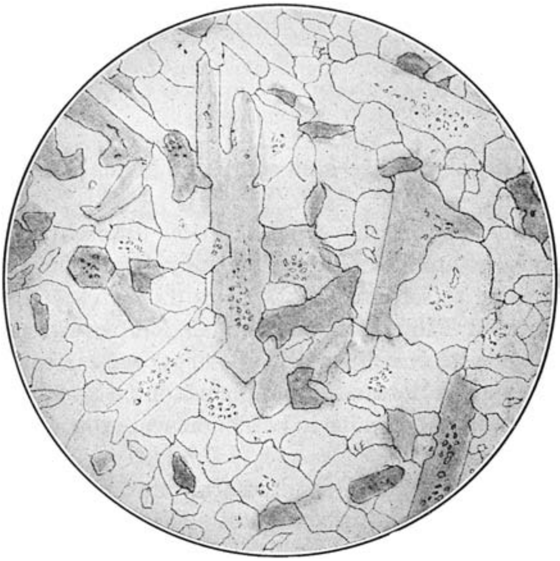

FIG. 63.—Silicified limestone ("jasperoid"). Aspen, Colorado. All quartz. Small inclusions of calcite in some of the grains. Crossed nicols. Magnified 30 diameters.

replaced by sulphides are mentioned by S. F. Emmons[1] and J. M. Boutwell.[2]

In many replacements gangue minerals like quartz and barite may crystallize first, while pyrite comes next and other sulphides later.

Irving[3] has pointed out that in some cases replacement begins from a great number of points in the rock where metasomes or metacrysts may develop (Fig. 62) and by continuation of the same process (Fig. 63) the remainder of the rock is finally replaced; the contact is then indefinite. In other cases the complete change occurs rapidly, advancing like a wave over the country rock; the contacts are then sharp and the process probably consisted in replacement of the original rock by colloidal silica. The replaced rock is usually dense and compact; in places, however, drusy cavities occur in it.

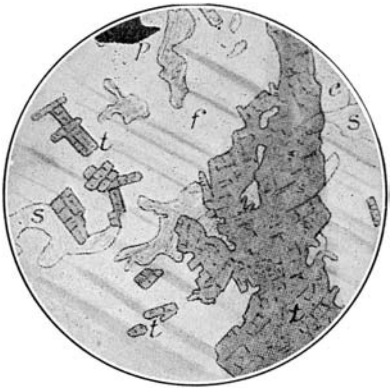

Fig. 64.—Replacement veinlet of tourmaline in fresh andesine grain. Keystone Mine, Meadow Lake, Nevada County, California. *t*, tourmaline; *f*, andesine; *e*, epidote; *s*, sericite; *p*, pyrite. Magnified 50 diameters.

Under the influence of the same solution different results may be produced in different rocks. Limestone may be silicified while diorite may be transformed to sericite.

**Replacements at High Temperature.**—Complete recrystallization, development of silicate minerals with little or no water, and coarse texture are typical of deposits formed by replacement at high temperatures, probably above 400°. Mineralizers like fluorine, boron, or phosphorus are frequently introduced.

The best examples of such textures are found in the replacement of limestones in contact-metamorphic deposits (Figs. 289 and 290) resulting in coarse aggregates of metacrysts of andradite garnet with metasomes of quartz, calcite, epidote, and pyroxene. The limestone may be recrys-

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[1] S. F. Emmons, *Trans.*, Am. Inst. Min. Eng., 23, 1893, p. 602.

[2] J. M. Boutwell, *Prof. Paper* 38, U. S. Geol. Survey, 1905, p. 193.

[3] J. D. Irving, *Jour.* Canadian Min. Inst., 14, 1911, pp. 395–471.

tallized in part to coarse calcite. Magnetite and sulphides develop in large grains.

Adjoining tin-bearing veins the rocks are recrystallized to greisen, consisting of coarse metasomes of muscovite, topaz, quartz, tourmaline, fluorite, and cassiterite (Fig. 262). Calcareous rocks or greenstones containing much lime are recrystallized to aggregates of axinite, actinolite, garnet, etc.

Replacement by apatite (containing phosphorus), scapolite (containing chlorine), and pyroxene occur adjoining certain high temperature veins.

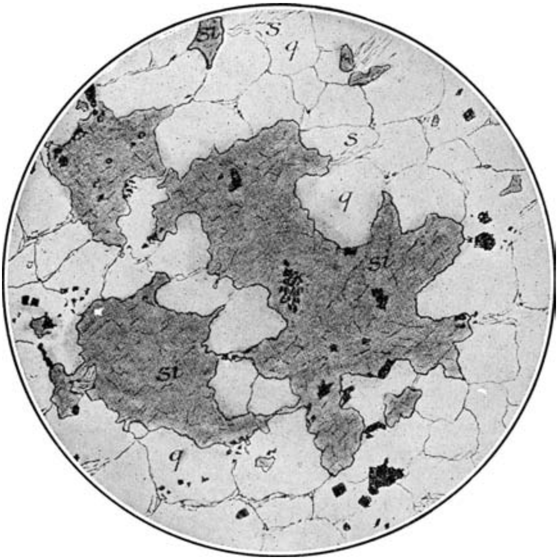

FIG. 65.—Siderite with pyrite and galena, replacing quartzite. Helena and Frisco Mine, Coeur d'Alene, Idaho. *q*, quartz grains; *s*, sericite; *si*, siderite; black, galena and pyrite. Magnified 100 diameters.

Along some gold-quartz veins albite, biotite, and zoisite may develop in the wall rock.

**Replacements at Intermediate Temperature.**—Replacements at more moderate temperature are likely to result in textures of finer grain, and hydrous silicates like chlorite and sericite are abundant (Fig. 66). There are exceptions to this where barite or fluorite replaces limestone for both of these minerals easily develop as perfect metacrysts (Figs. 67 and 69). Silicification of limestone, argillaceous shale, and rhyolite is a very common process taking place frequently with preservation of texture. The quartz will usually be fine-grained. Silicified limestones are called jasperoids (Figs. 62 and 63).

Limestone may be replaced by massive sulphides (Fig. 68). Alteration by hot waters of granular and porphyritic igneous rocks as well as of schists of similar composition results in sericitization of the femic and salic minerals, sometimes also of the quartz, with development of fibrous aggregates. Pyrite, secondary quartz, rutile, and albite are sometimes found in these rocks. In some classes of such metasomatic rocks carbonates of calcium, magnesium, and iron also occur.

Serpentine is altered near some gold-quartz veins to coarse aggregates of ankerite, quartz, and mariposite (chromiferous sericite).

Quartzite and quartzitic slates in some lead deposits may be extensively replaced by siderite (Fig. 65).

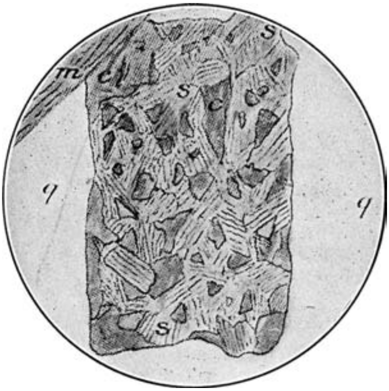

FIG. 66.—Andesine crystal in granodiorite, replaced by sericite and calcite. Pinetree vein, Ophir, Placer County, California. *q*, quartz; *m*, muscovite; *c*, calcite; *s*, sericite. Magnified 80 diameters.

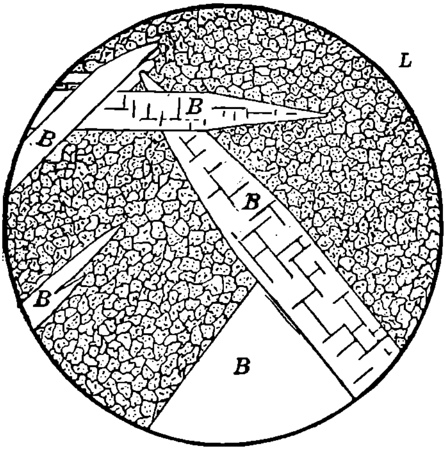

FIG. 67.—Barite (*B*), replacing gray, fine-grained limestone (*L*), Ouray, Colorado. (*After J. D. Irving, U. S. Geol. Survey.*)

**Replacement at Low Temperature.**—Under the influence of cooler solutions the intensity of replacement is diminished. The minerals formed are strongly hydrated, the texture fine-grained. In igneous rocks chlorite, epidote, quartz, and calcite, in part also sericite, may form by the action of cool weak solutions. Replacement by sulphides such as pyrite, galena, and sphalerite may take place. Limestone may be silicified to fine-grained jasperoids. To a limited extent sulphides may replace other sulphides. Chalcocite, for instance, replaces pyrite, chalcopyrite, and bornite. Replacement of effusive rocks by alunite, pyrite, and kaolinite is characteristic of some deposits formed near the surface.

In acid waters, kaolinite replaces sericite and other silicates. A certain pH concentration appears to be necessary for the formation of kaolinite.

Heated alkaline waters are not believed to be capable of developing kaolin minerals from the aluminum silicates of the rocks; alkaline silicates like sericite will result. On the other hand the ordinary dilute ground waters will develop kaolin minerals in the rocks.

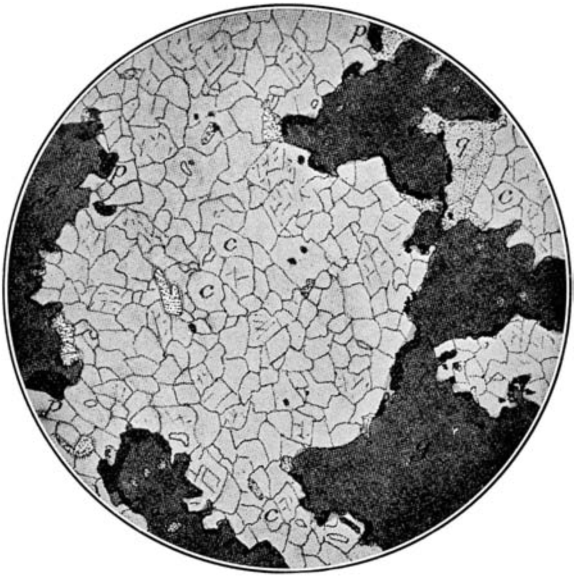

Fig. 68.—Galena, replacing crystalline dolomite. Elkhorn Mine, Montana. *g*, galena; *p*, pyrite; *c*, calcite grains of limestone; *q*, secondary quartz. Magnified 15 diameters.

*A* *B*

Fig. 69.—*A*, Fluorite replacing limestone. Florence Mine, Judith Mountains, Montana. *f*, fluorite; *l*, limestone; *q*, secondary quartz. Magnified 7 diameters. *B*. Illinois fluorite veins. Fluorite, dark, replacing calcite. Crossed nicols. (*B*, *after L. W. Currier*.)

In other words, the kaolin minerals are confined to the uppermost metamorphic zone and rarely venture far below the zone of weathering.[1]

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[1] W. Lindgren, The origin of kaolin, *Econ. Geol.*, 10, 1915, pp. 89–93.

H. C. Boydell, Operative causes in ore deposition, *Trans.*, Inst. Min. and Met., London, 37, 1928, pp. 50–177 (with discussion).

**Criteria of Replacement.**—F. Posepny first established replacement as a mode of origin of mineral deposits. Shortly afterward, S. F. Emmons[1] demonstrated it to be a common mode of origin and illustrated it by the description of many ore-bodies in Colorado and elsewhere. About 1900, W. Lindgren described the principal modes of metasomatism.[2] In 1911, J. D. Irving[3] published a paper of great value in which the criteria of replacement ore-bodies were summarized.

Some of these criteria in favor of replacement have already been mentioned but they may be briefly recalled here:

1. Form of ore-body, more or less irregular. Gradually fading limits. Not always conclusive.
2. Presence of unsupported residual rock masses. Sometimes the orientation of bedding may be proved parallel with the surrounding rocks.
3. Absence of crustification. A *banding* may be observed in places due to preservation of bedding or shearing planes.
4. Absence of concave contacts; in limestone, for instance, solution of cavities tends to produce flat concave depressions; a filled cave would show this whereas replacement proceeds with convex outlines toward the unaltered rock.
5. Preservation of textures and structures of original rock. The last-named criterion is the most conclusive.

The criteria for the determination of replacement are sometimes difficult to establish; many mistakes have been made along this line.[4] Replacement veinlets crossing the older minerals and dependence of the replacing mineral on minute fissures and cracks constitute good evidence. The projecting of crystals of one mineral into another is not always a safe proof of replacement. The apparent host may possibly be a later mineral molded about the crystals. In many cases adjoining minerals may have developed practically simultaneously. A peculiar type of replacement results in pseudo-eutectic texture simulating an intergrowth (Fig. 60).

Replacement may attack a mineral grain from fractures throughout its mass. When it begins along the periphery and works inward (Fig. 70) it may be called "centripetal." Such replacements are common in supergene enrichments. A less common form of replacement begins in the central part of the host mineral and works outward, leaving only a

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[1] S. F. Emmons, The genesis of certain ore deposits, *Trans.*, Am. Inst. Min. Eng., 15, 1887, pp. 125–147.

S. F. Emmons, Structural relations of ore deposits, *idem*, 16, 1888, pp. 804–839.

S. F. Emmons, On the origin of fissure veins, *Proc.*, Colorado Sci. Soc., 2, 1888, pp. 189–208.

[2] W. Lindgren, Metasomatic processes in fissure veins, *Trans.*, Am. Inst. Min. Eng., 30, 1901, pp. 578–692.

[3] J. D. Irving, Some features of replacement ore-bodies and the criteria by means of which they may be recognized, *Jour.* Canadian Min. Inst., 14, 1911, pp. 395–471; *Econ. Geol.*, 6, 1911, pp. 527–561.

[4] E. S. Bastin *et al.*, Criteria of age relations of minerals with especial reference to polished sections, *Econ. Geol.*, 26, 1931, pp. 561–610.

shell of the original substance; this may be called "centrifugal" replacement (Figs. 71, 72). In other cases replacement begins from cracks and cleavage planes all through a grain.

**Repeated Replacements.**—As later minerals, as a rule, may replace any of the earlier minerals, it is not always a safe conclusion that the guest has directly replaced the host, for there may have been intermediate stages. Thus if a veinlet of chalcocite intersects and replaces pyrite, there exists a possibility that the chalcocite may actually have replaced, not pyrite, but an intermediate mineral like bornite. A rock may be partially replaced by some carbonate, ankerite, for instance, and the latter may have been replaced by quartz.[1] Such intermediate replacements are very rarely complete; care must be exercised not to exaggerate their importance.

FIG. 70.—Polished section showing rims of argentite replacing sphalerite from the periphery. Crown King, Arizona. Magnified 90 diameters.

**The Reaction Series in Hydrothermal Replacements.**—Butler[2] has extended Bowen's reaction series (p. 110) by suggesting that by the action of magmatic extracts the mineralization may continue from pyromagmatic to deuteric and to hydrothermal minerals. Crystallization in a closed system will always move in a certain, normal direction; but as ore deposits, contrary to conditions in a magmatic melt, are not developed in a closed system, many difficulties are met in the interpretation of mineral succession. Lowering temperature conditions will produce minerals low in the magmatic series, but never those high in the series;

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[1] See L. C. Graton and H. E. McKinstry, Outstanding features of Hollinger geology, *Econ. Geol.*, 28, 1933, in press.

[2] B. S. Butler, Influence of the replaced rock on replacement minerals associated with ore deposits, *Econ. Geol.*, 27, 1932, pp. 1–24.

the latter have a more restricted range of temperature and are likely to be more diagnostic of thermal conditions. Butler applies this to the nickel deposits of Sudbury, and to the copper deposits of Engels and Bingham. At the latter place, the augite of the monzonite changes by deuteric processes to biotite and the ground mass becomes an aggregate of quartz and orthoclase; further hydrothermal alteration develops sericite and disseminated grains of pyrite and chalcopyrite.

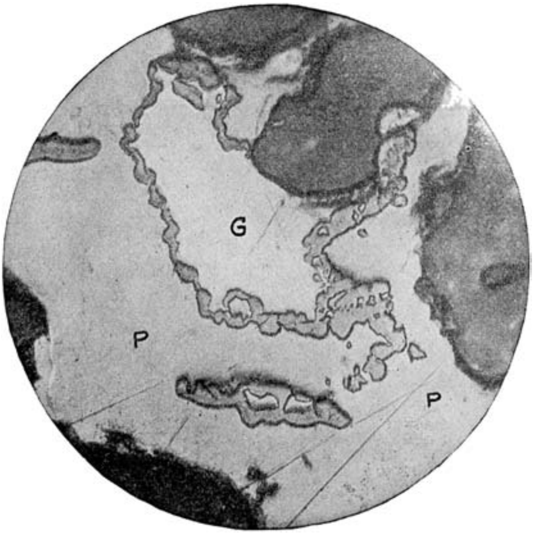

Fig. 71.—Polished section showing sphalerite rim replaced internally by galena and protecting galena from replacement by pyrargyrite. Premier Mine, British Columbia. Magnified 60 diameters. (*After W. D. Burton, "Economic Geology."*)

**Unmixing.**[1]—If a mineral has separated out at a high temperature it often contains, in solid solution, various other constituents; upon slow cooling these may suddenly crystallize at a given temperature so that the original mineral may contain small inclusions of the ejected substance. There are many examples of this in metallic alloys and the process is

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[1] A. E. Nissen and S. L. Hoyt, On the occurrence of silver in argentiferous galena ores, *Econ. Geol.*, 10, 1915, pp. 172–179.

H. Schneiderhöhn, Entmischungserscheinungen, etc., *Metall und Erz.*, 19, 1922, pp. 501–526.

L. P. Teas, The relation of sphalerite to other sulphides in ores, *Trans.*, Am. Inst. Min. Eng., 59, 1918, pp. 68–97.

G. M. Schwartz, *Econ. Geol.*, 22, 1927, pp. 44–61 (cubanite); *idem*, 23, 1928, pp. 381–397 (bornite); *idem*, 26, 1931, pp. 739–763.

A. Wandke, *Econ. Geol.*, 21, 1926, pp. 166–171 (chalcocite).

W. H. Newhouse, *Econ. Geol.*, 22, 1927, pp. 289–299 (pentlandite).

S. F. Emmons, J. D. Irving, and G. F. Loughlin, *Prof. Paper* 148, U. S. Geol. Survey, 1927 (lillianite).

P. Ramdohr, Beobachtungen an Magnetit, Ilmenit, etc., *Neues Jahrb.*, Beil. Bd. 54, 1926, pp. 320–379.

undoubtedly also of considerable importance in mineral deposits. Thus sphalerite contains iron and copper in solid solution which may separate out in dots and gashes of chalcopyrite and pyrrhotite (Fig. 73). Again bornite may contain dots and gashes of chalcocite (Fig. 74), or of chalcopyrite. Many investigators believe that these two are evidence of this process which is called *unmixing.* Sometimes these new products separate out along cleavage faces (Fig. 75). It is fair to add that no complete unanimity reigns in regard to the interpretation of many of these phenomena. Other examples are chalcopyrite in stannite, argentite grains in galena, gashes of pentlandite in pyrrhotite, intergrowths of bismuthinite and argentite, finally, laths of cubanite in pyrrhotite and chalcopyrite. If the unmixed material is heated and suddenly quenched it would be expected that they would

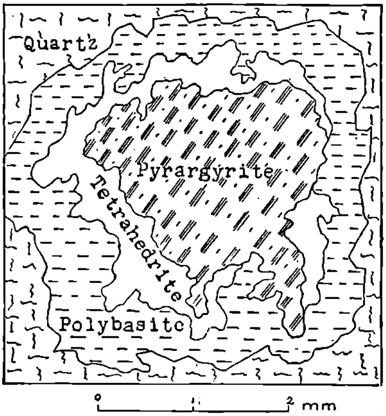

FIG. 72.—Drawing from polished section showing centripetal replacement of tetrahedrite by polybasite and centrifugal replacement of same mineral by pyrargyrite, leaving shell of tetrahedrite. Smuggler-Union Mine, Colorado. (*After M. E. Hurst, "Economic Geology."*)

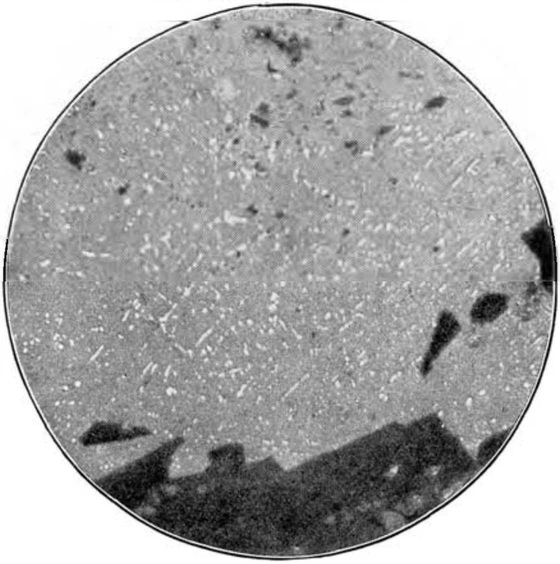

FIG. 73.—Polished section showing dots and rods of chalcopyrite in sphalerite. Black, siderite. Siegen district, Germany. Magnified 340 diameters. (*After H. Schneiderhöhn.*)

again form a homogeneous solid solution and this appears to have been proved in the case of galena, cubanite, and pentlandite. Quite naturally

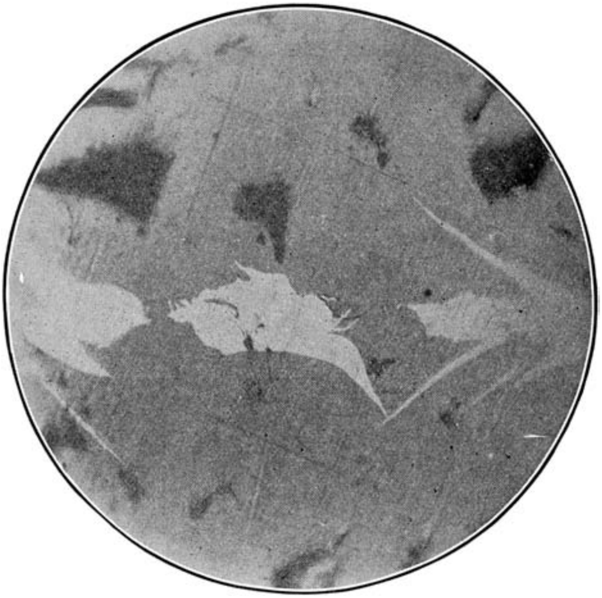

FIG. 74.—Polished section showing gashes of chalcocite in bornite, Bristol, Connecticut. Magnified 480 diameters. (*After W. L. Whitehead.*)

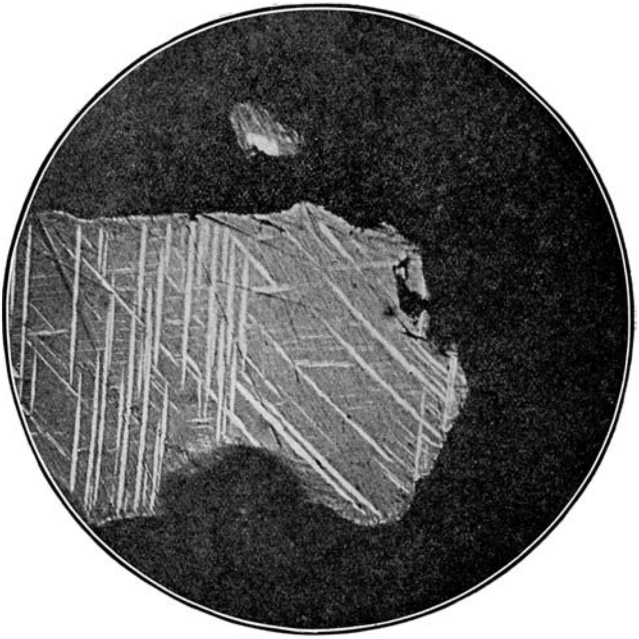

FIG. 75.—Lines of chalcopyrite following cleavage planes in bornite. Replacement or unmixing, Engels Mine, California. Magnified 246 diameters. (*After W. L. Whitehead.*)

these products of unmixing are more common in high temperature deposits.

**Rôle of Colloids in Filling and Replacement.**[1]—It is well known that colloidal deposits, for instance, of silica, iron hydroxide, and aluminum hydroxide play an important part in mineral deposits formed at or near the surface (Fig. 76).

Colloidal silica or silica gel is also of considerable importance in the origin of deposits formed relatively near the surface by ascending waters. Some of the quartz filling in such veins is extremely fine-grained and bears evidence of having been deposited as a stiff jelly which soon afterward was crystallized in chalcedonic or cryptocrystalline form.[2] Clear evidence of this is seen in some filled veins from the Tintic district, Utah,[3] where the original delicate banding by deposition is still seen though the substance is now microcrystalline quartz.

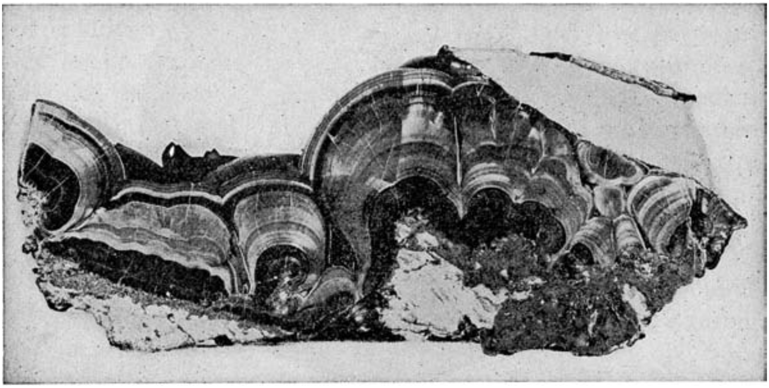

FIG. 76.—Psilomelane with colloform texture on quartzite, Virginia. Three-fourths of natural size. (*Specimen from F. L. Hess.*)

In some deposits formed at moderate temperatures, limestone and dolomite may be replaced by silica gel which afterward crystallized to chalcedony. This type of replacement is characterized by sharp contacts with the unaltered rock; it does not proceed from crystal nuclei of quartz

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[1] A. Scott, *Fourth Rept.* on colloid chemistry, Brit. Assoc. Adv. Sci., London, 1922.

W. Lindgren, The colloidal chemistry of minerals and ore deposits. In Theory and application of colloidal behavior, R. H. Bogue, 1924.

H. C. Boydell, The rôle of colloidal solutions in the formation of mineral deposits, *Trans.*, Inst. Min. and Met. (London), 34th session, 1924, pp. 145–337, including discussion.

W. Lindgren, Metasomatism, *Bull.* Geol. Soc. Am., 36, 1925, pp. 247–262.

[2] W. Lindgren, Geology and mineral deposits of the National District, Nevada, *Bull.* 601, U. S. Geol. Survey, 1915. Similar evidence is found in the quartz veins of Waihi, New Zealand, and in some tin veins of Bolivia.

[3] W. Lindgren, Processes of mineralization and enrichment in the Tintic mining district, *Econ. Geol.*, 10, 1915, pp. 225–240.

starting at numerous points but advances like a wave and stops with sharp contacts (see p. 178). Later metalliferous solutions penetrated this gel and deposited sulphides in it. Sometimes a banding has been produced which strongly recalls the so-called Liesegang rings[1] in artificial gels and indicates a rhythmical precipitation of sulphides (Fig. 77).

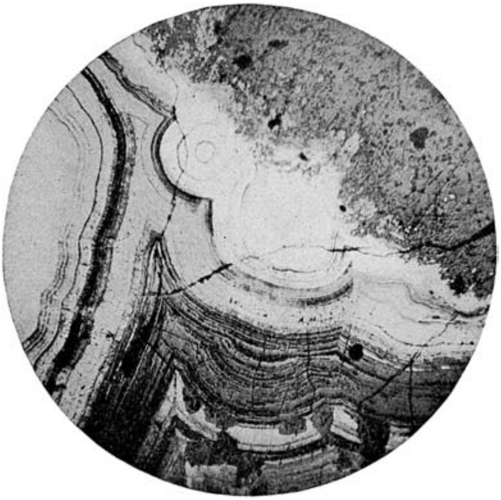

FIG. 77.—Iron sulphide (white) and sphalerite (gray in various shades) with colloform texture. Polish Silesia. Magnified 25 diameters.

Many ore minerals were originally deposited as colloidal precipitates and have later acquired crystallinity. Cassiterite, pyrite, marcasite, sphalerite, chalcopyrite, bornite, chalcocite, and others sometimes show this, particularly in deposits originating at moderate or intermediate temperatures.

[1] R. E. Liesegang, Geologische Diffusionen, 1913, p. 180. Reviewed by A. Knopf in *Econ. Geol.*, 8, 1913, p. 803.

S. G. Lasky, Colloidal origin of some of the Kennecott ore minerals, *Econ. Geol.*, 25, 1930, pp. 737–757.

See also footnotes on p. 189.

# CHAPTER XV

## ORE-SHOOTS[1]

**Form of Primary Ore-shoots.**—Commercial ore or mineral does not ordinarily occupy the whole volume of a deposit. The ore is in most cases surrounded by material of poorer grades, sometimes fading into the country rock, or again sharply separated from it. In replacement deposits the disseminated grains of galena, for instance, or sphalerite, may gradually become so few that the mass can no longer be treated with profit. In veins, only certain parts of the sheet-like body can be extracted, while the remainder of the vein material may consist of gangue minerals only, or of clayey attrition masses or breccias.

Those parts of a deposit in which the valuable minerals are so concentrated that their utilization becomes possible are called ore-shoots. Their occurrence and form are exceedingly variable, and it is often most difficult to ascertain the causes which have guided their development.

In deposits of sedimentary origin the ore-shoots have, of course, the general tabular form; but admixture with gangue materials or valueless matter may so dilute the ore that only certain parts of the body can be extracted. Various assortments of detritus and complex conditions of precipitation from waters of seas, lakes, and rivers have influenced the concentration of the richer ore masses. In addition, alterations by meteoric waters are common; in the case of phosphate deposits and beds of siderite they have resulted in enrichment.

In deposits of igneous origin the general form of the deposit is also that of the ore-shoots. In some deposits, such as the magnetite deposits

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[1] T. A. Rickard, The formation of bonanzas in gold veins, *Trans.*, Am. Inst. Min. Eng., 31, 1902, pp. 198–220.

The localization of values in ore-bodies, etc. Discussion by J. D. Irving, F. C. Smith, Reno Sales, F. L. Ransome, H. V. Winchell, H. Sjögren, and W. Lindgren, *Econ. Geol.*, 3, 1908 pp. 143–154; 224–229; 326–330; 331–336; 425–427; 637–642. *Idem*, 4, 1909, pp. 56–61.

C. W. Purington, Ore horizons in the San Juan Mountains, *Econ. Geol.*, 1, 1905, pp. 129–133.

H. C. Hoover, The valuation of gold mines, *Eng. and Min. Jour.*, May 19, 1904.

R. A. F. Penrose, Jr., Some causes of ore-shoots, *Econ. Geol.*, 5, 1910, pp. 97–133.

C. D. Hulin, Structural control of ore deposition, *Econ. Geol.*, 24, 1929, pp. 15–49.

Adolph Knopf, The Mother Lode system of California, *Prof. Paper* 157, U. S. Geol. Survey, 1929.

W. H. Newhouse, Some relations of ore deposits to folded rocks, *Trans.*, Am. Inst. Min. Met. Eng., General Volume, 1931, pp. 224–246.

of northern Sweden and the dike-like deposits of ilmenite at Iron Mountain, Wyoming, there is practically no waste material and the whole igneous body constitutes ore. The tabular form of certain chromite deposits is remarkable.

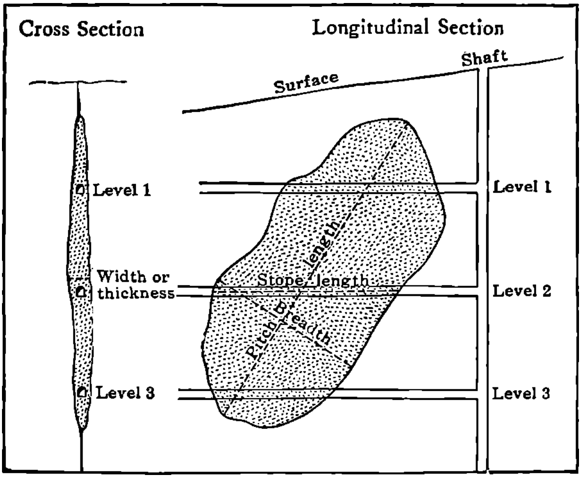

FIG. 78.—Diagram illustrating the terms used to describe the dimensions of ore-shoots. (*After W. Lindgren and F. L. Ransome, U. S. Geol. Survey.*)

More commonly the irregularly lenticular or tabular masses of igneous rocks in which ore minerals have developed by magmatic segregation (for instance, gabbro containing chalcopyrite) have nuclei of richer material gradually fading into more normal rock.

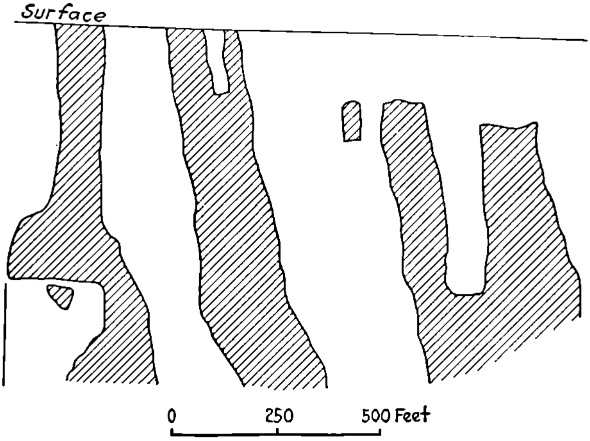

FIG. 79.—Longitudinal section showing ore-shoots of veins, at Nevada City, California.

In the epigenetic deposits the outlines of the ore-shoots are exceedingly variable. In those deposits which are formed by replacement this is particularly true, and few rules can be laid down for their occurrence. The form is determined by the fissures giving access to the solutions, by the presence of impermeable rocks, and by the varying susceptibility to replacement of the original rocks.

Most attention has been given to the shoots in fissure veins. Although the ore in the main follows the fissure and therefore has a tabular or sheet-like form, it rarely occupies the whole space along this fissure, but is concentrated in bodies of varying size, shape, and continuity. Smaller bodies are known as bunches, pockets, or kidneys; in gold-quartz veins these may be exceedingly rich. Narrow ore-shoots, greatly elongated in the vertical direction, whether occurring in fissure veins or independently of them (for instance, in volcanic necks), are called *chimneys, pipes,* or *necks* (Fig. 82).

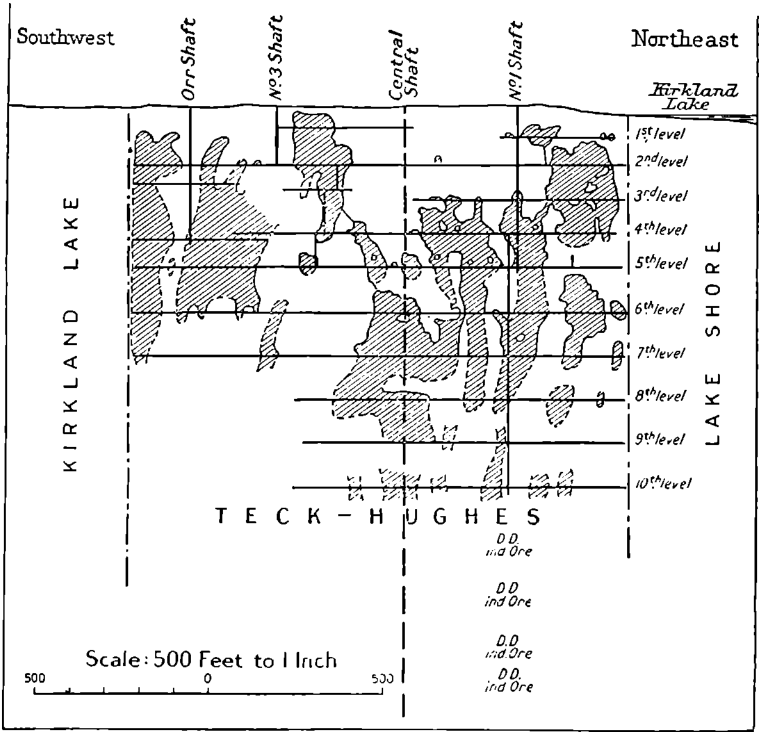

FIG. 80.—Longitudinal stope section, Teck-Hughes Mine. (*From D. L. H. Forbes.*)

Ore-shoots may be entirely irregular, but commonly have a more or less well-defined columnar, steeply pitching shape, best shown in projection upon the plane of the vein. Figure 78 shows the terminology proposed[1] for various dimensions of an ore-shoot in a vein. The pitch length, or axial length, is the distance between the two extreme ends of the shoot.

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[1] W. Lindgren and F. L. Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906, p. 206.

The pitch is the angle which the pitch length makes with the strike of the vein, and is measured on the plane of the vein. The stope length is the horizontal length of the ore-shoot on any particular level. The thickness or width is measured perpendicularly to the plane of the vein. The breadth of the ore-shoot is the stope length, multiplied by the sine of the pitch.

Figure 80 shows the ore-shoots of a gold-quartz vein at Kirkland Lake, Ontario. Flat-dipping shoots are not so common. Figure 81 shows an excellent example of a flat shoot in the Eureka-Idaho vein at Grass Valley, California. Figures 196 and 201 show the tendency of many shoots to cease in depth.

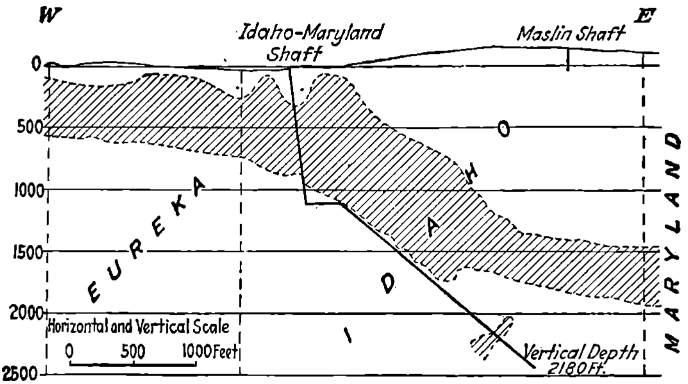

FIG. 81.—Approximate outline of the Eureka-Idaho ore-shoot, Grass Valley, California, in projection on the plane of the vein.

In parallel veins the shoots are often, roughly speaking, coextensive. Sometimes the shoots in a series of parallel veins persistently recur where the veins cross a certain stratum or dike, as, for instance, where the gold-quartz veins of Gympie, Queensland, intersect certain carbonaceous strata, or as at Thames, New Zealand, where the veins intersect certain softened and altered andesites. Many shoots follow intersections of veins or of veins with fissures.

Shoots, however large, do not continue indefinitely, but end in depth, usually with gradual deterioration. Small masses or kidneys are likely to be found below the termination of a large ore-shoot. Exploration may find another shoot below the first, either on the same fissure or imbricating on a parallel vein. When great depth is attained the grade of the ore usually decreases in the deeper levels, but this rule is not without exceptions. Many shoots are lenticular, that is, they contain a rich nucleus, outward from which the ore gradually decreases in tenor. H. C. Hoover, from an examination of 70 mines, concluded that ore-shoots are generally lenticular and that the probable minimum extension of an ore-shoot below any given level would be a factor of not less than a radius of one-half of its breadth.

At Cripple Creek, Lindgren and Ransome found that the shoots which begin distinctly below the surface have a marked elongated form, the ratio between pitch length and breadth varying from 1½:1 to 5:1.

Primary ore-shoots rarely continue for more than 2,000 feet along the strike, or for more than 2,000 feet along the pitch length. There are exceptions, such as the Morro Velho shoot (Brazil), which has a pitch length of 10,000 feet.

In a given district the pitch of the ore-shoot is often predominantly in one direction; thus at Nevada City and Grass Valley the shoots pitch to the right of an observer who looks down the dip of the vein. In another district the opposite may be true. In some places the tenor varies directly, in others, inversely with the swelling of the vein. According to a rule often quoted, the shoots follow the directions of the striations on the vein walls; but this rule is not infallible.

**Shoots of Successive Mineralizations.** While in some veins the whole width consists of uniform ore, it is exceedingly common, especially in thick veins, to find that there are certain streaks which are far richer than the rest. They may follow footwall or hanging wall, or the center of the vein, or may switch from one side to another. Such phenomena indicate reopening of the vein or brecciation with later enrichment.

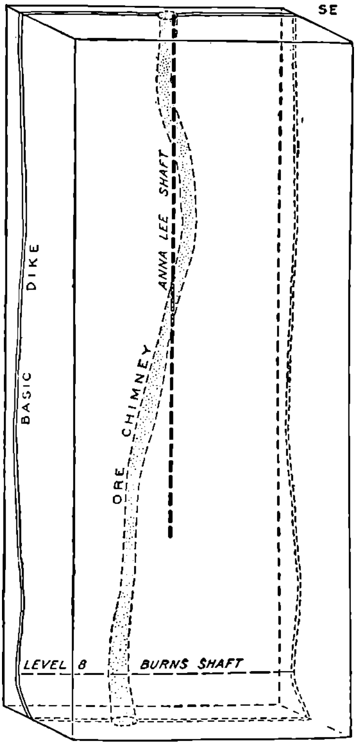

FIG. 82.—Stereogram of Anna Lee ore chimney, Cripple Creek, Colorado. Shoot probably determined by intersection of the basic dike with a fissure. (*After V. G. Hills.*)

**Superficial or Secondary Shoots.**—Descending surface waters decompose and often enrich the upper part of veins or other deposits. Such enriched superficial portions of an ore deposit are dependent upon the ground-water level and, when projected upon the plane of the vein, follow the surface of the ground and terminate below along an irregular and jagged line. Oxidized ores, as well as sulphides due to enrichment, are found in them, usually at different levels. The surface shoots are, in fact, characterized by horizontal extension, in contradistinction to the predominance of the vertical direction in the primary shoots. The mineralogical characteristics of superficial shoots will be discussed in detail in a later chapter. Their tendency is to spread along the strike of the vein, often also out into the wall rock. Thus pay ore may be found for a long distance along the trend of the vein and its appearance will be that of the oxidized outcrops of a long primary shoot, when in fact deeper explorations may prove the existence of only a few narrow primary ore-bodies underneath the continuous surface ore. Frequently, oxidized silver ores will be found in croppings along a vein which are simply concentrations of a primary vein filling that contains no workable shoots. To this class belong also the horizontal or flat shoots of secondary copper sulphides (chalcocite and covellite) formed by descending solutions in copper deposits at or near the water level. The primary material may or may not constitute commercial ore. If spread over wide mineralized areas such shoots are often called *chalcocite blankets.*

Descending metal solutions may wander out in the country rock and here form new lateral bodies of oxidized ore.

**Causes of Primary Ore-shoots.**—Ore-shoots are due to the abundant precipitation of valuable minerals from their solutions. The causes are in part physicochemical and in part mechanical:

1. Decrease of pressure and temperature.
2. Chemical character of wall rock.
3. Mingling of solutions.
4. Physical character of wall rocks.
5. Structural conditions.

**Decrease of Pressure and Temperature.**—The fundamental reason for the deposition of ores in veins and allied epigenetic deposits in the upper crust is probably that the metals were in solution in hot waters which during their ascent gradually encountered conditions favorable for precipitation. First among these conditions is decreasing temperature. If this is true the deposits should gradually become poorer or barren in depth.[1] In a general way this is doubtless true, but for many substances the vertical space through which deposition can take place is very large. We know that gold-bearing quartz was deposited in California over a vertical distance of 5,000 feet, and in Alaska and at Bendigo, Australia, the interval is about the same. This deposition took place at considerable depth below the surface, probably several thousand feet

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[1] T. A. Rickard, Persistence of ore in depth, *Trans.*, Inst. Min. and Met., London, 24, 1915, pp. 3–46, with discussion.

W. Lindgren, Ore deposition and deep mining, *Econ. Geol.*, 1, 1905, pp. 34–46.

F. L. Garrison, Decrease of value in ore-shoots with depth, *Trans.*, Canadian Min. Inst., 15, 1912, pp. 192–209.

J. F. Kemp, The influence of depth on the character of metalliferous deposits, Compte Rendu, XII$^{ème}$ session, Congrès géologique internat., Canada, 1914, pp. 253–260.

Malcolm Maclaren, *idem*, pp. 295–304.

below it; and as it is known that gold-bearing quartz may also be deposited within the upper zone, we have thus a total vertical range of at least 9,000 feet. In the lowest levels at the places mentioned the ore is of low grade, but in Alaska at least there is a large quantity available. The richest ore was doubtless deposited close to the surface, where we find the bonanzas of the Tertiary gold and silver veins. The most persistent gold-bearing ore-shoots known are those in veins formed at intermediate or high temperatures. Such are, for instance, the North Star vein at Grass Valley, California, which with very slight impoverishment has been followed for 9,000 feet on a dip of 20° (p. 548). The Kolar veins in India have been mined to a vertical depth of 7,000 feet in shoots of considerable regularity (Fig. 83), with little change in tenor of ore. The most persistent ore-body known is that of Morro Velho mine in Brazil, where a pitching ore-shoot has been worked to a vertical depth of 7,000 feet and a pitch length of 10,000 feet (Fig. 277). For copper ores the vertical range of deposition is likewise great, though unlike gold and silver they seem to be deposited in greatest quantity at lower levels and high temperatures. Lead, on the other hand, appears

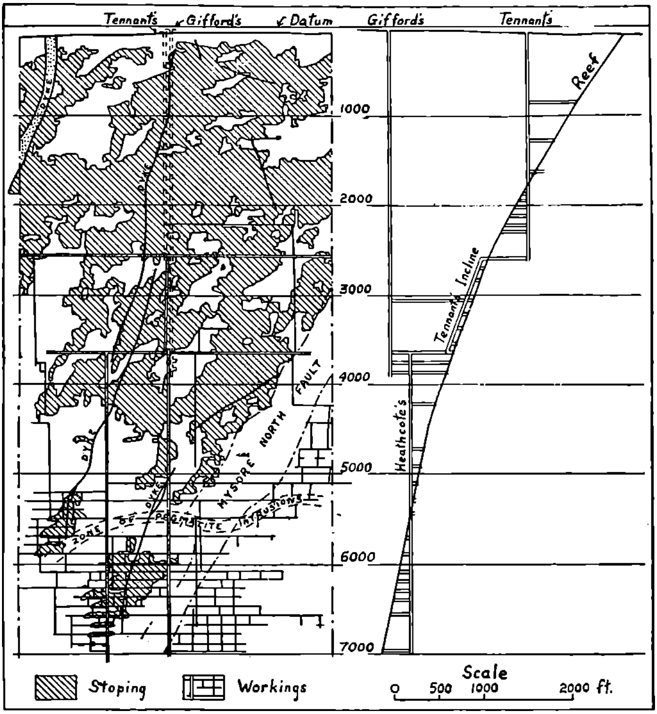

FIG. 83.—Longitudinal section and vertical section of Champion Reef Mine, Mysore, India. (*After H. M. White, "Mining Magazine" (London).*)

to be precipitated nearer the surface and at lower temperatures; while zinc in this respect stands between copper and lead.

The relations set forth explain why so little decisive evidence of vertical succession in deposition is available from observations at any one mine.

In the Cornwall veins tin and tungsten prevail in the lower levels in granitic country rock, while copper was deposited in the cooler region of the slates covering the granite batholiths; the lead ores are found some distance away from the intrusive granite. In many lead mines it has been noted that within a distance of 700 to 3,000 feet from the surface the lead minerals give way to pyrite and sphalerite. In quicksilver mines the ore often becomes impoverished within 1,000 feet below the surface.

The dependence of the deposition of various metals upon temperature and therefore also upon the vertical and horizontal distance from the place of origin of the mineralizing solutions has been emphasized by several investigators.[1]

**Character of Wall Rock.**—The character of the wall rock has sometimes a decided influence on the ore-shoots, particularly in replacement deposits, where large masses of limestone, dolomite, lime-shale, and other rocks have been replaced by ore. At Freiberg, Saxony, the gray gneiss is the favorable rock, while the veins split or become unproductive in the red gneiss or in the mica schists.

Carbonaceous rocks are believed to influence deposition favorably by their reducing action; the gold-quartz shoots of Gympie, Queensland, are often quoted, as well as the supposedly carbonaceous "indicator" at Ballarat, Victoria. The well-known replacement of fossil wood by chalcocite in a certain class of copper deposits may be added to these examples, as well as the supposed influence of certain oil shales on the deposition of lead ores in Wisconsin. The importance of precipitation by carbonaceous material has been overestimated, but in many cases the hydrocarbons have certainly favorably influenced the deposition of ores.[2]

Rocks containing pyrite or other sulphides often enrich traversing veins. Examples of this are known from Kongsberg, Norway, where the silver veins are productive when crossing certain schists with disseminated sulphides. At Ophir, California, gold-quartz veins are enriched when crossing "iron belts" of pyritic amphibolites.

Where a vein cuts through a thick series of sedimentary rocks it often widens and contains rich ore in the limestones, while poor or barren in

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[1] J. E. Spurr, A theory of ore deposition, *Econ. Geol.*, 2, 1907, p. 790.

L. de Launay, La métallogénie de l'Italie, Congrès géologique internat., Mexique, 1, 1906, p. 571. Also in Gîtes Minéraux, 1, Paris, 1913.

W. Lindgren, Processes of mineralization and enrichment in the Tintic mining district, *Econ. Geol.* 10, 1915, p. 228.

[2] W. P. Jenney, The chemistry of ore deposition, *Trans.*, Am. Inst. Min. Eng., 33, 1903, pp. 445–498.

shale or sandstone. Similarly, where a thick series of igneous rocks, as in the San Juan region, Colorado, is intersected by veins ore horizons will develop in rocks which by their physical and chemical character are most favorable to continuous fissures or to replacement (Figs. 196 and 201). Rhyolites are unfavorable because fissures tend to split in such rocks; tuffs likewise, because the solutions tend to disperse through great masses of rock. Local brecciation may create favorable loci.

On the other hand, rocks like andesites and latites are usually favorable. Purington[1] has shown that in the San Juan Mountains the andesitic breccias which contain abundant ferromagnesian silicates are most favorable to ore deposition.

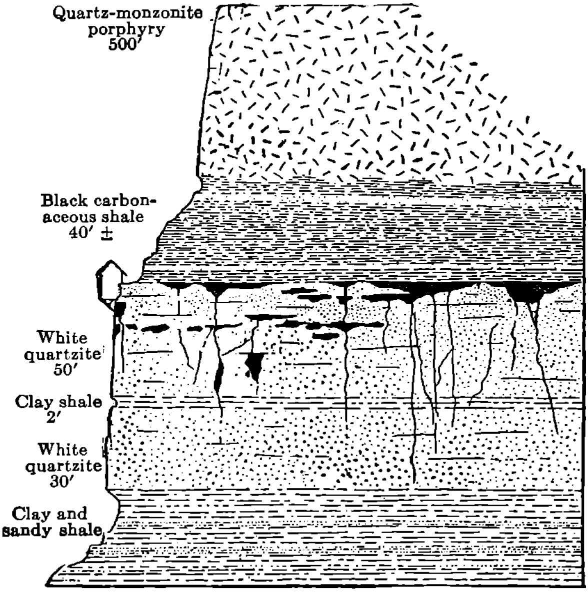

FIG. 84.—Cross section of the American Nettie Mine showing shoots of ore (black) underneath impervious barrier of shale. (*After J. D. Irving.*)

**General Structural Conditions.**—For the development of an ore-shoot the solutions must gain access to the locus of deposition by means of fissures. Newhouse has shown that in folded rocks the anticlines are favorable places because more fractured and permeable.

Open spaces available for ore are created along fissures by change of strike and dip and by junctions and intersections. Shoots may be suddenly cut off by gouges on pre-mineral or inter-mineral fractures.

In many fissures the open spaces are sealed by an early deposition of barren quartz gangue. Sulphide minerals and gangue are deposited

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[1] C. W. Purington, Ore horizons in the San Juan Mountains, *Econ. Geol.*, 1, 1905, pp. 129–133.

later, so that in order to obtain space inter-mineral fractures must be opened in the barren material. Hulin has emphasized the fact that a productive vein is usually the result of a long-continued deposition of gangue with many inter-mineral fractures and brecciations in which the ore minerals are successively formed.

**Impermeable Barriers.**—More or less horizontal ore-bodies are often encountered where impervious rocks interpose barriers to the solutions.

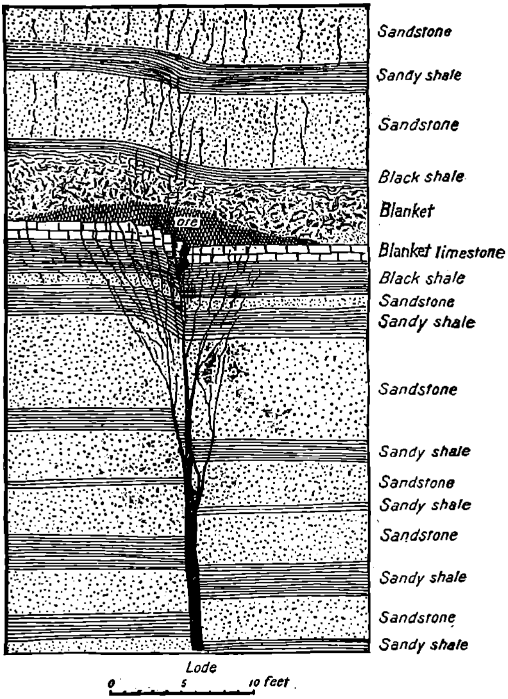

FIG. 85.—Diagrammatic section across a lode, and ore-body formed beneath an impervious stratum (blanket) of black shale, Rico, Colorado. (*After F. L. Ransome, U. S. Geol. Survey.*)

The occurrence of ore in horizontal extension *below* such barriers is, in fact, one of the best indications that the solutions have been ascending.

The blanket veins of Rico, Colorado (Fig. 85), present another good illustration of this principle, as do also the ores of the American Nettie mine near Ouray, Colorado (Fig. 84), and the siliceous gold ores replacing dolomite in the Black Hills, South Dakota (Fig. 228). The impermeable stratum is not necessarily shale; it may be a fault gouge, which is impervious to solutions, or a sheet of volcanic rock which, for some reason, the fissures failed to penetrate. The same principle of impermeable barriers serves to explain why the vein material is often confined between the clay seams of hanging and footwall without entering the adjacent country rock by replacement.

Where one fissure is faulted by another, deposition may occur because the circulation becomes impeded at the fault and partial stagnation follows, producing more complete precipitation.

Where the solutions have moved downward, as in the concentration of hematite ore from poorer "iron formations," it is often observed that ores occur on *impervious basements* and in *troughs* caused by shales, clayey fissures, or dikes.

**Intersections.**—Enrichment and ore-shoots along intersections of two veins or of a vein and a fissure are very common phenomena, well exemplified at Freiberg, Saxony (Fig. 86), and at Cripple Creek, Colorado. Van Hise attributes the shoots at such intersections to the mingling of two solutions and consequent precipitation of some constituents. In part they may be due to the shattering of the rocks at the intersection; and Penrose notes that shoots are more likely to occur where the intersection takes place at acute angles, forming wedge-shaped blocks that are easily broken along their edges.

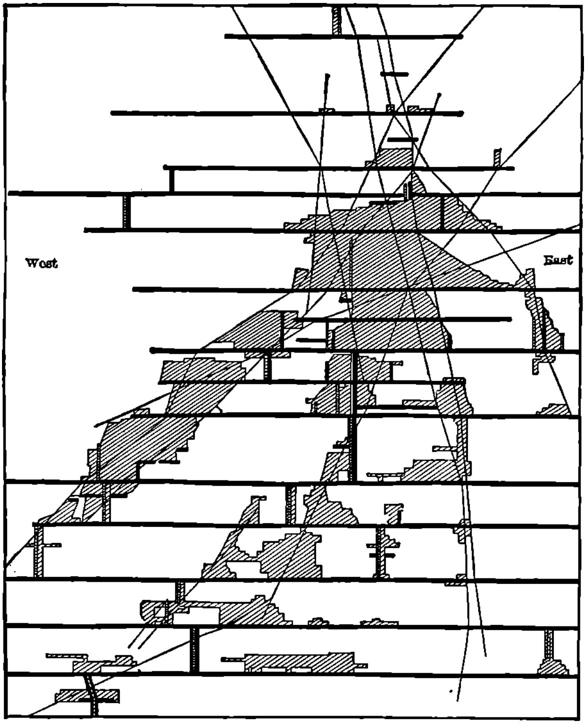

Fig. 86.—Longitudinal section along the Neu Hoffnung vein, Freiberg, Germany, showing ore-shoots along intersection with several other veins. (*After R. Beck.*)

Though enrichment at intersections is common it is by no means a universal rule, and indeed sometimes a vein is impoverished at the intersection with a barren fissure. The influence of intersections is often attributed to the mingling of solutions of different types, producing precipitation of some constituent.

The occurrence of many large shoots such as those in the gold-quartz veins of California and Ontario, at Cripple Creek, Colorado, and in the Coeur d'Alene, Idaho, lead mines cannot be fully explained by intersections or by the influence of the wall rock.

Such shoots are generally considered as the result of decrease in temperature of ascending solutions in channels of circulation.

The importance of structure for the localization of ore shoots cannot be too strongly emphasized. The general geology, the mode of deformation, differences in chemical character, hardness or permeability of the rocks, depth of ore formation, all are most important factors which must be carefully studied to allow conclusions as to the tracing or discovery of ore shoots, a task which will confront all those engaged in the study of mineral deposits. The many and complex features are well exemplified in the ore shoots of the Mother Lode of California.

# CHAPTER XVI

## CLASSIFICATION OF MINERAL DEPOSITS

**Classification by Form and Substance.**—A genetic classification of deposits of useful minerals is really equivalent to the classification of "geological bodies" as defined in Chapter I; and is, therefore, naturally beset with all the difficulties connected with an imperfect knowledge of geological processes. The early attempts in the way of systematic treatment, however, avoided this troublesome path by the simple expedient of classifying by substance or uses, or by form. These schemes are followed in many textbooks, even among those of recent date; undoubtedly they have some advantages, especially for the miner, the industrial chemist, and the metallurgist, who are principally interested in the form of the deposit or in the study of ores of certain metals.

By substance and uses mineral deposits may be classified as follows:

| | |
|---|---|
| 1. Structural materials | Stone, glass sand, cement rock, clay, asphaltum. |
| 2. Fuels | Coal, petroleum, natural gas, peat. |
| 3. Abrasives | Corundum, garnet. |
| 4. Fertilizers | Potash salts, phosphates, green-sands. |
| 5. Precious stones | Diamond, opal, tourmaline. |
| 6. Various industrial uses | Graphite, barite, borax, asbestos, sulphur. |
| 7. Metallic ores | Iron ores, copper ores, gold and silver ores, tin ores, aluminum ores, etc. |

However convenient, it is evident that this classification cannot lead to a thorough appreciation of the manifold processes by which mineral deposits are formed in nature.

The early and not yet entirely abandoned schemes refer to the form of the geological bodies. But form is closely connected with genesis and even in one of the earliest classifications on this basis, that of Bernhard von Cotta,[1] the difficulty of avoiding genetic conceptions is felt in his definition of a vein as a "filled fissure." He divided ore deposits as follows:

I. Regular deposits.
  *A*. Beds.
  *B*. Veins.
    *a*. Ordinary fissure veins (true fissure veins).
    *b*. Bedded veins.

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[1] Die Lehre von den Lagerstätten, Freiberg, 1859.

        *c.* Contact veins.
        *d.* Lenticular veins.
- II. Irregular deposits.
    - *C.* Stocks (irregular masses with distinct limits).
        - *a.* Recumbent.
        - *b.* Vertical.
    - *D.* Impregnations (irregular masses, fading into country rock).

With variations this plan of classification is followed in many of the older textbooks. Not unlike it is a classification by J. A. Phillips in his treatise on ore deposits, revised in 1896 by H. Louis.

L. de Launay[1] arranges the deposits according to the principal elements contained. This logical, though not genetic, plan presents difficulties, and it seemed wise to avoid it.

Form and structure, though highly important, are largely accidental and offer no safe basis for a scientific classification.

**Genetic Classifications.**—A genetic classification is the most desirable both theoretically and practically. In exploring and exploiting ore deposits, the miner is almost forced to form an idea of its origin in order to follow up the ore-bodies to best advantage. Von Groddeck and Stelzner were really the first mining geologists who appreciated and applied the genetic principle in classification. Of course, the time was hardly ripe for its introduction until the conceptions of genesis had crystallized into fairly definite form. Stelzner remarks, with good reason, that it is only by standing upon the ground of a genetic theory that the miner finds courage to sink deep shafts or drive long tunnels.

We are still in doubt as to the true mode of origin for many deposits. But, as von Groddeck and Stelzner have pointed out, this applies to any classification; and this very uncertainty is a stimulus to further investigations.

The different classifications proposed will not be given here in detail. An excellent account is found in Kemp's "Ore Deposits of the United States and Canada," Appendix I. Von Groddeck, Stelzner, Posepny, Wadsworth, Monroe, Kemp, Crosby, Hoefer, Spurr, Van Hise, Weed, and several others have more or less successfully attacked the problem of a consistent genetic classification.

Von Groddeck, followed by Stelzner and Beck, makes the primary distinction, whether the useful minerals were originally formed in or with the rock in which they now occur or whether they were introduced into pre-existing rocks. Stelzner called the former syngenetic, the latter epigenetic.

J. F. Kemp divides the deposits into (I) those of igneous origin, (II) those precipitated from solutions, and (III) those deposited from suspen-

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[1] L. de Launay, Gîtes Minéraux et Métallifères, 3 vols., Paris, 1913.

sion, or residues after the decomposition of rocks. Difficulties appear here too, for what are igneous magmas but solutions?

Beck's classification is in part based on that of Stelzner. In the first edition of his handbook, "Die Lehre von den Erzlagerstätten," the syngenetic or epigenetic origin was made the principal basis of classification. In the edition of 1909 this is changed and the deposits are classified as follows, on the basis of the various phases of their genetic history:

1. Magmatic segregations.
2. Contact-metamorphic ore deposits.
3. Fissure veins.
4. Bedded deposits. } Morphologic facies of a single genetic group.
5. Stocks.
6. Secondary alterations.
7. Sedimentary ore deposits.
8. Detrital deposits.

While this is a decided improvement upon the first classification adopted by Beck, the description of the various deposits shows that many genetically different types are forced into one and the same subdivision.

Weed[1] goes further and states the origin of the ore-forming solutions. His first class includes igneous deposits, segregated in a magma; his second, igneous emanations, including contact deposits, and tin veins; his third, gas-aqueous or pneumato-hydato-genetic deposits formed by magmatic waters mingled with ground waters. His fourth and smallest division includes those mineral masses formed by surface waters.

Beyschlag, Krusch, and Vogt (1914) follow a scheme in which there is little of a truly genetic classification. They divide deposits into (1) magmatic; (2) contact (contact-metamorphic); (3) veins, cavity fillings and metasomatic deposits; and (4) ore beds (mostly sedimentary). Under each heading are numerous "groups," not necessarily related.

The classification of Beck and Berg (1922) does not seem to differ very much from the previous classification and still groups many incongruous items under one heading, for instance under ore-"stocks" and ore "beds."

P. Niggli (1925) and Schneiderhöhn (1925) divide the deposits of magmatic affiliations as follows: (1) liquid magmatic; (2) plutonic, pegmatitic to pneumatolytic to pyrometasomatic; (3) plutonic, hydrothermal, (4) volcanic (extrusive) deposits. This attempt to separate the "plutonic" from the "extrusive" deposits will hardly stand, but the plan is an improvement on the classifications just mentioned. Schneiderhöhn adds under (4) the new type of the submarine exhalation deposits containing magnetite, pyrite, and siderite. He also extends his classification to include the deposits of sedimentary and metamorphic derivation.

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[1] W. H. Weed in "Ore deposits," a discussion republished from the *Eng. and Min. Jour.*, New York, 1903, pp. 20–23.

Mineral deposits must have been formed by geological processes. Recognizing this, Van Hise[1] classifies ores as follows: Those produced (1) by processes of sedimentation; (2) by igneous processes; (3) by metamorphic processes. But he includes under this heading practically all veins and allied geological bodies, conceiving them to be deposited by the circulating ground water.

It is probably impossible to produce a classification which will win the approval of all. In the ultimate analysis by far the larger number of mineral deposits have been formed by physicochemical reactions in solutions, whether these were aqueous, igneous, or gaseous. According to this view the only consistent division that can be made is that between deposits formed by mechanical concentration of pre-existing minerals and those formed by reactions in solutions.

A genetic classification should not be confined to a general indication of the relative time of ore deposition—whether at the same time or later than the country rock. Nor should it confine itself to a statement of the agents of ore deposition—whether aqueous, igneous, or gaseous solutions, or whether sedimentary, igneous, or metamorphic processes. The statement of the place of ore deposition—at the surface or below it; in shallow waters or in deep seas—is important but not sufficient.

Some authors have attempted a classification by mode of deposition—whether by replacement or by filling of open cavities—but all such attempts have been failures, for the two processes are so closely associated that separation is impossible.

The genetic classification should ultimately determine the limits of ore deposition in each class by temperature and pressure. Each deposit should be considered as a problem in physical chemistry; and the solution of this problem, with the necessary geological data, will suffice to fix the mode of formation of the deposit.

We are far from having the complete material for such a classification, but we have at least a few starting points. It is necessary to determine, by experiment or by observation in nature, the limits of existence of each mineral species. Some will be found to be "persistent" under widely differing conditions of temperature and pressure—like fluorite, quartz, or gold. For others a far more limited range will be established. By collecting the data of mineral association, sequence of deposition, and stability range of the component parts of the deposit, it will be possible to ascertain the conditions prevailing at the time of ore deposition.[2]

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[1] C. R. Van Hise, A treatise on metamorphism, *Mon.* 47, U. S. Geol. Survey, 1904.

[2] T. Crook, The genetic classification of rocks and ore deposits, *Mineralog. Mag.*, London, 17, 1914, pp. 55–85.

G. F. Loughlin and C. H. Behre, Jr., Classification of ore deposits, Lindgren Volume, Chap. 4, Am. Inst. Min. Eng., 1933, in press.

Perhaps it is well not to expect too much from physical chemistry, magnificent as its services have been. The complications, even in simple systems, become great when, besides temperature and pressure, concentration, mass action, and time must be considered. In multicomponent systems the difficulty increases enormously. At the same time it is believed that the direction indicated is the only safe one to take in classifying the complex phenomena of ore deposition.

**Geological Thermometry.**[1]—In the classification here proposed the temperature of origin of a deposit is thus a most important item. Some minerals develop over a very wide range of temperature, but there are others which are much more limited or which undergo identifiable changes at certain temperatures. By such means the temperature of origin can sometimes be determined, at least within given limits. Among such data are the melting points which, however, only indicate maxima. A crystal of bismuth must, for instance, have been formed below 271°C. (the melting point of the metal). More useful are the inversion points. Below 575°C., for instance, quartz changes to a form with a recognizably different symmetry. Another inversion point is that between orthorhombic and isometric chalcocite at 91°C. Conclusions as to temperature may also be drawn from the change of color of minerals at certain temperatures, or from the concentration of liquid inclusions determined from relations of fluid and separated crystals, or from the disappearance of "unmixed" material in certain minerals. Ramdohr gives a list of 49 minerals which offer significant temperature data.

## OUTLINE OF PROPOSED CLASSIFICATION

**Detrital and Sedimentary Deposits.**—In the scheme followed in this book there are two major divisions. The first includes deposits formed by mechanical processes of concentration. This includes the detrital deposits such as placers and quartz sand formed at moderate temperature and pressure.

The second division contains the great majority of mineral deposits which have been produced by chemical processes of concentration. Many important processes, such as those productive of iron ores and phosphates, for instance, take place by interactions of solutions in bodies of surface waters. These processes may be of inorganic origin or they may take place through the medium of living bodies, almost always at moderate temperatures. The products are usually mingled with detrital

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[1] N. L. Bowen, Geologic thermometry, in E. E. Fairbanks's "The Laboratory investigation of ores," New York, 1928, pp. 172–199.

H. Seifert, Geologische Thermometer, *Fortschritte d. Min. Krist. u. Petrog.*, 14, 1929, pp. 167–291. (Bibliography.)

P. Ramdohr, Neue Beobachtungen ueber die Verwendbarkeit opaker Erze als "Geologische Thermometer," *Zeitschr. prakt. Geol.*, 39, 1931, pp. 65–73; 89–91. (Bibliography.)

matter. They may be enriched by secondary processes in the unconsolidated strata or by processes of weathering after their exposure to air.

Another class of deposits is formed in bodies of surface waters by their evaporation and consequent precipitation of the salts dissolved in them; these are frequently termed the "saline residues." Common salt, gypsum, and borates are among the substances found in these deposits.

**Concentration of Substances Contained in the Rocks.**—Instead of at the surface or in bodies of surface waters, the processes of concentration of useful substances may go on in the rocks themselves. We may distinguish two cases: the substances were originally contained in the same geological body in which the deposit is found, or they have been introduced from the outside.

The apparent objection to this basis of subdivision, namely, the difficulty of deciding the source of the mineral or metal, is met in many cases by the knowledge acquired during late years. There may be deposits for which the question cannot be decided, but I believe that in the near future we shall in most cases have sufficiently good evidence. No one seriously maintains that the gold in the quartz veins of California, for instance, has been leached from the surrounding country rock; and surely no one denies that the oxidized nickel silicate ores of certain peridotites were originally contained in minute distribution in these rocks.

In the case of substances contained in the geological body itself, the concentration may be effected by (1) rock decay and residual weathering—that is, by oxygenated surface waters; (2) by the ground water of the deeper circulation; and (3) by processes of dynamic and regional metamorphism.

**Residual Weathering.**—Rock decay tends to destroy the rocks as units; to break them down, mechanically and chemically, and to re-assort their constituents in new combinations. In the decaying mass certain constituents are concentrated or precipitated; its detritus is swept away and deposited in rivers, lakes, and oceans; its soluble constituents are carried into the larger reservoirs and there perhaps precipitated in various forms.

It is true that not quite all the sedimentary deposits are derived from the decaying rocks; the fossil coals are indirectly made from the carbon of the atmosphere; volcanic ashes contribute a share to the sediments; the exhalations of eruptive magmas, as well as ascending waters, contribute some dissolved matter from the lower part of the earth's crust.

Processes of sedimentation and rock decay take place at moderate temperatures and pressures and the new minerals formed are, as a rule, characterized by high hydration. Below 0° C. mineral deposits do not form, except in so far as freezing of water is retarded by rapid motion or dissolved salts. Few of the deposits have been formed at temperatures above 50°, and this only exceptionally during eruption, evaporation in shallow desert lakes, or oxidation of pyritic rocks. The pressure is in general little different from that of the normal atmosphere, but in deposits in deep seas or lakes considerably higher pressures prevailed. This increased pressure, at low temperature, appears to have had little influence on the mineral associations formed.

**Deep Circulating Waters.**—Under the influence of the ground water of the deeper circulation many ore deposits are formed, concerning some of which there may be room for differing opinions. Copper may be leached from greenstones and the ores of the metal may be deposited in veins in the same rock. Hematite, like that of the Lake Superior region, may be concentrated from the surrounding low-grade "iron formation." Barite, magnesite, and sulphur are other instances.

**Regional Metamorphism.**—Again, the agency may be metamorphism under stress or regional metamorphism; in such cases the change takes place with very little water and it is not considered probable that a great concentration of the metals contained can be effected. Other materials may form, such as slate from shales, or useful minerals like garnets, cyanite, or graphite may develop in the rock. During static metamorphism, temperature and pressure are likely to be somewhat higher than at the surface. Regional metamorphism takes place under heavy pressure and at fairly high temperatures at great depth. It may merge into igneous metamorphism.

**Introduced Ores not Connected with Igneous Rocks.**—Much more common is the case where the valuable minerals have been introduced into the rock from without, and to this class belong the majority of the metal deposits. Deposits of this kind occur along fissures, or form replacements along fissures, or are found in general where opportunity is offered for vigorous circulation of the depositing waters. For a long time it was held by many that the metallic contents of fissure veins were derived from the surrounding rock, but it is now generally admitted that such a view in most cases is erroneous.

Certain metallic ores occur apparently independent of igneous rocks; the mineral associations in these indicate a deposition at moderate pressure and temperature, the latter probably rarely reaching 100° C. Of this kind are certain lead-zinc deposits in limestone or the copper deposits in sandstone which are so common in various parts of the world. Most geologists agree that such deposits have been formed by surface waters, at moderate depths; and that the metals have been leached from neighboring strata and, after a comparatively short wandering, deposited in fractured rocks in their present resting places. These deposits are generally poor in gold and silver.

**Deposits Genetically Connected with Igneous Rocks.**—There is also another and larger class which appears only in or near igneous rocks and whose epoch of formation usually can be shown to have followed closely after the eruption. This class has been clearly recognized by almost all geologists. There is also general agreement that these deposits have been laid down by heated, ascending waters, although there is no unanimity as to the source of the water or the source of the metal. To some, the water and the dissolved metals are simply igneous emanations from a cooling magma; to others, the waters are of atmospheric origin and, heated by their passage through still warm igneous rocks, have dissolved the metals contained in them. The majority hold to an igneous origin with gradually increasing admixture of meteoric water and substances dissolved on the upward path.

Nearly all metal deposits of the American Cordilleran region belong to this division. It is subdivided into several groups, according to the evidence of mineral association and geological relations. The first group includes ores deposited at slight depth below the surface; the temperature is here relatively low, perhaps from 50° to 200° C., and the pressure will scarcely exceed 100 atmospheres. Examples of this group are found in the gold and silver veins of Tonopah, Nevada, the Cripple Creek, Colorado, gold telluride veins, and the California quicksilver veins. These are the epithermal deposits. They are never found at great depths.

A second group comprises the deposits formed by hot ascending solutions at moderate depths, say from 5,000 feet to 10,000 feet below the surface, at temperatures of perhaps from 200° to 300° C. and correspondingly increased pressure. The present outcrops are exposed by deep erosion and they almost always appear in or close to intrusive bodies. The gold-quartz veins of California and the metasomatic pyritic deposits of Leadville, Colorado, may serve as examples. These are the mesothermal deposits.

A third, deep-seated group includes veins and replacement deposits. During the genesis of these the temperature was high, but in most cases below 575° C., the crystallographic inversion point for quartz. The pressure was probably very high. The cassiterite veins, some gold-quartz veins of the Appalachian type, and the tourmaline-copper veins belong to this group, which with great confidence may be ascribed to emanations from magmas. When occurring in limestone they carry lime-iron silicates. These are the hypothermal deposits.[1]

The deposits unquestionably formed by direct igneous emanations are the contact-metamorphic ores appearing in carbonate rocks along igneous contacts. They contain oxide ores, such as magnetite and specu-

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[1] L. C. Graton in a paper presented at the Tulsa meeting of the Geological Society of America, 1931 (in press) emphasizes the great depth of the mesothermal and hypothermal zone. He also proposes to introduce a new term, "the leptothermal zone," to cover deposits intermediate between the mesothermal and the epithermal. For the deposits formed very near the surface, above the epithermal ores, he proposes the name of "telethermal."

larite, together with sulphides of copper, zinc, and iron, and present an association of other minerals characteristic of pyrometasomatism.

The emanations from effusive bodies are deposited as sublimates of little economic importance.

It must be remembered that these classes are connected by transitions; also that high temperature is not always proportional to depth for by igneous action the isothermal lines may be carried high up towards the earth's surface.

**Products of Magmatic Differentiation.**—The last class is that of the deposits formed by concentration in igneous magmas; of all types these have formed at the highest temperature and pressure. They include oxides or sulphides segregated in the magmas, like the iron ores of Kiruna in northern Sweden, the titanic iron ores of the Adirondacks, or the copper-nickel ores of Sudbury, Ontario. They also include the pegmatite dikes, which contain many gems and rare metals and which are regarded as segregations from cooling granitic magmas. The pegmatites were formed at comparatively low temperatures—probably from 500° to 800° C.—but during the differentiation of the other deposits mentioned considerably higher temperatures probably prevailed. The pressure must, of course, have been very high.

**Metamorphism and Surface Enrichment of Deposits.**—In the proposed classification the mineral deposits are supposed to have suffered no change from their original condition. This is, of course, rarely strictly true, for chemical changes as a rule begin soon after the cessation of the agency which caused the deposition. In sedimentary beds this is particularly the case, for cementation and hardening and various chemical reactions begin almost from the time of deposition. It is, however, not the custom to refer to these changes as metamorphism.

Many mineral deposits have undergone great changes from their original conditions. They may have been reached by igneous metamorphism; and, thus, a coal bed transformed into anthracite or a bed of limonite into magnetite. Or they may have been sheared or crushed during regional metamorphism. Or, most common of all cases, they may have been altered by surface waters. Such oxidizing surface waters, as well as similar waters at somewhat greater depth, when they have parted with their free oxygen, produce peculiar modifications and often most important enrichments.

## A CLASSIFICATION OF MINERAL DEPOSITS[1]

I. Deposits produced by mechanical processes of concentration. (Temperature and pressure moderate.)

II. Deposits produced by chemical processes of concentration. (Temperature and pressure vary between wide limits.)

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[1] Presented before the Geological Society of Washington, May 10, 1911.

A CLASSIFICATION OF MINERAL DEPOSITS.—(*Continued*)

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|---|---|
| A. In bodies of surface waters. | |
| 1. By interaction of solutions.<br>*a.* Inorganic reactions.<br>*b.* Organic reactions.<br>2. By evaporation of solvents. | Temperature, 0° to 70° C. ±<br>Pressure, moderate to strong. |
| B. In bodies of rocks. | |
| 1. By concentration of substances contained in the geologic body itself. | |
| *a.* Concentration by rock decay and residual weathering near surface. | Temperature, 0°–100° C. ±<br>Pressure, moderate. |
| *b.* Concentration by ground water of deeper circulation. | Temperature, 0°–100° C. ±<br>Pressure, moderate. |
| *c.* Concentration by dynamic and regional metamorphism. | Temperature up to 400° C. ±<br>Pressure, high. |
| 2. Concentration effected by introduction of substances foreign to the rock. | |
| *a.* Origin independent of igneous activity. | |
| By circulating atmospheric waters at moderate or slight depth. | Temperature, to 100° C. ±<br>Pressure, moderate. |
| *b.* Origin dependent upon the eruption of igneous rocks. | |
| a. By hot ascending waters of uncertain origin, but charged with igneous emanations. | |
| 1. Deposition and concentration at slight depth. Epithermal deposits. | Temperature, 50°–200° C. ±<br>Pressure, moderate. |
| 2. Deposition and concentration at intermediate depths. Mesothermal deposits. | Temperature, 200°–300° C. ±<br>Pressure, high. |
| 3. Deposition and concentration at great depth or at high temperature and pressure. Hypothermal deposits. | Temperature, 300°–500° C. ±<br>Pressure, very high. |
| b. By direct igneous emanations. | |
| 1. From intrusive bodies. Contact metamorphic or pyrometasomatic deposits. | Temperature, probably 500°–800° C. ±<br>Pressure, very high. |
| 2. From effusive bodies. Sublimates, fumaroles. | Temperature, 100°–600° C.<br>Pressure, atmospheric to moderate. |
| C. In magmas, by processes of differentiation. | |
| *a.* Magmatic deposits proper. | Temperature, 700°–1500° C. ±<br>Pressure, very high. |
| *b.* Pegmatites. | Temperature, about 575° C. ±. Pressure, very high. |

# CHAPTER XVII

# DEPOSITS FORMED BY MECHANICAL PROCESSES OF TRANSPORTATION AND CONCENTRATION; DETRITAL DEPOSITS

## INTRODUCTION

Weathering tends to destroy rocks and mineral deposits by disintegration and chemical decomposition. In part, new minerals, like kaolin and limonite, form; in part, the more resistant minerals, like quartz, gold, platinum, magnetite, cassiterite, and garnet, are set free in individual grains. Erosion now steps in and the detritus is swept down the slopes and into the water channels. Mechanical separation in running water or along sea or lake beaches sorts the detritus according to specific gravity and size of grains. The heaviest particles, as those of gold, magnetite, and garnet, tend to collect in the lower part of the assorted detritus; the quartz grains are carried farther; the minute and easily moved scales of clayey substance are ultimately deposited as sedimentary beds; the colloids are coagulated by the electrolytes in the sea water.

## DETRITAL QUARTZ DEPOSITS

The quartz grains are often accumulated as beds of almost pure quartz sands. These are used extensively as ingredients in pottery and glass[1] also for abrasive purposes in sawing soft rocks, such as marble. Such sands should contain 99 per cent silica. At Ottawa, Illinois, for instance, such high-grade silica sand is mined by the hydraulic method and classified before using. It occurs in the St. Peter sandstone (Ordovician) and is of the well-rounded wind-blown dune type (Fig. 87). Somewhat argillaceous quartz sands without carbonates and carrying 80 to 90 per cent silica are used as molding sands and are mined on a large scale, though occurring in thin beds.[2] When compacted by pressure and by cementation the quartz sands are transformed into siliceous sandstones and quartzites which are used for millstones, whetstones, and grindstones.[3] Comparatively few localities furnish good material. With

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[1] E. F. Burchard, Requirements of sand and limestone for glass-making, *Bull.* 285, U. S. Geol. Survey, 1906, pp. 473–475.

R. M. Weigel, Technology and uses of silica and sand, *Bull.* 266, U. S. Bur. Mines, 1927.

[2] L. Heber Cole, The occurrence and testing of foundry molding sands, *Trans.*, Canadian Min. Inst., 20, 1917, pp. 265–291.

[3] R. B. Ladoo, The non-metallic minerals, New York, 1925, pp. 3–13.

the development of modern methods of grinding the importance of millstones has greatly decreased. Technical and statistical information on these subjects is contained in *Mineral Resources of the United States*, part 2, published annually by the U. S. Bureau of Mines, under "Abrasive Materials" and "Sand and Gravel."

In the case of very fine-grained whetstones a doubt may exist whether the material is of detrital origin or formed by chemical agencies. The so-called novaculite of Arkansas, the best whetstone known, is a good example of this. It occurs in the Devonian beds of Garland and Saline counties in that State, and is classified according to color and quality as Washita and Arkansas stones. The latter are snow-white and are the harder. The rock is much jointed and only small pieces are obtainable. Branner considers this material a metamorphosed chert, while Griswold[1] believes it to be a fine-grained sediment.

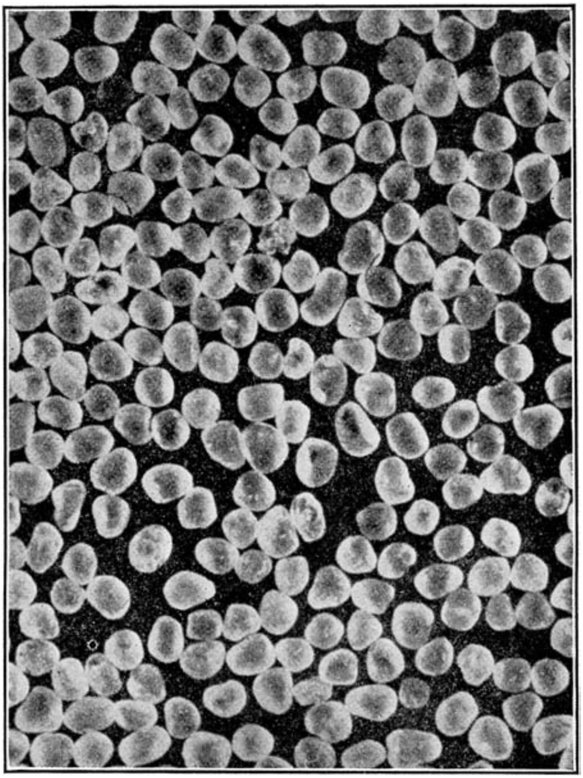

FIG. 87.—Quartz sand, St. Peter sandstone, Ottawa, Illinois, after screening. Magnified 5 diameters.

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[1] L. S. Griswold, Whetstones and the novaculites of Arkansas, *Ann. Rept.* Arkansas Geol. Survey, 3, 1890. H. D. Miser and A. H. Purdue, *Bull.* 808, U. S. Geol. Survey, 1929.

Quartz sand and pure quartzite (ganister) are used extensively in the manufacture of refractories.[1] Such material should contain 97 per cent silica and not over 0.40 per cent alkalies.

### DETRITAL CLAY DEPOSITS[2]

The fine material resulting from the decay of rocks is carried away, suspended in water, and deposited in river beds, lakes, and seas as sedimentary clay. The nature of clays is a much discussed subject. Perhaps the best definition is given by G. P. Merrill, who says[3] that the clays are widely diverse in origin and in mineral and chemical composition, but have the common property of plasticity when wet and that of induration when dried. Clays are finely comminuted aggregates of hydrous aluminous silicates, detrital quartz and other mineral fragments, often also, iron hydroxide and calcic and magnesic carbonates. The sedimentary clay is therefore to be regarded rather as a rock than as a mineral. Its principal use is for structural purposes. The detailed description of these deposits, therefore, does not fall within the scope of this book. The material is largely used for bricks, refractories, and other ceramic products.

The larger part of the clays is derived from decomposition and hydration of feldspathic minerals; other silicates, however, contribute their share. The mineral kaolinite ($H_4Al_2Si_2O_9$) is one of the principal constituents of clay.[4] The formation of kaolinite from feldspars can easily be traced in decomposing rocks at the surface, but in the clays the mineral is so comminuted that it cannot always be readily identified. It is known that colloidal hydrous silicates of aluminum exist, such as halloysite and allophane. The sedimentary clays rarely approach kaolinite in composition. Kaolinite should contain 46.5 per cent $SiO_2$, 39.5 per cent $Al_2O_3$, and 14 per cent $H_2O$; but by reason of admixture of quartz and undecomposed silicates, the sedimentary clays usually contain much more silica than the amount indicated.

Clays without carbonates generally contain more magnesium than calcium, and potassium exceeds sodium. Titanium often exceeds 1 per cent. Much of the titanium and potassium is probably present in colloidal state. Traces of copper, nickel, lead, zinc, and vanadium are sometimes found.

Regarding residual clays derived from the decomposition of rocks in place see page 351. The clays formed by the action of sulphuric acid on silicates in the oxidized part of ore deposits are described on page 457.

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[1] J. S. McDowell, A study of the silica refractories, *Trans.*, Am. Inst. Min. Met. Eng., 57, 1918, pp. 3–61.

D. W. Ross, Silica refractories, *Tech. Paper* 116, U. S. Bur. Standards, 1919.

[2] For more details in regard to the important clay industry the reader is referred to H. Ries, Clays, New York, 1927. Also Building stones and clay products, New York, 1912. P. M. Tyler, Clays, *Information Circ.* 6155, U. S. Bur. Mines, 1929, 63 pp.

[3] G. P. Merrill, Rocks, rock-weathering, and soils, New York, 1897, p. 135.

[4] C. S. Ross and P. F. Kerr, The kaolin minerals, *Prof. Paper* 165, U. S. Geol. Survey, 1931, pp. 151–180.

## FULLER'S EARTH[1]

Fuller's earth is the name given to certain sediments of clay-like material, originally used in England by fullers for cleansing cloth of grease. At present this substance is extensively used for deodorizing, decolorizing, and clarifying fats and oils; much of it is employed in the refining of petroleum. Its value thus depends upon its absorbent qualities.

The material occurs in sedimentary beds of Mesozoic, Cenozoic, and Quaternary age, but a similar material is also derived from the weathering of basic igneous rocks. Microscopic examination gives little evidence of its origin; in color it ranges from gray to dark green; it possesses little or no plasticity. The chemical analysis also has little value in determining its quality. J T. Porter believes, and probably justly, that the material owes its quality to the adsorbent power of colloid hydrous aluminum silicates.

The analyses show that the silica varies between 47 and 75 per cent, alumina from 10 to 19 per cent, lime from 1 to 4 per cent, magnesia from 2 to 4 per cent, ferric oxide from 2 to 10 per cent, and combined water from 5 to 21 per cent.

In Gadsden County, Florida, and Decatur County, Georgia, it occurs in Tertiary strata and is mined in open pits; in Arkansas it is obtained from weathered basic dikes. The further preparation includes drying, grinding and bolting to sizes from 30 to 100 mesh per inch. Very fine material clogs the filter presses. The rapidly rising domestic production amounted to 336,000 short tons in 1930. About 7,000 tons are imported. Georgia and Florida yield most of the total domestic production. The price of the Florida material is about $13 per ton. Texas, Illinois, and Massachusetts also contribute to the production.

*Bentonite*[2] is another highly absorbent variety of clay containing about 60 per cent $SiO_2$, 23 per cent $Al_2O_3$, 10 per cent $H_2O$, and several per cent of alkalies. It occurs in Cretaceous and Tertiary beds in the Cor-

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[1] T. W. Vaughan, Fuller's earth of Florida and Georgia, *Bull.* 213, U. S. Geol. Survey, 1903, pp. 392–399.

J. T. Porter, Properties and tests of fuller's earth, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 268–290.

E. H. Sellards and H. Gunter, *Second Ann. Rept.*, Florida Geol. Survey, 1908–1909, pp. 255–290.

Charles L. Parsons, Fuller's earth, *Bull.* 71, U. S. Bur. Mines, 1913.

T. P. Maynard and L. C. Mallory, Commercial preparation and use of fuller's earth, *Chem. and Met. Eng.*, 26, June 7, 1922.

[2] H. S. Spence, Bentonite, *Rept.* 626, Canada Dept. Mines, Mines Branch, 1924.

C. S. Ross and E. V. Shannon, The minerals of bentonite and related clays and their physical properties, *Jour.* Am. Ceramic Soc., 9, 1926, pp. 79–95.

C. W. Davis and H. C. Vacher, Bentonite, *Tech. Paper* 438, U. S. Bur. Mines, 1928, 51 pp.

dilleran region and is believed to be a decomposed volcanic ash. It is used for the bleaching of oils, as a filler in various processes, for the purification of water (on the principle of permutite), and other purposes. Ross and Shannon state that it is mainly composed of a crystalline clay mineral of micaceous habit probably montmorillonite $(MgCa)O.Al_2O_35SiO_2.xH_2O$, or in part beidellite, $Al_2O_3.3SiO_2.xH_2O$. These minerals belong to the same class as leverrierite and halloysite. The water is high, ranging up to 24 per cent.

## PLACER DEPOSITS

**Origin and Distribution.**—The heavier and less abundant minerals in the rocks are the most resistant to decomposition, and when the weathered rock is eroded and sorted by water they usually become concentrated in the lower parts of the sand and gravel beds. The gold-bearing gravels, which form an important source of supply of this metal, were called placers[1] by the early Spanish miners of this continent, and this name is probably the best that can be adopted for deposits of this class. Instead of gold the valuable mineral may be cassiterite, magnetite, monazite, diamonds, or other precious stones. Other terms have been employed, as "gravel deposits" or "gold-bearing gravels," or "alluvial deposits"—all equally objectionable, for the material may be sand instead of gravel, and it may be deposited along the ocean beach instead of in watercourses.

The processes of erosion and concentration have been active since earliest geologic time, hence we may have detrital deposits or placers of differing ages. Land deposits are, however, usually thin and easily removed; thus placers of pre-Tertiary age are comparatively rare.

In the formation of placers nature simply employs in her own leisurely way the processes of crushing and concentration which we use in ore dressing. The rocks are broken and comminuted by the expansion due to alternating heat and cold; by the growth of plants; or by the impact of sliding and water-carried rocks; or by the grinding action of ice; or finally by chemical decomposition and hydration. The products are concentrated in water courses by running water or along shores by ocean currents by motion similar to that on tables and jigs. Spherical particles of different substances fall in water at a rate proportional to their weight divided by the resistance. As the resistance is proportional to the area exposed, a fragment of quartz the size of a pea will fall much more slowly than a piece of gold of the same size. It will in fact be carried along easily in a current of water in which a piece of gold of the same size will sink instantly. Thus the specific gravities of the valuable minerals play

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[1] Derivation uncertain: Placer, pleasure; Plaza, place. Stelzner (Die Erzlagerstätten, p. 1261) says placer is a local Spanish term for sand bank. The Germans use "Seife," meaning washings. In French the word "alluvions" is often used.

a prominent part in the formation of placers. The specific gravity of the more important substances is as follows: Quartz, 2.65; feldspar, 2.55 to 2.75; ferromagnesian silicates, 2.9 to 3.4; garnet, 3.14 to 4.13; diamond, 3.52; corundum, 4.0; monazite, 5.0; magnetite, 5.1; cassiterite, 6.4 to 7.1; gold, 15.6 to 19.33; platinum, 14.0 to 19.0 (21 to 22 when chemically pure).

The shape of the particles is also of importance. Flaky minerals, like molybdenite, scaly gold, or specularite, are difficult to concentrate in spite of their high specific gravity.

## GOLD PLACERS

**Introduction.**—Gold is the most important placer mineral. Roughly speaking, about $25,000,000 out of a world's production of about $400,000,000 are derived from Tertiary or Quaternary placer deposits. Gold placers as a rule are easily discovered and worked; the supplies of old and long-settled countries were generally long ago exhausted. Bohemia, Italy, Spain, and Hungary, now almost barren of placers, once furnished their share. New deposits are usually discovered on the outskirts of civilization, as in Brazil in the eighteenth century, in Australia and California during the middle of the last century, and in Alaska and Siberia to-day. The production of placer gold in the United States, including Alaska, in 1897 was $7,800,000; in 1916 it was $22,882,000, the increase being due to the discovery of placers in Alaska and to the development of the dredging fields in California; in 1930 it was $8,400,000 or 20 per cent of the gold production. Eighty-five per cent of this gold was recovered by dredging. Practically all this gold comes from Quaternary and Tertiary placers, some dating back as far as the Eocene. A small quantity is obtained from Cretaceous conglomerates in Oregon and northern California. Permian gold-bearing conglomerates occur in Bohemia, according to Posepny.[1] Permo-Carboniferous conglomerates containing detrital gold have been described by Wilkinson from New South Wales.[2] In most cases the gold content of these older conglomerates is small and they can rarely be profitably worked. An example of ancient placers is furnished by the Cambrian basal conglomerate of the Black Hills, South Dakota, which unconformably covers the pre-Cambrian schists and gold-bearing quartz veins. It was first described by W. B. Devereux[3] and later by J. D. Irving.[4] This conglomerate, which is from 2 to 30 feet

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[1] Genesis of ore deposits, 1902, p. 163.

[2] *Idem*, p. 162.

[3] W. B. Devereux, *Trans.*, Am. Inst. Min. Eng., 10, 1882, pp. 465–475.

[4] J. D. Irving, Economic resources of the northern Black Hills, *Prof. Paper* 26, U. S. Geol. Survey, 1904, pp. 98–111.

thick and is overlain by quartzite, carries in places gold of unquestionably detrital origin, as indicated by the rounded grains, and has been profitably worked in several mines. While there may have been some later mineralization the facts seem satisfactorily established. If the ancient gold-bearing conglomerates of the Transvaal are classed as placer, then over one-half of the world's production would be derived from detrital deposits.

**Origin of Placer Gold.**—In primary deposits gold is mainly contained in veins, lodes, or shear zones; and these appear in rocks of many different kinds. It is often stated that gold is distributed as fine particles in schists and massive rocks and that placer gold in certain districts is derived from this source. Most of these statements are not supported by evidence, though it is not denied that gold may in rare instances be distributed in this manner. Even in the Yukon region, concerning which such statements have often been made, the origin of the gold from veins, lodes, and shear zones is now recognized.[1]

The great majority of gold placers have been derived from the weathering and disintegration of auriferous veins, lodes, shear zones, or more irregular replacement deposits. These primary deposits were not necessarily rich and may not be profitable to work. In many regions the rocks contain abundant joints, seams, or small veins in which the gold has been deposited with quartz.

**Eluvial Deposits.**—Gold placers may be formed by rapid erosion of hard rocks, but such placers are not often rich and highly concentrated. In the great placer regions the concentration has generally been preceded by an epoch of deep secular decay of the surface. It has been supposed by many that this deep rock decay is peculiar to the tropics, but this is not correct. The process has been active in the southern Appalachian States, in California, and even in Alaska, as well as in countries like the Guianas and Madagascar. When the outcrops of gold-bearing veins are decomposed a gradual concentration of the gold follows, either directly over the primary deposits or on the gentle slopes immediately below. The vein when located on a hillside bends over (Fig. 88) and disintegration breaks up the rocks and quartz, the latter as a rule yielding much more slowly than the rocks; the less resistant minerals weather into limonite, kaolin, and soluble salts. The volume is greatly reduced, with accompanying gold concentration. The auriferous sulphides yield native gold, hydroxide of iron, and soluble salts. Some solution and redeposition of gold doubtless take place whenever the solutions contain free chlorine. The final result is a loose, ferruginous detritus easily washed and containing easily recovered gold. This gold consists of grains of rough and irregular form and has a fineness but slightly greater than that

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[1] A. H. Brooks, The gold placers of parts of Seward Peninsula, Alaska, *Bull.* 328, U. S. Geol. Survey, 1908, pp. 110–135.

of the gold in the primary vein. Stelzner has applied to such residual concentrations, which may be worked like ordinary placers, the term *eluvial* gold deposits.

In the gold region of the southern Appalachian States the decomposition of the country rock, which generally is a schist, may reach a depth of 100 feet or more.[1] The decomposed material of the auriferous veins slides downhill, mixing with the weathered rock; and during this process the gold in part sinks deeper into the detritus. This has given rise to a peculiar system of mining by which the whole mass is washed by the hydraulic method and the more resistant quartz boulders crushed in a stamp mill with coarse screen. This has been practiced at Dahlonega, Georgia, and is often called the Dahlonega system. Similar deposits were worked in California, particularly in Eldorado county, and are here called "seam diggings" from the fact that the gold occurs disseminated in quartz seams traversing a certain belt of schists. Such deposits frequently occasion legal contests owing to the uncertainty whether they should be considered as placers or as mineral-bearing veins.

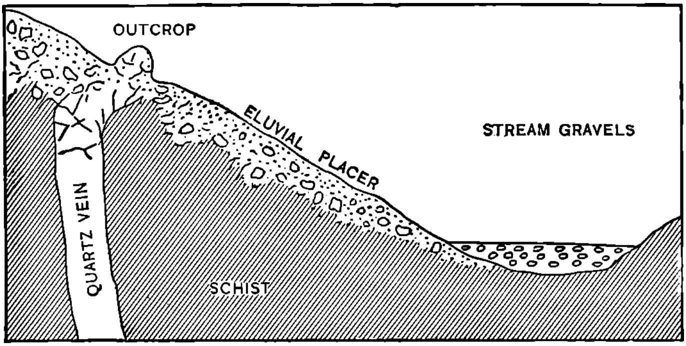

FIG. 88.—Diagram showing development of eluvial and stream placers.

In certain regions of Brazil[2] the schists and gneisses are covered by auriferous detritus accumulated in place. Another example is the "Tapanhoancanga" of the same country. This is a bed of residual, or lateritic, iron ore up to 10 feet thick covering the underlying hematite schist and containing gold throughout. The gold probably occurred in veinlets in the schists and the gold-bearing detrital material was concentrated from a considerable thickness of schist weathering in place.

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[1] G. F. Becker, Reconnaissance of the gold fields of the southern Appalachians, *Twenty-sixth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1895.

[2] O. A. Derby, Peculiar modes of occurrence of gold in Brazil, *Am. Jour. Sci.*, 3d ser., 28, 1884, p. 440.

O. Derby, Notes on Brazilian gold ores, *Trans.*, Am. Inst. Min. Eng., 33, 1892, pp. 282–283.

A. J. Bensusan, *Trans.*, Inst. Min. and Met., London, 1929, pp. 450–483.

Excellent examples of eluvial deposits are reported from Dutch, British, and French Guiana,[1] though ordinary stream placers are the most common deposits in these countries. Over a great part of this gold-bearing territory secular decay of crystalline rocks has resulted in a deep mantle of ferruginous clayey earth—laterite—and in places the gold has been concentrated in this material below outcrops of gold-bearing veins. Many of the stream beds are also worked for placer gold, the detritus usually resting on the clayey surface of the compact laterite.

It is stated that many rocks in the Guianas contain gold and that the placer gold is derived from such material; particularly are the basic rocks, diabases and amphibolites, said to be auriferous. This conclusion should probably be accepted with reservations. It seems more probable that the gold contained in the greenstones is of secondary origin and that here, as elsewhere, granitic intrusions have caused the formation of a series of gold-bearing veins in the surrounding rocks.

**Processes of Concentration.**—In most cases the cycle has been carried further and the material is not only decomposed, but eroded, transported, and redeposited. This can be effected by wind, by streams, or by the surf of the sea.

**Eolian Deposits.**—Deposits concentrated by eolian agencies can, of course, be formed only in dry countries where long subaerial decay has paved the way for the work of the dust storms; from the decomposed and crumbled outcrops of the lodes the winds blow away the lighter sands, leaving a mass of coarser detritus which contains the gold. Such wind-born placers have been noted by H. C. Hoover[2] and T. A. Rickard[3] near the outcrops of the Western Australian gold veins. No examples of this kind are known from the Cordilleran States of America.

**Stream Deposits.**—Running water is by far the most important agency in the formation of gold placers. The high specific gravity of gold explains many of the puzzling features of the placers. Placer gold is six or seven times as heavy as the most common accompanying minerals —feldspar and quartz—and settles to the bottom in flowing water

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[1] C. G. Dubois, Beiträg zur Kenntniss der surinamischen Laterit, etc., *Tschermak's min. u. petr. Mitt.*, 22, 1903, pp. 1–61.

E. D. de Levat, Guide pratique, etc. de l'or en Guyane française, Paris, 1898. See also *Mineral Industry*, 7, 1899.

A. Bordeaux, *Trans.*, Am. Inst. Min. Eng., 41, 1910, pp. 567–593. See also *Mines, Carrieres*, February, 1930, pp. 17–27.

J. B. Harrison, The geology of the gold fields of British Guiana, London, 1908.

J. B. Harrison, in the *Reports* of the Institute of Mines, British Guiana.

E. E. Lungwitz, Die Goldseifen von British Guiana, *Zeitschr. prakt. Geol.*, 1900, pp. 203–218.

[2] H. C. Hoover, The superficial alteration of Western Australian ore deposits, *Trans.*, Am. Inst. Min. Eng., 28, 1898, pp. 762–763.

[3] T. A. Rickard, The alluvial deposits of Western Australia, *idem*, pp. 480–537.

with surprising rapidity. It is almost impossible to lose a particle of gold, of the value of one cent, in a miner's pan; it sinks immediately to the bottom of the gravel and sand after one or two preliminary shakes in water. Once lodged at the bottom it stays there, in spite of shaking and rotating. This illustrates the fundamental fact that the gold is mainly on the bed-rock. The rapid settling of the gold accounts for the partial failure of some devices for placer mining, particularly the clam-shell and the suction dredges.

The ease with which such concentration, according to the specific gravity, is effected is shown by the well-known fact that in powdered samples of ore, as well as in dumps at the mine, a settling of the heavier ore particles toward the bottom can often be observed.

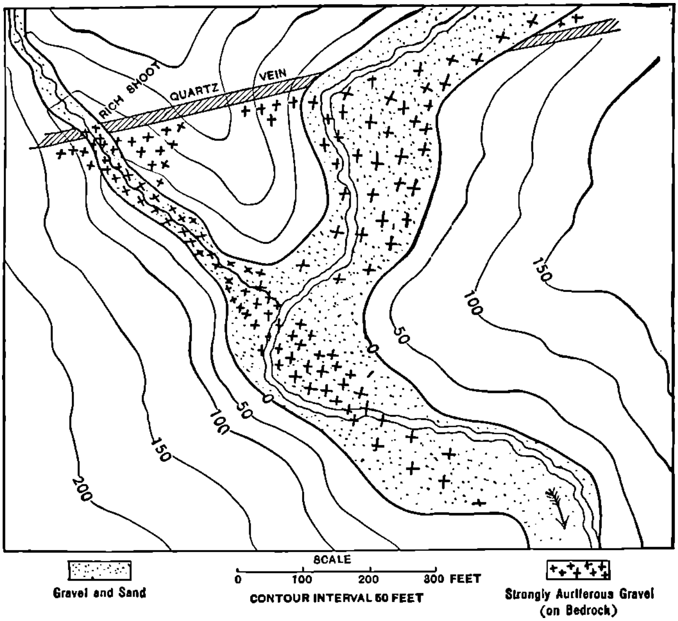

FIG. 89.—Plan of quartz vein and placers below it, illustrating the development of pay streaks.

Suppose we have a gold-bearing quartz vein deeply altered by rock decay; now let the region be raised, say 500 feet, by one of these slow oscillations which so commonly affect the crust. A river has excavated a valley to the corresponding depth in this elevated plateau, and this valley—under the influence of a pause in the elevating movement—becomes filled with gravels to a depth of about 100 feet. Let a tributary gulch with steep grade be cut back into the plateau to the gold deposit (Fig. 89); when the gulch reaches it the eluvial deposit will be carried down by sliding and washing; the clay and limonite are rapidly removed in suspension; the angular gravel of quartz and rock, grinding the fragments of gold between them and on the bed-rock, will be moved downward, the fine grains in suspension, the coarser ones dragging and rolling on the bottom. There is little deposition; the transporting power is great and in flood time the whole gravel mass, of no great depth, will probably be in motion. Heavy gold nuggets may lodge in the lee of little ridges. The gold settles rapidly; most of it, continually hammered and slowly shaping itself in flat, smooth grains, will be dragged down stream and finally reach the edge of the flood-plain in the river. At this place the larger part of the gold stops. It is not washed out with the sand and gravel but stays on the bed-rock near the margin. The finer particles will, of course, be carried out a little distance, but they soon sink into the water-filled gravel after the manner of grains of heavy ores in concentrating jigs. Just as in the gulch the whole mass of detritus is transported, so it is thought that in larger streams the body of water-soaked gravel and sand works downstream very slowly. During this process the lighter gold contained in the detrital material also works forward and downward, gradually accumulating with the nuggets or coarser pieces, which have already reached their final resting ground.

This mode of operation contains the key to the genesis of the placers. It is not to be expected that the coarse and ordinary fine gold will be carried out into the middle of wide flood-plains. As the flood-plain widens it will cover the accessions of gold along its margin, and the final result will be a streak of rich gold-bearing gravel, resting on the bed-rock and extending downstream deep underneath the surface. When this is traced upstream the primary deposit, the vein, will be found. The actual occurrences of course show infinite variation. Let us assume that, as happens in the Creswick district in Victoria, Australia, a broad stream with moderate grade crosses a deeply decomposed belt of soft slate containing an abundance of small veins or stringers of quartz with native gold, and that, in addition, a fair balance between transportation and deposition persists for a long time. The result will be a gravel deposit, only a few feet deep, but with an abundance of gold concentrated on the bed-rock over the whole width of the stream. Each freshet is sufficient to churn up and move forward the whole mass of gravel, continually adding to the concentrated gold on the clayey bed-rock.

Again, we may assume extremely active erosion, as is the case in the Sierra Nevada of California. Canyons several thousand feet in depth have been cut in an uplifted plateau, veritable trenches or sluice boxes, the grade of which is from 60 to 150 feet per mile. Stretches of wild gorges with polished bottoms alternate with stretches of less grade where shallow gravel accumulates. These canyons receive for long distances an abundant supply of gold, of all sizes, from older hill gravels or from decaying quartz veins. The result will be that but little gold will lodge in the gorges, while extremely rich shallow gravel bars will accumulate in the convex stream curves (Fig. 90). Gradient, volume, and load usually vary in the same stream so that deposition may be going on in one part of its valley and erosion in another. Continued corrasion of the stream-bed results in deepening the canyon and leaving the bars as elevated benches. The miners of 1849 first found these bars and worked them. In searching for the source of the gold they soon found a trail of metal leading up the gulches to great masses of older gravels on the hills, 2,000 to 3,000 feet above. These gravels were washed by the hydraulic method; and immense masses of tailings with a little gold were carried down to the rivers, totally overloading them. After the prohibition of hydraulic mining the streams gradually resumed active transportation. The whole gravel mass moved slowly downstream and a gradual reconcentration on the bed-rock took place. The tailings deposited became enriched and will ultimately be reworked.[1]

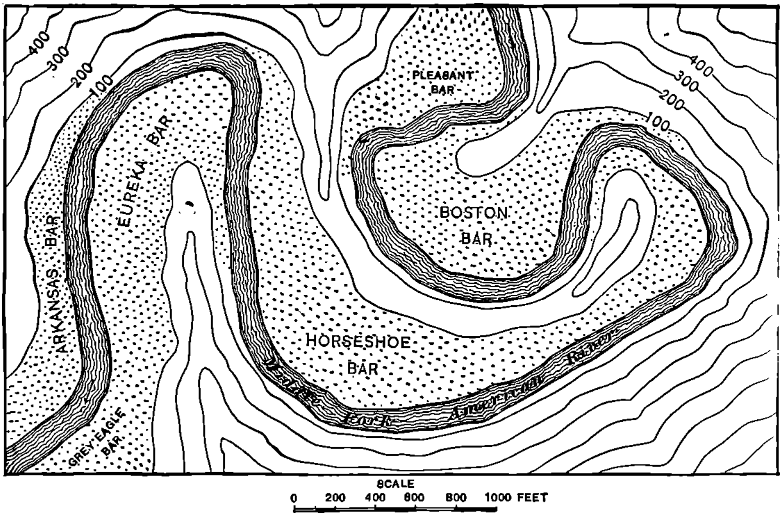

FIG. 90.—Low gravel bars, American River, California, showing placer deposits on inner side of bends. (*After R. L. Dunn.*)

The torrential floods of the canyons scarcely permitted the lodgment of fine gold. This was swept out through the narrow portals into the Sacramento Valley, where the grade of the streams suddenly diminishes. The most minute particles may have been carried as far as San Francisco Bay, but the bulk of the fine gold lodged in the flood-plains within a few miles of the mouth of the canyons. Easily caught upon the clayey "*false bed-rock*" of volcanic tuff, this gold, the average particles of

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[1] G. K. Gilbert, Hydraulic-mining débris in the Sierra Nevada, *Prof. Paper* 105, U. S. Geol. Survey, 1917. Report to the legislature of the hydraulic mining commission on the resumption of hydraulic mining in California, Sacramento, 1927, 85 pp.

which are about 0.3 millimeter in diameter, formed meandering pay streaks at the base of a sandy gravel bed from 10 to 60 feet in depth. Such deposits have been worked at Oroville, Butte County, California, by dredging.

By an odd paradox, gold is at the same time the easiest and the most difficult mineral to recover. It is divisible to a high degree and owing to its insolubility the finest particles are preserved. A piece of gold worth one cent is without trouble divisible into 2,000 parts, and one of these minute particles can readily be recognized in a pan. In extreme subdivision the gold acquires a scaly, flat form, being known as flour gold or flake gold, is carried away very readily by water, and does not sink easily in sand or gravel. In part the flour gold is suspended by air films, and can be carried away in rivers of moderate grade for hundreds of miles. The gold occurring in the sand bars of Snake River, Idaho, is a good example of this.[1] It will settle in thin pay streaks at bars and other favorable places, but the next freshet will probably destroy the sand bars and sweep the gold away. This accounts also for the distribution of fine gold in great masses of gravel beds—for example, in the wash 600 feet thick deposited by glacial streams at Tacoma and other places on Puget Sound. Almost every pan of this gravel will show a "color," but the material contains only a fraction of a cent per cubic yard. The fine colors along the Columbia River in northeastern Washington range in value from less than 0.0005 to 0.02 cent, the average being about 0.002 cent.[2]

The much-discussed concentration of gold on the bed-rock seems, then, to be due partly to the natural jig-like movement in moderately deep gravels,[3] during long-continued conditions of fair balance between loading and erosive power; partly to slow forward and downward motion of heavier gravel masses,[4] of which exact measurement as yet is lacking; and last and largely, to the fact that heavier gold will not be carried out into the gravel flats of rivers of gentle grade—the only ones that have extensive flood-plains—but is immediately deposited on the marginal bed-rock of the gradually deepening and widening gravel plains.

The best conditions for the concentration of gold are found in moderately hilly countries where deep secular decay of rocks has been followed by slight uplifts. Subsequent slight elevations would easily produce re-sorting and enrichment of the gravels. In regions of gold placers the richest material is usually produced by repeated reworking of gold-bearing gravels by nature. Each reworking increases the richness of the gravels,

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[1] J. M. Hill, Gold of the Snake River, *Bull.* 620, U. S. Geol. Survey, 1916, pp. 271–294.

[2] A. J. Collier, *Bull.* 315, U. S. Geol. Survey, 1907, p. 61.

[3] F. Posepny, Genesis of ore deposits, New York, 1902, p. 154.

[4] T. A. Rickard, *Min. and Sci. Press*, Aug. 15, 1908.

eliminates easily decomposed pebbles, and finally results in a gravel of the hardest, most resistant rock—quartzite or quartz. Quartz is the common gangue mineral in gold regions; hence the prevalence of "white gravels" or "white channels," almost exclusively composed of white quartz pebbles.

## CLASSIFICATION OF FLUVIATILE AND MARINE PLACERS

According to their occurrence the placers may be conveniently divided as follows:[1]

PLACERS CLASSIFIED

| Present topographic cycle | Past cycles, elevated | Past cycles, depressed |
|---|---|---|
| 1. Gulch and creek gravels. | 1. High creek gravels. | 1. Deep creek gravels. |
| 2. River and bar gravels. | 2. Bench gravels.<br>Hill gravels or high river gravels. | 2. Deep river gravels. |
| 3. Gravel plains. | 3. Elevated gravel plains. | 3. Depressed gravel plains. |
| 4. Beaches. | 4. Elevated beaches. | 4. Depressed beaches. |

Examples of present gulch, creek, and river gravels are not difficult to find; they occur in all gold-bearing regions where erosion is active and where precipitation is abundant enough to cause the sorting and carrying forward of the gravels in the stream beds. In the upper parts of the stream courses the gravel will be coarse and semiangular; in the lower parts the sands increase and the pebbles are smoother. Where the rivers emerge from their narrow valleys and spread with gentle grade over flood-plains, more extensive sand and gravel beds will accumulate, generally, however, with less gold than in the more confined part of the course. Some of the fine gold may reach the sea and be concentrated by the surf and the oblique shore currents into thin pay streaks on the sandy beach.

**Marine Placers.**—Beach placers occur along many shores and are often produced by concentration from a sea bluff or elevated gravel plain. The beach at Nome, Alaska (Fig. 91), is a narrow strip about 200 feet wide, from which over $2,000,000 in fine gold has been washed; the flaky gold averaged 70 or 80 colors to the cent.[2] Two older elevated beach lines are found farther inland. The beach gold of the Oregon and

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[1] See also A. H. Brooks, The gold placers of parts of Seward Peninsula, Alaska, *Bull.* 328, U. S. Geol. Survey, 1908, p. 115.

[2] A. J. Collier and F. L. Hess, *Bull.* 328, U. S. Geol. Survey, 1908, pp. 140–228.

California coasts is much finer, the colors ranging from 100 to 600 to the cent.

**Buried Placers.**—Subsidence or overloading may cause the placers to be deeply covered by barren detritus. Many of the streams of Alaska, particularly in their lower reaches, are thus covered; the process of concentration is stopped, the present watercourses having insufficient grade to effect the transportation of detritus. Figure 92 shows the dredging operations on the Solomon River, Alaska. The depth of the gravel in the river bottom is about 20 feet. Figure 93 shows a diagrammatic section of the Oroville dredging ground, Butte County, California.

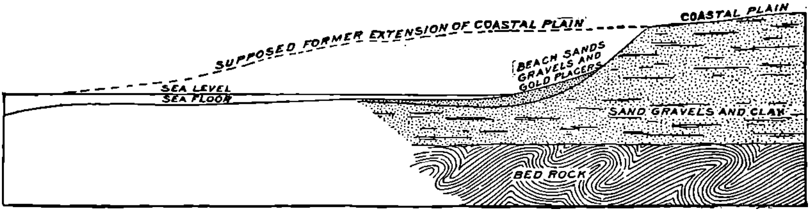

FIG. 91.—Diagrammatic section illustrating development of beach placer. (*After A. J. Collier and F. L. Hess, U. S. Geol. Survey.*)

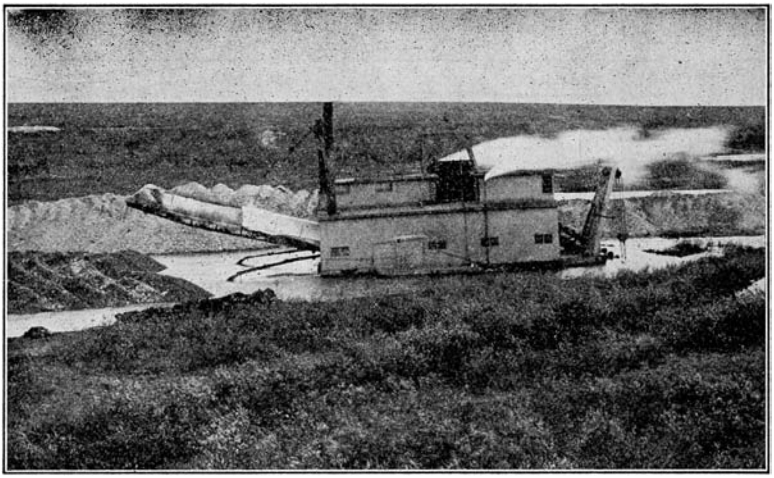

FIG. 92.—Gold dredging on the Solomon River, Alaska. (*After P. S. Smith, U. S. Geol. Survey.*)

The depth of the gravel is about 30 feet. At Fairbanks, Alaska, according to Prindle,[1] the placers occur in tributaries of moderate length, which flow in open valleys; some of the deposits are as much as 300 feet deep. The pay gravels, in part subangular, lie on the bed-rock and are from a

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[1] L. M. Prindle, The Fairbanks and Rampart quadrangles, *Bull.* 337, U. S. Geol. Survey, 1908.

few inches to 12 feet in thickness; these are covered by 10 to 60 feet of angular wash, evidently accumulated rapidly without opportunity for concentration, and above this rests a thick deposit of muck over which the sluggish streams pursue their way. The richest gravel worked in 1905, containing from $5 to $10 per cubic yard, occupied pay streaks on the bed-rock 150 to 200 feet wide, considerably less than the average width of the valley bottom. All the gravel on the bed-rock is, however, more or less auriferous. The gold is moderately coarse. Near the head of the stream deposition closely follows cutting and there the deeply buried, more or less permanently frozen pay streaks of the lower valleys merge into the deposits of the present stream activity.

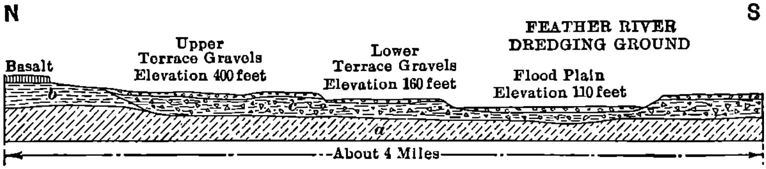

Fig. 93.—Diagrammatic section across Feather River below Oroville, California. *a*, Bed-rock; *b*, Ione formation; *c*, tuffs of Oroville.

On a large scale similar conditions prevailed in Victoria, Australia.[1] Here there existed in Pliocene time an extensive river system with shallow, well washed, and locally extremely rich gravels which were formed during a prolonged time of nice balance between erosion and deposition. The region was then depressed and covered by thick beds of sand and clay. Above this were poured out basalt flows, in places several hundred feet thick (Figs. 94 and 95). The broad valleys remain on the whole as before, but the present streams are weak and have

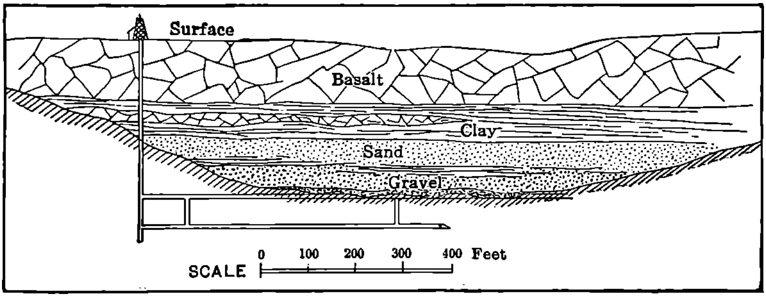

Fig. 94.—Diagram illustrating buried gravel channels (deep leads) of Victoria, Australia, and method of mining these deposits.

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[1] W. Lindgren, *Min. Mag.*, 2, 1905, p. 33.
W. Lindgren, *Eng. and Min. Jour.*, Vol. 79, Feb. 16, 1905, pp. 314–316.
H. L. Wilkinson, *Trans.*, Inst. Min. and Met., London, 1907, p. 9.
Stanley Hunter, *Mem.* 7, Geol. Survey Victoria, 1909.

little power of transportation and concentration. The discoveries of gold were made near the sources of the old rivers, where their gravels are near the surface; they were followed upward into the gullies of the slate hills, and downward below the level of the basalt flows. Such were the conditions, for instance, at Ballarat. South of Ballarat certain of the Pliocene stream gravels merge into coastal gravel plains, soon becoming marine in character. Such coastal gravel beds are opened in the Pitsfield mines, where the pay streaks of fine gold, resting on an almost level bed-rock, are worked beneath several hundred feet of sands and gravels.

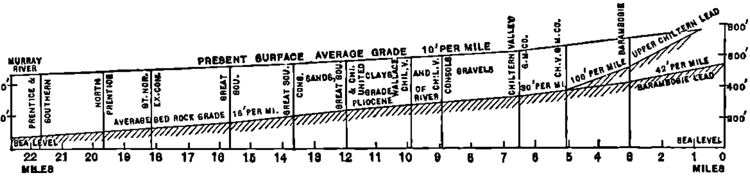

FIG. 95.—Longitudinal section of Chiltern Valley and Rutherglen deep leads, Victoria, Australia, showing steeper grade of Tertiary river beds.

The Sierra Nevada of California,[1] on the other hand, offers an excellent instance of the result of elevation on gravel deposits. In the early Tertiary the surface slope of this range was comparatively gentle, and during long periods of rock decay and well-balanced conditions gold from the quartz veins became strongly concentrated on the bed-rock of the streams. The deeper gravels were then covered by a considerable thickness of more rapidly accumulated and poorer, but well-washed material, and this in turn by heavy masses of rhyolitic tuffs and andesite breccias so that the old channels were sealed in places by as much as 1,500 feet of superincumbent barren material. The range was elevated by mountain-building disturbances; new rivers were laid out and rapidly eroded cañons to a depth of 2,000 or 3,000 feet. Eventually the old gravels were exposed and now rest as more or less connected remnants on the summits of the ridges between the modern cañons. The heavy gravel masses are worked by the hydraulic method, or the pay streak

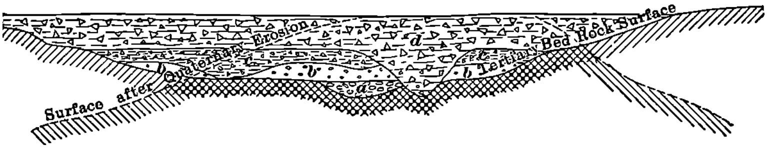

FIG. 96.—Schematic representation of the four principal epochs of Tertiary gravels in the Sierra Nevada. *a*, deep gravels (Eocene); *b*, bench gravels (Miocene); *c*, rhyolitic tuffs and interrhyolitic channel; *d*, andesitic tuffs and intervolcanic channel.

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[1] W. Lindgren, The Tertiary gravels of the Sierra Nevada, *Prof. Paper* 73, U. S. Geol. Survey, 1911.

on the bed-rock is extracted by tunneling operations in the "drift mines" (Figs. 96, 97, 98).

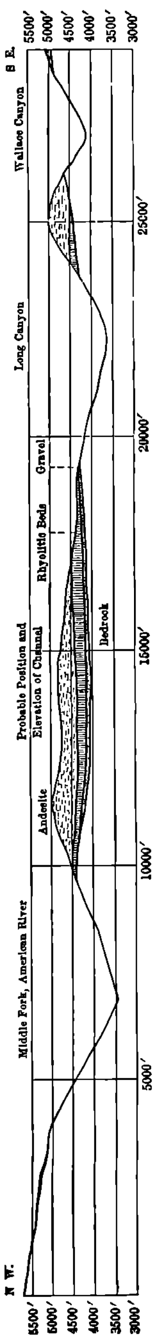

FIG. 97.—Section across the Tertiary Yuba River on the Long Canyon divide, Placer County, California.

The gold from the destroyed portions of the old channels, together with more set free from the quartz veins during the erosion, accumulated in the modern cañons. Along their slopes benches remain in places, indicating transient accumulations of gravel during the process of cañon cutting.

Somewhat similar conditions exist in some parts of Alaska. Near Nome on the ridges surrounding Anvil Creek are "high gravels" 600 to 700 feet above the present rivers. These gravels, some of which are rich, are the remnants of an old, now almost wholly eroded system of drainage.

In the Klondike, also, high gravels occur a few hundred feet above the present creeks, the most conspicuous instance being the "White channel," described by McConnell[1] (Fig. 99).

Elevated beaches have been mined, for instance, at Nome, where there are two old beach lines 37 and 70 feet above the present level of the ocean. In Santa Cruz County, California, a similar elevated beach was mined for some time. Gold-bearing beach sand occurs all along the Pacific coast from San Diego to Alaska, and in many other parts of the world.

**Size and Mineral Association of Placer Gold.**—In placers gold occurs in all sizes, from masses weighing 200 pounds to the most minute flakes. Large nuggets are recorded from California; still larger specimens, weighing as much as 2,280 ounces, were obtained in Victoria, Australia. It is often stated that heavier masses occur in placers than in quartz veins. This is decidedly erroneous. A mass of native gold found in the Monumental mine of Sierra County, California, weighed 1,146 troy ounces, and a quartz vein at Hill End, New South Wales, yielded a specimen which contained about 3,000 ounces. Any one who has had much experience in gold mining has noted the occurrence of thick sheets and masses of gold in deposits of certain kinds—for instance, in the pockety quartz veins of Alleghany, California.

[1] R. G. McConnell, Klondike gold fields, *Ann. Rept.*, Canada, Geol. Survey 14, 1905, 71 pp.

R. G. McConnell, Report on gold values in the Klondike high-level gravels, Canada, Geol. Survey, 1907, 34 pp.

Almost all the so-called placer nuggets of unusual size have been obtained from superficial deposits at or just below the outcrops of rich veins. This applies to the Ballarat nuggets, weighing from 80 to 160 pounds, which occurred in small, steep gulches underneath the basalt flows, but immediately below the extremely rich outcrops of the quartz veins. It also applies to the nuggets of Carson Hill, California, the Poseidon nugget of Victoria (found in 1906 and weighing 953 troy ounces), and other occurrences. Some very rich placer deposits—for instance, those of the Klondike, Yukon Territory, and the Berry mines in Victoria, Australia—contain no specially large pieces of gold. The heaviest nugget found in the Klondike is said to have weighed 85 ounces.

The angularity of the gold is inversely proportional to the distance traveled; the final product is usually a flat, rounded grain from a fraction up to 1 millimeter in diameter. Occasionally crystallized gold is found in placers, but this is unusual and indicates close proximity of the primary deposit.

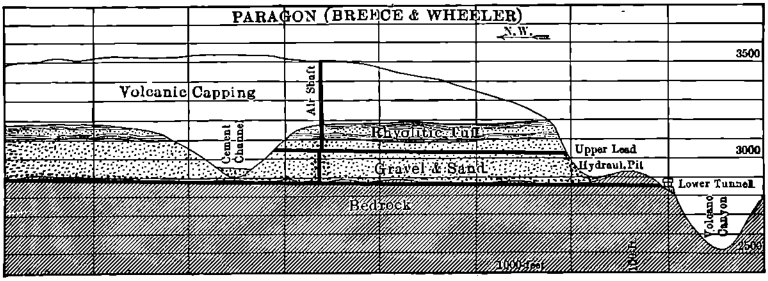

FIG. 98.—Longitudinal section of "Blue gravel channel," at Breece and Wheeler Mine, Forest Hill divide, Placer County, California. (*After R. E. Browne.*)

There is probably no authenticated case of crystallized gold occurring in the gravels of larger water courses where there has been long transportation, and this is assuredly a strong argument against the assumption that such crystals are formed by secondary processes in the gravels.

Fragments of quartz often adhere to the gold or form part of the rounded nugget. While the quartz pebbles so abundantly found in gold-bearing gravels do not ordinarily contain visible gold, there are many instances of such occurrences—for example, at Elk City and Idaho City, Idaho, and at Dutch Flat and Nevada City, California. Some placer gold, more frequently the scaly variety, is covered by a thin film of silica, manganese dioxide, or limonite, and does not amalgamate easily.

The most abundant minerals associated with the gold in placers are magnetite and ilmenite ("black sand"), garnet, zircon ("white sand"), and monazite ("yellow sand"), as well as many others of the heavy minerals occurring in the rocks which contain the primary gold deposits. Cassiterite is common in placers, and some deep gold placers in Victoria contain enough to make it a valuable by-product. Gray platinum and silvery foils of iridosmine are present in small quantities in many California placers adjacent to areas of serpentine. None of the minerals mentioned are ordinarily derived from the gold-bearing veins, but from the surrounding rocks.

Pyrite or marcasite may form in the gravels; sometimes this pyrite contains a little gold, but contamination of the assay samples by the placer gold itself is always a possibility. Again the pyrite may be clastic and derived from the surrounding rocks, for pyrite does not seem to oxidize readily in running water; or, as near Nevada City, California, in the Harmony channel, the gravel may contain undecomposed pyrite, rich in gold, and derived directly from the primary veins over which the water course flowed.

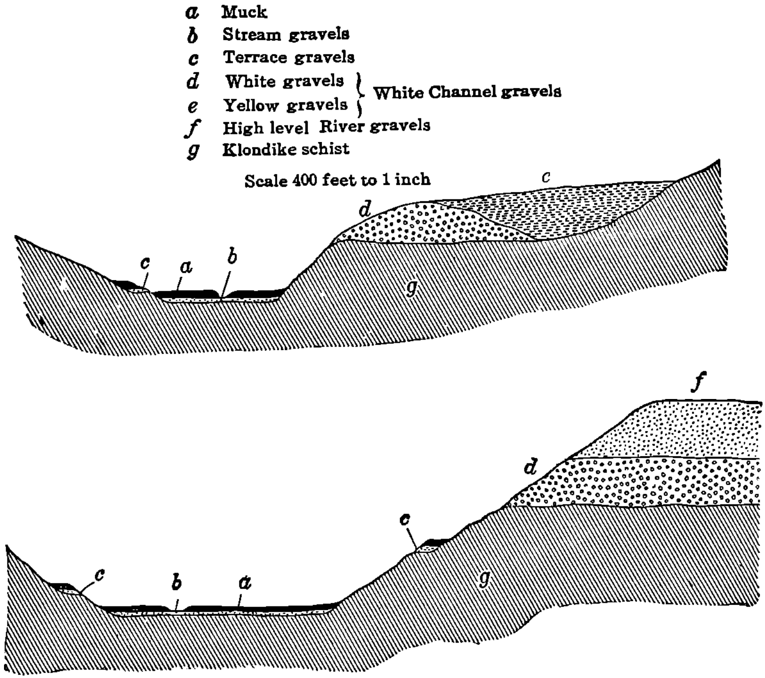

FIG. 99.—Sections across Bonanza Valley, Yukon Territory, showing several types of gravel deposits. (*After R. G. McConnell, Canada Geol. Survey.*)

Other occasional associates of gold, probably derived from its primary deposits, are silver in nuggets (Alaska), native bismuth (Queensland, Bolivia,[1] and Alaska), native amalgam, palladium-gold, native copper, and cinnabar. The presence of native lead has usually been explained by accidental admixture of hunter's shot, but J. Park[2] asserts that there

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[1] Perfect round smooth nuggets of bismuth are plentiful in some placers near Cherolque, Bolivia. Some of them are 1 inch in diameter.

[2] J. Park, Mining geology, London, 1907, p. 18.

is an instance of its undoubted presence in gravel, the lead containing a skeleton of native gold.

**Fineness and Relation to Vein Gold.**—The fineness of placer gold (or parts of gold per thousand) varies from about 500 to 999. Silver is always alloyed with the gold, but other metals are rarely prominent; copper is occasionally present. While vein gold may have a fineness of 997 to 999, this is exceptional; far more commonly its fineness ranges from 500, which corresponds to electrum, to about 800 or 850. The placer gold in any district will usually be of higher grade than the vein gold, and its fineness increases with the distance transported and with the decreasing size of the grains. Thus, while in California the vein gold averages 850 fine, the transported placer gold in the Tertiary channels averages 930 to 950. It has been shown that this increase in fineness is due to the solution of the silver in the alloy in the outer layer of the grains by the action of surface waters. McConnell has proved that in the nuggets from the Klondike the outside actually has a greater fineness than the inside. The loss of silver in the outer part was from 5 to 7 per cent. This interesting result well illustrates the relative insolubility of gold.[1]

**Gold in Relation to Bed-rock.**—While the bulk of the gold usually rests on the bed-rock or within a foot or two of it, this is not an invariable rule. In some gravels the coarser gold is occasionally scattered through the lower 4 to 20 feet. But it is never, except in minute quantities, distributed equally through a great thickness of gravels. An excellent instance is McConnell's section of the "White channel" deposit in the Klondike. (See Fig. 99.) The washed gravel is here 150 feet thick. The gold content of the gravel is as follows:

0– 6 feet above bed-rock, $4.13 per cubic yard.
6–12 feet above bed-rock, $0.18 per cubic yard.
12–18 feet above bed-rock, $0.047 per cubic yard.
18–24 feet above bed-rock, $0.04 per cubic yard.
24–30 feet above bed-rock, $0.034 per cubic yard.
30–36 feet above bed-rock, $0.032 per cubic yard.
36–42 feet above bed-rock, $0.032 per cubic yard.
42–48 feet above bed-rock, $0.045 per cubic yard.
48–54 feet above bed-rock, $0.025 per cubic yard.

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[1] W. B. Devereux, The occurrence of gold in the Potsdam formation, *Trans.*, Am. Inst. Min. Eng., 1882, pp. 465–475.

Ross E. Browne, Colorado placer gold, *Eng. and Min. Jour.*, 59, 1895, pp. 101–102.

W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, p. 637. See also *Prof. Paper* 73, U. S. Geol. Survey, 1911.

R. G. McConnell, Report on gold values in the Klondike high-level gravels, Canada Geol. Survey, 1907, 34 pp.

From 54 feet above bed-rock the quantity of gold contained per cubic yard gradually and steadily diminished to $0.006 at the top. There is only fine gold in the upper gravels. A local enrichment has taken place on false bed-rock, a clayey stratum 42 feet above the real bed-rock; here the gold is much coarser than directly above or below, but finer than on the bed-rock. Occasionally rich gravel may be found a few feet above bed-rock while it is less rich immediately on it.

Coarse and moderately coarse gold moves very slowly. McConnell found, for instance, that the "White channel," where intersected and eroded by gulches, has left almost the whole amount of its gold in these gulches immediately below the place where the trenching has occurred. In some cases the horizontal movement scarcely equaled the vertical.

Smooth, hard bed-rock is poorly adapted to retain the gold; when it is somewhat clayey and decomposed much better results are obtained. Schists and slates make good bed-rock when decomposed, especially when they strike parallel to the channel. Serpentine forms a smooth and unsatisfactory bed-rock.

Gold works down into bed-rock in a most surprising way. In hard rock it settles into the most minute crevices. In soft rock it burrows to a depth of 1 to 5 feet, so that it is always necessary to mine this amount of the bed-rock. In limestone, irregular solution cavities may contain the detrital gold, and these sometimes descend to a depth of 50 feet or more. Compact clay is good bed-rock, also clayey sandstone and clayey volcanic tuffs, the occurrence of the latter being exemplified in the Oroville dredging grounds, in California.

In glacial till and moraines there has been little opportunity for concentration, and unless the primary vein deposits were unusually rich, these gravels are of little value; the gold contained in them may, of course, be concentrated by glacial streams working over the morainal detritus.

**Grade of Auriferous Watercourses.**—All degrees of slope occur in watercourses containing gold-bearing gravels. In steep creeks the grade may be many hundred feet per mile, but the placers in these are usually poor. California rivers, in the Sierra Nevada, have grades of 50 to 100 feet or more per mile. Many of these have been extremely rich where gravel bars have had an opportunity to accumulate. The "White channel," in the Klondike, has a grade of about 30 feet per mile. Many of the present Alaskan streams have grades of 100 to 150 feet per mile. In the principal Tertiary channels of Victoria, Australia, low grades down to 20 feet prevailed.

In depressed or elevated channels of past epochs, as in California, Victoria, and the Klondike, changes of original grade must be considered. This is best established in the California channels, which now have grades of 100 to 150 feet, whereas the original streams had much less, the increase being due to the westward tilting of the Sierra Nevada. The best results of gold concentration are probably obtained in rivers of moderate grades, perhaps 30 feet per mile, under nicely balanced conditions of corrasion and deposition. Whenever overloading and active deposition take place concentration of coarse gold ceases. On the other hand, where erosion is rapid conditions for rich placers are less favorable, unless, as in the present streams of the Sierra Nevada, the gold supply is unusually abundant.

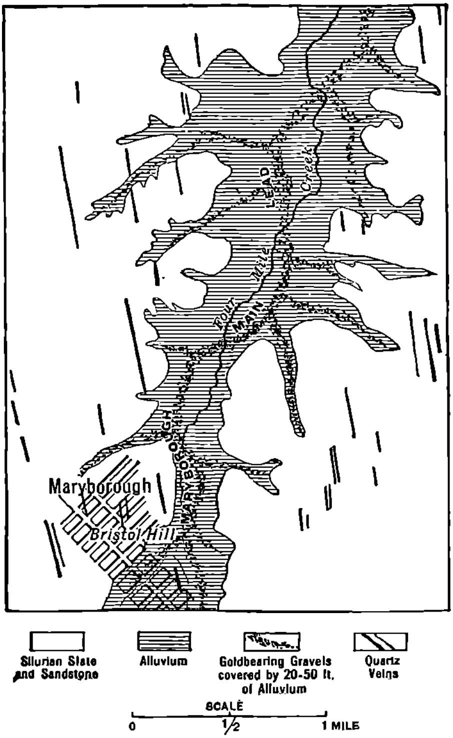

FIG. 100.—Map showing position of pay streak in alluvial gravels of Maryboro, Victoria. (*After S. B. Hunter.*)

**The Pay Streak or "Run of Gold."**[1]—Except in smaller creeks the distribution of the gold in a gravel bed is far from regular. There is usually gold on the bed-rock over the whole area of the stream bed; but the richer part makes a narrower streak which follows a devious course, distinctly affected by the character of the bed-rock, sometimes splitting and re-forming, following first one side, then crossing diagonally to the other side. It is not necessarily in the deepest depression or gutter. Figure 100 shows this devious course of the pay streak in comparatively

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[1] J. B. Tyrrell, The law of the pay streak in placer deposits, *Trans.*, Inst. Min. and Met., London, May 16, 1912; *Min. and Sci. Press*, June 1, 1912.

shallow gravels at Maryboro, Victoria. It is clearly independent of the course of the present small stream. In broader gravel plains, of which the Homebush and Pitsfield Tertiary placers of Victoria are examples, the "run of gold" follows a distinct and well-defined course on an almost level country rock. All this shows clearly enough the impossibility of the view that the gold was first uniformly distributed through the gravels and then gradually settled to the bottom under the influence of gravity. These pay streaks assuredly indicate epochs of well-balanced and long-maintained conditions during which the gravels could accumulate to only moderate depths and were at all times water-soaked and in a condition of slow movement. With more abundant loading of detrital material the gold-transporting power of the stream diminishes at a very rapid rate.

**Solution and Precipitation of Gold.**—Many of the earlier observers, such as Genth, Lieber, Selwyn, Laur, Egleston, C. Newbery, and Daintree, concluded from observations in various parts of the world that placer gold, particularly the large nuggets, has been deposited by circulating solutions. At present the mechanical derivation of the gold seems established beyond all doubt, although under exceptional circumstances some solution and redeposition may have taken place.[1] Even now, however, some writers, like J. M. Maclaren[2] and H. C. Boydell,[3] are inclined to place emphasis on this secondary and probably colloidal deposition. It is probable, nevertheless, that this process is absolutely insignificant from an economic point of view. Nuggets, when cut and polished, almost always show a granular structure perfectly in accordance with vein gold. Liversidge, in a long series of experiments, found only two specimens (both from New Guinea) which showed a concentric structure indicative of concretionary deposition. Very rare instances are quoted of quartz pebbles with dendritic films of gold[4] or of nuggets with minute gold crystals on their surface.[5] The collection of J. Edman, of San Francisco, contained a small crystal of magnetite coated with a thin film of gold. This came from the Tertiary deposits at Providence Hill, Plumas County, California; and Mr. Edman stated that he had never seen similar occurrences in the modern gravels. It seems to be well established that pyrite reduced by organic material in the gravels may contain some gold and also that the metal is occasionally found at the roots of trees or in the grass roots.

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[1] Regarding the older literature, see the text-books of Stelzner, Bergeat, and Beck. In more detail, see Liversidge, *Jour.* Roy. Soc. New South Wales, 27, 1893, p. 343; 31, 1897, p. 79; 40, 1906, p. 161.

[2] J. M. Maclaren, Gold, London, 1908, pp. 80–86.

[3] The rôle of colloidal solutions in the formation of mineral deposits, *Trans.*, Inst. Min. and Met., London, 1924, p. 205.

[4] R. G. McConnell, *Ann. Rept.*, Canada Geol. Survey, 14, 1905, *B*, 71 pp.

[5] H. A. Gordon, *Trans.*, Am. Inst. Min. Eng., 25, 1895, p. 294.

The gold crystallized in minute octahedrons in the clay of Kanowna, Western Australia, is, as Maitland[1] pointed out, immediately above or adjacent to the decomposed outcrops of the veins; and the occurrence can scarcely be called a placer. The gold which works down into the soft bed-rock of the placers is in all cases, where observed, of clearly detrital origin.

It is stated that the ashes of trees in the gold-bearing region of the Guianas contain an appreciable quantity of gold. Originally asserted by Lungwitz, this has been denied by Dubois and Kollbeck and then reasserted by Harrison on the basis of careful investigations.[2] From widely separate parts of the world gold has been reported in the ash of coal, but in this case it may be detrital and contained in admixed sand and clay.

Gold is easily brought into the colloidal state and as such it may be transported in solutions of colloidal silica. It is very readily precipitated by electrolytes and this mode of solution may account for some cases of secondary gold in placers. According to Emmons' investigations,[3] gold is soluble in superficial waters only when free chlorine becomes liberated by the interaction of sulphuric acid, sodium chloride, and manganese dioxide, a combination that must sometimes occur in ore deposits subject to oxidation; in the presence of oxidizing pyrite some gold may therefore be taken into solution, as chloride; but it would probably not remain long before encountering reducing substances. While gold is slightly soluble in sodium carbonate, sodium sulphide, and other similar compounds, these would not ordinarily be encountered in the waters of the zone of oxidation.

**Relation to Primary Deposits.**—That placer gold is directly derived by mechanical processes from vein deposits or analogous occurrences is absolutely certain, and examples of convincing character are present everywhere. This does not imply that the primary deposit can be worked at a profit. In most cases the placer gold is traceable up to the deposit. On this principle the pocket hunter proceeds, panning the detritus on the slopes and working up hill until the source of the scattered gold has been found. The area in which the detritus occurs has the shape of a triangle, the apex of which is the pocket.

It is a common experience that rivers or creeks crossing a vein or a mineral belt are enriched immediately below it, the coarseness of the gold increasing upstream to the place where the outcrops are crossed. As examples may serve the great accumulations of placer gold in the Neocene gravels of Eldorado County, California, where the Mother Lode

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[1] J. M. Maclaren, *op. cit.*, p. 83.

[2] J. B. Harrison, Geology of the gold fields of British Guiana, London, 1908, p. 209.

[3] W. H. Emmons, Enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917, p. 305.

crosses them, and the rich channels in upper Nevada County, just below the belt of quartz veins at Washington and Graniteville. There are fine examples in Victoria, where the gravels are rich only where they cross or follow systems of veins or "reef lines." The White channel of the Forest Hill divide, California, follows a belt of quartz stringers in clay slate. The Idaho Basin,[1] Idaho, presents an excellent instance of large gravel bodies the gold content of which is traceable up to certain auriferous vein systems.

**Economic Notes.**—The world's annual production of placer gold was about $25,000,000 in 1930. To this the Alaska and Yukon districts contributed $4,000,000, California $4,000,000, Victoria $500,000, New Guinea $800,000, and Siberia $13,000,000 (?).[2] While placers are found in almost all gold- and silver-producing regions, Brazil, the Ural Mountains, Siberia, California, Alaska, and Victoria have had by far the greatest total production.

Gold-bearing gravel is often measured by the ton, but more commonly by the cubic yard. Still another measure is by surface area, sometimes by the square foot, in Australia commonly by the square fathom; this is especially applicable to deep mining when only the richest bottom layer is mined; at least 2 feet of gravel and 1 foot of soft bed-rock are extracted, making one square fathom equivalent to a minimum of 4 cubic yards.

In river bars gravels are worked by wing dams and pits kept dry by simple pumping devices. On a large scale they may be ground-sluiced or washed by the hydraulic method, with the aid of elevators when the natural fall is insufficient.

The elevated gravels of earlier periods are worked in California by tunnels and drifting operations on the bed-rock. The minimum cost of working under the most favorable conditions is 50 cents per cubic yard, but is commonly $1 to $2 per cubic yard; most of the gravels actually worked contain at least $1.50, and often much more. The whole gravel body may be washed by the hydraulic method, when the expense may be reduced to 2 to 5 cents per cubic yard; of course, the cost of preliminary work like ditches, etc., is often great. Some creek gravels in the Seward Peninsula, Alaska, contain from $2 to $6 per cubic yard; the width of the deposit may be about 50 feet, the depth 3 to 6 feet.

The depressed gravels of earlier periods are worked by drifting from shafts, as in Victoria, where, however, the preliminary pumping, to permit access, is an extremely heavy expense, often indeed prohibitory. Some of these Australian channels have been extremely rich, the workable portions ranging from $2 to $15 per cubic yard. Some of the channels

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[1] W. Lindgren, The mining districts of Idaho Basin and the Boise Ridge, *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, pp. 617–744.

[2] The decline is marked since 1910, when Alaska and the Yukon produced $20,000,000, California $9,000,000, Victoria $2,000,000.

are in places several hundred feet in width. In later years production has declined steadily.

Of late, gravels have been extensively worked in California, Alaska, the Klondike, and elsewhere by the dredging process. In California, where this method has reached its highest development, $7,769,000 was obtained from 58 dredges in 1916 from the flood-plains of the rivers at the foot of the Sierra Nevada; and the cost has been reduced from about 10 cents to 3 or 4 cents per cubic yard handled. In 1930 this amount had been reduced to about $3,600,000. In Alaska the cost is, of course, much higher; but gravels containing 17 cents per cubic yard are workable under favorable conditions. The dredge will probably prove to be the most efficient placer-mining machine of the future, replacing the hydraulic method which offers difficulties in the disposition of the tailings. In 1917 Alaska yielded $2,500,000 from dredging; in 1931 more than $3,000,000, or about 80 per cent of total placer products.

The largest dredging operations at present are conducted by the Fairbanks Exploration Company, at Fairbanks, Alaska, where five dredges, with 10-cubic foot buckets, are working gravels 50 to 80 feet thick. It is necessary to thaw the frozen ground in advance by circulation of cold water.

Certain gravels in the dry regions of Arizona and northern Mexico are treated by pneumatic concentration in so-called dry washers, but the output of these placers is insignificant.[1]

Yields of placer deposits are often calculated in dollars per lineal foot of channel. Good channels for drifting may produce from $70 to $500 per foot. The richest drift mine worked was probably "Madame Berry" in Victoria, with average width of 450 feet, yielding $1,293 per foot along channel. The two claims below this produced, respectively, $843 and $443 per foot, the last-named channel being mined 1,000 feet wide. The "White channel" in the Klondike gave $380 per foot; the Red Point channel in Placer County, California, $72, the width being 120 feet; the American Hill hydraulic mine, Nevada County, 1,000 feet wide, $414; the Nome creeks, Alaska, 50 feet wide, about $100. By drifting operations alone, only a part of the gold will be extracted, say one-fifth to one-half, dependent upon the thickness of overlying gravels.

At present, Siberia appears to contain the largest reserves of placer ground, divided between the Urals and western Siberia and eastern Siberia, including the Amur and the Lena drainage basins. Before the World War these areas, particularly the last named, yielded large productions, the total for 1913 being about $40,000,000, of which $34,500,000 came from placers. After the Russian Revolution the output again gradually increased, until in 1930 it was estimated at $19,000,000, of which $13,000,000

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[1] F. J. H. Merrill, *Min. and Sci. Press*, July 13, 1912.

came from placers.[1] The larger part of this came from the Lena drainage. The deposits at Bodaibo had a high production. Other placers are found on the Aldan, Kolar, Vitim, and Kolim rivers in the same region. According to Obrutchev[2] it is a glaciated mountain region, with summits rising 2,500 feet above the valleys. The rocks are pre-Cambrian limestone and crystalline schists, with large areas of intrusive granite. Quartz veins are abundant but do not appear to contain gold. The deposit consists of pre-glacial, eluvial-alluvial gold placers and heavy interglacial, partly auriferous sands and gravels. The workings are now in upstream terrace gravels and also in recent alluvium and low terraces.

### THE GOLD-BEARING CONGLOMERATES OF SOUTH AFRICA

Of the extensive literature the following principal papers are quoted:

G. F. Becker, The Witwatersrand banket, etc., *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 5, 1896.

F. H. Hatch and G. S. Corstorphine, The Geology of South Africa, London, 1905.

J. W. Gregory, The origin of the gold in the Rand banket, *Trans.*, Inst. Min. and Met., London, 17, October, 1907. Also *Econ. Geol.*, 4, 1909, pp. 118–129. Discussion, *idem*, by G. F. Becker and G. A. Denny.

R. B. Young, The Rand banket, London, 1917, 125 pp. Best description.

C. B. Horwood, The Rand banket, *Min. and Sci. Press*, October to December, 1913. Also, Gold deposits of the Rand, 1917.

E. T. Mellor, The upper Witwatersrand System; the East Rand, *Trans.*, Geol. Soc. South Africa, 18, 1915, pp. 11–71.

E. T. Mellor, The conglomerates of the Witwatersrand with discussion, *Trans.*, Inst. Min. and Met., London, 25, 1916, pp. 226–348.

Hugh F. Marriott, Mining on the Rand, *Trans.*, Inst. Min. and Met., London, 27, 1918; *Min. and Sci. Press*, July 20, 1918.

A. L. Du Toit, The geology of South Africa, London, 1926.

L. Reinecke, The location of payable ore-bodies in the gold-bearing reefs of Witwatersrand, *Trans.*, Geol. Soc. South Africa, 30, 1928, pp. 89–119.

L. C. Graton, Hydrothermal origin of the Rand gold deposits, Pt. 1, *Econ. Geol.*, 25, 1930; Supplement to No. 3, pp. 1–185.

This paper constitutes the most complete presentation of facts in favor of the hydrothermal origin. It was followed by papers by Reinecke, Rogers, Keep, Gregory, Homersham, Young, Mellor, Macadam, Cowles, and Cameron, all of whom vigorously deny the hydrothermal introduction of the gold. See *Trans.*, Geol. Soc. South Africa, 33 and 34, 1930 and 1931. Abstracts in Annotated Bibliography of Economic Geology, 3 (2), 4 (1), and 4 (2).

The development of the gold-bearing conglomerates of the Witwatersrand district, in the Transvaal, is one of the most wonderful chapters in the history of mining. From an inconspicuous beginning in 1887, the production of these unique deposits has steadily increased. The

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[1] V. M. Karmashov, The gold mining industry in the Soviet Republic, *Eng. and Min. Jour.*, Aug. 10, 1931, p. 113.

[2] V. A. Obrutchev and A. P. Gerasimov, Geologic map of the Lena gold region, Sheets $VI_1$ and $VI_2$ Com. geol., Leningrad, 1929; abstract, *Geol. Zentralbl.*, 43 (6), 1931, p. 379.

increase in production continued to 1916; after a slight decline the output of gold is again increasing. In 1931 the ore production amounted to about 32,000,000 tons, with a record yield of $225,000,000 which is over one-half of the total production of the world for that year. The total production to the end of 1930 exceeded $4,500,000,000, which is more than the total gold production of California, Colorado, and Alaska. The average content of the ore has decreased, probably mostly on account of reduction in mining and metallurgical costs, from $12 to $6 or $7 per ton. It is possible that in the future ore of $5 per ton will be utilized. It is thought that a slow decline in the production will soon begin. A depth of 8,000 feet has now been reached in the Village Deep Mine, and, owing to a favorable geothermic gradient (p. 101), it will be possible to go considerably deeper. Many of the mines exceed 5,000 feet in depth. The ore is first crushed in stamp mills, then ground in tube mills and the pulp concentrated on corduroy blanket tables; the concentrate is amalgamated and the tailings treated by the cyanide process.

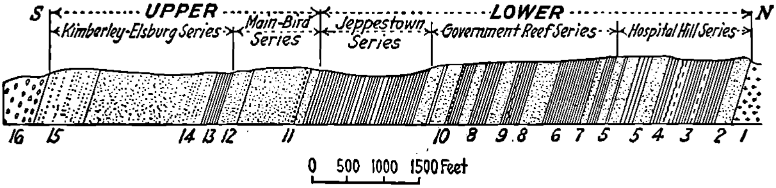

FIG. 101.—Section of the Witwatersrand system on the central Rand. 1, Granite; 11, Main Reef; 12, Bird Reef; 14, Kimberley Reef; 16, Amygdaloid. (*After E. T. Mellor.*)

South Africa is in the main a plateau of sedimentary beds resting on a basement of granite and crystalline schists.

The oldest rocks known are the Swaziland crystalline schists and the granites intruded in them. On their eroded surface rest the upper and lower Witwatersrand system of slates, quartzites, and conglomerates (Fig. 101), attaining a maximum of 25,000 feet in thickness; and on top of these in turn a thick series of volcanic flows, called the Ventersdorp system.

The age of the Witwatersrand system is not definitely known; it is probably Cambrian or pre-Cambrian. Next higher in the succession of rocks is the Transvaal system, including the Black Reef (oldest), Dolomite, and Pretoria series. This is covered by the Waterberg system and this again by the definitely Devonian Cape system (Table Mountain sandstone of the Cape) and the most recent Karroo system, which is coal-bearing and considered to be of Permo-Carboniferous age. Each system is separated by an unconformity from the next.

The Witwatersrand system is folded in a syncline extending about 120 miles east to west and 45 miles north to south. At Johannesburg, on the north side of the syncline, the dip is to the south, steep near the surface, but flattening in depth to about 30°. Faulting is common and there are a number of intrusive diabase dikes, thought to belong to the overlying Ventersdorp volcanic system.

Auriferous conglomerates occur at several horizons in the Witwatersrand system and also in the Black Reef series. The productive beds are, however, in the upper part of the Witwatersrand, including a thickness of about 7,000 feet of quartzites and conglomerates, among which the following are distinguished, beginning from the top: Kimberley group, Bird Reef group, Livingstone Reef group, and Main Reef group. The first two are each about 500 feet thick but the conglomerates contained are of low grade, rarely exceeding $3 per ton in gold. The Main Reef group, about 90 feet thick, includes several conglomerate beds more or less persistent. The position of the Main Reef group with reference to overlying beds is seen in Fig. 101.

The usual subdivision of the Main Reef group includes from top to bottom:

South Reef (3 feet).
Quartzite (20 to 80 feet).
Main Reef Leader (2 to 6 feet).
Quartzite (2 to 20 feet), including in places the "Bastard Reef" (scattered pebbles).
Main Reef (4 to 8 feet).

Of these the Main Reef Leader is the most productive and persistent member. The pebbles in the conglomerate are small, averaging 2 inches in diameter, and consist of well-rolled fragments of glassy quartz with a few pebbles of more angular quartzite, banded chert, and slate. The pebbles lie in a matrix of sandy material, which has become hardened by infiltration of silica. Other detrital minerals are iridosmine, zircon, and chromite. There is no magnetite. Pyrite occurs in abundance in the cement, averaging about 3 per cent of the rock. It is present both in crystalline form and as rounded colloform replacements after quartz probably representing two generations, both subsequent to the sedimentation. The latter may be marcasite. Chloritoid,[1] chlorite, sericite, calcite (scarce), rutile, tourmaline (scarce), and carbon (in grains and veinlets) are other authigenetic minerals. The gold is not contained in the pebbles, but only in the cement, and forms minute angular crystalline aggregates, apparently never rounded particles. It is usually closely connected with pyrite, either enclosed by it or covering the surface of pyrite aggregates. As a rule it is not visible to the naked eye.

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[1] A prismatic colorless mineral usually described as chloritoid is common (Fig. 102), but its identification appears questionable.

In spite of a long-continued discussion there is no unanimity among geologists as to the genesis of these remarkable deposits.

The first suggestion that the conglomerate may be simply an alluvial or littoral placer is refuted by the character of the gold and its close association with the pyrite and the absence of magnetite and ilmenite. It is clear that if this is a placer deposit there have been extensive recrystallization and some migration.

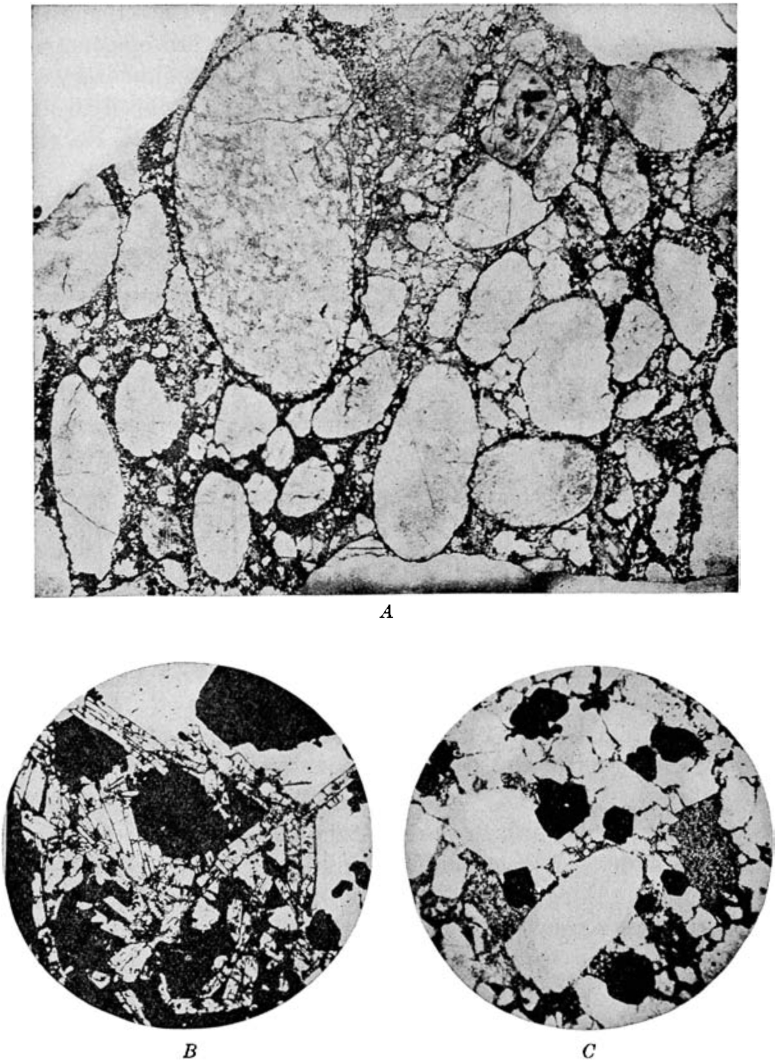

Fig. 102.—*A*, section of Witwatersrand banket. Four-fifths of natural size. *B*, pyrite replacing quartz and molded on chloritoid. Magnified 56 diameters. *C*, matrix of banket showing crystals of pyrite, replacing quartz grains. Magnified 12 diameters. (*A, B, and C after R. B. Young.*)

The advocates of the placer theory, among whom are G. F. Becker, J. W. Gregory, G. A. Denny, R. B. Young, E. T. Mellor, and many others, are compelled to admit a recrystallization of the gold and a transformation of magnetite and ilmenite into pyrite.

Many geologists and engineers, impressed with the difficulties confronting the placer theory, hold that the deposits are epigenetic—that the gold and pyrite have been introduced by a post-sedimentary infiltration, perhaps after the intrusion of the diabase dikes. This hydrothermal theory is held by H. Louis, J. H. Hammond, R. Beck, F. H. Hatch, G. S. Corstorphine, C. B. Horwood, and L. C. Graton.

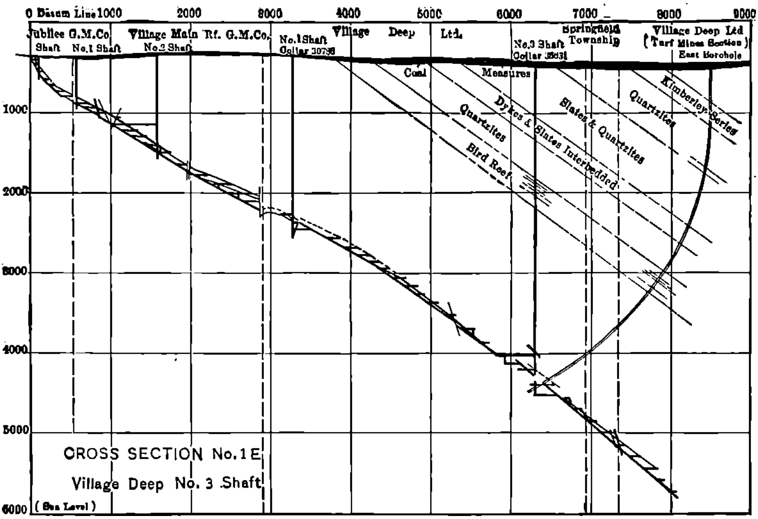

FIG. 103.—Section through Village Deep No. 3 shaft. (*After H. F. Marriott.*)

Small and irregular quartz veins which in some places contain a little gold and sulphides intersect the Witwatersrand series. Whether there is any enrichment along the few diabase dikes is a disputed question. No doubt these quartz veins are related to the dikes.

E. T. Mellor has approached the subject from the wider geological viewpoint and his papers contain very strong arguments in favor of the original deposition of the gold in alluvial gravels. He considers the quartzite and conglomerate series as large delta deposits rather than shore gravels and shows the existence of many horizons of gold-bearing conglomerates.

Against the hydrothermal theory stands a long array of strong arguments: (1) The absence of channels followed by the solutions; (2) the regular distribution of the gold in the conglomerate; often it is concentrated in its lower layers; (3) the practical confinement of the gold to the conglomerates, though the quartzites are equally permeable; (4) the conglomerates were deposited in an alluvial plain skirting the deeply eroded Swaziland schists with their lenticular gold-quartz veins and would thus certainly contain some gold.

There are many strong arguments supporting the placer theory. The conglomerates were deposited in a plain, skirting the deeply eroded Swaziland schists and granites, with their abundant gold-quartz veins. The deposits may be alluvial, deltaic, littoral, or even (possibly) submarine, but they must have contained detrital gold and black sand, and the gold must have been in minute particles. The presence of water-worn particles of iridosmine is demonstrated by Young.[1] If the placers contained water-worn iridosmine they must have contained a great deal more detrital gold, whatever their present condition may be.

Practically only the conglomerates are now gold-bearing, though the sands were also permeable. There are in the Witwatersrand series and in the overlying Transvaal system a great many auriferous conglomerates, generally of low grade. The ore-shoots are large, irregular, and patchy, very rarely presenting any similarity to the well-known forms shown in veins.

If the gold were of hydrothermal origin it would seem that it must have been deposited in other structural planes as well as in the pebble beds; and this has been found only in small quartz veins near the banket and near diabase dikes. The uniform testimony seems to be that such veins are of low grade.

On the other hand, evidence for hydrothermal action is also strong. The conglomerates have been deeply buried with attendant, rather high temperature. Pyrite, sericite, chlorite, rutile, chloritoid (?), and, in places, tourmaline have been introduced. There is much secondary quartz in veinlets in and near the ore. A great deal of sulphur has been introduced; the gold now shows no trace of detrital rounding but has evidently been recrystallized. We must admit that black sand was once present in the deposits and 2 to 3 per cent does not seem unreasonable. Volcanic activity during and since Witwatersrand time has been intense and gold-bearing veins of the hydrothermal type occur at Pilgrims Rest and other places in the Transvaal. It seems probable that much water, partly or wholly of magmatic origin, certainly containing hydrogen sulphide and alkaline sulphide and possibly gold and other metals, must have accompanied these igneous outbursts. These waters found their way into the auriferous conglomerates and there transformed the iron oxides to sulphide and added some gold to that of detrital origin, but they were never rich enough to transform the quartzites into ore. The very fine detrital gold may have been dissolved by the sodium sulphide

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[1] In 1930 several thousand ounces of iridosmine were saved at the mills.

undoubtedly present in the thermal waters and rapidly re-precipitated by the pyrite.

The conclusion might, therefore, be justified that the Rand banket is a detrital gold deposit slightly enriched by hydrothermal processes.[1] The difficulty, not yet fully explained, lies in the abnormal richness and extent of the conglomerates. It is pointed out, however, that large areas of the conglomerate are practically barren. The rich beach sands of Nome, Alaska, have been cited as an analogous case but the analogy is by no means perfect.

Similar conglomerates of considerable geological antiquity are found in West Africa at Tarkwa and Abosso[2] and these have been worked on a fairly large scale. Instead of pyrite these contain hematite, and ilmenite with chloritoid or ottrelite.

## PLATINUM PLACERS[3]

It is known that platinum occurs as a primary constituent of peridotites (p. 778), and specimens showing its intergrowth with olivine and chromite have been described. Almost the entire world's production is obtained from placers and 95 per cent of it was formerly extracted from the placers on the eastern slope of the Ural Mountains, where detrital platinum occurs in the gravels of the stream courses, which head in certain Paleozoic intrusions of peridotite and pyroxenite, in part altered to serpentine. It is associated with iridosmine, iridium, chromite, and often also with gold. The crude platinum forms small rounded grains, very rarely nuggets up to 20 pounds in weight; and its fineness (per thousand) ranges from 750 to 850, the remainder being iron, copper, and various metals of the platinum group, particularly iridium. In part the placers are worked by dredging.

Platinum-bearing gravels occur also in the Choco district Colombia,[4] South America, in river beds and Tertiary conglomerates, but the pro-

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[1] It seems to be incumbent on the advocates of the placer theory to find some conglomerates which contain no pyrite but which do contain detrital gold. The size of the iridosmine grains range up to 1 millimeter, and therefore the detrital gold where preserved should be rather easily distinguished. In other words, it was fine but not so fine as to be classed as *flour gold.*

[2] O. A. L. Whitelaw, The Tarkwa-Abosso goldfield, *Mem.* 1, Gold Coast Geol. Survey, 1929. Gold in the banket is considered detrital.

A. Bray, Notes on the banket reefs of the Gold Coast Colony, *Trans.*, Inst. Min. and Met., London, 38, 1929, pp. 21–69. Bray suggests a hydrothermal origin. During discussion of the paper many dissented from this view.

[3] J. F. Kemp, Geological relations and distribution of platinum and associated metals, *Bull.* 193, U. S. Geol. Survey, 1902.

Louis Duparc, Le platine et les gîtes platinifères de l'Oural, Genève, 1920.

C. W. Purington, The platinum deposits of the Tura River System, Ural Mountains, *Trans.*, Am. Inst. Min. Eng., 29, 1899, pp. 3–16.

[4] *Mining Mag.*, London, 33, 1925, p. 135.

J. G. Kellner, Die Platinlagerstätten Kolumbiens, *Zeitschr. prakt. Geol.*, 36, 1928, pp. 1–7.

duction from them is stationary. The deposits are worked by dredges.[1] A considerable production is reported from Ethiopia. In the United States the metal occurs in small quantities together with gold in almost all the gold-bearing districts in northern and central California and in southwestern Oregon, where serpentine or peridotite is found. In 1931, 446 ounces of crude platinum were reported from the United States and Alaska, chiefly from the black sands of the dredges. Platinum also occurs in the beach sands of southern Oregon, together with more or less gold; a small quantity of this is recovered. The Tulameen district, British Columbia, formerly yielded some production. The normal world's production of crude platinum before the World War was about 300,000 troy ounces, but it is now much less—about 150,000 troy ounces in 1930. At least 100,000 ounces came from Russia and 42,382 ounces from Colombia. For a long time the price of platinum was less than that of gold; a gradual rise increased the value to $20 per ounce, and in 1911 it reached $45. Increasing scarcity forced the price up to $120 in 1925, only to drop again to $60 in 1927 and to $24 in 1931, but recovered before the end of the year to $40. Crude platinum with 70 to 85 per cent Pt is sold from $30 to $40 per ounce.

Placers yielding osmiridium at the rate of up to several thousand ounces per annum have been worked in Tasmania since 1914.[2] They occur in peridotite and serpentine, which also contain osmiridium and a little gold.

### CASSITERITE PLACERS[3]

The original home of cassiterite ($SnO_2$) is either in the granites, in pegmatite dikes, or in quartz veins. From any of these sources it may be set free by weathering and disintegration, and, on account of its high specific gravity, it easily becomes concentrated in gravel deposits of different types. Among the accompanying minerals, tourmaline, topaz, and wolframite are the most common. Grains of metallic tin are reported to occur with cassiterite in Nigeria and Australia. Eluvial deposits immediately below the outcrops are numerous and are worked

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[1] James W. Neil, Recovery of platinum in gold dredging, *Min. and Sci. Press*, Dec. 8, 1917.

[2] A. McIntosh Reid, Osmiridium in Tasmania, *Bull.* 32, Geol. Survey Tasmania, 1921.

[3] H. W. Kayser and R. Provis, The Mt. Bischoff tin mine, *Proc.*, Inst. Civil Eng. (London), 123, 1896, pp. 377–387.

O. H. Van der Wyck, The occurrence of tin ore in the islands of Banca and Billiton, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, 1896, pt. 3, pp. 227–242.

F. D. Adams, Tin Mining in Malaya, *Trans.*, Canadian Inst. Min. and Met., 31, 1929, pp. 115–146.

L. C. Graton, Reconnaissance of some gold and tin deposits of the southern Appalachians, *Bull.* 293, U. S. Geol. Survey, 1906.

W. R. Jones, Tinfields of the world, London, 1925.

on a large scale at Mount Bischoff, in Tasmania. A small deposit of this kind resting in a shallow gully immediately below a pegmatite dike was mined near Gaffney, South Carolina, in 1905. The earliest production of stream tin came from gravels below the tin-bearing lodes of the Erzgebirge, Saxony, and of Cornwall, both sources now practically exhausted. About 70 per cent of the world's production of about 174,000 (1930) short tons of tin is still obtained from alluvial placers, mainly in the Malay Peninsula (see p. 644) and the islands of Banka and Billiton, near Sumatra, and in Nigeria and China. The Malay peninsula furnished 60 per cent of the world's production from gravels averaging 0.5 pound of cassiterite to the cubic yard. New South Wales and Victoria furnish minor amounts. In the latter state some cassiterite is saved in working Pliocene auriferous stream channels. In this case the tin ore appears to be sparsely disseminated in granite and is liberated after its disintegration. At the Briseis mine, in Tasmania (now closed), the deposit worked consisted of 14 to 45 feet of river gravel, covered by 20 to 40 feet of decomposed basalt and containing from 2 to 4 pounds of cassiterite per cubic yard.

In the United States small amounts of stream tin are recovered in Alaska near the extreme western point of the American continent, in the Black Hills of South Dakota, and in North and South Carolina.

As tin is worth from 25 to 90 cents per pound and the easily reduced cassiterite contains 78.6 per cent of the metal, it is clear that a small quantity, say 0.5 pound per cubic yard of gravel, might suffice for profitable working. The placers in the Malay states are worked by sluicing, by dredges, and by gravel pumps.

## MONAZITE PLACERS[1]

Monazite, an anhydrous phosphate of cerium, lanthanum, and other cerium metals, usually contains also from 3 to 8 per cent of thoria, making it valuable for the production of nitrate of thorium, which is utilized in the manufacture of incandescent gas mantles.

The mineral has a specific gravity of 5.203, a resinous luster, and a yellow to brown color; when occurring in placers it is found together with gold, zircon, magnetite, ilmenite, garnet, etc., after concentration in sluices. From its associated minerals it is cleaned in electromagnetic separators, the final product being about 90 per cent pure. The source of the monazite is in the granites, gneisses, and pegmatites, where it occurs as a primary mineral. As its value (changing with the percentage of thoria) is about 8 cents per pound, monazite gravels may in places form workable deposits, especially where, as often happens, gold is present.

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[1] J. H. Pratt and D. B. Sterrett, Monazite and monazite mining in the Carolinas, *Trans.*, Am. Inst. Min. Eng., 40, 1909, pp. 488–511.

Monazite is now obtained from marine and fluviatile placers in Brazil and India, but it also occurs in similar deposits in North and South Carolina. It occurs in Idaho, where a large intrusive batholith of granite or quartz monzonite evidently carries the mineral sparsely distributed throughout. The principal occurrence in Idaho is at the old placer district of the Idaho Basin. In 1910 about 100,000 pounds of monazite were mined in the United States, chiefly from placer deposits in the Carolinas. Since 1910 there has been no production in the United States, the supply (600 tons in 1929) being obtained from the beach deposits in Brazil, India, and Ceylon. At Travancore, India, much ilmenite is also recovered.

## OTHER PLACERS

Magnetite, or "black sand," has been frequently mentioned above as a product of concentration in gravels and sands and is usually derived from the disintegration of igneous rocks. Along the beaches and the bars of some rivers it may accumulate in considerable masses—for instance, on the lower St. Lawrence River, Canada, and along the Columbia River, Oregon—but it is exceptional that such deposits have been utilized.[1] More or less ilmenite is usually mixed with the magnetite.

There are several examples of eluvial deposits of iron ore (magnetite, hematite, or limonite), formed below outcrops of iron deposits, and also of such detrital masses in the débris slopes of older formations. At Iron Mountain, Missouri,[2] Paleozoic rocks rest upon a deposit of boulders of iron ore and porphyry, which in turn lie upon pre-Cambrian porphyry. The porphyry itself also contains deposits of hematite (Fig. 104).

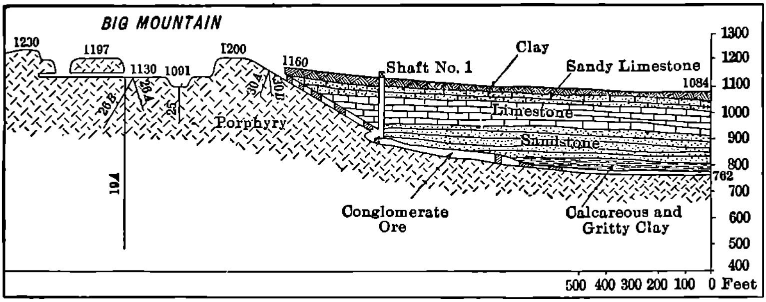

FIG. 104.—Section of Iron Mountain, Missouri, showing mining of detrital ore underneath Paleozoic limestone and sandstone, and of hematite ore in the porphyry. (*After G. W. Crane.*)

Similar eluvial masses of copper and lead ores are found in places. We may recall the great débris mass of chalcocite below the outcrops of

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[1] The magnetite sands of Japan appear to have been rather extensively utilized; also those occurring along the coast of New Zealand.

[2] G. W. Crane, The iron ores of Missouri, Geol. Survey Missouri, 2d ser., 10, 1912, pp. 107–145.

the Bonanza mine[1] in the Copper River region, Alaska, and galena beds on the slopes below the Elkhorn mine, Wood River, Idaho.[2]

## PRECIOUS STONES

Diamond placers are of considerable importance, and until 1871 the whole world production came from such sources in India and South America. The parent source is probably always a basic igneous rock (p. 774). In small amounts diamonds are widely distributed. In the gold belt of California many small diamonds have been found in the gold-bearing gravels. Scattered diamonds have been found in the glacial drift areas in Indiana and Ohio. In India they occur in the gravels of the Deccan mines, now practically unproductive. In British Guiana they are mined from the gravels of the Mazaruni River, original source unknown. In Brazil, State of Minas Geraes, they occur in sands or gravels derived from conglomerates; here also the original source is not determined. The black *carbonado* has a granular texture, without visible cleavage, and is used for diamond drilling. It is mined from river gravels in the State of Bahia, Brazil, and has considerable economic importance.

Most of the diamond placers are in South Africa, and the yield is large and valuable.[3] For many years diamond washings have been operated on the Vaal River and along the west coast at Lüderitz Bay and other places. In 1926 a great rush took place to the Lichtenburg placers in western Transvaal, and in 1927 a similar excitement carried thousands of people to other placers near the mouth of the Orange River, on the west coast, where the diamonds occurred in rich shore terraces, in several horizons identified by certain shells. The diamond-bearing ground probably extends for 200 miles south of that river along the sea shore. The original source is surely in peridotite-pipes of the interior similar to those of Kimberley (p. 775). In 1929, the production of this district was 900,000 carats; 1,000,000 carats came from the Lichtenburg placers. The Belgian Congo yielded 1,000,000 carats, and Angola 200,000. Since 1930 the output has dropped.

Some placers yield precious stones; most of the rubies, sapphires, zircons, spinels, aquamarines, chrysoberyls, etc., are washed from gravels and occur in regions of highly metamorphosed rocks intruded by granite and pegmatite. The island of Ceylon[4] is rich in such deposits and has

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[1] F. H. Moffit and S. R. Capps, *Bull.* 448, U. S. Geol. Survey, 1911, p. 89.

[2] W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, p. 210.

[3] A. F. Williams, Diamond-bearing alluvial gravels of the Union of South Africa, Third Empire Mining and Metallurgical Congress, South Africa, 1930.

P. A. Wagner and H. Merensky, The diamond deposits on the coast of Namaqualand, *Trans.*, Geol. Soc. South Africa, 31, 1928, pp. 1–41.

E. Kaiser, Die Diamantenwüste S. W. Africa, Berlin, 1926, 2 vols.

E. Reuning, Der Ursprung der Küstendiamanten, *Neues Jahrbuch*, Beil. Bd. 64, 1931, pp. 775–828.

[4] Frank D. Adams, A visit to the gem districts of Ceylon and Burma, *Bull.* Canadian Inst. Min. and Met., February, 1926.

been renowned for these occurrences since ancient times. Their original source has not always been determined, but they probably came from pegmatites. A zircon, which is colorless, but turns blue upon heating, occurs in granite and gneiss near the south point of the island. The blue sapphires occur in several districts between Kandy and the south shore. They occur in gravel lying at the base of alluvial accumulations. The deposits are worked on a small scale and gems to the value of a few hundred thousand dollars are extracted annually.

The ruby mines of Mogok, Burma,[1] are likewise of ancient renown. The gems occur in gravel on a tributary of the Irawaddy River, and their source is a crystalline limestone intruded by granite and alkaline rocks.

Sapphires of pale blue or yellow color occur in the gravel bars of the Missouri River,[2] near Helena, Montana. This deposit was worked for several years. According to G. F. Kunz, the gems are derived from dikes of a mica-augite andesite, but the primary deposits are not worked (p. 797).

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[1] Frank D. Adams, *idem.*

[2] D. B. Sterrett, *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1910, p. 877.

## CHAPTER XVIII

# DEPOSITS PRODUCED BY CHEMICAL PROCESSES OF CONCENTRATION IN BODIES OF SURFACE WATER BY REACTIONS BETWEEN SOLUTIONS

### CHANGES DURING SEDIMENTATION

For tracing the chemical changes which take place during the transformation of igneous rocks to sediments the fine silts and muds deposited near the shore are particularly interesting, for they form about 95 per cent of the sediment veneer of the crust.

Below is arranged side by side the average percentage composition of all igneous rocks, of the silts of the Mississippi delta, and of the blue and green "terrigenous" marine muds.

The analyses given below carry considerable weight, for each column is a composite of a large number of samples. It may be of interest to compare them in detail. After the cooling of the igneous rocks come decomposition, weathering, erosion, and sedimentation, all processes involving a great amount of solution and mechanical transportation. Of the original rock the larger part remains in analyses 2 and 3. All of the original constituents have, of course, been more or less leached and removed.

The most striking gain in comparing 1 and 2 is in silica and represents mainly mechanical concentration of quartz. Quite decided are the gains in chlorine, fluorine, and sulphur, suggesting that if by some chance such sediments were assimilated by magmas they would not be lacking in mineralizers.

Among the losses all the common rock-forming bases are conspicuous, particularly alumina. There is a relative loss of magnesia and lime both to the extent of 57 per cent of the amounts in the original rock. Soda loses 61 per cent, or more than any other of these constituents; potash shows a loss of only 27 per cent. Losses prevail among most of the rarer metals; but, of course, their expression in percentages would not mean much.

The comparison between 1 and 3 brings out the gain in alumina and water, which, of course, was to be expected; but there is, also, a very decided gain in iron oxides, and in sulphur (fluorine and chlorine not determined). Chromium is almost constant; quite certainly there is an apparent increase in vanadium, nickel, barium, strontium, copper, and zinc, while lead diminishes. These facts are of considerable interest in connection with the biologic processes referred to later. Magnesia, lime, and alkalies show losses similar to those in the silts, less potash being lost than soda.

COMPARATIVE RESULTS OF SEDIMENTATION

| | 1 | 2 | 3 |
|---|---|---|---|
| $SiO_2$ | 59.12 | 69.96 | 57.05 |
| $Al_2O_3$ | 15.34 | 10.52 | 17.22 |
| $Fe_2O_3$ | 3.08 | 3.47 | 5.07 |
| $FeO$ | 3.80 | | 2.30 |
| $MgO$ | 3.49 | 1.41 | 2.17 |
| $CaO$ | 5.08 | 2.17 | 2.04 |
| $Na_2O$ | 3.84 | 1.51 | 1.05 |
| $K_2O$ | 3.13 | 2.30 | 2.25 |
| $H_2O$ | 1.15 | 3.78−<br>1.96+ | 7.17 |
| $CO_2$ | 0.102 | 1.40 | ............ |
| $TiO_2$ | 1.050 | 0.59 | 1.27 |
| $ZrO_2$ | 0.039 | 0.05 | ............ |
| $P_2O_5$ | 0.299 | 0.18 | 0.21 |
| Cl | 0.048 | 0.30 | ............ |
| F | 0.030 | 0.07 | ............ |
| S | 0.052 | 0.07 | 0.13 |
| $SO_3$ | ............ | 0.03 | ............ |
| $(Ce, Y)_2O_3$ | 0.020 | ............ | ............ |
| $Cr_2O_3$ | 0.055 | 0.01 | 0.05 |
| $V_2O_3$ | 0.026 | 0.02 | 0.03 |
| $MnO$ | 0.124 | 0.06 | 0.12 |
| $NiO$ | 0.025 | 0.017 | 0.063 |
| $BaO$ | 0.055 | 0.08 | 0.06 |
| $SrO$ | 0.022 | tr | 0.03 |
| $Li_2O$ | 0.007 | ............ | ............ |
| Cu | 0.010 | 0.0043 | 0.0120 |
| Zn | 0.004 | 0.0010 | 0.0053 |
| Pb | 0.002 | 0.0002 | 0.0004 |
| $As_2O_5$ | ............ | 0.0004 | trace |
| Organic Matter | ............ | 0.66 | ............ |
| Carbon | ............ | ............ | 1.69 |
| Totals | 100.000 | 100.6229 | 99.9907 |

1. Average composition of igneous rocks. Clarke and Washington, *Prof. Paper* 127, U. S. Geol. Survey, 1924.

2. Mississippi delta silts, 235 samples. Geo. Steiger, Analyst. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 509.

3. Marine terrigenous clays, 52 samples. Geo. Steiger, Analyst. Clarke, *idem.*, p. 508.

It may be argued that these analyses do not represent the direct change from igneous rocks to silt and clay for the material represented in columns 2 and 3 is undoubtedly largely derived from sedimentary rocks. This does not lessen their value for comparison; it simply emphasizes more the changes by successive sedimentary cycles. That these comparisons in many instances are not far from indicating the actual gains and losses would appear from the calculations made by Leith and Mead.[1] These investigators point out quite correctly that during sedimentation considerable gains are made of water, carbon dioxide, oxygen, and other constituents derived from the atmosphere and the hydrosphere. Their main result is that 100 grams of igneous rock yield 114 grams of end products consisting approximately of 87.8 grams of shale, 12.9 grams of sandstone, 6.7 grams of limestone, and 6.6 grams of ocean salts.

Total weight of sediments derived from 100 grams of igneous rock is therefore 107.4 grams.

Expressing it in another way, 100 grams of igneous rocks contribute approximately 97 grams to the sediments, the remaining 3 grams going to the salts of the sea.

**Biochemical Processes.**—Plants and animals play an intense and dramatic part in the concentration of certain elements. Biochemical processes involve the concentration of certain elements. They, also, involve a certain amount of dissipation during life. After the death of the being, dissipation is the rule unless counteracted by certain other processes. The living matter collects elements, which it needs; after death it gives them up to the media in which it lived.

Often we observe consecutive stages in concentration. In the sea water the blue-green algæ concentrate phosphorus, certain mollusks, or crustaceans feed on the algæ, and other meat-eating mollusks devour the vegetarians. Small fishes eat the mollusks, large fishes eat the small, finally seals and birds swallow the fishes, and so in about six transformations the phosphorus originally contained in the sea water may come to rest in deposits of guano on desert islands or in accumulations of bones of vertebrate denizens of the sea.

Plants are continually accumulating carbon from the atmosphere. The carbon is returned by decay, or under protective cover is indefinitely preserved until the coal seams, exposed to the air, slowly oxidize or struck by lightning burn to red bands along the hillside, or again man steps in and by combustion returns vast amounts to the atmosphere. Plants also accumulate potash, phosphorus, nitrogen, silica, and calcium carbonates; their chlorophyll demands iron and magnesium. The calcium carbonate of algæ, the silica of diatoms may accumulate to important deposits. The ashes of plants give us a clue to many of these complex processes. Take sea-weed, for instance; in its ashes have been found potassium, iodine, nickel, cobalt, barium, and strontium, and even copper, zinc, and

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[1] C. K. Leith and W. J. Mead, Metamorphic geology, New York, 1915, pp. 64 and 69.

boron; in this case the concentration of iodine has had a certain economic importance. Whence came the iodine, whence the boron? Undoubtedly set free by gas-fluxing operations in magmas, thence discharged into the sea.

Molybdenum as well as chromium and selenium doubtless play some rôle in the organic cycle, and such a part is certainly also played by vanadium and uranium; this will be referred to farther on.

Even the lowest forms of vegetable life are of importance for the concentration of some deposits. Think of the iron and manganese bacteria which no doubt are essential for the formation of certain bog-ore deposits; consider the sulphur bacteria that may have contributed to some sulphur deposits.

No less important is the part played by animals, particularly those of the sea. For the animals that roam over the earth are likely to return their bodies to the air and the soil, while in the sea the more or less continuous sedimentation may cover and preserve their remains. Take the case of barium and strontium, originally contained in igneous rocks—quite abundantly in some alkaline varieties—and undoubtedly also given off by igneous emanations. The solutions from the land carry a continuous stream of these elements into the sea. Their presence in the sea water is well established. It has been shown above that barium and strontium are actually present in silts and muds in larger amounts than in igneous rocks. In agreement with this, most of our deposits of sulphates and carbonates of these metals are found in sedimentary rocks; such circumstances indicating concentration by later searching meteoric waters. Are there any organic agencies which may effect a slight concentration in the sediments of the barium found in the sea? The biologists answer this by telling us that certain protozoans segregate barium sulphate in minute crystals in their tissues and that certain radiolaria, like Acantharia, form their skeletons, in part, of strontium sulphate. Is there not a source here for meteoric waters to work upon when millions of their bodies were covered by rapid sedimentation?

The substances which are concentrated by the inhabitants of the sea include silica, sulphur, and potassium, the carbonates of calcium and magnesium, and the carbono-phosphates of calcium. These produce directly or indirectly such deposits as phosphate beds, glauconite sands, and limestones. We observe also the accumulation of heavy metals, such as copper, lead, zinc, and vanadium, and apparently also chromium, cobalt, nickel, and manganese. The three last-named metals have been determined in analysis of globigerina ooze, and are probably of organic origin.

Nickel and cobalt are present in all arable soils and are also found in small though essential quantities in the pancreas of animals.[1]

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[1] G. Bertrand, The importance of minute chemical constituents of biological products, *Science*, Dec. 24, 1926.

In the blood of molluscs[1] copper appears as an important constituent, also zinc, and probably lead. Zinc, copper, nickel, and cobalt certainly occur in sea water, and their biochemical concentration becomes one of the important marine processes.

Crustaceans are rich in zinc. Lead is more doubtful but in some form or other this element is probably contained in the molluscs. Vanadium takes the place of copper in the blood of ascidians and holothurians;[2] to what extent by other animals is, as yet, not ascertained. Anyway, vanadium must be an important constituent of sea water, though no analysis thus far confirms this.

Collectively these concentrations of heavy metals in sea animals are enormous. Naturally much of the metal is returned to the sea when the animal dies; but if for some reason or other the death rate is high, and the remains are buried in sediments, much of the metals must be retained in the mud; most probably in the state of sulphides. Minute particles of sulphides, mostly sphalerite or galena, are often found in fossilized molluscs.

## SEDIMENTARY MINERAL DEPOSITS

The sedimentary mineral deposits originating by the sorting of detritus have been described in Chap. XVII, and the salt deposits resulting from the evaporation of the solvent are treated in Chap. XX.

In the present chapter will be described a number of deposits formed by chemical reactions in the sea water resulting in limestone, dolomite, chert, iron ore, manganese ore, and phosphates. Purely sedimentary deposits of sulphides and barite are rare. Iron, manganese, calcium, magnesium, silica, and phosphorus are the principal elements concentrated on a large scale. They are usually the products of reactions between sea water and fluviatile water near the shore.

## LIMESTONE

**Definition and Origin.**—The limestones are sedimentary rocks, predominantly composed of carbonate of calcium (CaO, 56 per cent;

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[1] Molluscs, crustaceans, many insects (*e.g.*, *Blatta orientalis*), and fishes contain copper to the amount of from 1 to 14 milligrams per 100 grams of living animal. Zinc, a characteristic constituent of the human liver, is present in gasteropods; often there is more zinc than copper. The ashes of some gasteropods contain about 12 per cent ZnO and 7.8 per cent CuO. A. P. Matthews, Physiological chemistry, 3d ed., New York, 1920. R. H. A. Plimmer, Practical organic and biologic chemistry, London, 1926.

[2] A. H. Phillips found 0.123 per cent of the weight of the entire animal dried at 110°, *Am. Jour. Sci.*, 4th ser. 46, 1918, p. 473. M. Henze found 18.5 per cent $V_2O_5$ in the chromogen in the blood of ascidians. Some species of these animals do not appear to carry vanadium. *Zeitschr. f. physiol. Chemie*, 72, 1911, pp. 494–501.

$CO_2$, 44 per cent), usually calcite, but in recent deposits also aragonite.[1] They contain minor amounts of magnesium and iron, also varying amounts of alumina and silica; and by the increase of these constituents transitions to shale or sandstone result. Phosphate of calcium and organic matter also enter into the composition of most limestones. The rocks contain the crystalline carbonate of calcium for the amorphous form only occurs as a transition phase; the grain varies between the widest limits.

When water containing bicarbonate of calcium is discharged into the ocean or bodies of fresh water the calcium carbonate is often precipitated because of changing equilibrium in the solutions.[2] This is exemplified by deposits along the shore of the Great Salt Lake in Utah. Such limestones often form "oölitic" beds of small, rounded concretions.

Generally, organic life plays a most important part in the deposition of calcite either indirectly by precipitation by ammonium carbonate generated by decaying organisms, or directly by life processes. Bacteria[3] may be the agents, very often also algæ,[4] the latter both in the sea and in fresh-water deposits. Molluscs, corals, crustacea, and echinoderms segregate calcite and aragonite[5] in their shells which accumulate on the bottom at moderate depths.

Many organisms, such as sponges, secrete silica from the sea water and thus cherty deposits may be admixed with the limestones. Many limestones are almost wholly made up of shell remains, but in others no trace of organic structure may be visible. Metamorphism tends to increase the grain and destroy the fossils. Evaporation of ordinary surface waters in dry climates may produce thick beds of porous limestone. This is known in Mexico as "caliche."

Calcite and aragonite are often deposited in large masses by hot springs containing bicarbonate of calcium, and such deposits may closely simulate limestones. Certain beautiful banded and translucent spring deposits are called onyx and are used for ornamental stones.

Among the many varieties of limestone the following may be mentioned:

**Chalk.**—This is a white, fine-grained, loosely coherent pure limestone of comminuted shells of molluscs and also of foraminifera. Its occurrence in the Cretaceous along the English coast is well known. Exten-

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[1] J. Johnston, E. M. Merwin, and E. D. Williamson, The several forms of calcium carbonate, *Am. Jour. Sci.*, 4th ser., 41, 1916, pp. 473–512.

[2] J. Johnston and E. D. Williamson, The rôle of inorganic agencies in the deposition of calcium carbonate, *Jour. Geol.*, 24, 1916, pp. 729–750.

[3] T. W. Vaughn, Chemical and organic deposits of the sea, *Bull.* Geol. Soc. Am., 28, 1917, pp. 933–944.

For a modern review of the origin of limestone see W. H. Twenhofel, Treatise on sedimentation, 2d ed., Baltimore, 1932, pp. 289–329.

[4] W. H. Bradley, Algae reefs and oölites of the Green River formation, *Prof. Paper* 154-*G*, U. S. Geol. Survey, 1929, pp. 203–223.

[5] Aragonite is the unstable form of calcium carbonate and always tends to change to calcite. It is most common in recent or Tertiary deposits.

sive beds are reported from Texas, New Mexico, Arkansas, and Kansas but are rarely of satisfactory grade. Chalk is used as fertilizer, for whiting, for marking, for polishing powder, and for many other purposes. "Paris white" is a pigment made by grinding "cliffstone," a hard variety of chalk. Much of this is imported.

"Whiting" substitutes are manufactured from limestone in many parts of the United States. Such material should contain 95 per cent or more $CaCO_3$ and be ground to 200-mesh size. It should contain less than 0.3 per cent $Fe_2O_3$ and less than 4 per cent $SiO_2$.[1]

**Lithographic Stone.**—The variety of limestone used for engraving and the reproduction of colored plates is a fine-grained rock with imperfect conchoidal fracture, gray or yellowish in color, and uniform in texture. It must be porous, to absorb the grease in the printer's ink, and soft enough to work readily under the engraver's tool. Lithographic stone of good quality is difficult to find. The product from the Solenhofen quarries in Bavaria is a Jurassic limestone of unusual excellence. The material is variable in composition and its value is ascertained only by trial.[2] The price ranges up to 20 cents per pound.

Lithographic stone is reported to occur in several states of the Union, but none of it appears to be as good as Solenhofen rock. The plates used are 22 or 28 by 40 inches, and 3 inches thick. The better grades are expensive, selling at about 22 cents per pound. The best grade quarried in the United States comes from Brandenburg, Kentucky. Little lithographic stone is now used.

**Hydraulic Limestone.**—Certain argillaceous limestones or dolomitic limestones are used for the manufacture of natural cement. Such rock, crushed and burned, hardens or "sets" when mixed with water, owing to hydration and crystallization of the silicate and aluminate of calcium formed during the burning. "Portland" cements are mixtures of limestone and argillaceous rocks, subjected to a similar process of grinding and burning.

**Lime.**—Pure limestones are changed by burning and consequent decarbonation at 898° C. to quicklime which is usually shipped as "lumplime"; this "slakes" to a calcium hydroxide when mixed with water and is used, with the addition of sand, as mortar in brick constructions. The "slaking" is retarded by the presence of magnesia and argillaceous impurities.[3] Slaked lime, finely powdered is a product used extensively for waterproofing concrete.

[1] R. B. Ladoo, Non-metallic minerals, New York, 1925, pp. 123–130.

[2] S. J. Kübel, *Mineral Resources*, U. S. Geol. Survey, 1900, pp. 869–873.

G. P. Merrill, The non-metallic minerals, New York, 1910, pp. 147–152.

[3] E. F. Burchard and W. E. Emley, The source, manufacture and use of lime, *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1913, pp. 1509–1593.

E. C. Eckel, Cement, limes and plasters, New York, 1922.

W. E. Emley, Manufacture of lime, *Tech. Paper* 16, U. S. Bur. Standards, 1913.

**Uses.**—While limestone is mainly a structural material it is also used as a flux in smelting operations. The pure varieties are most acceptable, though magnesian limestones are also used in iron furnaces. Burnt lime generally contains from 90 to 95 per cent CaO, but magnesium limes with 30 per cent or over of MgO are also used, though less well liked.

Burnt lime is probably employed for more purposes than any other natural product. It is used for the manufacture of bleaching powder, ammonia, calcium carbide, fertilizers, wood alcohol, soap, glycerine, glue, glass, pottery, paints, paper, and sugar. Also for tanneries and as insecticide and fungicide. An amount of 2,000,000 to 3,000,000 tons is produced annually in the United States.

## DOLOMITE[1]

Pure dolomite contains 54.35 per cent $CaCO_3$ and 45.65 per cent $MgCO_3$. Beds of dolomite and dolomitic limestone are common in sedimentary deposits. They may often be distinguished by a fine-grained sugary texture, due to a development of uniform rhombohedral crystals. Dolomite is somewhat harder than limestone and is insoluble in dilute hydrochloric acid. Magnesium carbonate is much more soluble in water than calcium carbonate. Some travertines from mineral springs are rich in $MgCO_3$ and may contain up to 29 per cent of this compound. Dolomite is doubtless deposited by direct precipitation in sea water, but most of it has been formed by alteration of the limestone by sea water, or by subsequent dolomitization by cold or hot waters.

Deep borings in coral reefs have shown that the limestone, somewhat magnesian at the surface, passes into dolomite in depth. Certain algæ deposit much $MgCO_3$ with $CaCO_3$; some shells also contain magnesium carbonate but seldom more than 7 per cent. In warm waters the percentage of $MgCO_3$ in shells tends to increase.

Dolomite may be used for lime burning, and it is said that it can be used as a substitute for magnesite in the manufacture of refractory bricks.

Instances are known of the deposition of thin beds of pure magnesite in bodies of water, but larger masses of magnesite are usually formed from limestone by metasomatic processes (p. 388).

## IMPORTANCE OF CARBONATE ROCKS AS RELATED TO ORE DEPOSITS

Within the zone of oxidation the carbonate rocks are often dissolved, residual clays being then developed. Accessory constituents such as

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[1] F. M. Van Tuyl, The origin of dolomite, *Ann. Rept.*, Iowa Geol. Survey, 25, 1914, pp. 251–422; *Am. Jour. Sci.*, 4th ser., 42, 1926, pp. 249–260.

For an extended discussion of dolomite see F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 565–580.

E. Steidtmann, Origin of dolomite, etc., *Bull.* Geol. Soc. Am., 28, 1917, pp. 431–450.

zinc and lead in the form of sulphides, or admixed phosphates may then become concentrated and acquire economic importance.

Limestone is easily silicified by waters containing silica; the silica usually appears as irregular masses of fine-grained quartz or chert. It is quite as easily dolomitized by dilute waters containing some magnesia, and this is often observed near ore deposits formed at slight or moderate depth. Limestone and dolomite, under the influence of heated waters, are subject to replacement by quartz, dolomite, barite, and fluorite or by metallic ores such as pyrite, sphalerite, and galena. At high temperature and pressure pure limestones recrystallize to marble. Silicates, such as garnet, diopside, or wollastonite, form in argillaceous or siliceous limestone from the impurities contained or from the introduction of solutions rich in silica and iron. Lastly, the limestones are easily soluble and caves develop along fractures, forming receptacles for the deposition of ores.

## CHERTS AND DIATOMACEOUS EARTH

The silica accumulated by detrital processes as sandstone and quartzite has already been mentioned. Silica may, however, also be extracted from water and deposited as a sediment by means of organisms, such as radiolarians, diatoms, and sponges. In part this silica forms cherty masses included in limestone; in part it is deposited as distinct beds. Diatomaceous earth[1] is a deposit formed in lakes and swamps, as well as in the sea; and, when pure, consists of the delicate tests of diatoms, a class of algæ. Such deposits accumulate abundantly where siliceous volcanic tuffs were deposited in lakes, as occurred at many places in the Cordilleran region during the Tertiary period. Thick beds are found in the Miocene of Santa Barbara County, California.[2] The diatomaceous earth is frequently more or less admixed with rhyolitic glass and other detritus; the tests consist of hydrated opaline silica. The earth forms light-colored beds of extremely fine texture. It finds extensive use for polishing powder, steam-pipe packing and insulation, absorbent and filtering medium for various liquids, and for mixing with cements. The price is about $20 per ton for material crushed to 200-mesh size. It contains up to 87 per cent $SiO_2$ and 5 to 9 per cent $H_2O$.

## SEDIMENTARY SULPHIDE DEPOSITS

As the sedimentary rocks largely consist of the detritus of the continents, it is self-evident that they may contain the metals of the rocks and ore deposits of the land areas. Iron is, of course, abundant, also in a

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[1] W. C. Phalen, and F. J. Katz, *Mineral Resources*, U. S. Geol. Survey, 1908–1925, under heading Abrasives.

For the preparation and uses, see W. M. Weigel, Technology and uses of silica and sand, *Bull.* 266, U. S. Bur. Mines, 1927.

V. L. Eardley-Wilmot, Diatomite, *Publ.* 723, Canada Dept. Mines, 1930.

[2] R. Arnold and P. Anderson, *Bull.* 315, U. S. Geol. Survey, 1906, pp. 438–447.

lesser degree manganese; concretions of hydrous oxides of manganese are found in the deep-sea deposits and analysis shows that these contain notable amounts of nickel, cobalt, copper, zinc, lead, and molybdenum. Many limestones have been shown to contain minute amounts of zinc, lead, and copper. It is extremely unlikely that a sufficient quantity of pyrite and other sulphides would escape oxidation to form important detrital deposits.

The deep-sea deposits accumulate so slowly that they are of little importance in this discussion. The sediments nearer to the shores contain much more terrigenous material and are, therefore, of special interest.

Pyrite, and infrequently other sulphides, may be precipitated by chemical reactions in sediments. Beds of oölitic pyrite are known (p. 269); iron disulphide is formed, in places, in bogs and streams or in oceanic sediments where hydrogen sulphide developed by decaying organic matter reacts on the sulphates of iron. If these sediments are brought to the surface by orogenic movements and slightly metamorphosed, the sulphide, originally in fine dissemination, may recrystallize in more prominent form. As a matter of fact, the deep-sea muds thus far analyzed contain little or no pyrite. In the special case of the Black Sea, often quoted of late from N. Androussof's description,[1] micro-organisms assist in liberating hydrogen sulphide, part of which, by reaction with iron from the sediments, develops colloidal ferrous sulphide. Such colloids may crystallize as disulphide to pyrite or marcasite. Regarding melnikowite, supposedly a black disulphide forming in sediments, see the literature.[2]

Although pyritic clays are abundant in the unmetamorphosed sedimentary formations, there is little evidence of extensive sedimentary beds of pyrite.

Barite when appearing in sandstone is probably deposited by hot springs. For barite and manganese carbonates as marine shore deposits see page 281.

Evidence is scant as to the sedimentary deposition on a large scale of sulphides other than pyrite or marcasite. "The Kupferschiefer" of Mansfeld, in which the sulphides may be of syngenetic origin, will be described elsewhere (p. 415).

## SEDIMENTARY IRON ORES

It is conceded that iron ores, such as magnetite, can be deposited by mechanical concentration as placers along rivers or the seashore (p. 249),

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[1] N. Androussof, Guide des excursions du VII$^{ème}$ Congrès géologique internat., 29, 1907, p. 6.

[2] B. Doss, *Zeitschr. prakt. Geol.*, 20, 1912, p. 460.

E. C. Harder, Iron-depositing bacteria and their geologic relations, *Prof. Paper* 113, U. S. Geol. Survey, 1919, p. 61.

or again we may easily conceive hematite or limonite derived from deep decay of rocks along the littoral, or from the oxidation of pyrite deposits, as at Rio Tinto, Spain, swept out into the sea and deposited close to the shore. Iron ores are also formed by chemical reactions in bodies of water, and these yield a notable proportion of the iron production of the world. In the latter cases the iron has been supplied from the land areas in form of solutions. In many instances both dissolved iron salts and detrital minerals of iron contribute to the genesis of the deposits.

The surface waters extract iron from ferromagnesian silicates as well as from oxides or other minerals; this extraction proceeds most energetically in regions covered by a deep mantle of decayed rock. Both iron and manganese are contained in springs and streams. An example of such spring water, rising underneath a deposit of bog iron ore in Holland, is quoted by Clarke:

ANALYSIS OF SPRING WATER AT EDERVEEN, NETHERLANDS[1]
(Analyst, G. Moll van Charante)
(Parts per million)

| | | | |
|---|---|---|---|
| Ca | 107.6 | $Al_2O_3$ | 3.3 |
| Mg | 5.6 | Cl | 15.2 |
| Fe | 19.6 | $H_3PO_4$ | 10.9 |
| Mn | 11.4 | $SO_4$ | 0.9 |
| K | 0.9 | $CO_3$ | 207.6 |
| Na | 10.0 | $SiO_2$ | 18.0 |
| | | Organic | 56.0 |
| | | | 467.0 |

Tropical rivers often carry several per cent of $Fe_2O_3$ in the dissolved salts, the maximum being about 9 per cent in some Guiana (South America) streams.

A larger part of the iron dissolved by the surface water is precipitated after a short journey,[2] but some of it is carried down by the streams to lakes and seas, in which it then may be deposited on an extensive scale. The sea water contains about 0.6 milligram per liter of iron and probably more at some places near the shore.

## LIMONITES IN SWAMPS AND LAKES (BOG IRON ORES)

**Occurrence.**—The bog iron ores are found in swamps, lakes, or even sluggish water courses, and are especially abundant in the recently glaciated regions of northern Europe, Asia, and North America. They consist of dark-brown, rough and cellular masses or loose particles, sometimes oölitic in texture and then designated as "shot ore," and form a

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[1] Cited by F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 536.

[2] The clogging of water-supply pipes by hydroxides of manganese and iron is a common occurrence.

layer of varying thickness at the bottom of the swamp or lake. Plants and roots may be replaced by limonite. Such ores are usually mined by means of primitive dredges or scoops.

The ore occurs mainly in shallow waters along the shore, to a depth of about 12 feet. After removal a new layer is formed within a few years; according to A. Geikie[1] several inches of limonite accumulated in 26 years in a Swedish deposit. The rate would naturally be subject to great variations according to local conditions.

The bog iron ores are now of slight importance to the mining industry, but the easily traceable processes of their formation give us a most welcome key to the origin of other and more obscure deposits.

**Composition.**—These ores are always mixed with sand and clay and rarely contain as much as 50 per cent of iron. The principal mineral contained is limonite, but carbonate of iron is commonly present, also phosphate of iron as vivianite; soluble silica is sometimes recorded. In some low-grade ores from the Netherlands, the analyses of which are quoted by Clarke,[2] there is much more ferrous carbonate than limonite. Varying quantities of manganese are present in ores from Sweden, Finland, and Holland. The Swedish ore contains traces of vanadium, molybdenum, copper, lead, zinc, arsenic, nickel, and cobalt. All bog iron ores contain phosphorus, but there is rarely much sulphur.

According to Svanberg, cited by Zirkel,[3] the average of 30 analyses of Swedish bog iron ores gave:

| | | | |
|---|---|---|---|
| $Fe_2O_3$ | 62.57 | MgO | 0.19 |
| $Mn_2O_3$ | 5.58 | $P_2O_5$ | 0.48 |
| $SiO_2$ | 12.64 | $SO_3$ | 0.07 |
| $Al_2O_3$ | 3.58 | Ignition | 13.53 |
| CaO | 1.37 | | |
| | | Total | 100.01 |

**Origin.**[4]—The agents by which iron is carried into solution are (1) carbon dioxide from the air and decomposing organisms; (2) sulphuric acid from the weathering of pyrite; and (3) organic acids derived from decomposing vegetable matter. In the absence of air ferric oxide is reduced to the ferrous state and forms soluble double salts with ammonia

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[1] A. Geikie, Textbook of geology, 4th ed., 1903, p. 187.

[2] F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 538.

[3] F. Zirkel, Lehrbuch der Petrographie, 3, 1894, p. 574.

[4] F. M. Stapff, *Zeitschr.* deutsch. geol. Gesell., 18, 1866, p. 86.
J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 2, 1894, p. 30; and 3, 1895, p. 38.
J. M. van Bemmelen, *Zeitschr. anorg. Chemie*, 22, 1906, p. 313.
Ossian Aschan, *Zeitschr. prakt. Geol.*, 15, 1907, pp. 56–62.
E. S. Moore, *Econ. Geol.*, 5, 1910, pp. 528–537.
N. J. Harrar, Solvent effects of certain organic acids upon oxides of iron, *Econ. Geol.*, 24, 1929, pp. 50–61.

and humic acid. The importance of organic acids was emphasized long ago by A. A. Julien and by Sterry Hunt.

Precipitation is effected in bicarbonate solutions by the escape of carbon dioxide in the air or through its absorption by plant cells. The ferrous carbonate is easily oxidized to ferric hydroxides. In the presence of much organic matter ferrous carbonate remains in the precipitate.

From ferrous sulphate solution iron is precipitated as limonite by oxidation and hydrolysis, or by reaction with calcium carbonate solution, in which case siderite and gypsum will result, the former oxidizing to limonite, or the iron may be precipitated by ammonium humate, always present in swamp waters, or finally by soluble calcium phosphate, in which case vivianite or other iron phosphates result. Less commonly the iron is precipitated as pyrite by alkaline sulphides or hydrogen sulphide.

From soluble humates iron is also precipitated by organisms, called iron bacteria,[1] which take up these humates, as well as ferrous carbonate, and coat their cell walls with the segregated limonite; but regarding the real importance of this process we have few data.

In these, as in so many other surface reactions, the ferric hydroxides are probably precipitated as colloidal complexes of indefinite composition, or "gels," which in time tend to change to crystalline bodies. Much of the ferric hydroxide is doubtless transported for considerable distance in colloid form. Five "species" of ferric hydroxide have been recognized. Arranged by increasing water they are:

| | | |
|---|---|---|
| Turgite | $2Fe_2O_3.H_2O$ | 94.6 per cent $Fe_2O_3$ |
| Göthite | $2Fe_2O_3.2H_2O$ | 89.9 per cent $Fe_2O_3$ |
| Limonite | $2Fe_2O_3.3H_2O$ | 85.5 per cent $Fe_2O_3$ |
| Xanthosiderite | $2Fe_2O_3.4H_2O$ | 81.6 per cent $Fe_2O_3$ |
| Limnite | $2Fe_2O_3.6H_2O$ | 74.7 per cent $Fe_2O_3$ |

Turgite is red and may be mistaken for hematite. Several complex ferric silicates and sulphates look somewhat like limonite. Posnjak and Merwin[2] admit only one crystalline form—a monohydrate, crystallizing as göthite or lepidocrocite. They define limonite as a colloidal, amorphous ferric oxide monohydrate with variable adsorbed and capillary water.

**Examples.**—At Radnor and Drummondville, Three Rivers district, Quebec, the occurrences of bog iron are extensive. This iron ore was utilized until 1911, being dug in the swamps or dredged in the lakes; in

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[1] E. C. Harder, Iron-depositing bacteria and their geologic relations, *Prof. Paper* 113, U. S. Geol. Survey, 1919.

[2] E. Posnjak and H. E. Merwin, The hydrated ferric oxides, *Am. Jour. Sci.*, 4th ser., 4, 1919, pp. 311–348; The system $Fe_2O_3 - SO_3 - H_2O$, *Jour.* Amer. Chem. Soc., 44, pt. 2, 1922, pp. 1965–1994.

H. B. Weiser, The hydrous oxides, New York, 1926, 482 pp.

1911 the operations ceased. The production in this district began in 1733.

One of the most famous deposits formerly mined is at Katahdin, Maine. Small deposits are found at very many places in New England and have been worked on a small scale.

## THE SIDERITES OF MARINE AND BRACKISH-WATER STRATA

**Occurrence.**—Siderite ($FeCO_3$) is an iron ore of some importance, in both epigenetic and syngenetic deposits. It occurs in fissure veins and as a replacement of limestone, but is also found in the sedimentary rocks as a product of the sedimentary processes. The sedimentary siderite ores are called clay ironstone, spherosiderite, or black band. A dense or fine-grained concretionary or spherulitic structure is characteristic of the "clay ironstone" occurring in clays or shales; and these concretions, more or less admixed with clay and sand and often inclosing vegetable remains, are found abundantly at certain horizons. The variety called "black band" forms continuous beds of dark-colored, compact appearance in the shales of the coal measures, often directly underneath or above the coal beds.

These ores contain less than 48 per cent of iron and must be calcined before smelting. Both sulphur and phosphorus are present, sometimes in considerable quantities.

Marcasite, pyrite, arsenopyrite, millerite, galena, sphalerite, and chalcopyrite are sometimes found along cracks in the concretions of siderite, indicating that the iron solutions carried small amounts of the less common metals, probably as sulphates. An analysis of siderite ore from Maryland[1] showed in percentage 36.05 Fe, 13.53 $SiO_2$, 6.47 $Al_2O_3$, 0.94 Mn, 0.08 P, and 0.42 S.

The economic importance of these ores, formerly great, is now small. Near the surface they are sometimes changed to limonite.

The origin of sedimentary siderites is explained along the same lines as that of the bog-iron ores. Solutions of ferrous bicarbonate or sulphate were supplied to the marshes along the sea coast or to the shallow sea where organic matter was abundant. Precipitation of the normal insoluble carbonate took place through absorption of the solvent $CO_2$ by vegetation. Free oxygen was absent, for otherwise the carbonate would have been transformed into limonite. But even if the iron had originally been deposited as limonite, a reduction and carbonation to siderite may have been effected by the limonitic precipitate being covered by mud containing organic matter.

[1] J. T. Singewald, Jr., *Econ. Geol.*, 4, 1909, pp. 530–543.

J. T. Singewald, Jr., Report on the iron ores of Maryland, Maryland Geol. and Econ. Survey, 9, pt. 3, 1911.

The concretionary masses are not the products of primary precipitation, but are, probably in all cases, segregated into nodular form by the action of percolating solutions around a suitable nucleus while the sediments were still soft.

**Examples.**—In the United States sedimentary siderites are known in Pennsylvania, Ohio, West Virginia, Maryland, and Kentucky. Their present industrial importance is slight, but they were formerly mined on a more extensive scale. The production in 1930 was only 1,000 long tons from Pennsylvania, and this was used for the manufacture of red paint.

In Pennsylvania and adjacent States the upper barren Coal Measures contain abundant nodules of siderite in the shales and sandstones, but no valuable deposits. In the upper productive Coal Measures, or Monongahela River series, black band ore occurs, for instance, just below or above the Pittsburgh coal bed. In the lower Coal Measures the siderites are especially abundant; in Ohio 12 horizons of black bands and concretionary ores are distinguished by Orton.[1]

Siderite ores also occur in the Tertiary Claiborne formation of Mississippi.

The black bands are common in Germany. They were formerly actively worked in Westphalia and near Saarbrücken, where the ore formed flat lenticular masses as much as 1½ meters thick and sometimes several hundred meters in extent.

In England the black bands were formerly of the highest importance and 50 years ago furnished four-fifths of the total iron output. They are now mined only in North Staffordshire and in Scotland. In Wales the black bands occur in the lower Coal Measures. Kendall[2] enumerates 75 horizons of siderite ore.

In Scotland (Ayrshire) the black bands occur both in Coal Measures and in Carboniferous limestone. The ores contain 25 to 40 per cent Fe, and occur as thin strata, 1½ feet or less thick; several of them are usually close together.

### THE OÖLITIC MARINE IRON ORES

Oölites[3] (named from semblance to fish roe) are rounded, concretionary grains, commonly showing concentric deposition bands and radial fibrous crystal structure. Their size varies up to a diameter of 2 millimeters; a kernel of a sand grain or fossil fragment is often present.

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[1] Ohio Geol. Survey, 5, 1884, p. 378.

[2] J. D. Kendall, The iron ores of Great Britain and Ireland, London, 1893, pp. 145–199.

[3] For best review and literature see W. H. Twenhofel, Treatise on sedimentation, Baltimore, 2d ed., 1932, pp. 430–460; 757–769.

W. H. Bucher, On oölites and spherulites, *Jour. Geol.*, 26, 1918, pp. 593–621.

G. Linck, Die Bildung der Oölite und Rogensteine, *Neues Jahrbuch.*, Beil. Bd. 16, 1903, p. 495.

Rocks with abundant oölites are said to have an oölitic texture. Oölitic limestones (Fig. 105) occur in many sedimentary series, and are sometimes silicified. The oölites of the iron ores consist of hematite, siderite, iron-rich chlorite, or more rarely of limonite. Frequently all of these occur together. The matrix consists of calcite, siderite, or argillaceous substance. Oölites of hematite, siderite, or iron silicate usually contain some colloidal silica which is left as a skeleton when the concretions are dissolved in acid.

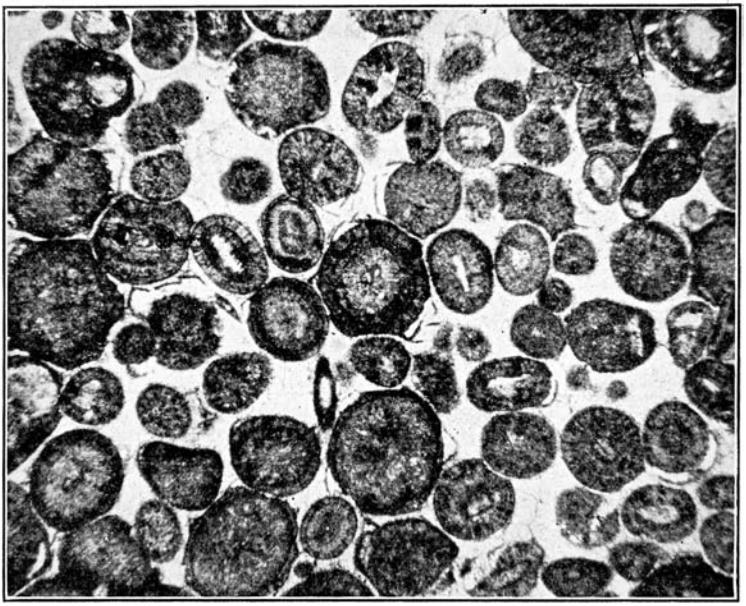

Fig. 105.—Thin section of calcite oölite with calcite cement, Short Creek oölite, Missouri. (*After P. W. George.*)

Undoubtedly the oölites are formed by colloidal processes; this is almost self-evident and has been proved by the investigations of Schade, quoted by Bucher and others.

The iron oölites are shallow water products and their development is dependent upon a nearby source of iron from decomposing rocks, upon wave action and currents, and upon rapidly succeeding fluctuations of level. Their marine origin is abundantly proved.

In the past the view has often been expressed that the oölitic iron ores have developed from oölitic limestone by the later introductions of iron. Few hold to this opinion at the present time; the iron was supplied when the oölites were formed. Many complex changes took place, however, during deposition. Conditions were alternatingly oxidizing and reducing, resulting in hematite, or ferrous silicate, or ferrous carbonate.

In later ages, when exposed to weathering the ferrous minerals may change to limonite, or calcite may be dissolved, causing enrichment of the ore.

**The Hydrous Iron Silicates of Sedimentary Origin.**—The minerals to be considered are glauconite, chamosite, thuringite (daphnite), and greenalite. Their approximate composition is shown in the following table. The composition is not constant and the formulas are more or less uncertain, as is natural because all are probably of colloidal origin:

APPROXIMATE COMPOSITION OF SEDIMENTARY IRON SILICATES

| | Glauconite | Chamosite | Thuringite | Greenalite |
|---|---|---|---|---|
| $SiO_2$ | 48 | 25 | 23 | 30 |
| $Al_2O_3$ | 7 | 17 | 17 | |
| $Fe_2O_3$ | 24 | 6 | 15 | 35 |
| FeO | 2 | 39 | 33 | 26 |
| CaO | 1 | | | |
| MgO | 3 | 3 | | |
| $K_2O$ | 7 | | | |
| $H_2O$ | 8 | 10 | 11 | 9 |

Glauconite ($KFeSi_2O_6.H_2O$) occupies a separate place on account of its constant tenor of potassium, which, it seems, is higher in the older than in the modern varieties. It does not occur in the oölitic ores, but is, in part, concretionary; it forms dark-green granules often crystalline and often deposited in the interior of shells. Glauconite, according to Caspari, Hummel, and others,[1] is colloidal while others point out that it has a definite crystalline structure and occasionally forms single crystals. The extensive literature on the subject is quoted in the publications referred to below. The mineral is scattered in marine sands of all ages; some contain so much that they are referred to as "greensands," *e.g.*, in the Cretaceous beds of New Jersey. These rocks contain much phosphorus besides potassium and are sometimes used as fertilizers, but never as iron ores because there is always too much quartz present. By weathering they may be enriched to limonitic ores.

According to Murray and Renard, glauconite is formed just beyond the limits of wave and current action, where the muddy deposits begin. It is not a littoral deposit. The colloidal silica and clay in the mud

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[1] Besides the general literature quoted above see: W. A. Caspari, Contrib. chemistry marine glauconite, *Proc.*, Edinburgh Royal Soc., 30, 1909, p. 364.

K. Hummel, Die Entstehung eisenreicher Gesteine durch Halmyrolyse, *Geologische Rundschau*, 13, 1922, pp. 40–136.

L. Cayeux, Contrib. étude microgr. des terres sed., Lille, 1897.

A. Lacroix, Minéralogie de la France, 1, 1893–1895, pp. 406–408.

C. S. Ross, The optical properties and chemical composition of glauconite, *Proc.*, U. S. Nat. Mus., 69, 1926, pp. 1–15.

react with ferric iron set free by oxidation on the sea bottom, and the precipitated gel of iron silicate absorbs potassium from the sea water. Caspari claims to have made glauconite by synthesis. Glauconite is probably a metacolloid.

Hummel points out that the glauconite is the product of cool waters or where cool and warm currents meet.

Chamosite, thuringite, and greenalite are iron-rich chlorites.

Greenalite, which evidently is related to thuringite, occurs abundantly, according to Leith,[1] in the Huronian sideritic cherts of the Mesabi and other iron districts in the Lake Superior region. The greenalite in part yielded the material from which the hematites were derived.

Chamosite ($Al_2O_3.3FeO.2SiO_2.xH_2O$ (Hallimond)) and thuringite[2] are dark green, chloritic minerals, which when occurring in the sedimentary iron ores nearly always show an oölitic development.

The beds first described occur in Thuringia (Germany) and in Bohemia. They were formerly mined extensively and are still exploited in the latter region. In Bohemia the Silurian series comprises shale, graywacke, and diabase tuffs, and contains beds of oölitic iron ore, one bed being 16 feet thick, while other beds, also thick, consist mainly of oölitic chamosite. The latter show a matrix of siderite or chamosite in which are embedded oölites of chamosite. The ores are rich in phosphorus and carry a little magnetite.

**The Oölitic Limonite Ores.**—Primary marine limonite with oölitic texture is apparently rare, probably on account of the dehydrating effect of sea water. Hallimond describes oölitic limonite in sandstone and clay from the Frodingham district, England; it is, however, of no great importance. Possibly the deposits can be accounted for by an unusual supply of ferric hydrate from the adjoining shores.

**The Oölitic Pyritic Ores.**—While pyrite and marcasite are common in many sedimentary beds, an oölitic development is rare. The deposits at Meggen, Germany,[3] have been worked for pyrite on a fairly extensive scale, the bed being from 12 to 20 feet thick and enclosed within a thin limestone bed in folded Devonian slates (Fig. 106). Barite and sphalerite also occur here. Bergeat holds to the sedimentary origin of the oölitic pyrite but this opinion is not shared by Krusch who regards the deposit as a replacement of limestone. The mine produces 330,000 metric tons of pyrite and 170,000 tons of barite annually.

However this may be, it is certain that true oölitic beds of pyrite (or marcasite) are found at various places (Wabana, Newfoundland, Cleveland Hills, England), though they are small and have no economic value.

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[1] C. K. Leith, *Mon.* 43, U. S. Geol. Survey, 1903, pp. 237–279.

[2] E. R. Zalinsky, *Neues Jahrbuch*, Beil. Bd. 19, 1904, pp. 40–84.

[3] A. Bergeat, *Zeitschr. prakt. Geol.*, 22, 1914, pp. 237–249.
Beyschlag, Krusch, and Vogt, Die Lagerstätten, 2, 1913, p. 414.

**Distribution.**—Hummel discusses the peculiar distribution of the oölitic, marine iron ores and points out that they are apparently not being deposited at the present time, in contrast to the glauconite beds which are being formed in the open sea at many places. He suggests that they require warm waters in sheltered seas free from cold currents. The oölitic hematites are found in many formations from pre-Cambrian to Paleozoic, inclusive. The oölitic siderite-chamosite ores are abundant in the Jurassic, *e.g.*, in England and France, though they are also known in older formations. In many deposits, particularly of Paleozoic age, the three minerals occur together. The Cretaceous while rich in glauconite beds contains no oölitic iron ores. The opinion is gaining ground that siderite is not generally deposited in oölitic form but develops by replacement of chamosite very soon after the oölites are formed.

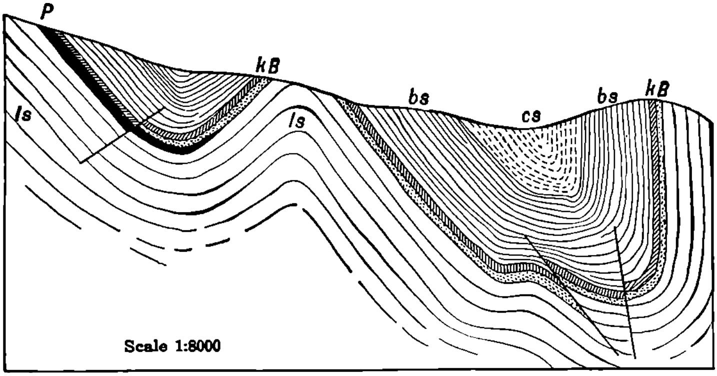

FIG. 106.—Vertical section through the pyritic deposit at Meggen, Germany. *P*, pyrite; *B*, barite; *k*, limestone; *bs*, *cs*, *ls*, Devonian slates. (*After Strauss.*)

Hallimond thought that the pyrite might replace chamosite, but this appears doubtful. At least, the Wabana, Newfoundland, oölites show every indication of a primary deposit naturally formed under reducing conditions.

That either siderite or pyrite can be deposited in large bodies in the open sea must be considered very unlikely.

**Part Played by Organisms.**—The part played by micro-organisms is, as yet, difficult to evaluate. The scattered literature and much original work are found in a publication by E. C. Harder.[1] Bacteria, particularly the "thread bacteria," but also protozoa, algæ, and fungi are concerned in the precipitation of ferric hydroxide. Bacteria are most important. In the presence of soluble iron salts, bacteria of the Crenothrix type segregate iron hydroxide by oxidation of $FeCO_3$; and it seems that this process can go on also in brackish and salt waters.

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[1] *Prof. Paper* 113, U. S. Geol. Survey, 1919.

Thus far this process seems most important for the bog-iron ores. Certain other bacteria seem to promote the formation of calcareous oölites in the sea. The blue-green algæ develop oxygen in their life process and this would, of course, promote oxidation, say of siderite to hematite. This may be of some importance in the formation of oölitic ores.

## THE OÖLITIC HEMATITE ORES

**The Clinton Ores.**[1]—The most important oölitic ores in the United States are those of the Clinton formation in the Appalachian States. They persist with remarkable regularity wherever this formation appears. The Clinton (Silurian) lies between the Trenton limestone and the Devonian shale, and invariably contains one or several beds of hematite ore alternating with limestone and shale. The succession of sedimentary rocks in the Birmingham district is as follows. In a general way the section applies to the entire southern Appalachian region.

| | | Feet |
|---|---|---|
| Carboniferous: | | |
| Pennsylvanian: Pottsville formation ("Coal Measures") | | 2,600 to 7,000 |
| Unconformity. | | |
| Mississippian: | | |
| Parkwood formation | | 0 to 2,000 |
| Pennington shale (30–300 feet)<br>Bangor limestone (670 feet) | Floyd shale | 1,000 ± |
| Fort Payne chert | | 200 to 250 |
| Unconformity. | | |
| Devonian: | | |
| Chattanooga shale<br>Frog Mountain sandstone | | 1 to 25 |
| Unconformity. | | |
| Silurian: Clinton (Rockwood) formation | | 250 to 500 |
| Unconformity. | | |
| Ordovician: Chickamauga (Pelham) limestone | | 200 to 1,000 |
| Unconformity. | | |
| Cambro-Ordovician: Knox dolomite | | 3,300 |
| Cambrian: | | |
| Conasauga (Coosa) limestone | | 1,000+ |
| Rome (Montevallo) shale (great thickness). | | |

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[1] C. H. Smyth, Jr., On the Clinton iron ore, *Am. Jour. Sci.*, 3d ser., 43, 1892, p. 487.

E. F. Burchard, The Clinton iron ore deposits of Alabama, *Trans.*, Am. Inst. Min. Eng., 39, 1908, pp. 997–1055.

D. H. Newland and C. A. Hartnagel, Iron ores of the Clinton formation, *Bull.* 123, New York State Mus., 1908.

Burchard, Butts, and Eckel, The Birmingham district, Alabama, *Bull.* 400, U. S. Geol. Survey, 1910.

E. F. Burchard, The red iron ores of East Tennessee, *Bull.* 16, Geol. Survey Tennessee, 1913.

W. R. Crane, Red ore mining methods in the Birmingham district, *Trans.*, Am. Inst. Min. Met. Eng., 72, 1925, pp. 157–186; also pp. 187–225.

Iron ore mining practice in the Birmingham district, *Bull.* 239, U. S. Bur. Mines, 1926.

The Clinton ores extend from western New York, through Pennsylvania, Virginia, West Virginia, Kentucky, Tennessee, and northwestern Georgia into Alabama, where, near Birmingham, they attain their greatest development. The ores constitute beds or lenses at various horizons in the Clinton formation, which forms a striking unit of red shallow water deposits underlain and covered disconformably by great thicknesses of limestones of the Cambrian and Mississippian ages, respectively (Fig. 107). Thin beds of ferruginous sandstone, shale, and oölitic hematite make up the formation, with frequent cross bedding and some conglomerates. The ores contain calcite and in some places show gradual transition to limestones.

The average thickness of the ore beds is only two or three feet, but in Alabama they reach 20 feet of merchantable ore with occasional thin shale or sandstone partings. Single ore beds may extend for many miles. In the Birmingham district the Clinton beds outcrop on the east flank of an anticline and can be traced continuously northward into Tennessee, but the actively working mines extend only for 15 miles along the outcrop. Good ore beds have been found by drilling for several miles eastward but toward the west the formation becomes more calcareous.

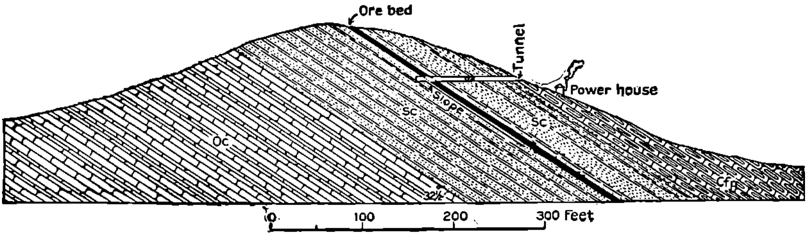

Fig. 107.—Section showing Clinton iron ores, Birmingham, Alabama; *Sc*, Clinton (Rockwood) formation, Silurian; *Oc*, Chickamauga (Pelham) limestone, Ordovician; *Cfp*, Fort Payne chert, Mississippian. (*After E. F. Burchard.*)

Four beds are known within 80 feet in the upper part of the formation, two of which are worked, with a thickness of from 9 to 20 feet. The iron ores are generally sharply bounded by shale or sandstone; in places they form transitions into ferruginous sandstone.

An important iron industry is based upon the deposits in Alabama and the annual production of ore has now attained about 6,500,000 tons, or about 11 per cent of the total output of iron ore in the United States. Mining has been carried 4,000 feet on the dip in some of the properties; and entirely similar ore has been shown to exist by borings at a vertical depth of 2,000 feet, 2 miles eastward from the outcrop. Large reserves of ore are available in this district.

Clinton ores are also mined north of Alabama in Tennessee though the operations are generally confined to the enriched surface ore. North of Tennessee the mining is profitable in few places.

There are several types of Clinton ores; most of them are fairly rich in calcium carbonate.

One common type is a fine-grained pebbly conglomerate or sandstone, each pebble or grain coated with hematite and the rock cemented with that mineral and with calcite. Another type consists largely of fragments of bryozoa, shells, trilobites, etc., partly coated or replaced by ferric oxide, besides an abundance of oölitic grains, usually with a grain of sand as the center (Fig. 108). Still another type consists entirely of oölites of hematite in calcite matrix, averaging 1 or 2 millimeters in diameter. A fourth type shows small flattened hematite concretions with fragments of fossils changed to hematite; this is the "flax seed ore" which is very common at Birmingham. There is very little siderite or chlorite. Fragments of quartz and other minerals are common. The beds vary along the strike in their calcareous or siliceous admixtures. Phosphorus is present in considerable amounts.

FIG. 108.—Clinton ore, Wolcott, Wayne County, New York. Magnified 20 diameters. Ore essentially formed of remains of bryozoans and crinoids. *a*, fragments of bryozoans, ferric oxide; *b*, fragment of bryozoan encrusted with ferric oxide; *c*, bryozoan structure almost obliterated by ferric oxide; *d*, crinoid stalk replaced by ferric oxide, cells filled with calcite of uniform optical orientation; *e*, same, almost entirely replaced; *J*, calcite cement. (*After L. Cayeux.*)

At the surface and down to a depth of about 200 feet the calcium carbonate is in part dissolved and the ore correspondingly enriched. Such ore is called "soft," in contrast to the unaltered or "hard" variety. The poorest ores used carry 25 to 30 per cent iron.

ANALYSES OF CLINTON ORES

(E. C. Harder, *Mineral Resources*, U. S. Geol. Survey, 1908)

| | Hard ore | Soft ore |
|---|---|---|
| Fe | 37.00 | 50.44 |
| $SiO_2$ | 7.14 | 12.10 |
| $Al_2O_3$ | 3.81 | 6.06 |
| CaO | 19.20 | 4.65 |
| Mn | 0.23 | 0.21 |
| S | 0.08 | 0.07 |
| P | 0.30 | 0.46 |

That the hematite was formed by direct sedimentation is held by C. H. Smyth; similar views are advocated by Newlands, Eckel, and Burchard. Smyth believes that the iron was carried out into shallow marine basins and there slowly oxidized and precipitated mechanically around the shells or replaced them. There is little iron silicate. According to G. W. Stose, siderite occurs in some of the Virginia ores.

## THE JURASSIC IRON ORES OF ENGLAND[1]

The English iron industry is now largely dependent upon the sedimentary ores occurring in the Jurassic. These ores now furnish the larger part of an annual production of 15,000,000 tons. The greatest yield comes from the Cleveland Hills in the Yorkshire district. While the ores are of low grade, they are cheaply mined, largely in open cuts.

The recent memoirs of the Geological Survey have furnished long awaited information. The ores occur in the Middle Lias and a section would be about as follows; the figures being in feet:

| | | |
|---|---|---|
| Upper Lias | Shale | |
| Middle Lias | Shale | 0–6 |
| | Main ore seam | 0–11 |
| | Shale | 2½–5 |
| | Pecten seams | 1½–16 |
| | Shale | 3–6 |
| | Two-foot ore seam | 1½–2½ |
| | Shale | 20–30 |
| | Avicula seam | 0–3 |
| | Sandy sediments | 70–90 |
| Lower Lias | Shale, etc. | |

The main seam is divided in several sections, separated by partings. Accessory constituents are pyrite, barite, sphalerite, and galena, the last often formed in shells. Sphalerite is often seen. There is also,

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[1] Special reports on the mineral resources of Great Britain, *Mem.*, Geol. Survey, 8–13, 1919–1920.

Also A. F. Hallimond, Bedded iron ores of England and Wales, *idem*, 29, 1925, pp. 134.

W. Barnes, Mining iron ore in the Midlands, *Mining Mag.*, London, March, 1918, pp. 120–126.

in the upper part a thin bed of oölitic pyrite, containing 0.11 per cent Ni and Co, 0.015 per cent CuO and 0.15 per cent PbO.

The common types of ore are designated as "chamositic mudstone" and "chamositic, siderite mudstone:" a marine deposit composed of a matrix of fine-grained chamosite with oölites of the same mineral and rhombs and grains of siderite. The minerals are substantially in the condition in which they were deposited. Chamosite is probably the earliest mineral; siderite followed and is rarely oölitic. There is no hematite or original limonite. Hallimond pictures the development of chamosite as follows: Along a muddy, shallow shore the waters were charged with colloidal clay material; as the ferrous iron increases by detritus or organic action, the solubility product of chamosite will be reached and chamosite precipitated. There will take place a progressive transformation of clay to chamosite.

Hallimond gives an interesting analysis of the average composition of the ore.

ANALYSIS OF CLEVELAND HILLS IRON ORE

| | | | |
|---|---|---|---|
| $SiO_2$ | 8.51 | S | 0.05 |
| $Al_2O_3$ | 6.12 | $P_2O_5$ | 1.30 |
| $Fe_2O_3$ | 1.77 | $TiO_2$ | 0.36 |
| FeO | 36.91 | $Cr_2O_3$ | 0.03 |
| CaO | 5.54 | $V_2O_3$ | 0.08 |
| MgO | 3.75 | As | 0.02 |
| $CO_2$ | 20.70 | $K_2O$ | 0.03 |
| $H_2O$ combination | 4.05 | $Na_2O$ | 0.05 |
| $H_2O$ moisture | 10.00 | C | 0.27 |
| | | MnO | 0.42 |

The metallic iron is said to vary between 29 and 30 per cent. A few grains of magnetite are found in the ore.

CALCULATED COMPOSITION

| | |
|---|---|
| $CaSO_4$ | 2.83 |
| $CaCO_3$ | 7.13 |
| $MgCO_3$ | 7.88 |
| $MnCO_3$ | 0.68 |
| $FeCO_3$ | 34.70 |
| Chamosite | 34.24 |

## THE OÖLITIC IRON ORES OF LORRAINE

The so-called "minettes,"[1] or oölitic "limonites" of Lorraine and of Luxembourg, are of the highest importance as present and future

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[1] L. van Werveke, *Zeitschr. prakt. Geol.*, 1895, p. 497; 1901, pp. 396–403.

F. Vilain, Ann. Mines, *Mem.* 1, 1902, pp. 113–290.

P. M. Nicou, in "Iron ore resources of the world," Int. Geol. Congress, Stockholm, 1910.

G. Berg, Struktur und Entstehung der lothringischen Minette Erze, *Zeitschr.* deutsch. geol. Gesell., *Monatsber.*, 1920, p. 77.

L. Cayeux, Les minérais de fer oölitiques de France, 2, Paris, 1922.

resources of European iron. Before the World War there were in France at least 50 mines with an annual ore production of nearly 20,000,000 metric tons (1913). The proved reserves are estimated at 3,000,000,000 tons. In German Lorraine, now ceded to France, the production attained similar figures and the estimated reserves are over 2,000,000,000 tons. The total production in France attained 48,820,526 metric tons in 1930, nearly all of which was derived from these ores. Dipping gently westward the strata attain a depth of 3,000 feet or more, though farther west the iron ore tends to thin out. The area covered is about 113,000 hectares. The principal area consists of the Longwy-Briey basin. The present mining is done at a depth of 800 feet or less, and in part by tunneling or open cuts.

The water to be pumped is heavy, in places reaching 10 cubic meters per minute. The ores lie in the Middle (Dogger) part of the Jurassic

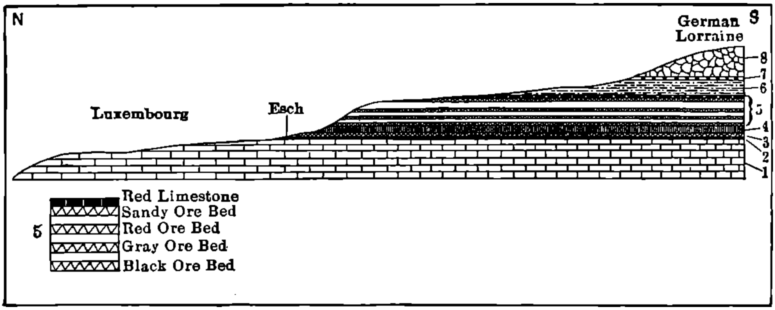

FIG. 109.—Section through the minette measures at Esch; 8, calcareous layer with *Harpoceras humphriesianum;* 7, calcareous layer with *Harpoceras sowerbyi;* 6, marl with *Harpoceras murchisonae;* 5, the minette measure group (see legend); 4, sandstone with *Trigonia navis;* 3 and 2, upper and lower clays with *Harpoceras striatulum;* 1, Lias (micaceous marl). (*After W. Branco.*)

system and occur with shales, sandstones, and marls as distinct beds within a vertical distance of 75 to 150 feet (Fig. 109). The strata are not absolutely persistent at the same level, but are local accumulations, thinning out in lenticular manner. The several beds known as the red, gray, and black beds are of different thickness, the maximum being 15 feet. The minimum exploitable limit is 3 feet. A low percentage of iron, varying from 31 to 40, is characteristic, likewise a high percentage of phosphorus, varying from 1.6 to 1.8, the latter making the ores available for the basic process. From 5 to 12 per cent CaO and from 7 to 20 per cent $SiO_2$ are present. As the iron decreases calcium increases. The ores are earthy and soft and are of brown, gray, or yellow tints.

The detailed and beautifully illustrated monograph of Cayeux is the latest contribution to the study of the Lorraine oölites. The author states that limonite is the most abundant, but probably latest mineral. The other minerals also very important are siderite, chlorite, and hematite.

Oölites of magnetite are known in the Longwy-Briey basin and are probably caused by a slight metamorphism; pyrite is rarely present. The oölites often show fracturing and redeposition; concentric deposition in narrow bands is common; some consist entirely of hematite or limonite.

As to the origin Cayeux still holds to the view expressed in 1909 that the iron oölites replace calcite oölites. The order of minerals is siderite (oldest); chlorite; hematite and limonite; but hematite oölites may also be directly derived from siderite (Fig. 110). The iron is derived from the adjacent land areas. No detailed examination of the chlorite is given, nor any chemical aspect of the process. The results of Cayeux seem to be in conflict with those of Hallimond on the British iron ores.

FIG. 110.—Ore from Silurian beds at La Ferrière-aux-Étangs, France. Magnified 22 diameters. The oölites are chlorite with a kernel of siderite; the fine-grained cement is chlorite and siderite. *a*, oölite of chlorite, in center of lighter color; partly converted into hematite on the outside; *b*, nucleus of corroded pure siderite; *c*, same of yellow altered siderite; *d*, grains of siderite in the cement; *e*, chloritic oölite, partly crushed and invaded by cement; *f*, blackish cement of chlorite and siderite. (*After L. Cayeux.*)

Less important oölites occur in other parts of the Jurassic system in Europe.

## THE OÖLITIC HEMATITE-CHAMOSITE-SIDERITE ORES OF NEWFOUNDLAND

While ores of this type have been described from many places, the monograph by A. O. Hayes on the Wabana ores in Newfoundland is of particular interest.[1] The ores occur in the upper 1,000 feet of flat-dipping

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[1] A. O. Hayes, The Wabana iron ore of Newfoundland, *Mem.* 78, Geol. Survey Canada, 1915; *Econ. Geol.*, 26, 1931, pp. 44–64.

J. B. Gilliott, Folding and faulting of the Wabana ore deposits, *Trans.*, Can. Inst. Min. Met., 27, 1925, pp. 616–634.

Ordovician sandstone and shale and contain several workable beds from 10 to 30 feet in thickness, one of which has been mined for a distance of 1½ miles under the sea. Very large ore reserves are indicated. The ores are fine-grained brown or green oölites and contain some fragments of marine shells, but there is little calcite and no limestone. They average in per cent, 50–70 hematite, 15–25 chamosite, 0–50 siderite, 0–1 calcite, and 1–10 quartz. The oölites consist often of concentric shells of hematite and chamosite such as shown in Fig. 111, and are frequently embedded in a matrix of siderite. The hematite concretions, upon treatment with HCl, yield a residual skeleton of silica.

Fig. 111.—Oölitic chamosite-hematite ore of Wabana with siderite. *a*, spherule of hematite and chamosite in concentric layers; *b*, outer border of siderite, replacing spherule; *c*, spherule of hematite-chamosite partially replaced by siderite; *d*, siderite replacing spherule and matrix. Magnified 110 diameters. (*After A. O. Hayes.*)

It is shown that borings of algæ penetrate both oölites and matrix and that thus the ore was practically in its present condition when covered by later sediments. Oxygen given off by these algæ may have caused oxidation of chamosite to hematite. Direct precipitation of all three iron minerals is, therefore, advocated, though siderite is believed to be the latest and may replace chamosite.

Of exceptional interest are thin beds of pyritic oölite above the "Dominion" bed. They contain graptolites, and the small pyrite concretions lie in an argillaceous matrix with some crystalline quartz.

Similar chamosite-hematite ores have been described from many places.[1]

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[1] L. Cayeux, Les minerais de fer oölitiques de France, 1, Paris, 1909.

W. T. Dörpinghaus, Eisenerzlagerstätten vom Chamosittypus (Spain), Archiv für Lagerstätten Forschung, Heft 16, Berlin, 1914, pp. 53–87.

## REVIEW OF THE SEDIMENTARY IRON ORES

The descriptions given above show that in marshes, lakes, and rivers the hydroxides of iron, mainly limonite, are deposited and that smaller quantities of siderite, iron sulphide, and iron phosphates may be precipitated. Siderite is also deposited in brackish water or under littoral conditions where there is much organic matter present.

Regarding the marine ores, it is certain that glauconite and allied iron silicates are deposited in the sea and that under special reducing conditions siderite and iron disulphide may also form. The probability is also very strong that hematite is developed, in part from oxidation of siderite and glauconite, in part by detrital processes. Whether limonite is ever formed in sea water is more doubtful for the salt solutions have a strong dehydrating effect. Many of the "marine" limonites are products of oxidation of siderite and iron silicates.

The marine iron ores are all shallow water deposits and the frequent oölitic structure is in part at least due to accompanying action of waves and currents. Many of the replacements observed have certainly occurred immediately after deposition. Some geologists, like Cayeux, hold that the ore was a limestone of organic origin which has been later transformed into hematite and siderite by successive replacements, but there seems to be little to support this view.

Wherever iron disulphide is formed reducing conditions prevailed; and the sulphide was precipitated as a colloid.

Any of these oölitic deposits may, of course, have been enriched, after uplift and erosion, by solution of calcite but the iron was certainly not introduced by atmospheric waters.

So we arrive at the conception of shallow bays in which coral reefs flourished or the detritus of older fossiliferous limestone was spread. Into these bays were swept, at intervals, masses of finely divided detritus from the deep mantle of decayed rock of adjacent tropical land areas, undoubtedly rich in hematite as such products always are. The water discharged from the land certainly contained ferrous bicarbonate. In this mud agitated by the waves progressed numerous and complicated reactions. Colloidal iron silicates were formed which quickly oxidized to hematite or were replaced by siderite. Oölites and shells of calcite were replaced by hematite or by siderite but all these complex processes proceeded almost simultaneously. Somewhat similar conditions are found to-day, for instance, on the south side of Molokai, Hawaiian Islands, where such hematite mud is spread out over a large area of shallow coral reef.

Naturally there are extensive sedimentary beds which contain iron but not in sufficient amount to be called iron ores. Among these are, for instance, the glauconitic beds of many formations and the "iron formations" of the Lake Superior region.

In the origin of oölites, colloidal precipitates play an important part. The calcite oölites are believed to have been formed by successive layers of calcium carbonate gel which almost instantly was converted into fibrous calcite. Likewise the concentric structures of siderite and chamosite passed through a gel stage accompanied by adsorption of phosphorus.

## SEDIMENTARY MANGANESE ORES[1]

There is much less manganese than iron in the earth's crust, the average analyses of igneous rock calculated by Clarke showing but 0.078 per cent of manganese. Deposits of manganese ore are also much less common than those of iron ore. Nevertheless, many spring waters carry manganese and a minute amount of it is contained in sea water. Sedimentary deposits of manganese are known, marine and lacustrine as well as fluviatile.

According to experiments by E. C. Sullivan[2] the manganese in rocks is taken into solution more easily than iron, both by carbonated water and by dilute sulphuric acid. He also finds that from mixed ferrous and manganese sulphates almost all of the iron is precipitated by carbonate of calcium before any manganese is thrown down. Fresenius, many years ago, also found that from spring water iron is precipitated first as limonite, while the manganese remains in solution much longer. This accounts for the very general separation of the two metals in the oxidized zone.

Manganese is dissolved mainly as bicarbonate, more rarely as sulphate (p. 363), possibly also as phosphate. It is easily precipitated by oxidation, generally as $MnO_2$ in the form of pyrolusite (63.2 per cent Mn), or as slightly hydrous psilomelane or wad (an impure mixture of manganese oxides), or more rarely as manganite ($Mn_2O_3.H_2O$). The precipitate is generally a "gel," which crystallizes in time, but which has a tendency to adsorb certain oxides, especially those of barium and potassium. Manganese dioxide is, like limonite, precipitated by bacterial action.[3]

**Bog Manganese Ore.**—It has been stated above that many bog iron ores contain manganese; pure bog manganese ores are also known, though the deposits are not abundant. The material is generally earthy and soft, approaching wad in composition. In part the bog manganese consists of a skeleton of hard and glossy black ore containing cavities filled with a black powder. The deposits are rarely more than a few feet in thickness;

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[1] For excellent review see D. F. Hewett in Twenhofel, Treatise on sedimentation, 2d ed., Baltimore, 1932, pp. 565–581.

[2] E. C. Sullivan, quoted by W. H. Emmons in *Bull.* 46, Am. Inst. Min. Eng., 1910, p. 803.

[3] G. A. Thiel, *Econ. Geol.*, 20, 1925, pp. 301–310.

a small occurrence near Wickes, Montana, described by Harder,[1] lies in the flat bottom of a gulch covered by soil and underlain by ochery bog limonite.

A much larger and thicker deposit occurs at Hillsborough, New Brunswick; it is said to extend over 17 acres with a thickness of 6½ feet. An analysis shows Mn, 45.81; Fe, 9.95; S, 0.03; P, 0.05, and $SiO_2$, 5.36 per cent.[2]

**Manganese in Lacustrine and Marine Beds.**—Many sedimentary beds in all parts of the world contain manganese derived from the degradation of old land areas; it occurs as carbonate and stains or concretions of dioxide in tuffs, quartzites, sandstones, clays, shales, and limestones. It is frequently contained in beds of jasper or radiolarian chert. Strongly manganiferous sediments may recrystallize to crystalline schists, the manganese assuming the form of rhodonite, rhodochrosite, or manganese garnet (spessartite). The presence of manganese nodules in deep sea deposits is well known; they are considered to be rather a submarine product of segregation from the red pelagic mud than a chemical precipitate from the ocean. Very rarely, however, do these sedimentary rocks contain manganese of economic importance; and it is only by subsequent concentration by oxidation, especially effective in regions of deep secular decay, that valuable deposits are developed (pp. 362–369).

An excellent example of an undoubtedly sedimentary and practically unaltered deposit is described from Newfoundland by N. C. Dale.[3] It is of little economic importance. The metal occurs as carbonate, with some $MnO_2$, in nodular form, in shaly and calcareous beds of Cambrian age and is associated with calcium phosphate in nodular form, hematite spherules, and barite in crystals and blades; the psilomelane in the deposit also contains barium. Such deposits could probably only form in shallow water mud near land areas subjected to secular rock decay.

There are many such low-grade sedimentary manganese beds. We may recall the extensive manganiferous "iron formation" of Cuyuna Range, Minnesota (p. 302).

The great manganese deposits of the province of Kutais, in Trans-Caucasia (Georgian Republic),[4] are apparently sedimentary, and marine; but it is not impossible that here, too, enrichment by decomposition has taken place. These deposits, said to be the largest in the world, are beds in Eocene clays, marls, and sandstones, the last resting on Creta-

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[1] E. C. Harder, Manganese deposits of the United States, *Bull.* 427, U. S. Geol. Survey, 1910, p. 137.

[2] E. C. Harder, *idem.*, p. 171.

[3] N. C. Dale, The Cambrian manganese deposits of Conception and Trinity Bays, Newfoundland, *Proc.*, Am. Philos. Soc., 54, 1915, pp. 371–456.

[4] C. F. Drake, The manganese ore industry of Caucasus, *Trans.*, Am. Inst. Min. Eng., 28, 1898, p. 191.

E. C. Harder, *op. cit.*, p. 208.

ceous limestone, on the top of an extensive plateau. The ore beds, at the base of the Eocene, are 5 to 8 feet thick, and consist of several strata of oölitic pyrolusite with cementing earthy manganese ore. They are said to extend over an area of 10 square miles. The ores average 40 to 50 per cent Mn and 0.16 per cent P. Drake gives a complete analysis of an ore containing, $MnO_2$, 86.25; $Mn_3O_4$, 0.47; $Fe_2O_3$, 0.61; NiO, 0.3 per cent, and a trace of copper. Barium is present as usual in these ores. The production in 1930 was 1,421,000 tons, even larger than before the World War. Reserves are given as 73,000,000 tons. Similar large deposits of Oligocene age occur at Nicopol, Jekaterinoslaw. Costs of production are said to be about $10 per ton.[1]

## SEDIMENTARY PHOSPHATE BEDS[2]

**Composition of the Calcium Phosphates.**—Phosphorus enters in the average composition of igneous rocks, according to F. W. Clarke, to the extent of only 0.13 per cent, and the analyses of sediments show smaller percentages. Nevertheless, it plays a most important part in the life processes of plants and animals, in the sea and on the land; and in places its compounds accumulate in large masses. Its most common salt is a calcium phosphate; the phosphates of iron, aluminum, lead, and other metals are entirely subordinate.

Apatite, the most common calcium phosphate, also contains $CaF_2$ or $CaCl_2$. The formulas may be written $Ca_5(PO_4)_3F$ and $Ca_5(PO_4)_3Cl$, or $3Ca_3(PO_4)_2.Ca(F,Cl)_2$, the first part of the latter formula being the tri-basic calcium phosphate. Fluorine apatite contains 42.3 per cent $P_2O_5$; chlorine apatite, 41.0 per cent. The pure tri-basic phosphate, which is used as a standard to express the tenor of phosphate rocks, contains 45.8 per cent $P_2O_5$. The phosphate in sedimentary rocks approaches more or less closely the tri-basic phosphate, but sometimes is almost identical with a fluorine apatite.

In deposits of guano a considerable number of acid hydrous phosphates, such as monetite ($CaHPO_4$) and brushite ($CaHPO_4.2H_2O$), have been found, but they have little practical importance. In the same deposits various complex phosphates of iron, magnesium, sodium, and ammonium occur, but these also are unimportant.

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[1] K. V. Markov, *Trans.*, U. S. S. R. Prospecting Service, 27, 1931, 41 pp.

[2] F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 523–534.

O. Stutzer and W. Wetzel, Die wichtigsten Lagerstätten der Nicht-Erze, Berlin, 1932, pp. 1–291.

Eliot Blackwelder, The geologic rôle of phosphorus, *Am. Jour. Sci.*, 4th ser., 42, 1916, pp. 285–298.

*Mineral Resources*, U. S. Bur. Mines, 1930, pt. 1, p. 326.

The mineralogical composition of the marine and residual phosphates is complex.[1] Apatite is essentially a high temperature mineral and has not been recognized in the marine phosphates; in the latter hydrous carbono-phosphates play the principal part. The latter are amorphous and doubtless hardened colloidal precipitates (Fig. 112*A*); there are two species: collophanite ($9CaO.3P_2O_5.CaO.CO_2.H_2O + nH_2O$) and fluocollophanite. The crystalline minerals which in part are altered colloids, in part crusts and mammillary structures, comprise dahllite and francolite both of which are similar carbono-phosphates with or without fluorine.

The marine phosphate rocks, aside from detrital impurities, thus contain calcium carbonate and calcium phosphate; shell fragments and glauconitic granules are frequently present. The poorer kinds may be classified as phosphatic sands, marls, or limestones. The richer varieties are usually oölitic, dark-colored rocks, occasionally with a peculiar whitish efflorescence, and may carry large amounts of organic matter. They are inconspicuous and in places difficult to recognize. The specific gravity, averaging 2.9 in 70 per cent phosphate rock, is considerably higher than that of limestone and may be used to aid in the identification. A rapid field assay with ammonium molybdate is the best test.

**Other Phosphates.**—Among the iron phosphates, the blue vivianite, $Fe_3(PO_4)_2.8H_2O$, is the best known, and appears frequently in bog iron ores. Of the aluminum phosphates, wavellite, $4AlPO_4.2Al(OH)_3 + 9H_2O$, and turquoise, $AlPO_4.Al(OH)_3 + H_2O$, are the best known, the former locally used as a source of phosphorus, the latter as a blue semi-precious stone; both are usually products of the uppermost zone of the crust, sometimes even forming in the zone of oxidation. In a similar geological position occur the lead phosphate, pyromorphite, corresponding in formula to chlorine apatite. Other phosphates, like amblygonite, a fluo-phosphate of lithium and aluminum, monazite, and other phosphates of the rare earths, find their home in the pegmatite dikes. This illustrates the variety of occurrence of the phosphates.

**Phosphate Deposits.**—The many kinds of deposits in which calcium phosphate is of economic importance are shown by the following list:

1. Disseminated in igneous rocks or in their differentiation products of metallic ores.
2. Apatite veins, closely allied to pegmatitic dikes.
3. Marine concretionary beds.

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[1] A. Lacroix, Sur la constitution minéralogique des phosphorites françaises, *Compte Rendu*, 150, 1910, p. 1213.

H. S. Gale and R. W. Richards, *Bull.* 430, U. S. Geol. Survey, 1910, p. 464.

W. T. Schaller, *Bull.* 509, U. S. Geol. Survey, 1912, pp. 89–100.

A. F. Rogers, A review of the amorphous minerals, *Jour. Geol.*, 25, 1917, pp. 515–541.

A. F. Rogers, *Am. Jour. Sci.*, 5th ser., 3, 1922, p. 269.

4. Sub-aërial accumulations of animal excrement—bat caves, guano islands.

5. Metasomatic deposits by replacement of limestone by means of phosphate solutions, from Nos. 3 and 4.

6. Residual concretions, by action of atmospheric waters on No. 3.

**Use.**—The principal use of calcium phosphate is for soil fertilization; all the classes enumerated above are so utilized. Under No. 1 comes, for instance, the apatite concentrate from the Adirondack magnetite ores; under No. 2 the apatite veins of Canada and Norway; the occurrences of the remaining classes are described below.

For utilization it is necessary to transform the insoluble tribasic phosphate into soluble form. This is generally effected by a 60 per cent solution of sulphuric acid;[1] hence the dependence of the phosphate industry on an abundant and cheap supply of sulphuric acid, illustrated, for instance, in the establishment of large sulphuric acid plants at the pyritic copper deposits of Ducktown, Tennessee, for the treatment of the sedimentary phosphates of the Southern states. The treatment with $H_2SO_4$ results in a partial decomposition, with the formation of soluble calcium phosphate, also called super-phosphate or mono-calcium phosphate ($CaH_4(PO_4)_2.H_2O$), and also some di-calcium phosphate, which is much less soluble. The standard is 77 per cent of the tri-basic calcium phosphate with less than 3 per cent of alumina plus iron, but not all of the production reaches this grade.

Experiments show that even the tri-calcium phosphate or apatite is soluble, particularly in water containing carbon dioxide; its solubility in solutions of $CaCO_3$ or in pure water is slight, but the presence of sodium chloride increases the solubility.[2] The marked absorption of phosphoric acid by clays and soils is held to be due to the presence of colloidal bodies.

**Production.**—Though some phosphates are obtained from apatite deposits and from basic slags, the greater part comes from sedimentary and residual beds. In the United States, the bulk of the production comes from Florida, Tennessee, and South Carolina, in the order named; by far the most is mined in Florida.[3] The yield of the United States in 1930 was 3,951,363 long tons. Large quantities were exported. The average price of phosphate was $3.56 per long ton in 1930.

Of other countries Algeria, Morocco, and Tunis produced about 5,951,000 metric tons, and the oceanic islands about 600,000 metric tons

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[1] The reaction is expressed by the following formula: $Ca_3(PO_4)_2 + 2H_2SO_4 = CaH_4(PO_4)_2 + 2CaSO_4$.

[2] H. E. Patten and W. H. Waggaman, Absorption by soils, *Bull.* 52, Bur. Soils, U. S. Dept. Agriculture, 1908.

O. Schreiner and G. H. Failyer, Absorption of phosphates and potassium by soils, *Bull.* 32, Bur. Soils, U. S. Dept. Agriculture, 1906.

[3] R. W. Stone in J. E. Spurr and F. S. Wormser, The marketing of metals and minerals, 1925, pp. 467–485.

in 1929. Africa is now supplanting the United States in the markets of Europe.

**Origin of the Phosphate Rocks.**—As all land animals absorb phosphoric acid and segregate it as calcium phosphate in their bones and excrements, it is not difficult to understand the accumulation of phosphates wherever animal life is particularly abundant and undisturbed. Besides phosphates, such deposits contain much ammonia and nitrogen, except where subjected to leaching by heavy precipitation. Of this kind are the bone beds which are found occasionally in various formations and in caves.

The guano of commerce is deposited by sea birds congregating in enormous numbers on desert coasts and oceanic islands, for instance, along the Peruvian and Chilean coasts, on Christmas Island in the Indian Ocean, and in the West Indies. Some of these deposits cover whole islands and, in places, may accumulate to a depth of 100 feet; and it is stated that under favorable circumstances the rate of deposition is rapid. The guano of dry climates varies greatly in texture and color, but generally is granular, light colored, and porous. It contains on an average 10.90 per cent nitrogen, 27.60 per cent phosphates, and 2 to 3 per cent potash.[1]

The West Indian deposits—for instance, those on Navassa[2] and Sombrero islands—have been leached and are in part hard and compact, in part porous and friable. The phosphate has been concentrated to 70 or 75 per cent. The material contains from 21 to 40 per cent of phosphoric acid, 1 to 2 per cent sulphuric acid, 20 to 45 per cent lime, usually also much ferric oxide and alumina. The underlying limestone or igneous rock may be locally replaced by the phosphatic solutions.

The marine phosphate beds also derive their material from animal life. Sea water contains phosphoric acid, though the quantity is extremely small, and likewise some fluorine, each amounting to about a little less than one part per million.

Many shells, particularly those of the older formations, are rich in phosphorus and fluorine.[3] A Cambrian *Obolus* contained 36.54 per cent $P_2O_5$ and 2.78 per cent F; a recent *Lingula* yielded 23.20 per cent $P_2O_5$ and 1.52 per cent F.[4] The shells of crustaceans contain up to 26 per cent $Ca_3P_2O_8$. Pteropods, lamellibranchs, gastropods, corals, and protozoans also carry phosphorus though in most cases the quantity is small. Phos-

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[1] R. A. F. Penrose, Jr., *Bull.* 46, U. S. Geol. Survey, 1888.

F. D. Power, Phosphate deposits of the Pacific, *Econ. Geol.*, 20, 1925, pp. 266–281.

[2] E. V. D'Invilliers, Phosphate deposits of the Island of Navassa, *Bull.* Geol. Soc. Am., 2, 1891, p. 71.

[3] F. W. Clarke and W. C. Wheeler, The inorganic constituents of marine invertebrates, *Prof. Paper* 124, U. S. Geol. Survey, 1922.

[4] Andersson and Sahlbom, Ueber den Fluorgehalt schwedischer Phosphorite, *Bull.* 4, Geol. Inst. Upsala, 1900, p. 79. *Neues Jahrbuch,* Ref., 1, 1903, pp. 195, 197.

phorus and fluorine are found in the bones and teeth of fishes. The marine sediments, then, all hold more or less phosphates; and it is a matter of some surprise that fluorite does not more commonly occur in sedimentary rocks.[1]

In some beds the phosphates occur disseminated in small quantities, in part as small concretions, in part remaining in the shell fragments. In the more valuable deposits the phosphates appear in more concentrated form and characteristically assume the forms of nodules, or concretions (sometimes of large size), or oölitic rocks, built up of small oölites in part of concentric and fibrous structure. The nodules have often a shell nucleus and, as a result of enrichment, may contain more phosphate in the peripheral than in the central parts (Fig. 112*B*).

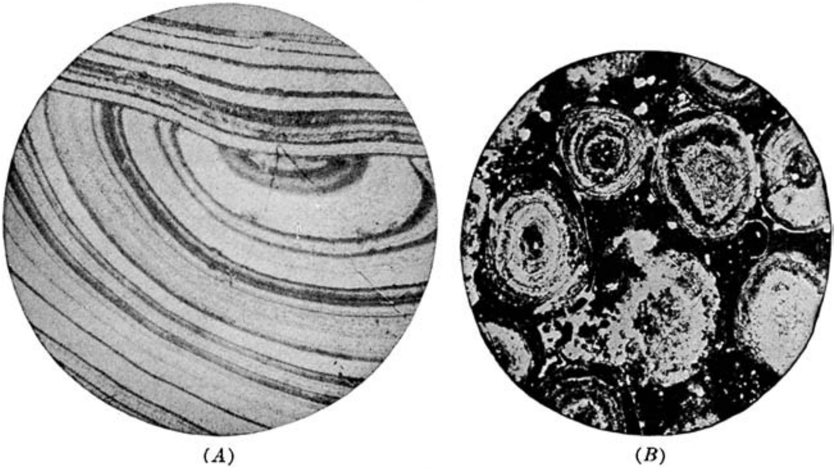
(*A*) (*B*)

FIG. 112.—*A*, colloform, isotropic phosphate, Ocean Island, South Pacific. Magnified 30 diameters. *B*, oölitic phosphate, Cokeville, Wyoming. Magnified 30 diameters. (*Reprinted by permission from "Economic Geology," H. Ries, 6th ed., published by John Wiley & Sons, Inc.*)

While phosphate nodules have been brought up by the dredge from great oceanic depths, the conditions for their formation are probably best at moderate depths, near shores, where the marine life is most abundantly developed or, as pointed out by some authors, where sudden changes of temperature, owing to conflicting currents, kill large numbers of marine organisms.

The origin of the oölitic and nodular phosphate rocks, in some of which recognizable organic remains are scarce, has been discussed extensively, but is as yet not fully explained. It is believed that ammonium phosphate may form in the organic matter and that this reacts on shell

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[1] K. Andrée, Ueber einige Vorkommen von Flusspath in Sedimenten, *Tschermak's min. u. petr. Mitt.*, 28, 1909, pp. 535–562.

H. S. Gale and R. W. Richards, *Bull.* 430, U. S. Geol. Survey, 1910, p. 463.

remains, replacing them with calcium phosphate, which eventually accumulates in larger concretions.[1] These processes are likely to continue for some time at least after the sedimentation, in the yet soft sediments.

After the beds have been uplifted and exposed to weathering, enrichment takes place easily by the removal of calcium carbonate. This is especially effective in regions of deep rock decay, as in the Southern states. The rock phosphates of Utah and Idaho have remained almost unaltered. Humus acids and $CO_2$ appear to be the best solvents for phosphates.[2]

The cycle of migration of the phosphates is a fascinating study. From their original home in the igneous rocks they are dissolved by surface waters and absorbed by all living things, vegetable and animal, on land and sea. After the death of the organisms the phosphates return to the soil or to the sedimentary beds to be dissolved and used anew by other generations.

**Occurrences of Phosphate Rocks.**—Deposits of phosphate rock are found in the marine beds of all ages and in almost all countries, at least from the Cambrian, when the segregation of phosphoric acid by the inhabitants of the sea appears to have begun, to the Tertiary; and in the present oceans such deposits certainly continue to form. In description it is impracticable to separate the primary marine deposits from those altered by weathering.

Large deposits, enriched by weathering, are worked in the Cretaceous beds of northern France. In the southwestern part of that country, in the departments of Lot and Lot-et-Garonne, phosphates occur in irregular fissures with clay in Jurassic limestone.[3] These deposits are probably formed by replacement effected by solutions descending from low-grade sedimentary phosphate beds.

Phosphate beds are now mined on a large scale along the frontier of Algeria and Tunis, also in Morocco.[4] The beds occur in the lower Eocene, which covers Cretaceous strata, and consist in part of large

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[1] Renard and Cornet, *Bull.* 21, sér. 3, Acad. Belgique, 1891, p. 126.

L. Kruft, *Neues Jahrbuch,* Beil. Bd. 15, 1902, pp. 1–65. Ref. in *Zeitschr. prakt. Geol.*, 10, 1902, p. 301.

R. Delkeskamp, *Zeitschr. prakt. Geol.*, 12, 1904, p. 299.

[2] W. A. P. Graham, Experiments on the origin of phosphate deposits, *Econ. Geol.*, 20, 1925, pp. 319–334.

[3] L. de Launay, Gîtes minéraux, 1, 1913, p. 679.

[4] M. Blayac, Description géologique de la région des phosphates du Dyr et du Kouif, *Ann. des Mines* (9) 6, 1894, pp. 319–330.

O. Tietze, Die Phosphatlagerstätten von Algier und Tunis, *Zeitschr. prakt. Geol.*, 1907, p. 229.

On the Russian phosphates, see W. Tschirvinski, *Neues Jahrbuch,* Bd. 2, 1911, p. 51. For reports by J. V. Samoiloff see *Miner. Mag.*, 18, 1917, p. 87.

A. C. Lawson, The phosphate deposits of Kourigha, Morocco, *Econ. Geol.*, 26, 1931, pp. 480–484.

concretions in marl, sometimes carrying the rich phosphate only as a crust; other beds are formed of a soft material consisting of small and smooth brown or yellowish grains of phosphate cemented by calcite and containing many fossils and much bituminous matter. The thickness of the richest phosphatic stratum is said to be 10 to 15 feet.

The deposits found in the United States are mainly in three regions—(1) the Atlantic coast belt of Tertiary rocks in the Carolinas and Florida; (2) the Tennessee area of Silurian and Devonian strata; (3) the Utah-Idaho region of Carboniferous beds.

The phosphates of the Utah-Idaho region[1] (production, in 1930, 67,276 long tons) are of great extent and prospective value. Owing to difficulties and cost of transportation, the production is restricted.

They extend north of Ogden, Utah, through Idaho, Wyoming, and Montana into Canada. The best deposits are in the ranges which constitute the northern continuation of the Wasatch. Their position is in the Phosphoria formation of the Permian, which has an average thickness of 600 feet and consists of limestones, cherty in part, phosphate beds, and shales. The phosphate horizon is in the middle of the formation and the beds have an average thickness of 200 feet (see Figs. 13, 14, and 113). The rocks are massive brown to gray phosphatic shales and beds of rock phosphate with some limestone. The richest bed mined at Montpelier, Idaho, carrying 70 per cent or more of $Ca_3P_2O_8$, lies at the base of the phosphate section and is 5 or 6 feet thick. It is a black to dull-gray oölitic rock, with cryptocrystalline ovules or concretions of all sizes up to one-half inch in diameter (Fig. 112). The oölitic phosphates contain small but definite amounts of chromium and vanadium doubtless of organic derivation.

Large sections of the phosphatic beds, in places a thickness of 75 feet, carry from 30 to 50 per cent of $Ca_3P_2O_8$. The beds are folded and locally have steep dips. The rock is hard and the mining is carried on by underground operations. Very little enrichment is noted.

The phosphates of western Tennessee[2] have been worked since 1894 and at present yield about 600,000 long tons per annum. They are of three classes: (1) Brown residual phosphates, resulting from leaching of Ordovician phosphatic limestones. The beds are from 3 to 8 feet thick and

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[1] F. B. Weeks and W. F. Ferrier, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 449–462.

H. S. Gale and R. W. Richards, *Bull.* 430, *idem,* 1910, pp. 457–535.

Eliot Blackwelder, *Bull.* 430, *idem,* 1910, pp. 536–551.

R. W. Richards and G. R. Mansfield, *Bull.* 470, *idem,* 1911, pp. 371–439; also Bull. 577, *idem,* 1914.

G. R. Mansfield, *Prof. Paper* 152, U. S. Geol. Survey, 1927, pp. 208–308, 361–373.

[2] C. W. Hayes, The Tennessee phosphates, *Seventeenth Ann. Rept.,* U. S. Geol. Survey, pt. 2, 1896, pp. 513–550.

R. W. Smith, Geology and origin of the phosphate deposits of Tennessee, *Eng. and Min. Jour.,* 132, July 27, 1931, pp. 58–62.

carry as much as 80 per cent of tricalcium phosphate. (2) The blue or black bedded phosphates occur in beds of Devonian age and show variations from oölitic through compact and conglomeratic to shaly forms. The high-grade rock is seldom more than 20 inches thick. The nodular variety, which is embedded in a green sand, carries about 60 per cent $Ca_3P_2O_8$. (3) The white phosphate, which is a post-Tertiary product of replacement or filling of cavities of limestone of Carboniferous age. None of it is now mined.

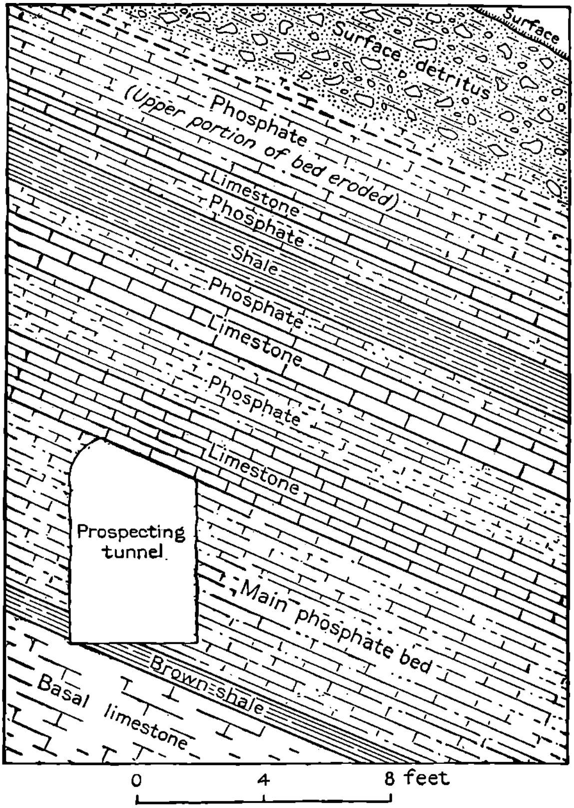

FIG. 113.—Section showing beds of phosphate, Montpelier, Idaho. (*After Weeks and Ferrier, U. S. Geol. Survey.*)

The phosphate beds of North and South Carolina,[1] discovered in 1867, extend along the coast for a distance of 60 miles. They are contained in loose beds of Miocene age, rich in fossils. The land deposits lie at a shallow depth and consist of so-called pebble rock, a solid mass from which the calcium carbonate has been leached and partly replaced by phosphate; the solution cavities give this material the appearance of a

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[1] G. S. Rogers, Phosphate deposits of South Carolina, *Bull.* 580, U. S. Geol. Survey, 1914, pp. 183–220.

mass of separate pebbles. The rock varies from 1 to 3 feet in thickness and is covered by a green sandy marl. Similar deposits have been dredged in the rivers, and consist essentially of water-rounded fragments of the land rock. The mining is carried on by steam-shovel or dredge operations.

The phosphate deposits of Florida[1] are next to those of North Africa the most productive in the world and large quantites are exported. In 1930, 3,261,539 long tons were produced.

The deposits follow in the main the northwestern coast of the State but lie some distance from the shore. There are several types, all contained in the Alum Bluff formation or above it, or in the underlying Vicksburg limestone, both formations of Oligocene age. The clays, marls and sandstones of the Alum Bluff contain in several horizons abundant smooth yellowish or brown nodules or ovules of phosphate which are considered by Matson and Sellards as of primary deposition and the source of all the other deposits. These beds are worked in some places but are not of great importance. The material is very similar to the phosphates of Gafsa in Tunis.

The so-called "land pebble" deposits of Polk and Hillsborough counties, which are the most important, are rudely stratified detrital and residual masses; they rest on the Alum Bluff formation and are believed to be derived from this source. They are considered to be of Miocene or Pliocene age. They contain concretions of white phosphate, averaging 65 to 70 per cent tribasic calcium phosphate, while the finer matrix often contains 20 to 30 per cent of the same. They lie at elevations of about 100 feet and form parts of gravel beds with casts of shells, sharks' teeth, and bones of mastodon, horse, and rhinoceros. The average depth of the deposit is about 12 feet, that of the sandy overburden up to 40 feet. The workable deposits average several acres, though some cover as much as 40 acres.

The type of deposit called "rock phosphate" occurs in the Vicksburg limestone of lower Oligocene age and rests in depressions on its surface. They are of secondary origin and are believed to be leached from the overlying Alum Bluff formation. The concentration was effected by chemical and mechanical means; the result is a mass of rounded or subangular phosphate concretions ("rocks" or "pebbles"), in places rudely sorted in layers of coarse and fine, contained in a matrix of sand, clay, or soft phosphate.

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[1] E. H. Sellards, *Fifth Ann. Rept.*, Florida Geol. Survey, 1913, pp. 23–80.

E. H. Sellards, *Trans.*, Am. Inst. Min. Eng., 50, 1914, pp. 901–916.

G. C. Matson, The phosphates of Florida, *Bull.* 604, U. S. Geol. Survey, 1915.

*Mineral Resources*, U. S. Bur. Mines, pt. 2, 1932, pp. 315–322; *idem*, for research on flotation and gravity concentration to improve output.

B. I. Johnson, Phosphate rock, U. S. Bur. Mines, *Information Circ.* 6256, 1930.

Pleistocene sand, up to 50 feet thick, covers the deposits. There are many deposits covering from 5 to 35 acres, while some are mere pockets in the limestone. The average thickness is 30 feet; transitions to the underlying limestone are sometimes observed. The concretions or nodules vary from a few inches to 10 feet in diameter, are close grained, light gray, and sometimes show cavities lined with secondary and mammillary phosphate. Shells are rare but fragments of bones and sharks' teeth occur here and there. The processes which have operated in the concentrations from the overlying Alum Bluff beds are replacement of limestone by phosphate and solution of residual limestone with resulting slumping; in places there has been mechanical transportation. The soft phosphates are thought to be formed by replacement of porous limestone. The percentage of "recoverable" phosphate in the deposits does not average much above 15 per cent. The higher grades of land-pebble phosphate (76 to 77 per cent) sell at $6.50 per ton.

ANALYSES OF PHOSPHATES

| | I | II | III |
|---|---|---|---|
| Insoluble | 1.82 | 6.69 | 3.05 |
| $SiO_2$ | 0.30 | .... | ...... |
| $Al_2O_3$ | 0.50 | 2.14 | 1.09 |
| $Fe_2O_3$ | 0.26 | 0.61 | 0.64 |
| $MgO$ | 0.22 | 0.33 | 0.57 |
| $CaO$ | 50.97 | 46.03 | 48.58 |
| $Na_2O$ | 2.00 | ...... | 0.09 |
| $K_2O$ | 0.47 | ...... | |
| $H_2O-$ | 0.48 | 0.79 | ...... |
| $H_2O+$ | 0.57 | 3.47 | ...... |
| $CO_2$ | 1.72 | 3.93 | 4.60 |
| $P_2O_5$ | 36.35 | 31.50 | 29.74 |
| $SO_3$ | 2.98 | ...... | 2.75 |
| Cl | | ...... | 0.11 |
| F | 0.40 | 1.86 | 2.12 |
| Organic | | ...... | 7.45 |
| | 99.04 | 97.35 | 100.79 |

I. Crawford Mountains, Utah. Geo. Steiger, analyst, *Bull.* 430, U. S., Geol. Survey, 1909, p. 465. No titanium, organic matter not determined, trace chlorine.

II. Florida. Land pebble, G. H. Eldridge.

III. Gafsa, Tunis. O. Tietze, *Zeitschr. prakt. Geol.*, 16, 1907, p. 248.

Analysis calculated on dry material, 3.81 per cent $H_2O$. $P_2O_5$ equivalent to 64.93 per cent $Ca_3P_2O_8$; 2.35 per cent $CaF_2$; 4.67 per cent $CaSO_4$; 10.45 per cent $CaCO_3$.

The so-called "river pebbles" are Pleistocene deposits in the present rivers but are not worked now.

Mineralogically, the Florida phosphates are held by Matson to consist of collophanite with francolite (p. 290).

The material is mined by steam shovel, by hydraulic method, or by dredge; it is then washed to remove the clay and afterward crushed, screened, and hand picked. Poorer materials are successfully concentrated by soap flotation.

The chemical composition of the marine and residual phosphates is shown in the analyses on page 291, of which I and III represent unaltered marine deposits and II a residual occurrence.

## CHAPTER XIX

# SEDIMENTARY IRON ORES, REGIONALLY METAMORPHOSED

### PROCESSES INVOLVED

Mineral deposits are usually formed during comparatively brief epochs, in which uniform conditions prevail, rendering a given set of minerals stable. In the development of the epigenetic deposits this is not invariably true, for we sometimes find evidence of successive changes in the mineral-bearing solutions; early minerals are dissolved and a new set formed. The replacement of calcite veins by silica offers an instance of this process, as do also the successive generations of minerals in many metalliferous veins and in pegmatite dikes.

After the epoch of mineralization has passed, the deposit will, as a rule, be subjected to different temperatures and different degrees of pressure, and solutions of various kinds will percolate through it. Consequently, in many deposits the minerals of their ores are now unstable and only the slowness of the changes may prevent them from being wholly altered. "Persistent" minerals remain unaffected except by mechanical deformation, but very few minerals are persistent in all zones.

In general, when by erosion, intrusion, or dynamo-metamorphism a mineral deposit is transferred to a new zone, the characteristic minerals of this zone will develop in it and become superimposed upon the original minerals. Some deposits have a complicated history, having been subjected to several changes, each of which has left its imprint on the ores.

It often happens that a deposit becomes involved in folding or dynamic metamorphism of general or local kind; there will then be mechanical deformation; veins and irregular masses will be squeezed out into lenses which may in places overlap or imbricate. The minerals of dynamic metamorphism, such as garnet, amphibole, and biotite of the lower zones, or chlorite, epidote, zoisite, muscovite, albite, and talc of the upper zones, will be formed from the old constituents. Hydrates may lose their water and carbonates their carbon dioxide. The quartz-sulphide veins are least affected, their minerals being comparatively persistent.

Most deposits have been exposed to static metamorphism at moderate temperature, during which chlorite and carbonates have developed. Increased temperature may leave some deposits unaltered, while others in the vicinity of igneous masses may be profoundly modified. Examples are known of sedimentary deposits of limonite or siderite which, close to intrusive rocks, change to magnetite and specularite and in which garnets and other silicates develop. Such deposits may simulate those of pyrometasomatic origin, but in the latter the ores did not exist in the sedimentary rocks but were introduced by solutions. Some of the deposits in the pre-Cambrian terranes owe their complex nature to successive changes, and their history may be most difficult to unravel.

**The Banded Quartz-hematite (Magnetite) Ores.**—In many parts of the world the pre-Cambrian beds contain conspicuously banded formations, often folded and corrugated, which are essentially made up of fine-grained quartz of various, often red, colors alternating with fine-grained hematite with which more or less magnetite is admixed. They are also known as banded, ferruginous chert or jasper, jaspilite, taconite, or "banded iron stones." Neither "jasper" nor "chert" is a strictly applicable name, though in places the silica is so fine-grained as to suggest chalcedony. Stronger metamorphism sometimes adds an amphibole, or other silicates, to the assembly. On account of high silica they are rarely suitable iron ores, and effective concentration is difficult to apply. Nature, however, sometimes provides methods by which high-grade bodies of hematite result from poor protore.

These "iron formations" occur abundantly in the Lake Superior region (p. 296) and are widely distributed in South Africa,[1] in the Dharwars of India,[2] in Western Australia, in Scandinavia, and in Brazil and Venezuela.

Although the typical concretionary and oölitic ores are absent from these terranes a sedimentary origin is in most cases proved and admitted, but opinions may differ as to the exact mode of genesis. Many of the deposits are chemical precipitates and probably all are shallow water marine deposits. Fossils are practically absent. The iron formations may have a thickness of many hundreds of feet.

In the Lake Superior district the original material has been proved to consist of siderite, an oölitic iron silicate, and silica, for of these original iron minerals a considerable portion is preserved. Wagner has shown that siderite was probably the original mineral in the Swaziland and the Witwatersrand systems. In the higher "iron stones" of the Pretoria series of the Transvaal System the iron ore is a partly recrystallized oölitic ore belonging to the magnetite-hematite-chamosite group. The

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[1] P. A. Wagner, The iron deposits of the Union of South Africa, *Mem.* 26, Geol. Survey South Africa, 1928. An important and complete account which every one interested in the Lake Superior Region should read. Wagner accepts the oxidation and leaching theory of Van Hise and Leith.

[2] C. M. Weld, The ancient sedimentary iron ores of India, *Econ. Geol.*, 10, 1915, pp. 435–452. Sir Thomas H. Holland, Twelfth Internat. Geol. Congress, Canada, 1913, p. 376, (Ottawa), 1914. J. A. Dunn, Origin of iron ores at Singhbhum, *Econ. Geol.* 30, 1935, pp. 643–654.

transformation to the typical "banded iron stone" requires, therefore, an oxidation of siderite and iron silicate, and the concentration to hematite ore requires in addition that the quartz should be removed by solution or else replaced by extraneous hematite. Regarding these oxidation and concentration processes the opinions are not unanimous.

### HEMATITE DEPOSITS OF THE LAKE SUPERIOR REGION

**General Character, Distribution.**—The iron ores mined in the Lake Superior region in Minnesota, Michigan, and Wisconsin amount to from 80 to 90 per cent of the total domestic output and in 1930 yielded 49,383,385 long tons. Since 1930 a rapid decline has set in, as in the production of all other metals. The world's production of iron ore in 1929 was about 200,000,000 metric tons. The ore is mainly hematite with small admixtures of limonite and magnetite. It occurs as masses, lenses, or flat deposits in pre-Cambrian sedimentary rocks. The deposits are believed to have been concentrated by the oxidizing and silica-dissolving effect of waters of meteoric origin, in original sediments, called "iron formations," which were originally rich in carbonate and silicate of iron. In their present form they are products of pre-Cambrian weathering which, probably under arid conditions, reached depths not approached elsewhere. Only to a small degree and near the surface does this ore-forming activity of the waters persist at the present time.

We owe most of our information concerning these deposits to the work of C. R. Van Hise, C. K. Leith, and many others recorded in a series of monographs of the United States Geological Survey. These and other papers are cited below.[1]

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[1] R. D. Irving and C. R. Van Hise, (Penokee district) *Mon.* 19, U. S. Geol. Survey, 1892. C. R. Van Hise and W. S. Bayley, (Marquette district) *Mon.* 28, 1897. J. M. Clements and H. L. Smyth, (Crystal Falls district) *Mon.* 36, 1899. C. K. Leith, (Mesabi district) *Mon.* 43, 1903. J. M. Clements, (Vermilion district) *Mon.* 45, 1903. W. S. Bayley, (Menominee district) *Mon.* 46, 1904.

C. R. Van Hise and C. K. Leith, The geology of the Lake Superior region, *Mon.* 52, *idem*, 1911.

S. Weidman, The Baraboo iron-bearing district, Wisconsin, *Bull.* 13, Wisconsin Geol. and Nat. Hist. Survey, 1904.

C. K. Leith, R. J. Lund, and A. Leith, Pre-Cambrian rocks of the Lake Superior region, *Prof. Pap.* 184, U. S. Geol. Survey, 1935, 34 pp.

E. C. Harder and A. W. Johnston, Notes on the geology of the Cuyuna district, *Bull.* 660, U. S. Geol. Survey, 1917, pp. 1–26; also, *Bull.* 15, Minnesota Geol. Survey, 1918.

J. F. Wolff (Mesabi range), *Trans.*, Am. Inst. Min. Eng., 56, 1917, pp. 142–169.

J. W. Gruner, The origin of sedimentary iron formations (Mesabi), *Econ. Geol.*, 17, 1922, pp. 407–460.

W. H. Collins, T. T. Quirke, and E. Thomson, Michipicoten iron ranges, *Mem.* 147, Canada Geol. Survey, 1926.

W. O. Hotchkiss, Lake Superior Region, *Guidebook* 27, 16th Internat. Geol. Congress, Washington, 1933, 101 pp.

There are seven principal districts in the United States and three or four in Canada, locally called ranges, as follows (Fig. 114):

1. The Mesabi, Vermilion, and Cuyuna ranges of northern Minnesota.

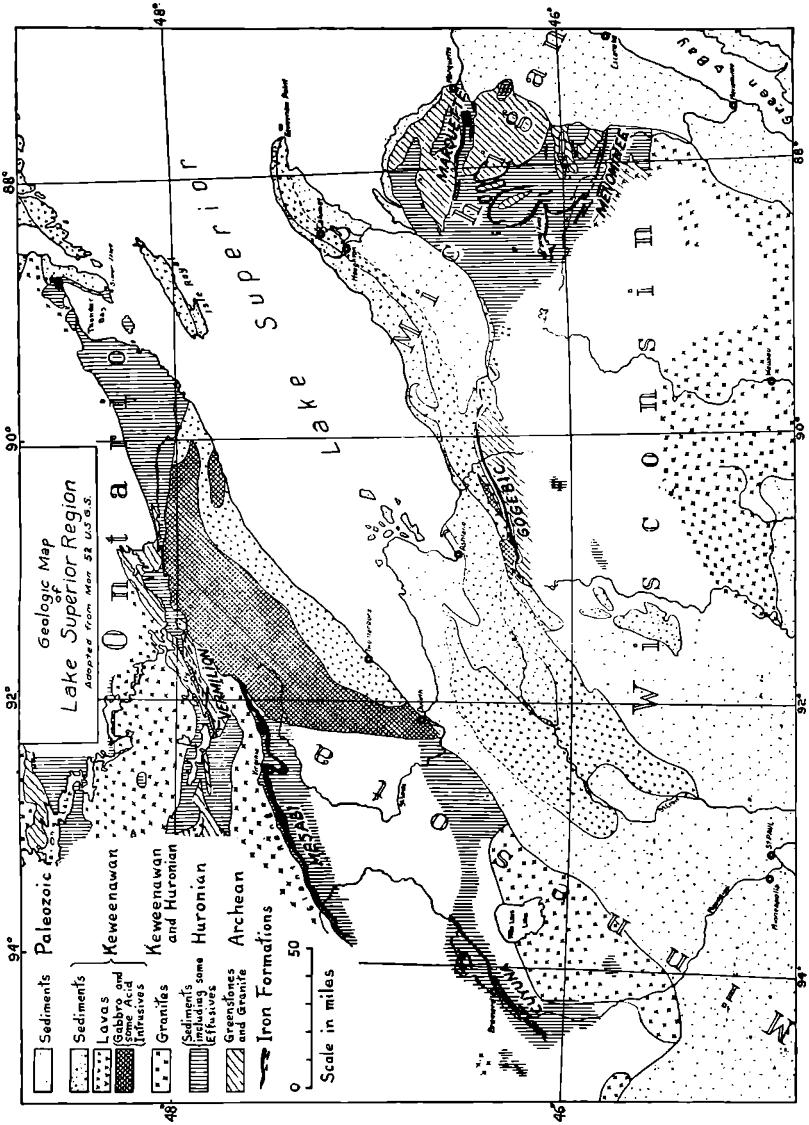

FIG. 114.—Geologic map of Lake Superior region. (*After C. R. Van Hise and C. K. Leith.*)

2. The Penokee-Gogebic, Marquette, Iron River, and Menominee ranges, mainly in northern Michigan.

The Baraboo range of southern Wisconsin and the Michipicoten, Gunflint Lake, and other districts in Canada are of minor importance.

### SUCCESSION OF PRE-CAMBRIAN ROCKS

| | LAKE SUPERIOR REGION UNITED STATES OF AMERICA | CANADA |
|---|---|---|
| | Cambrian: Potsdam sandstone. | Paleozoic beds. |
| | Great Unconformity and Peneplanation | |
| Keweenawan Series | Copper metallization.<br>Intrusions of gabbro, granite, etc. | Copper-silver-nickel metallization.<br>Intrusion of gabbro, diabase, and granite (Killarney). |
| | Sandstone.<br>Basic lavas. | Sandstone.<br>Basic lavas. |
| | Unconformity | |
| Huronian Series | Upper Huronian:<br>Virginia and Michigamme slates, Goodrich quartzite. Deerwood iron formation and upper iron formations of Crystal Falls and Iron River. | Animikie series at Thunder Bay (Whitewater series with tuffs).<br>Cobalt series (exact relation to Animikie and Whitewater unknown).<br>Quartzite and tillite 10,000 + feet. |
| | Unconformity. | |
| | Middle Huronian:<br>Quartzite, slate, and most iron formations: Biwabik, Vulcan, Negaunee, Ironwood. | Unconformity and erosion. |
| | Unconformity. | |
| | Folding. | Folding. |
| | Granite intrusions:<br>Giants Range granite. | |
| | Lower Huronian:<br>Quartzite, dolomite, and slate of Marquette, Menominee, and Gogebic ranges. Knife Lake slates. | Bruce series:<br>Quartzite, limestone, and graywacke 15,000 + feet. |
| | Great Unconformity and Erosion | |
| Archean or Basement Complex | Folding. | Basic intrusives.<br>Post-Timiskaming folding. |
| | | Gold-copper metallization: |
| | Laurentian:<br>Main granite intrusions. | Syenite and syenite porphyry.<br>Granite, granodiorite. |
| | Intrusive Contact | |
| | | Timiskaming series:<br>Graywacke and conglomerate (intercalated lavas).<br>Sudbury series.<br>Unconformity. |
| | Keewatin:<br>Lava flows with intercalated iron formation and other sediments. Soudan iron formation at Vermilion Range. | Keewatin series:<br>Basic flows and other effusives more or less metamorphosed. Some intercalated iron formations and other sediments (Dore series).<br>In eastern Canada the Keewatin is represented by limestone (Grenville) and amphibolite, intruded by granite, gneiss, and alkaline rocks with attendant pyrometamorphism. |

**Geology.**—Divisions of the pre-Cambrian geology in the Lake Superior region are given on page 297; on the right are the corresponding Canadian formations.

The Archean, or basement, complex consists of gneiss and granite with an extensive series of greenstones (basalt, amphibolite) which are largely surface lavas. These lavas are now regarded as the oldest formation exposed; the character of the basement upon which they were outpoured is unknown. The gneisses and granites are in large part intrusive into the Keewatin and Temiskaming series.

In the Vermilion and Michipicoten districts the productive formation is in the Keewatin series.

Unconformably overlying the Archean is the Huronian, which in the Lake Superior region consists of three parts separated by uncomformities. The principal iron-bearing formations are concentrated in the Middle Huronian, but the development differs materially in the several districts.

In the Marquette district all three divisions of the Huronian are present. The lower Marquette series consists of quartzite, dolomite, and slate 3,000 feet in maximum thickness. The middle Marquette series, 3,000 feet in maximum thickness, includes quartzite, slate, and the important Negaunee iron-bearing formation. The upper Marquette series includes quartzite, schist, slates, and fragmental basic volcanic rocks, accompanied by some iron-bearing formations

In the Crystal Falls and Menominee districts similar divisions appear.

In the Penokee-Gogebic district the Lower Huronian consists of quartzite and dolomite; the Middle Huronian includes a basal quartzite and the Ironwood "iron formation" which ranges up to 1,000 feet in thickness.

In the Mesabi district the Lower Huronian consists of conglomerates, graywackes, and slates standing vertically; it is intruded by the granite of the Giants Range, on the south slopes of which the iron deposits extend from east to west for a distance of 100 miles. The Middle Huronian comprises a basal quartzite, the Biwabik "iron-bearing formation"; the overlying Virginia slate is considered Upper Huronian. The total thickness is probably over 2,000 feet. The series dips gently at angles of 5° to 20° and is gently cross folded. Intrusive into these rocks at the east end of the district are Keweenawan granite and basic igneous rocks. Near these intrusives the sedimentary rocks are highly metamorphosed.

Above the Huronian rests the less metamorphosed Keweenawan series of sandstones, conglomerates, and igneous basic flows; the thickness is estimated to be over 35,000 feet. It contains deposits of native copper but no iron deposits.

The west end of Lake Superior consists of an eastward-pitching synclinorium of Keweenawan rocks. The next underlying series, the Upper Huronian, takes less part in this synclinal structure.

We have thus in the Lake Superior country six series, consisting, from top to bottom, of the Keweenawan, Upper, Middle, and Lower Huronian, Laurentian, and Keewatin, all but the last two separated by unconformities. Above them and separated by a marked unconformity rests the Cambrian Potsdam sandstone.

The work of R. C. Allen,[1] W. O. Hotchkiss, and others has lately changed the Huronian correlations to some extent. The latest conclusions are embodied in a guide book by Leith and Hotchkiss, now in press, for the International Geological Congress, to be held in 1933. The great iron formations are now considered Middle Huronian, except a few at the base of the Upper Huronian.

**The "Iron Formations."**—The iron ores of the Lake Superior region are believed to be derived through concentration by means of meteoric waters from lean "iron formations" containing about 25 per cent iron. The ores are products of enrichment of chemically deposited sediments, such as siderite and hydrated iron silicates, for the most part interbedded with normal clastic sediments, such as slate and quartzite.

The iron formations range from a few feet up to 1,000 feet in thickness and are sedimentary beds consisting, according to Leith:

> . . . mainly of chert, or fine-grained quartz, and ferric oxide segregated in bands or sheets, or irregularly mingled. Where in bands with the quartz layers colored red and the rock highly crystalline it is called jasper. Where less clearly crystallized and either in bands or irregularly intermingled the rock is known as ferruginous chert. The silica in these rocks varies from 32 to 80 per cent, the ferric oxide from 31 to 66 per cent. Other phases of the iron formation, subordinate in quantity, are (1) ordinary clay slates, showing every possible gradation through ferruginous slates into ferruginous cherts; (2) 'paint rocks,' oxidized equivalents of the slates; (3) cherty iron carbonate (siderite) and hydrous ferrous silicate (greenalite); (4) the iron ores themselves. Almost the entire bulk of the iron formations now consists of iron oxide and silica.

Spurr found that certain rocks of the Mesabi district contained, in a matrix of chert and iron carbonate, abundant round granules of a green chloritic substance at first called glauconite. Leith showed that potassium was absent and gave the mineral the name greenalite. Its composition is approximately 30 to 38 per cent $SiO_2$, 8 to 34 per cent $Fe_2O_3$, 25 to 47 per cent FeO, and 7 to 9 per cent $H_2O$ (p. 268). The greenalite rocks contain 50 to 80 per cent of this mineral, which is soluble in acids. The same mineral occurs in some of the siderite rocks of the more easterly districts.

Regional metamorphism (according to Grout and Broderick) or contact metamorphism, caused by Keweenawan intrusions of granites and gabbros, (according to Van Hise and Leith) have, in places, converted

[1] R. C. Allen, Correlations of formations of Huronian age in Michigan, *Trans.*, Am. Inst. Min. Met. Eng., 63, 1920, pp. 188–212.

the siderite and the greenalite rocks to magnetite-amphibole schists and the soft hematite to specularite; this is especially well observed in the Marquette and the east Mesabi ranges.

**The Iron Ores.**—The hematite ores are derived from the ferruginous cherts by a process of concentration, and both laterally and in depth gradually change into such rocks. The ores are admixed with enough magnetite to affect the magnetic needle and render possible magnetic surveys of the fields. The hard blue specular ores of the Marquette range contain more magnetite than the others. In other ranges, such as the Mesabi, Penokee, and Baraboo, the ore is soft, bluish, red, or brown in color, and partly hydrated. The average analysis of Lake Superior ores in 1909 was as follows:

| | Per Cent |
|---|---|
| Moisture (loss at 100° C.) | 11.28 |
| Analysis of dried ore: | |
| Iron | 58.45 |
| Phosphorus | 0.091 |
| Silica | 7.67 |
| Alumina | 2.23 |
| Manganese | 0.71 |
| Lime | 0.54 |
| Magnesia | 0.55 |
| Sulphur | 0.06 |
| Loss by ignition | 4.12 |

This corresponds to a composition as follows:

| | |
|---|---|
| Hematite (more or less hydrated) | 86.45 |
| Quartz | 4.89 |
| Kaolin | 5.25 |
| Chlorite | 1.01 |
| Dolomite | 0.81 |
| Apatite | 0.48 |
| Miscellaneous | 1.11 |
| | 100.00 |

The average analysis of natural (not dried) ore in 1925 was (in per cent) 51.74 Fe, 0.099 P, 8.42 $SiO_2$, 0.82 Mn, 10.58 moisture.

Some shipments run as low as 40 per cent Fe. Some ore from the Cuyuna range contains as much as 17 per cent manganese. The sulphur varied from 0.003 to 1.87 per cent, but it averages low. Phosphorus averages 0.07 per cent. Accessory, more or less rare minerals in the ore are apatite, wavellite, adularia, calcite, dolomite, siderite, pyrite, marcasite, chalcopyrite, tourmaline, ottrelite, chlorite, garnet, mica, rhodochrosite, barite, gypsum, analcite, goethite, and turgite. Most of them are probably later additions.

In 1929 the ore reserves of the Mesabi Range were estimated to be 1,178,855,601 tons; those of the whole region 1,411,490,291 tons.

The total yield of the Lake Superior ores from 1848 to 1930 has been 1,455,685,644 long tons, much the greater proportion having been extracted in the last four decades.

Carbonate ores have been mined in the Michipicoten district, Canada.

**Form of Ore-bodies.**—The ore forms irregular, often very large, but as a rule distinctly bedded or banded masses in the iron formations. The shape is commonly determined by impervious basements like clayey dikes, decomposed amphibolitic rocks, or folded sedimentary beds like slate, which have tended to guide the circulation of surface water into certain channels; the ores usually occur in pitching troughs caused by any or all of these factors.

In some ranges like the Gogebic, Marquette, and Iron River the strata are strongly folded and may dip at high angles; some of the ore-bodies have been followed to great depths. In the Marquette district ore is known to a depth of 3,000 feet. Good ore is mined at present in the Newport mine in the Gogebic district at 3,000 feet and is said to have been drilled at 4,000 feet. In the Mesabi Range the rocks lie at gentle angles; the alteration and concentration have extended over a wide area and few of the mines are deeper than 200 feet; the ore is known to extend to a depth of 900 feet. The shallow deposits of this range are mined on an enormous scale by steam shovels. The production was 31,300,000 long tons in 1930, which was about 80 per cent of iron ore output in the Lake Superior district.[1]

**Marquette Range.**—The mines of the Marquette Range are near Negaunee and Republic, south and southwest of Marquette, Michigan. The principal "iron formation," the Negaunee, is in the Middle Huronian; and the sedimentary rocks are intruded and metamorphosed by basic igneous rocks. Extensive folding has taken place and the strata are compressed into a great synclinal basin. The ores lie at the base of the Negaunee formation, where the underlying slates have been folded so as to form pitching synclinal basins, or where dikes have guided the concentrating waters. In part they occur also at the contact of the iron formation with basic intrusions—for instance, in pitching troughs between igneous masses and dikes branching from them. The surfaces of the igneous rocks are much altered, leached, and changed to clayey masses, called "soapstone" and "paint rock."

**Menominee Range.**—The iron-bearing district extends from western Michigan into Wisconsin, the principal mines being located at Iron Mountain, Norway, and Crystal Falls. The iron formation is chiefly in the Middle Huronian and is called the Vulcan formation; it is overlain by Upper Huronian slate and underlain by a Lower Huronian dolomite.

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[1] Regarding mining practice, methods, and costs see *Information Circ.*, 6380 and 6390, U. S. Bur. Mines, (Marquette Range) 1930; *Circ.* 6325 (Mesabi Range), 1930; *Circ.* 6369, 6348 (Gogebic Range), 1930.

Intricate folding characterizes the structure of the range, the ores of the different areas occurring in separate local basins. The deposits are large and consist of soft red hematite, considerably hydrated in places, and are generally found in pitching synclinal basins bottomed and capped by slate layers.

**Penokee-Gogebic Range.**[1]—This range is in northern Michigan and Wisconsin, the principal mines being at Hurley, Ironwood, and Bessemer. The ore appears in the Ironwood formation (Middle Huronian), which is overlain by slate and underlain by quartzite. The dip is steep and the sediments are in part metamorphosed by Keweenawan gabbro; for the most part the Ironwood formation is ferruginous chert. The ores are concentrated in large irregular bodies in the angles between the footwall quartzite or black slate and the basic dikes (Figs. 115 and 116), these rocks making an impervious trough, toward which the meteoric waters converged. Most of the deposits reach depths of 1,000 feet, and some attain 3,000 feet. Both soft, partly hydrated ore and hard slaty ore occur.

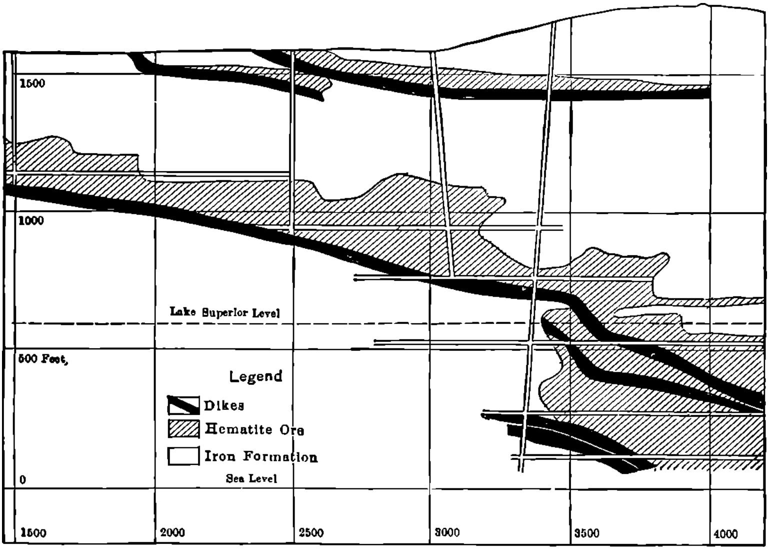

FIG. 115.—Longitudinal section of the Montreal Mine, Gogebic Range, Michigan, showing dependence of bodies of oxidized iron ore on dikes.

**Cuyuna Range.**—The Cuyuna district is situated near Brainerd, Minnesota, about 70 miles southwest of the Mesabi mines.[2] It extends

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[1] H. R. Aldrich, The geology of the Gogebic iron range, *Bull.* 71, Wisconsin Geol. and Nat. Hist. Survey, 1929, 279 pp.

[2] E. Newton, Manganiferous iron ores of the Cuyuna district, *Bull.* 5, Univ. Minnesota School of Mines Exp. Sta., 1918, 126 pp.

for 65 miles along the strike of the rocks in a northeast direction. The iron ore here is a partly hydrated hematite, in places accompanied by an unusual amount of manganese oxide (up to 18 per cent Mn in the shipped ores). It is contained in the Deerwood iron formation, probably of Upper Huronian age, which in depth appears to change to cherty iron carbonate, with some amphibole. The enclosing rocks are slates of various kinds compressed into steep folds, the details of which are difficult to trace owing to the covering glacial drift. The ore-bodies are elongated following the strike and while some cease at shallow depths others have so far been followed down for 400 feet. The phosphorus ranges from 0.1 to 0.5 per cent. In 1930 about 700,000 tons of manganese-iron ore were shipped, averaging 7.5 per cent Mn and 38 per cent Fe.

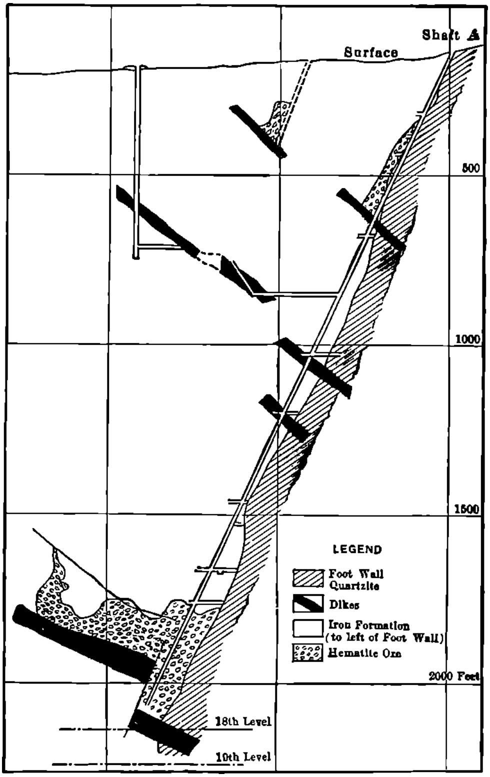

FIG. 116.—Vertical cross-section of the Newport Mine, Gogebic Range, Michigan, showing position of ore-bodies above dikes. (*Data from H. L. Smyth.*)

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E. C. Harder and A. W. Johnston, The geology of east-central Minnesota, *Bull.* 15, Minnesota Geol. Survey, 1918.

G. A. Thiel, *Econ. Geol.*, 19, 1925, pp. 132–145.

Carl Zapffe, *Trans.*, Am. Inst. Min. Met. Eng., 71, 1925, pp. 372–385.

The presence of manganite as the predominant manganese mineral indicates, according to Thiel, that the present ore-bodies have been enriched by solutions containing manganese from the overlying and now eroded portions of the iron formation. Simultaneous replacement of chert by hematite and manganite took place. Intimate intergrowth of manganite and pyrolusite with hematite in martite crystals are features observed.

**Mesabi Range.**—In northern Minnesota, the Mesabi Range extends from east to west for a distance of 75 to 100 miles on the south slope of a prominent ridge called the Giants Range. The principal mines are situated near the towns of Biwabik, Eveleth, Virginia, and Hibbing. The Huronian rocks here lie at gentler inclinations than elsewhere, dipping 8° to 10° S.E. so that the iron formation outcrops in a general northeast-southwest belt (Fig. 117).

The Biwabik iron formation of the Middle Huronian contains the deposits. It is underlain by the Pokegama quartzite and covered by the thick Virginia slate (Upper Huronian). Except at the eastern end of the range, where amphibole-magnetite rocks have developed, the iron formation is composed mainly of ferruginous chert in many complex layers. The iron ores cover large irregular areas along the outcrop of the Biwabik formation and descend to a maximum depth of 900 feet (Fig. 118). At the present time more or less ore is mined by the underground method. The deposits are most abundant at the synclines of the transverse folds of the formation. They are bedded and along the edges change rather abruptly to the ferruginous chert, from which they are derived by leaching of the silica. This relationship is clearly indicated by the slumping of the strata near the edges of the ore masses, as shown in Fig. 119. The iron formation is locally called "taconite."

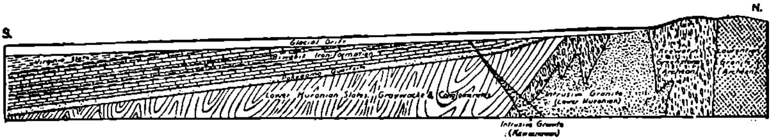

FIG. 117.—Generalized cross-section showing relation of iron-bearing formation to associated rocks in the Mesabi Range, Minnesota. (*After J. F. Wolff.*)

The secondary concentration of the iron ore has evidently taken place under surface conditions since the remote time of the post-Keweenawan folding, when the deposits first became exposed; it has also taken place below as well as above the present water level, which is about 75 feet underneath the surface.

Analyses show that the present surface water, containing about 20 parts per million of $SiO_2$, is slowly leaching silica, but removes little if any iron. The deposits do not appear to continue underneath the edge of the capping Virginia slate, perhaps because of the ponding of the water below that impervious formation. The amphibole-magnetite rocks in the eastern part of the district are more stable and have not suffered much alteration by oxidation.

During the development of the ore-bodies erosion has continually cut down the iron formation and this truncation has been accompanied by slow downward and lateral migration of the iron. Glacial erosion finally removed much material.

The ore is a soft and porous hematite, brown, red, or blue in color, averaging 55 to 58 per cent iron. It contains a little magnetite and some limonite. The mineral composition of the ore in 1909 was approximately, in per cent: hematite, 61.81; limonite, 25.95; quartz, 4.10; kaolin, 5.30; manganese dioxide, 1.30; miscellaneous, 1.54. Sulphur is low and phosphorus varies from 0.03 to 0.07 per cent. There is considerably more phosphorus in the ore than in the ferruginous chert; the greenalite and siderite rocks contain scarcely any phosphorus. Manganese up to 2 per cent is present.

**Vermilion Range.**—Northeast of the Mesabi, near the Canadian boundary, is the Vermilion Range, the principal mines being near the towns of Ely and Tower. The country rock is mostly the Keewatin greenstone, but infolded in it in synclinal basins or troughs is the iron formation, known as the Soudan. The ores are associated with ferruginous jaspers in these troughs and generally have a footwall of greenstone (Fig. 120). The ore is a dense and hard blue or red hematite which contains a little chalcopyrite, an unusual feature in this region.[1]

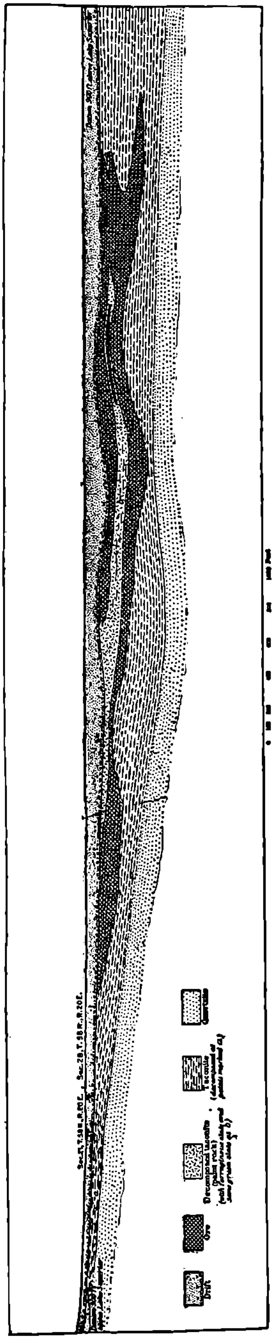

FIG. 118.—North-south cross-section through Biwabik iron formation, Mesabi Range, Minnesota. (After O. B. Warren and C. K. Leith, U. S. Geol. Survey.)

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[1] J. W. Gruner, The Soudan formation and a new suggestion as to the origin of the Vermilion iron ores, *Econ. Geol.*, 29, 1927, pp. 629–644.

**Origin of Lake Superior Iron Ores.**—It has been shown by Van Hise and Leith and their associates that the ferruginous cherts, jaspers, amphibolite-magnetite schists, and iron ores of the iron formations result from the alteration either of the cherty iron carbonate or of the greenalite. The small amounts of iron carbonate or ferrous silicate now found in the formations represent mere remnants left unaltered where protected by other rocks. The steps in the alteration may be observed; and, in the end products, the structures and textures of the original rock are often remarkably well retained. It is held that the ores and the ferruginous cherts or jaspers on one hand and the amphibole schists on the other hand represent alterations from the same original type. The source of the ore is not, as a rule, in the present ferruginous cherts; but it was developed from original lean siderite and greenalite rocks. It is held that in the

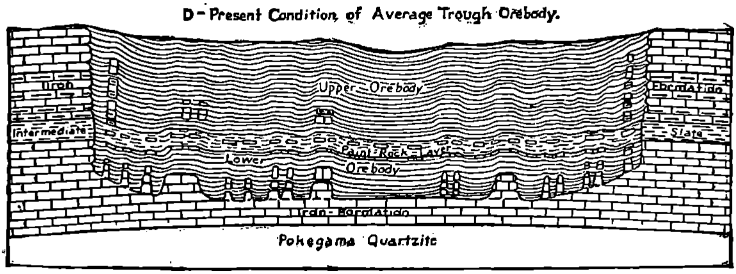

FIG. 119.—Cross-section showing slumping of ore-body at Mesabi Range. (*After J. F. Wolff.*)

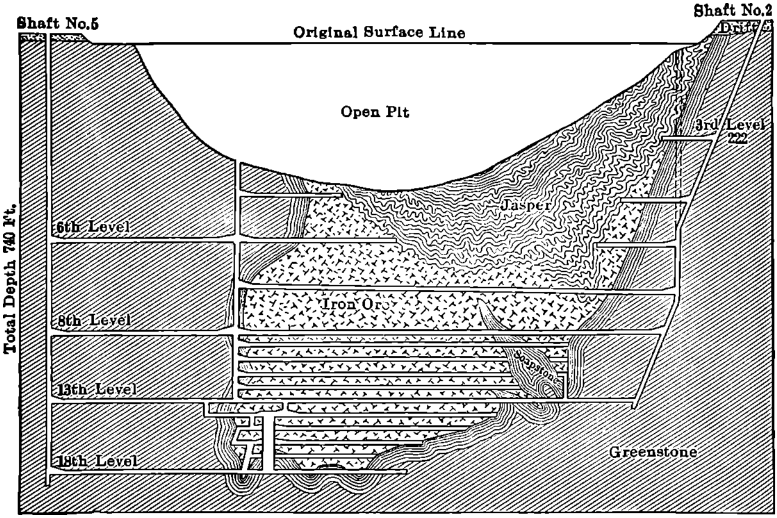

FIG. 120.—Vertical section through the Chandler mine, Vermilion Range, Minnesota. (*After J. M. Clements, U. S. Geol. Survey.*)

largest deposits ores and jaspers may have developed side by side, at the same time, from such original minerals. Iron carbonate prevailed in the Marquette, Gogebic, Vermilion, and Crystal Falls districts; greenalite in the Mesabi district (Fig. 121).

The concentration has been effected, according to the Lake Superior geologists, by water coming more or less directly from the surface, especially at places where such waters converge owing to the existence of impervious underlying formations, such as slate or "soapstone," that form pitching troughs, or owing to brecciation and fracturing of the iron formations.

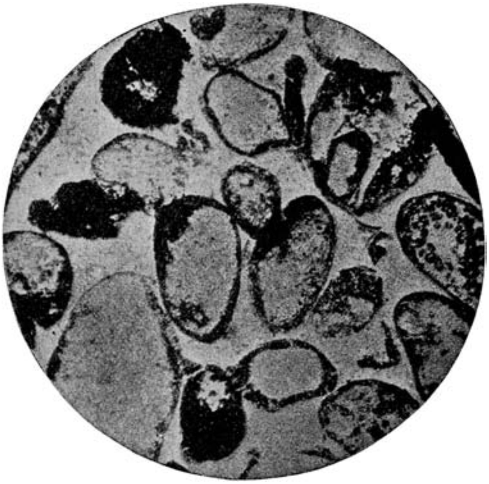

FIG. 121.—Ferruginous chert with greenalite granulites, in part replaced by ferric oxide (black). Magnified 40 diameters. (*After C. K. Leith.*)

The alteration of the iron formations, resulting in the concentration of the iron ores or in the development of ferruginous cherts, jaspers, and amphibolite schists, has taken place in different geologic periods under varying conditions. So far as the alteration has proceeded continuously under the influence of surface waters, without interruption by igneous activity or orogenic movements, soft ores and ferruginous cherts have resulted. So far as these products have been subjected to deep-seated alteration they have become dehydrated into hard red and blue specular ores and brilliant jaspers. So far as the alteration of the original iron formations has taken place within the sphere of influence of great intrusive masses, when waters were heated and oxygen not abundant, or under similar conditions, developed by deep submergence or by orogenic movement, ferrous silicates and magnetite resulted, as shown in the development of the grünerite schists.

The concentration of the ores was far advanced before Cambrian time, as shown by the fragments of ores in Cambrian conglomerates.

Most of the deposits were formed between the Keweenawan and the Cambrian deposition. At the close of pre-Cambrian time the ores were largely as we now find them, though some concentration has been going on since. During the Cretaceous period the region of the Mesabi Range, at least, was covered by the sea.

Regarding the origin of the cherty iron carbonates, Van Hise has held that they were derived largely from the more ancient basic volcanic rocks of the Lake Superior region. The iron was leached by underground waters and carried to the sea as carbonate, partly also as sulphate solution, and there deposited as limonite, from which through reduction by organic matter ferrous carbonate was formed.

About 1910 C. K. Leith expressed his views as follows: The iron was brought to the surface by igneous rocks and either contributed directly to the ocean by hot magmatic waters or later brought there by surface waters from weathered rocks. The iron-bearing minerals were then deposited as a chemical sediment in a conformable succession of sedimentary rocks and still later, under conditions of weathering, were locally enriched to ore by percolating surface waters.

> As first deposited the iron formation consisted essentially of iron carbonate or ferrous silicate (greenalite) with some ferric oxide, all minutely interlayed with chert, forming the ferruginous chert. When these were exposed to weathering the ferrous compounds, the siderite and greenalite, oxidized to hematite and limonite, essentially *in situ*, although some of it was simultaneously carried and redeposited. The result was ferruginous chert or jasper, averaging less than 30 per cent of iron. The concentration of the iron to 50 per cent and over has been accomplished essentially by the leaching of silica bands from the ferruginous chert and jasper. Infiltration of iron has been on a smaller and more variable scale. The leaching of the silica develops pore space and allows the iron layers to slump, thereby enriching the formation sufficiently to constitute an ore.

Only a small part of the volume of the iron formations—less than 2 per cent—has been altered to ore.

**Later Work.**—The literature of the Lake Superior iron ores is extensive and many different views have been expressed. T. B. Brooks and R. Pumpelly at one time considered them as dehydrated bog iron ores. This view has been adopted by S. Weidman in his description of the Baraboo ores of Wisconsin, where the ores appear to grade into dolomites, and where igneous rocks are absent.

Grout,[1] Broderick,[2] and Gruner[3] in later papers regard the iron formations as chemical precipitates which received their iron from deeply

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[1] F. F. Grout, The nature and origin of the Biwabik iron-bearing formation, *Econ. Geol.*, 14, 1919, pp. 452–464.

[2] T. M. Broderick, Detail stratigraphy of the Biwabik iron-bearing formation, *Econ. Geol.*, 14, 1919, pp. 441–451; also, Economic geology and stratigraphy of the Gunflint iron district, *idem*, 15, 1920, pp. 422–452. F. F. Grout and T. M. Broderick,

weathered land areas. They place less emphasis on cooperation by igneous agencies. The Biwabik iron formation has been studied in detail by Wolff, Grout, and Broderick. Gruner found that there is much magnetite in the unweathered parts of the formation exposed by many deep drill holes (Fig. 122) and that thus the magnetite is by no means restricted to the East Mesabi district. Gruner believes that the original iron minerals were siderite and greenalite, perhaps also hematite. Magnetite developed abundantly later by regional metamorphism and most of this magnetite was then oxidized to hematite. Magnetite is the oxide from which nearly all of the rich, so-called "blue ore" has been derived as may be seen from the abundance of hematite pseudomorphic after magnetite. Amphibole also occurs through the whole range and is not confined to the East Mesabi.

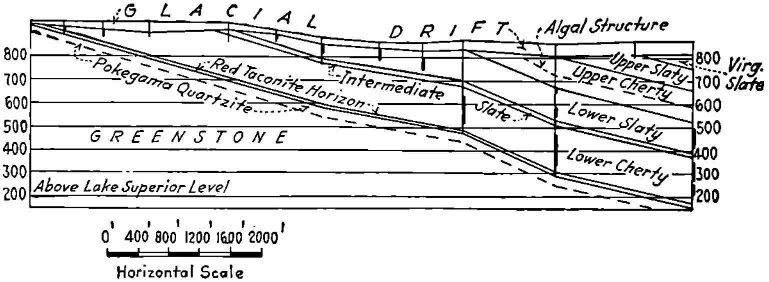

FIG. 122.—Cross-section through central part of Mesabi Range showing subdivisions of the Biwabik formation and abundance of magnetite. Heavy black lines along drill holes average 27 to 33 per cent magnetite. (*After J. W. Gruner, "Economic Geology."*)

Algæ and bacterial structures have now been found in the Biwabik formation, and Grout holds that the precipitation of iron minerals and silica was largely caused by low forms of organic life.

While it is generally agreed that the original substances of the iron formations were chert, siderite, and an iron silicate, opinions differ as to the widespread oxidation converting them to banded hematites and as to the process of concentration of rich hematite from the latter. There are difficulties in assuming a thorough oxidation of solid formations at a depth of several thousands of feet below the surface, even during an excessively dry climate and depressed water level in pre-Cambrian times. The leaching of silica (mostly quartz) from these massive formations by surface waters seems an equally difficult proposition. In defending this theory C. K. Leith[1] points to a probable rotation and tilting of erosion

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Organic structures in the Biwabik iron-bearing formation, *Am. Jour. Sci.*, 4th ser., 48, 1919, p. 199.

[3] J. W. Gruner, The origin of sedimentary iron formations, *Econ. Geol.*, 17, 1922, pp. 407–460; also Paragenesis of the martite ore-bodies and magnetites of the Mesabi range, *idem*, 17, 1922, pp. 1–14.

[1] *Econ. Geol.*, 26, 1931, pp. 274–288.

surfaces, thus increasing the apparent depth below the surface. With regard to the wholesale leaching of silica, Leith points to conditions during weathering where enormous masses of silica are undoubtedly removed by cold surface waters; but this silica is almost wholly derived from the decomposition of silicates and not from granular quartz.

Grout and Broderick think that the general oxidation may have followed very soon after deposition even before the burial of the beds, and analogies with many other deposits, *e.g.*, the iron ores of Wabana, Newfoundland, suggest that this may be true.

Moore and Maynard[1] in a series of articles maintain that the pre-Cambrian formations contained much organic matter, and that iron and silica were extensively dissolved and transported as colloids stabilized by organic material.

Gruner[2] believes that the oxidation of the iron formations, and the leaching of silica are chiefly the work of hot magmatic solutions derived from Keweenawan igneous rocks. This would need unlimited quantities of such water and a high temperature of about 200° C. In its present form Gruner's hypothesis appears improbable, but transportation of silica and iron has certainly taken place. If effected by magmatic ascending waters, they would surely have left more conspicuous traces of their passage than can now be detected. On the other hand, it seems certain that the iron formations have been exposed to a higher degree of heat than could be furnished by the ordinary meteoric waters. Note, for instance, the high-temperature veins which intersect the Cuyuna iron formation and the various silicates and sulphides often encountered elsewhere. The Keweenawan igneous rocks may well have contributed an increase in temperature in the underlying rocks. Replacement of chert by hematite must have been of much importance in places. The problem can not be regarded as solved as yet.

Another question of possible importance relates to the percentage of phosphorus. It is remarkably low for sedimentary deposits in the origin of which organic matter played a part. It is still more remarkable that the primary siderite-greenalite rocks at Mesabi are almost free from phosphorus.

**Michipicoten Range, Canada.**—The studies of Collins and Quirke[3] on the Michipicoten iron range in Ontario, seem to indicate a radically different origin. Collins holds that these steeply dipping beds of siderite, pyrite, and fine-grained silica are replacements by thermal solutions of beds of basic Keewatin lavas. The ore consists largely of siderite or

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[1] *Econ. Geol.*, 24, 1929, pp. 272–303; 365–402; 506–527.

[2] J. W. Gruner, Hydrothermal oxidation and leaching experiments, *Econ. Geol.*, 25, 1930, pp. 697–719; 837–867.

[3] W. H. Collins, T. T. Quirke, and Ellis Thomson, Michipicoten iron ranges, *Mem.* 147, Canada Geol. Survey, 1926.

ankerite, and has been mined for a number of years though the mines are now idle.

**Other Occurrences in the United States.**—Regionally metamorphosed banded iron ores are found in the pre-Cambrian of Wyoming[1] and Arizona;[2] both magnetite and hematite occur.

At Hartville, Wyoming, lenses of hematite occur in schist along a limestone footwall and have been followed to a depth of 900 feet. Ball shows that the deposit antedates the Guernsey formation, the lowest Paleozoic terrane present, and believes that the iron was leached by descending solutions from the upper part of the schist and deposited in its lower part by replacement.

Typical quartz-magnetite ores, described by Sidney Paige,[3] occur in the Llano region in Texas, but have not yet been utilized. The pre-Cambrian series of crystalline schists in this region consists of gneiss, mica schist, and quartzite with some limestone lenses. The ores are thinly bedded and occur in granular schists or gneisses. A specimen of lean ore consisted of magnetite 22 per cent, quartz 50 per cent, albite and albite-oligoclase 26 per cent. There is much more soda than potash. The iron was probably, according to Paige, deposited as glauconite; and contact metamorphism by later pre-Cambrian granite has effected the removal of potash and introduction of soda. Examples of adinole and other contact-metamorphic rocks are cited to support this view. Low-grade ore representing a bed 17 feet thick contained Fe 35.87 per cent, $SiO_2$ 34.57 per cent, Mn 1.05 per cent, P 0.07 per cent, S 0.04 per cent, and $TiO_2$ 0.15 per cent.

The Grenville series of pre-Cambrian metamorphosed sediments in northern New York contain, according to D. H. Newland,[4] similar deposits of magnetite. The ores are mostly enclosed in quartzose gneisses with hornblende and biotite (see p. 794).

## OTHER REGIONALLY METAMORPHOSED IRON ORES

**General Features.**—In metamorphosed sediments or in crystalline schists, the origin of which may be in doubt, bedded deposits of magnetite or specularite, or both, are often encountered. The well-known fact that iron ores such as limonite, siderite, hematite, or iron silicates (chamosite and thuringite) form integral parts of sedimentary series of all ages suggests strongly that the beds of these ores in metamorphosed rocks also had a sedimentary origin. As a rule this is no doubt true, but the

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[1] S. H. Ball, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 190–208.
T. S. Lovering, *Bull.* 811, *idem,* 1930, p. 219.

[2] Howland Bancroft, *Bull.* 451, *idem,* 1911.
Waldemar Lindgren, *Bull.* 782, *idem,* 1926, p. 35.

[3] *Bull.* 450, U. S. Geol. Survey, 1911.

[4] Geology of the Adirondack magnetic iron ores, *Bull.* 119, N. Y. State Mus., 1908, pp. 27, 40–41.

metamorphism may have gone so far that the original sedimentary nature of the surrounding rocks may be open to doubt; and some observers maintain an igneous origin for some such deposits. Indirectly, igneous rocks have often brought about the accumulation of bedded iron ores, either by the weathering and denudation of intrusive rocks or lavas rich in iron, or possibly by direct emanations from volcanic rocks.

Bedded metamorphic iron ores are accompanied by silicate minerals, like feldspar, actinolite, and garnet, usually also by quartz, and have assumed a thoroughly crystalline texture similar to that of other crystalline schists, the constituents being generally interpenetrating, indicating almost simultaneous development. Relic structure showing the sedimentary origin is rarely observed.

**Swedish "Dry Ores."**[1]—Sweden and Norway are rich in these bedded ores, which often appear in the vicinity of other iron deposits of different kind. Some are found near the great magmatic deposit of Kiruna, interbedded in tuff and shales of late pre-Cambrian age. Others, which are worked more extensively, appear near the metasomatic magnetites of central Sweden (p. 739) and form part of the complicated leptite series (p. 740). They are designated "dry ores" (torr-sten) and are usually siliceous, the accompanying beds averaging 84 per cent silica. The ores average 50 per cent iron, contained in micaceous fine-grained specularite with a little magnetite. The accompanying beds in places contain garnet, amphibole, or epidote, each mineral often forming a separate streak. They are markedly banded. Many of the beds are 10 or 15 feet thick, though some considerably exceed 15 feet, and have been followed with regular, steep dip to a depth of several hundred feet. These ores contain little phosphorus. An analysis of such ore from Striberg is as follows:

| | | | |
|---|---|---|---|
| Fe | 52.20 | CaO | 1.05 |
| $Fe_2O_3$ | 60.21 | $Al_2O_3$ | 0.89 |
| FeO | 13.93 | $SiO_2$ | 23.61 |
| MnO | 0.09 | $P_2O_5$ | 0.043 |
| Mg | 0.31 | S | 0.021 |

Until recently little doubt has been expressed about the sedimentary origin of these ores. H. Johansson has announced his opinion that the fine-grained leptites are simply a product of extreme magmatic differentiation and that the accompanying bedded iron ores are also of magmatic origin. He even believes that the metasomatic limestone and "skarn ores" (p. 740) have this origin. Hj. Sjögren does not share this opinion but holds that the bedded ores and limestone ores are caused by injection or replacement by "granitic extracts" while the differentiated

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[1] H. E. Johansson, *Geol. För. Förhandl.*, 32, 1910, pp 239–410.
Hj. Sjögren, *Trans.*, Am. Inst. Min. Eng., 38, 1908, pp. 766–835.
See also references on p. 739.

granulites were in the "anamorphic" zone. Holmquist thinks that by deep burial these originally sedimentary ores have been subjected to igneous metamorphism followed by slight regional metamorphism (p. 742).

It does not seem that the opponents to the sedimentary genesis of the ores have proved their case.

**Norwegian Ores.**[1]—Northern Norway is rich in deposits of the type here discussed. Banded magnetites, variously interpreted, occur on a large scale in thick beds that are traceable for several miles in South Varanger, near the Finland frontier. The ores are mined on a large scale and concentrated. Some of the larger bodies are 1,000 feet long and 25 feet thick and contain about 35 per cent iron. One hundred million metric tons are available for open-cut mining. An analysis given by Vogt shows 36.71 per cent $Fe_2O_3$, 15.40 per cent FeO, 43.92 per cent $SiO_2$, 0.07 per cent $P_2O_5$, and 0.04 per cent S. There is little alumina, lime, or magnesia. The ores are beautifully banded and, according to P. Geijer, are associated with fine-grained "leptites" (granulites) rich in quartz with some orthoclase and oligoclase; hornblende, garnet, and diopside accompany the ore (Fig. 123). While Vogt considers the ores to be due to igneous differentiation and Sjögren believes similar ores from the Lofoten Islands to be intrusive into an igneous rock, Geijer gives good

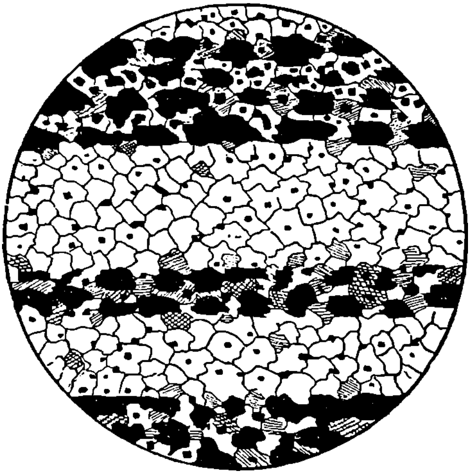

FIG. 123.—Thin section of typical Syd Varanger ore. Black, magnetite; white, quartz; striated, hornblende. Magnified 25 diameters. (*After J. H. L. Vogt.*)

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[1] J. H. L. Vogt, Norway, in Iron-ore resources of the world, Stockholm, 1910.

J. H. L. Vogt, Norges jernmalmforekomster, *Norges Geol. Undersök.*, No. 51, Kristiania, 1910.

Hj. Sjögren, Om jernmalmerna i granit på Lofoten, *Geol. För. Förhandl.*, 30, 1908.

Per Geijer, Contributions to the geology of the Sydvaranger iron-ore deposits, *idem*, 33, 1911, pp. 312–343.

reasons why they should be held to be of sedimentary origin and deposited as chemical sediments. It seems that the advocates of intrusive origin for these occurrences have few cogent arguments.

Ores of distinctly sedimentary origin are found at Dunderland and Naeverhaugen, also in northern Norway. They form beds traceable for many miles, with a thickness of 3 to 10 meters, or in places even 50 meters. They are intercalated in a thick series of mica schists and crystalline marbles believed to be of Paleozoic age. The closely banded ores carry mainly specularite and magnetite, but are of low grade. The concentration, attempted on a large scale and at great expense, proved difficult because of the scaly character of the specularite. The average content in iron is said to be 40 per cent. Hornblende, garnet, epidote, and feldspar are accessory minerals. There is little sulphur, but phosphorus is present in quantities as great as 0.3 per cent.

**The Brazilian Hematites.**—In the pre-Cambrian metamorphosed sediments of Minas Geraes in Brazil[1] there are thick beds of rich hematite in a formation of ferruginous sandstone (itabirite) underlain by heavy quartzite. The origin of this undoubtedly sedimentary hematite, which as yet has been mined only on a very small scale, is in doubt. There is no oölitic structure, nor are there fossils. Harder and Chamberlin state that "not having much confidence in the hypothesis that the iron oxide was precipitated directly from sea water by ordinary chemical means we prefer to turn to the iron bacteria as perhaps forming a better hypothesis."

Another suggestion would be to regard these ores as detrital, extremely fine-grained hematite, derived from the deeply oxidized mantle of neighboring continents and subsequently more or less metamorphosed.

The reserves are enormous, consisting of at least a billion tons of 50 per cent ore.

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[1] E. C. Harder and R. T. Chamberlin, The geology of Central Minas Geraes, Brazil, *Jour. Geol.*, 23, 1915, Nos. 4 and 5.

O. A. Derby, in The iron ore resources of the world, 11th Internat. Geol. Congress, II, Stockholm 1911, p. 819.

C. K. Leith and E. C. Harder, Hematite ores of Brazil, *Econ. Geol.*, 6, 1911, pp. 670–686.

# CHAPTER XX

## DEPOSITS FORMED BY EVAPORATION OF BODIES OF SURFACE WATERS

### THE SALINE RESIDUES[1]

#### INTRODUCTION

The deposits thus far described have been in the nature of insoluble residues, or chemical precipitates of relatively insoluble substances in lakes, rivers, and seas. There are, however, other deposits which also may be considered as chemical precipitates in surface waters but which consist of soluble salts formed by the evaporation of waters in closed or partially closed basins. They contain the easily soluble substances largely leached from the rocks, brought down by the rivers to oceans and lakes, and finally concentrated under certain characteristic conditions.

Closed basins are typical of dry climate and of deserts. The slow crustal movements tend to create them everywhere, by folding, subsidence, and uplift; but in the deserts the streams have not the power to cut outlets and to keep the drainage lines established. On the contrary, the movement of the débris from the mountain ranges in broad alluvial fans or aprons increases the tendency toward closed basins. The dry climate accelerates evaporation and the precipitation of the salts; dust storms transport vast masses of fine detritus; blinding salt flats extend between the barren mountain chains. Thus, at present, salt beds are found in the Cordilleran deserts along the western side of the whole American continent, in the Sahara, and in the arid, central part of Asia. Similar conditions existed in the past in different parts of the world: The Permian in central Europe, the Triassic in the Rocky Mountain region, and the Silurian in eastern North America—all these ages were at times characterized by arid wastes and deposition of salt and gypsum.

Saline deposits may then form: (1) in bays of the sea; (2) in lakes; (3) in playas or intermittent lakes; (4) on arid slopes by rapid evaporation of storm waters.

In the latter two classes capillary ascent of the solutions often helps to bring the salts to the surface to form "efflorescences" or saline crusts,

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[1] George P. Merrill, The non-metallic minerals, 1910.

F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 218–260.

A. W. Grabau, Principles of salt deposition, New York, 1920.

as illustrated by the crusts of "alkali" (carbonate, chloride, and sulphate of sodium, and sulphate of calcium) which so often interfere with agriculture in dry countries.

In regions of calcareous rocks, as in the undrained basins of Mexico, soft or compact beds of calcium carbonate locally called "caliche" or "tepetate" often cover the gentle slopes below the mountains. These belong to class 4. Minor saline deposits may result from evaporation at the surface of waters from ascending springs.

Bodies of soluble salts are rarely formed below the surface; but reactions may take place in a buried deposit by which new salts are formed or concentrations of disseminated substances are effected. In places it may be difficult to distinguish these, strictly speaking, epigenetic bodies from the syngenetic salts (p. 338).

Traces of copper are found in the saline residues. Minute amounts of gold and silver have been found in salt from sea water (p. 7). Regarding traces of gold in the potassium deposits in Germany the evidence is conflicting. The later data seem to show that no gold is present.

## TYPES OF WATER

From a geological standpoint there are two types of water in the seas and closed basins. The first, which may be called the oceanic type, contains dominant sodium chloride and is characteristic of the sea as well as of partly evaporated lakes in regions where sedimentary rocks prevail; the Great Salt Lake of Utah is an example. When such water is subjected to extreme evaporation, as in the Dead Sea, a "residual" type rich in magnesium chloride results. The second main type is that of generally smaller closed basins in regions of great volcanic activity; this type contains sulphate, carbonate, and borate of sodium, besides more or less chloride; it indicates the result of the first leaching of loose volcanic ejecta and also shows the influence of the discharge of hot springs containing sodium carbonate and borate.

Certain sedimentary series, such as the Cretaceous of the Western states, contain abundant alkaline sulphates. Leaching of these beds by atmospheric waters takes place and these products may be carried down into salt flats and small lakes. By chemical reactions (p. 49) carbonate of sodium forms from other sodium and calcium salts, and the lakes often contain much of this salt besides the sulphates. Such alkali lakes occur in Wyoming, for example. Borates characteristic of volcanic regions are generally lacking in these lakes.

The first, "oceanic" type of waters yields deposits of gypsum, common salt, and finally potassium and magnesium salts. The second and third, which may be called, respectively, the "volcanic" and the "sulphate" types yield sodium carbonate and sulphate, borates, probably also nitrates, as well as more or less sodium chloride.

COMPOSITION OF SALTS IN WATER OF SEA AND CLOSED BASINS

| | I | II | III | IV | V |
|---|---|---|---|---|---|
| Cl | 55.292 | 55.69 | 70.25 | 10.45 | 23.34 |
| Br | 0.188 | tr. | 1.55 | ........ | ........ |
| $SO_4$ | 7.692 | 6.52 | 0.21 | 54.07 | 12.86 |
| $B_4O_7$ | ...... | ..... | ..... | ..... | 0.32 |
| $CO_3$ | 0.207 | ..... | tr. | 4.24 | 23.42 |
| Li | ...... | 0.01 | ..... | ..... | ..... |
| Na | 30.593 | 32.92 | 6.33 | 25.88 | 37.93 |
| K | 1.106 | 1.70 | 1.70 | ...... | 1.85 |
| Ca | 1.197 | 1.05 | 5.54 | tr. | 0.04 |
| Mg | 3.725 | 2.10 | 14.42 | 5.36 | 0.10 |
| $Fe_2O_3$ | ...... | ..... | tr. | ...... | tr. |
| $Al_2O_3$ | ...... | ..... | tr. | ...... | tr. |
| $SiO_2$ | ...... | 0.01 | tr. | ...... | 0.14 |
| | 100 | 100 | 100 | 100 | 100 |

I. Oceanic type. Average of 77 analyses, W. Dittmar, Challenger Rept., 1, 1884. Salinity 3.5 per cent.

II. Oceanic type. Great Salt Lake, E. Waller, School of Mines, *Quart.*, 14, 1892. Salinity 23 per cent, subject to variations.

III. Residual type. The Dead Sea, *Comptes Rendus*, 62, 1866, p. 1329. A. Tereil, analyst. Salinity 20.7 per cent.

IV. Sulphate type. Devil's Lake, North Dakota, F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 165; H. W. Daudt, analyst. Salinity 1.1 per cent.

V. Volcanic type. Mono Lake, *Bull.* 60, U. S. Geol. Survey, 1890, p. 53. T. M. Chatard, analyst. Salinity 5.1 per cent.

The general relation of the salts dissolved in oceanic waters to those in rivers is as follows:

In ocean waters: $Cl > SO_4 > CO_3$; $Na > Mg > Ca$
In river waters: $CO_3 > SO_4 > Cl$; $Ca > Mg > Na$.

## NORMAL SUCCESSION OF SALTS

When water evaporates until precipitation of the dissolved salts begins, the least soluble salts will generally fall down first, while the most soluble salts will remain in the solution until the last. Experiments by J. Usiglio[1] on sea water showed that the carbonates of calcium and magnesium, with a little ferric oxide, were precipitated when one liter of the water was reduced from one-half to one-fifth. Gypsum was precipitated when the volume was one-fifth to one-seventh, but continued in lessening amounts until only 30 cubic centimeters of the original liter remained. Sodium chloride, the main constituent, was precipitated abundantly upon reduction of volume to 100 cubic centimeters, but continued until the volume of the water was only 16 cubic centimeters; even then some

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[1] *Annales chim. phys.*, 3d sér., 27, 1849, pp. 92–172.

of the salt remained in solution. Chloride and sulphate of magnesium fell down within the same limits but in increasing quantities, and the residual "bittern" contained mainly the chlorides of magnesium and potassium, bromide of sodium, sulphate of magnesium, and chloride of sodium.

The whole series of these salts is rarely represented in the saline deposits; the best known example of such complete evaporation is found in the great Prussian potash and salt deposits. Often, as in the "Red Beds" of the Western states, the process ceased after the gypsum was laid down; and changes of climate or invasion of the sea may have prevented the precipitation of sodium chloride.

Actually the results of precipitation are far more complex than the experiments mentioned would seem to show. Double or complex salts are often formed.

A saline solution containing the same salts as sea water but in different proportions would yield materially unlike results upon evaporation. In brief, temperature, concentration, and time are always factors in the origin of saline residues.

The study of the stability fields of these salts has received much impetus by the labors of J. H. van't Hoff and his numerous associates,[1] undertaken mainly to elucidate the problems of the potassium deposits of Prussia.

The occurrence of thick beds of anhydrite is explained by the work of van't Hoff and Weigert, who established that the mineral forms from gypsum in sodium chloride solutions at 30° C. In sea water the transformation takes place at 25° C. Crystals of gypsum, sinking through a salt solution at that temperature, are converted into anhydrite. This is an example of the more general rule of dehydration of minerals in contact with salt solutions, at temperatures considerably below their normal inversion temperature.

Many minerals are deposited in nature from solutions in a lower state of hydration than is produced at ordinary temperatures in the laboratory. Thus natron, the ordinary sodium carbonate ($Na_2CO_3.10H_2O$) rarely occurs as a natural product, although it is stable at temperatures below 37° C. The more common product is trona ($Na_2CO_3.NaHCO_3.2H_2O$), which is ordinarily stable only above 37° C.

From pure sodium sulphate solution mirabilite ($Na_2SO_4.10H_2O$) is ordinarily deposited, but in the presence of a sodium chloride thenardite ($Na_2SO_4$) is formed. From a solution of magnesium sulphate in the presence of magnesium chloride kieserite ($MgSO_4.H_2O$) is precipitated

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[1] J. H. van't Hoff, Zur Bildung der ozeanischen Salzablagerungen, Braunschweig, 1905 and 1909, Physical chemistry in the service of the sciences, University Chicago Press, 1903. H. Precht and E. Cohen, Ueber die Bildungsverhältnisse der ozeanischen Salzlager, Leipzig, 1912.

instead of the heptahydrate. The presence of a co-solute, by lowering the osmotic pressure, acts in the same direction as a rise of temperature. This principle is undoubtedly also applicable to minerals in rocks and veins.

### STRUCTURAL FEATURES

In desert valleys filled by temporary or permanent lakes we often find a succession of salt beds of no great thickness alternating with detrital matter of eolian or fluviatile origin. If the basin is large and deep a considerable mass of salt may accumulate. The laws governing the deposition of saline residues in the depressions of the deserts have been ably set forth by J. Walther.[1]

Salt deposits can be formed in marginal salt pans of the ocean subject to periodic flooding; such deposits would not ordinarily be very thick.

The salt deposits of marine origin are frequently of great thickness. In some cases they aggregate 1,000 to 2,000 feet; and it will be readily recognized that any theory based on a single cycle of evaporation of sea water, containing only 3.5 per cent of salts would meet with great difficulties.

The "bar theory," presented by C. Ochsenius,[2] in 1877, but already suggested by previously expressed views of Miller, Lyell, and Bischof, attempts to explain these thick salt beds. Ochsenius believed that salt deposits of the purity and thickness of those in central Germany could not have been formed by the flooding of a series of shallow sounds and lakes.

The bar theory premises a bay of the ocean separated from the open sea by a practically level bar which permits only about the same quantity of water to enter as is evaporated from the surface. A dry climate and absence of fresh-water tributaries to the bay are also premised. Under these conditions the sea water entering over the bar continuously carries a new supply to the bay; the surface layers, becoming denser, always sink and the concentration continually increases in the enclosed body of water.

The salt deposits on the bottom are increasing in thickness and the heavy "bittern" solution, with the remaining magnesium salts, correspondingly rises toward the surface.

When these dense solutions reach the surface of the bar, the movement is reversed and the residual "bittern" will flow outward into the ocean. No accumulation of potassium-magnesium salts will occur. Should, however, the bar have increased in height just at this time, the bittern would be retained in the now closed basin and the deposition of the potassium salts would follow.

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[1] J. Walther, Das Gesetz der Wüstenbildung, Leipzig, 1912.

[2] C. Ochsenius, Bedeutung des orographischen Elementes "Barre," *Zeitschr. prakt. Geol.*, 1893, pp. 189–201; 217–233.

The Gulf of Karaboghaz, on the eastern side of the Caspian Sea, is frequently referred to as an excellent illustration of the bar theory.[1]

Salt deposits may also be found in relic seas, previously enriched in salt and perhaps deepened by warping. Such conditions may have existed in the Permian of Kansas and in the Permian of central Germany.

## GYPSUM AND ANHYDRITE[2]

**Occurrence.**—Gypsum ($CaSO_4.2H_2O$) and anhydrite ($CaSO_4$) usually occur in sedimentary beds as saline residues. Both are also found occasionally as gangue minerals in ore deposits, and gypsum is in places a product of ascending springs or of reactions of acid waters on calcareous beds. Efflorescences of gypsum may be produced by capillary action over gypsiferous beds or along saline lakes. Anhydrite, on account of its slow transformation into the hydrous compound, has no economic value, while gypsum is one of the most important non-metallic minerals.

Gypsum in sedimentary deposits frequently forms almost pure beds of considerable thickness. It appears as snow white, fine-grained aggregates; characterized by softness (H: 2), low specific gravity (2.3), and solubility in dilute hydrochloric acid. Anhydrite likewise forms white, granular aggregates, but is easily distinguished from gypsum by its greater hardness, its greater specific gravity (2.9), its pseudocubical cleavage and resistance to weak HCl. Anhydrite slowly alters to gypsum, and many occurrences of apparently solid gypsum contain remnants of anhydrite.

Beds of gypsum and anhydrite occur in many water-laid formations all over the world. Gypsum predominates at the surface but alternating beds of the two are common. Beds of anhydrite up to 300 feet in thickness are found in the Permian of central Germany in connection with the potash salts (p. 336).

Anhydrite occurs with gypsum overlying thick salt beds in Louisiana and southern Texas; beds of anhydrite are also known from New Mexico, Texas, Oklahoma, Nova Scotia, and New Brunswick. Exceptionally thick beds are reported from Carlsbad, New Mexico, where 1,325 feet of anhydrite underlies 633 feet of salt.[3]

The gypsum beds of the United States are rarely more than 30 or 40 feet thick though there may be several in any one section. They are

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[1] W. H. Twenhofel, Treatise on sedimentation, Baltimore, 2d ed., 1932, p. 499.

[2] G. P. Grimsley, Michigan Geol. Survey, 9, pt. 2, 1904.

R. C. Wallace, Gypsum and anhydrite in genetic relationship, *Geol. Mag.*, 1, 1914, pp. 271–276.

R. W. Stone, Gypsum deposits of the United States, *Bull.* 697, U. S. Geol. Survey, 1920.

F. A. Wilder, *Bull.* Geol. Soc. Am., 32, 1921, p. 390. *Mineral Industry*, annual issues.

D. H. Newland, Geology of gypsum and anhydrite, *Econ. Geol.*, 16, 1921, pp. 393–404.

[3] N. H. Darton, *Bull.* 715, U. S. Geol. Survey, 1921, p. 221.

interstratified with limestone or shale; in places they are of great purity and snow white. The compact, translucent variety is called alabaster and is used for ornamental objects; gypsum in large plates or crystals is called selenite. Recent surface deposits, mixed with clay are known as "gypsite."

Economically important gypsum deposits are found in the Salina (Silurian) formation in northern New York and extend parallel to the south shore of Lake Ontario.

Gypsum beds are also extensively worked in Michigan where they are of Mississippian age (Lower Carboniferous). Equally important beds of the same age are exploited in Nova Scotia and New Brunswick. Iowa, Kansas, Texas, Oklahoma, New Mexico, and other states are rich in gypsum of Permian age; in the western part of this region gypsum occurs at several horizons in the "Red Beds" whose age ranges from Upper Carboniferous to Jurassic.

Exceptionally thick, but not easily utilized deposits of uncertain age overlie the "salt domes" (p. 333) of Louisiana and Texas. Tertiary deposits are known from California, and Quaternary "gypsite" is abundant in Kansas, Oklahoma, and Texas. The Tertiary beds in the basin of Paris, France, are rich in gypsum, hence the name "plaster of Paris."

**Uses.**—Experiments relating to the possible artificial conversion of anhydrite to gypsum by wet grinding are in progress. Gypsum finds extensive use in various industries. Ground in its natural state, it is employed as a fertilizer (land plaster) to counteract alkali in soils, to retard the setting of cement, and for numerous chemical purposes. It is often used as a "filler" or adulterant. Most important is, however, its use as structural material. For this purpose it is calcined at 350° F. when a large part of the water is expelled. After grinding and mixing with water gypsum forms again and the whole sets to a hard mass called stucco or plaster of Paris. The use of gypsum is increasing rapidly. In 1930 the production in the United States was 3,471,393 short tons. Most of it was mined in New York, Michigan, Iowa, and Texas. Average value of calcined product is about $11.50 per ton.

**Stability and Solubility.**[1]—As noted above gypsum is transformed to anhydrite in sea water at 25° C. In pure water it begins to change slowly to anhydrite at 63.5° C. At or above 25° C., a temperature often reached in salt lakes, alternating beds of gypsum and anhydrite may form, as indeed is often observed. To a large extent, anhydrite is the original mineral.[2]

The solubility of gypsum is a complicated problem owing to the existence of metastable forms—the hemi-hydrate and the soluble anhydrite—

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[1] C. Doelter, Handbuch der Mineralchemie, Bd. 4, Hft. 7, 8, 1927.

[2] O. Bowles and M. Farnsworth, Physical chemistry of the calcium sulphates, *Econ. Geol.*, 20, 1925, pp. 738–745.

and has been worked out by van't Hoff and Meyerhoffer.[1] The solubility of gypsum in water reaches a maximum of 0.21 per cent at 40° C., and decreases slightly above this temperature. At 63.5° C. the solubility of anhydrite is, of course, equal to that of gypsum, but beyond this point it decreases rapidly so that at 100° C. it is 0.06 per cent and at 200° C. only about 0.005 per cent.[2] Other calcium salts, having a common ion, depress the solubility of gypsum; but sodium chloride increases it about three times owing to formation of $CaCl_2$.

## SODIUM CARBONATE AND SODIUM SULPHATE

**Occurrence.**—Most of the soda of commerce is an artificial product from common salt, but both the carbonate and the sulphate of sodium are often contained in saline desert lakes or in residues from such lakes. The ordinary white efflorescence on the *playas* of the deserts consists of these salts together with more or less sodium chloride and a little of the chlorides and sulphates of potassium and magnesium; the soda lakes contain all these salts.[3] In fact, they occur widely scattered in the deserts of the Cordilleran region and Central Asia.

In the United States the commercial utilization of soda has been attempted at Owens Lake, in California, at the Ragtown lakes, in Nevada, and at the Wyoming soda lakes.

T. D. Chatard's work on Owens Lake, where sodium carbonate forms a little over one-third of the dissolved salts, showed that the order of deposition upon evaporation is: (1) trona ($Na_2CO_3.NaHCO_3.2H_2O$); (2) sodium sulphate; (3) sodium chloride, and (4) the easily soluble normal sodium carbonate. One or two of the Wyoming lake deposits are rich in soda. At Green River, Wyoming, borings in the Wasatch sandstone (Eocene?) at depths of 125 and 700 feet disclosed water forming an almost concentrated solution of sodium carbonate, which for a time was utilized for the manufacture of caustic soda.

Lakes containing sodium sulphate are widely scattered in the arid regions of the western Americas and Central Asia. In the United States

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[1] Summarized by Cameron and Bell, *Bull.* 33, U. S. Bur. Soils, 1906.

[2] A. C. Melcher, *Jour.* Am. Chem. Soc., 32, 1910, pp. 50–66.

[3] A. R. Schultz, Deposits of sodium salts in Wyoming, *Bull.* 430, U. S. Geol. Survey, 1910, pp. 570–589.

R. C. Wells, Sodium and sodium compounds, *Min. Res.*, U. S. Geol. Survey, II, 1918, pp. 159–198.

O. P. Jenkins, Verde River lake beds, *Am. Jour. Sci.*, 5th ser., 5, 1923, p. 65.

R. C. Wells, Sodium sulphate, its sources and use, *Bull.* 717, U. S. Geol. Survey, 1923.

F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 237.

L. H. Cole, Sodium sulphate of western Canada, Dept. Mines, Mines Branch, No. 646, 1926.

J. P. De Wet, Sodium sulphate deposits of Saskatchewan, *Can. Min. J.*, November, 1931.

they occur in Wyoming, Arizona, and New Mexico; and the salts have been leached from surrounding Mesozoic and Cenozoic sediments. The thickness of the salt beds amounts to 50 feet at most, and they may extend over hundreds of acres. The salts consist mainly of thenardite ($Na_2SO_4$), mirabilite ($Na_2SO_4.10H_2O$), glauberite ($NaSO_4.CaSO_4$), bloedite ($Na_2SO_4.MgSO_4.4H_2O$), epsomite ($MgSO_4.7H_2O$), natron, and halite. The largest known deposits are in the Saskatchewan lakes of Canada.

Sodium sulphate is much more soluble in warm than in cold water, but as the similar variation for sodium chloride is small, "a mere change of temperature between summer and winter in salt lakes may cause mirabilite to separate out or to redissolve." The Great Salt Lake, Utah, according to Gilbert, deposits sodium sulphate during winter. The United States production amounted to 32,000 tons in 1930, chiefly from Camp Verde, Arizona, and from Wyoming. The Canadian output was 32,000 tons. Sodium sulphate is used in paper and glass manufacturing and for many metallurgical and industrial purposes. The price is about $6 per ton.

## SODIUM NITRATE

The alkaline nitrates are very soluble salts which are found in larger masses only under exceptional conditions. Sodium nitrate is present in the soil and is produced by the so-called nitrifying bacteria[1] or by reactions between organic nitrogenous matter and alkaline salts. Sodium and potassium nitrates of great purity are sometimes found as efflorescences and veinlets on sheltered cliffs of various rocks and in caves and are in many cases produced by organic agencies. Calcium nitrate is known from limestone caves. Naturally, nitrate deposits are most common in arid countries. In minor quantities nitrates are widely scattered in the Western states[2] and very frequently they are associated with volcanic rocks, particularly rhyolite but also tuffs, basalts, and lake beds in regions of volcanic activity. The volcanic origin of these nitrates is not accepted by all writers but nevertheless it is the most probable theory advanced.

There are two sources of nitrogen which may be utilized by nature for the development of ammonia salts and nitrates. (1) The nitrogen in the air, which may be fixed by organisms or by electric atmospheric discharges and entrainment in rain water. (2) The nitrogen from the

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[1] H. S. Gale, Nitrate deposits, *Bull.* 523, U. S. Geol. Survey, 1912.

[2] G. R. Mansfield and Leona Boardman, Nitrate deposits of the United States, *Bull.* 838, U. S. Geol. Survey, 1932, 107 pp.

L. F. Noble, Nitrate deposits in southeastern California, *Bull.* 820, U. S. Geol. Survey, 1931, 108 pp.

interior of the earth, which possibly is contained in the magma as a nitride of boron or of some metal. At any rate the volcanic gases and exhalations frequently contain nitrogen and ammonia; it is held by many that a fixation of nitrogen from this source as nitrates is well possible.[1]

The only place where nitrates are present in abundance is in the Atacama Desert in northern Chile.[2] These wonderful deposits have long supplied the world with nitrates though this industry is now suffering keenly from competition of artificial nitrate made from the nitrogen of the air; the annual production in 1930 amounted to 3,000,000 metric tons. Improved processes for refining the caliche are now introduced by an American company. The deposits are situated in the provinces of Tarapaca and Antofagasta in the interior dry valleys between the Coast Range and the Andes, at elevations ranging from 1,000 feet to 3,000 feet, and extend for 300 miles parallel to the coast. The lowest depressions are often occupied by salt flats with a little nitrate. The nitrate deposits lie on the gentle slopes of the valleys. The nitrate bed is a superficial formation of considerable, though irregular, extent; it lies below an overburden of a few feet of loose crumbly material with subangular gravel, becoming harder toward the bottom. This overburden contains some nitrate and often much sodium chloride, sodium sulphate, and gypsum, as well as a little sodium iodate. The "caliche" or nitrate bed is a reddish brown sandy gravel cemented with salts; it averages a few feet in thickness. Below the "caliche" lies rudely stratified sand, gravel, or clay, often of considerable thickness. The "caliche" averages about 25 per cent sodium nitrate and the lower limit of workable material is placed at 15 per cent. Associated with the nitrate are a large amount of sodium chloride, more or less of the sulphate and borates of calcium and sodium, and a small but constant quantity of sodium iodate. Small quantities of the nitrates of potassium, calcium, and barium as well as a little calcium iodate and iodo-chromate (lautarite and dietzeite) are found.

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[1] F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 256.

[2] The literature is very extensive and only part can be quoted.

L. Darapsky, Das Department Tal-tal, Berlin, 1900. Ref. *Zeitschr. prakt. Geol.*, 1902, p. 153.

R. A. F. Penrose, Jr., *Jour. Geol.*, 18, 1910, pp. 1–32.

J. T. Singewald, Jr., and B. L. Miller, *Econ. Geol.*, 11, 1916, pp. 103–114.

Lorenzo Sundt, *Econ. Geol.*, 12, 1917, p. 89.

C. Doelter, Die Mineralchemie, 3, 1, 1918, pp. 267–281.

A. H. Rogers and H. R. Van Wagenen, The Chilean nitrate industry, *Bull.* 134, Am. Inst. Min. Met. Eng., Feb. 1918, pp. 505–522. Discussion, *Bull.* 136, *idem*, 1920, pp. 845–848.

W. L. Whitehead, The Chilean nitrate deposits, *Econ. Geol.*, 15, 1920, pp. 187–224.

W. Wetzel, Petrographische Untersuchungen an chilenischen Saltpetergesteinen, *Zeitschr. prakt. Geol.*, 1924, pp. 113–120.

F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 254–260.

The material mined is usually of the following composition:

| | Per cent |
|---|---|
| Sodium nitrate | 14–25 |
| Potassium nitrate | 2– 3 |
| Sodium chloride | 8–50 |
| Sodium sulphate | 2–12 |
| Calcium sulphate | 2– 6 |
| Magnesium sulphate | 0– 3 |
| Sodium biborate | 1– 3 |
| Sodium iodate | 0.05– 1 |
| Sodium perchlorate | 0.1 – 0.5 |
| Insoluble | 0–50 |

The origin of the nitrate deposits of Chile is a much debated question and few authors are in agreement.

Pissis, the Chilean geologist, followed lately, for instance, by Rogers and Van Wagenen, accounts for the deposits by fixation of atmospheric nitrogen by thunderstorms and its descent from the Andes in the underground circulation and ascent to the surface by capillarity.

Penrose and others hold that the nitrate came from beds of bird guano accumulated at the time when the Coast Range did not exist and that the nitrates were gradually leached and mingled with the salt waters of a closed basin. Others are inclined to consider the deposits caused by ordinary bacterial fixation or by oxidation of nitrogenous vegetable matter. Singewald and Miller think that the nitrates have been carried down by the ground water and emphasize that only the usual processes in operation everywhere, have been active. The accumulation is simply caused by the abnormally dry climate. Others like G. Steinmann, seek the origin in the Tertiary volcanic rocks. At Maricunga nitrates are said to occur in a recent volcanic tuff.

All these explanations appear inadequate or forced. The nitrate deposits are not marine or lacustrine. Their extent corresponds in a most remarkable way to the Jurassic and Cretaceous tuffs and lava flows which occupy so much space in this region, and there must be some causal connection. The nitrates may thus be of volcanic origin, and have been leached from the Mesozoic rocks under unusual climatic conditions. This view of the origin is supported by W. L. Whitehead[1] and F. W. Clarke.[2] The constant presence of borates is an additional suggestive fact.

## IODINE

The world's need of iodine is now mainly supplied by the nitrate region of Chile as a by-product. The production was 1,000 metric tons in 1929. Much more could be produced if needed. Smaller quantities are still

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[1] *Op. cit.*

[2] Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 258.

produced from seaweed in Japan, France, Norway, and England, and from volcanic springs in Java. Brines from oil wells often contain some iodine and may provide a future supply sufficient for the United States.[1]

## BORATES[2]

**General Occurrence.**—Borates and other boron compounds appear in nature under conditions indicating widely differing modes of origin. As complex and insoluble borosilicates like tourmaline and datolite they are disseminated in igneous and metamorphic rocks or in pegmatite dikes and fissure veins, but are here of no economic importance except that tourmaline occurring in this manner is sometimes utilized as a gem stone. As boric acid and borates of calcium and magnesium they appear in volcanic exhalations, of which the most famous are the "soffioni" of Tuscany, Italy,[3] from which large amounts of boric acid have been recovered. Borates, principally in the form of borax ($Na_2B_4O_7.10H_2O$), occur in hot springs and in lakes of volcanic regions. Borax was first obtained from such lakes situated in Tibet. The thermal waters of the California Coast Ranges and Nevada (p. 55) often contain boron, sometimes in large quantities. The borates from these springs are sometimes accumulated in little lake basins and there deposited by evaporation as borax crystals. About 50 years ago much borax was won from the Borax Lake, Lake County, California. The evaporated salts contained 62 per cent sodium carbonate, 20 per cent sodium chloride, and 18 per cent borax.

The borates occur abundantly in the *playas*, or shallow basins intermittently covered by water, or in Tertiary lake beds; they are as follows:[4]

Kernite. $Na_2O.2B_2O_3.4H_2O$. Synonym, rasorite.
Tincalconite. $Na_2O.2B_2O_3.5H_2O$.
Borax. $Na_2O.2B_2O_3.10H_2O$. Synonym, tincal.
Kramerite. $Na_2O.2CaO.5B_2O_3.10H_2O$.
Ulexite. $Na_2O.2CaO.5B_2O_3.16H_2O$. Synonym, boronatrocalcite.
Colemanite. $2CaO.3B_2O_3.5H_2O$.

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[1] P. M. Tyler and A. B. Clinton, Bromine and iodine, *Information Circ.* 6387. U. S. Bur. of Mines, 1930, 26 pp.

[2] M. R. Campbell, Reconnaissance of the borax deposits of Death Valley and the Mohave Desert, *Bull.* 200, U. S. Geol. Survey, 1902.

C. R. Keyes, Borax deposits of the United States, *Trans.*, Am. Inst. Min. Eng., 40, 1909, pp. 674–710.

H. S. Gale, The origin of colemanite deposits, *Prof. Paper* 85, U. S. Geol. Survey, 1913, pp. 3–9.

W. F. Foshag, The origin of the colemanite deposits, *Econ. Geol.*, 16, 1921, pp. 199–214.

[3] R. Nasini, I soffioni e i lagoni della Toscana, Rome, 1930, 652 pp.

[4] W. T. Schaller, Borate minerals from the Kramer district, California, *Prof. Paper* 158, U. S. Geol. Survey, 1930, pp. 137–170.

Meyerhofferite. $2CaO.3B_2O_3.7H_2O$.
Inyoite. $2CaO.3B_2O_3.13H_2O$.
Pandermite. $5CaO.6B_2O_3.9H_2O$. Synonym, priceite.
Howlite. $4CaO.5H_2O.5B_2O_3.2SiO_2$.

**Marine Borate Deposits.**—The marine deposits are mainly confined to the beds of potassium salts in central Germany, but borates have also been observed in sodium chloride, anhydrite, or gypsum. The principal occurrence is as boracite ($5MgO.MgCl_2.7B_2O_3$) and several other rare borates in the carnallite region (p. 336)—that is, in the deposits of the last mother liquors of evaporating sea water. The boracite usually forms small crystals or concretions, but one occurrence is recorded of a mass weighing about 1,400 pounds. A few hundred tons of borates are annually obtained by recrystallization. The boron compounds, then, remained with the most easily soluble salts and were finally precipitated as a magnesium salt because of the predominance of that metal over calcium in the sea water.

**Borax Marshes.**—The deserts of southern California, Nevada, and Oregon, are rich in borate deposits. Similar deposits occur in Argentina, Bolivia, and Chile. The desolate plains between the barren ranges of volcanic rocks contain shallow basins, which during the brief seasons of rain are covered with thin sheets of water. The evaporation of this water leaves dazzling expanses of white salt deposit or efflorescence, some of which may become covered by the fine sand carried by the desert storms. These deposits were discovered about 1870 and for many years yielded a large production of borax at Searles Marsh (60 miles north of Barstow, California), Death Valley, and other places. Though enormous quantities of these salts remain, they are now of little importance because richer and more easily worked deposits have replaced them. The crusts are rarely more than 1 foot thick; the percentage of borax varies considerably. According to Bailey, the crude salt from Searles Marsh yielded in per cent 50 sand, 12 sodium chloride, 10 sodium carbonate, 16 sodium sulphate, and 12 borax. Borings showed 20 feet of clay and sand with crystals of calcium borate, underlain by a bed of solid trona 28 feet thick, and below this 350 feet of clays impregnated with hydrogen sulphide. On ground that had been worked over, a new crust formed by capillary action that was thick enough to remove in 3 or 4 years. The area productive of borax amounts to about 1,700 acres, on which in wet seasons stands about 1 foot of water.

Though no ulexite was found at Searles Marsh, it is common in many other *playa* deposits, in both California and Nevada and in Argentina. It usually forms concretions of silky fiber, known as "cotton balls."

**Tertiary Lake Beds.**—The borates in the marshes and *playas* have undoubtedly been leached from the older deposits in the Tertiary lake beds, which have been recognized at many points in Inyo, San Bernardino, Kern, Los Angeles, and Ventura counties, California. These beds yield colemanite almost exclusively.

The colemanite deposits begin near the Pacific coast at Piru, Ventura County, and near Saugus, Los Angeles County, where, according to Keyes, they lie in a series of yellow clays and sandstones, probably Miocene in age and several thousand feet in thickness. The mineral is present as nodules in clay, and above the borate beds are strata of gypsum.

Other important beds are in the foot-hills of the Calico Mountains north of the Mojave River; they have a steep dip and have been mined by shafts to a depth of 400 feet. These large masses of low-grade colemanite shales, with 7 to 20 per cent boric acid, are not mined now; attention being confined to two solid beds of the mineral 7 to 10 feet thick. Rhyolite tuffs lie underneath the borate beds.

The richest colemanite beds are, however, in the Furnace Creek region of the Funeral Range, which overlooks Death Valley, in Inyo County, California. About 4,000 feet of Tertiary non-fossiliferous sediments are recognized here, which form a broad belt obliquely crossing the range and dipping 20° to 45° N. E. The lower and thicker part, according to Keyes, consists of conglomerates and sandstones, above which are olive-colored clays interbedded with basalts. The upper part of the clay series carries gypsum, colemanite, and thin layers of limestone. The borate beds are traceable for 25 miles. Within the colemanite-bearing beds, which may be as much as 50 feet thick, the bluish clays are thickly interspersed with milky white layers or nodules of the white, coarsely crystalline mineral, mingled with more or less gypsum, according to Keyes. The solid layers may be 15 feet thick. Near by the clays are impregnated with fine particles of colemanite and yield 10 to 25 per cent boric acid, but these low-grade deposits are not utilized at present.

In 1926 an important deposit of kernite was discovered on the Mojave Desert near Kramer, in Miocene "lake beds," mostly clay, overlying a basalt flow. The mineral here occurs abundantly and pure, associated with borax and a little stibnite and realgar. Near by the lake beds were known to contain deposits of colemanite and ulexite. Kernite occurs in clear crystals embedded in the clay, some of them are up to 8 feet in thickness; it is slowly soluble in cold, easily in hot water. The deposit is said to extend over an area of over 40 acres with an average thickness of 80 feet at a depth of 400 feet. The larger part of the borax produced in the United States now comes from this deposit.[1] As no other occurrences of kernite and kramerite are known, it is believed that very special conditions governed their deposition.

**Origin.**—The colemanite deposits, which in places occur with gypsum and limestone, are surely not of marine origin and can hardly be supposed to be saline precipitates from evaporating lake water. Replacement of

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[1] *Mineral Industry*, 1930, p. 545.

limestone has been suggested but Foshag points out that boric acid is not likely to replace carbonic acid and shows on the basis of van't Hoff's work that ulexite was the primary mineral which by leaching with NaCl would be converted to colemanite.[1] The strontianite deposits found in similar lake beds (p. 402) form a somewhat analogous occurrence.

The reaction $2NaCaB_5O_9.8H_2O$ (Ulexite) $\rightleftarrows$ $Na_2B_4O_7.10H_2O$ (Borax) + $Ca_2B_6O_{11}.5H_2O$ (Colemanite) + $H_2O$ is reversible.

Van't Hoff[2] has produced pandermite and colemanite from meyerhofferite and states that ulexite, pandermite, and probably colemanite can be formed at temperatures from 25° C. upward. The first salts precipitated in oceanic waters are calcium salts. For borates the saturation point is not reached until carnallite is precipitated.

W. T. Schaller and R. C. Wells obtained kernite and tincalconite artificially from borax by a special method and temperature of 150° C. It is therefore considered probable that it was formed by hot water acting on a pre-existing borax deposit, and that this hot water was evidently of volcanic origin as indicated by the presence of realgar and stibnite in the deposit.

**Production and Uses.**—The borate industry is now concentrated in southeastern California and has shown great expansion in the last years. In 1930, 177,400 short tons of crude borates were mined. The products are shipped direct to the sea board, where the material is manufactured into borax and boric acid.

Borax is extensively used in industrial chemistry, in metal enameling, in medicine, and in the household. Smaller quantities of borax are produced in Italy, Turkey, Peru, Argentina, and Chile, but practically the United States controls the industry.

## SODIUM CHLORIDE

**Occurrence.**[3]—Sodium chloride or common salt forms beds in sedimentary rocks, and in most cases its derivation by evaporation of saline solutions is clear. Of about 8,000,000 tons of salt produced in the United States in 1930, one-fourth was mined in solid form. Most of it is obtained from brines derived from solution of salt beds by natural waters or by water forced down into bore-holes to the saline strata; much also is produced by evaporation of sea water or water of saline lakes, such as the Great Salt Lake of Utah. The price in 1930 was $3 per short ton.

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[1] W. F. Foshag, *op. cit.*

[2] J. H. van't Hoff, Zur Bildung der ozeanischen Salzablagerungen, 2, 1909, pp. 45–75.

[3] J. O. von Buschmann, Das Salz, Leipzig, 1906 and 1909, 2 vols.

W. C. Phalen, Technology of salt making in the United States, *Bull.* 146. U. S. Bur. Mines, 1917 (with description of deposits).

A. W. Grabau, Principles of salt deposition, New York, 1920.

Salt beds are present in formations of different ages, but are perhaps most common in the Permian and Triassic strata; the oldest saline rocks in the United States are those of the Silurian in New York State. As may be easily understood from the general statements on previous pages, strata of calcium sulphate are ordinarily associated with salt beds and should appear below them; owing to recurrent and shifting epochs of desiccation the order may be reversed and gypsum beds appear above the salt. It is also very common to find crystals or streaks of anhydrite or gypsum with salt, as well as streaks of clay. In thickness salt beds vary enormously—from the thinnest strata to masses 1,500 feet or even

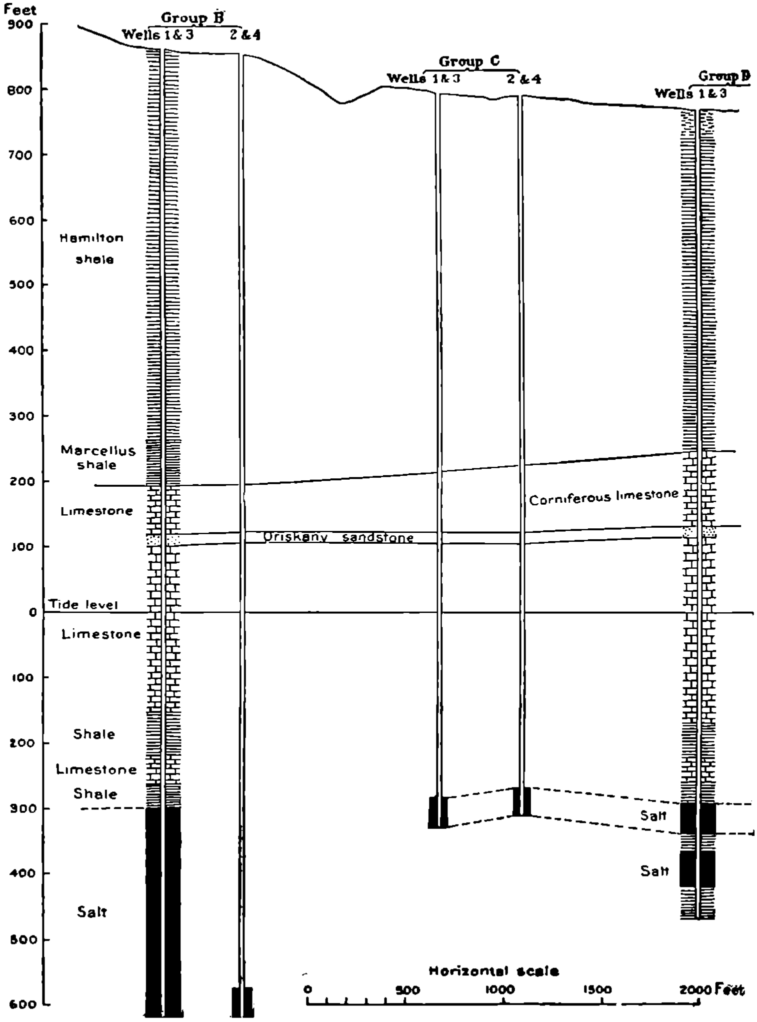

FIG. 124.—Sections of salt wells, Tully, New York. (*After F. J. H. Merrill.*)

more in thickness. A bore-hole near Sperenberg, in the German potash region, penetrated 3,900 feet of salt, but here, as in so many other places, the apparent thickness may be deceptive, being due to movements of folding and faulting. Besides, the plasticity of salt is remarkably great, much greater than that of the accompanying clays and anhydrite, and this, as the German geologists have found, leads to most astonishing and confusing stratigraphic relations.

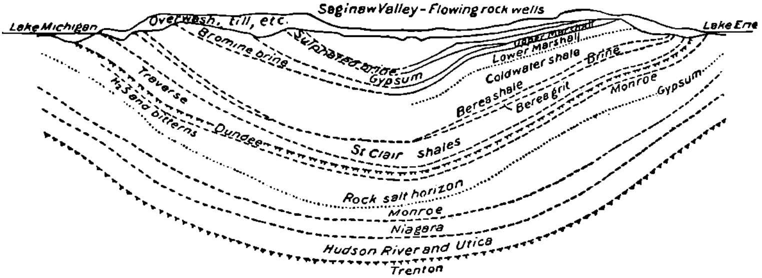

FIG. 125.—Section of lower Michigan basin. (*After A. C. Lane.*)

**Examples.**—Salt beds occur in New York State in the red Salina shales of the Silurian and underlie a considerable area. Much of the salt is recovered from artificial brines. The salt forms pure lenticular masses and layers interbedded with soft shales, limestone, and gypsum; the salt-bearing formation having a variable thickness up to 470 feet (Fig. 124). At Ithaca several beds of salt occur at a depth of 2,244 feet with a total thickness of 248 feet. A magnesian limestone, containing gypsum, lies above. Rock salt is mined at several places, one shaft opened at Cuylerville being 1,100 feet deep and reaching a salt bed 21 feet in thickness. Salt has also been mined near the outcrops of the beds at Livonia. Similar beds are worked in Ohio by bore-holes and brines.

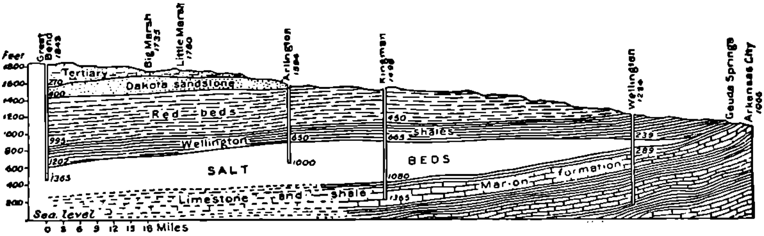

FIG. 126.—Section of Permian salt formation in Kansas. (*From Mineral Resources of Kansas.*)

The greatest salt production in the United States is derived from Michigan.[1] The salt occurs as thick beds at different horizons in the Salina formation and also in the sandstones of the Mississippian, or Lower Carboniferous (Fig. 125). The salt is recovered by means of natural and

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[1] A. C. Lane, *Water-supply Paper* 30, U. S. Geol. Survey 1899.

artificial brines; bromine, in which these brines are unusually rich, is obtained as a by-product of the final mother liquor. Deep mining has been undertaken under considerable difficulties near Detroit.

Kansas is likewise among the great producers.[1] Some salt is obtained from salt springs in the Carboniferous and on the "salt plains" leached from Permian beds. From the latter the principal product is derived; it occurs interstratified with shales, the total thickness of the salt beds being at most 500 feet. Some of the beds are said to be over 200 feet thick, but generally they are much less (Fig. 126). The deposits extend into Oklahoma (Fig. 127).

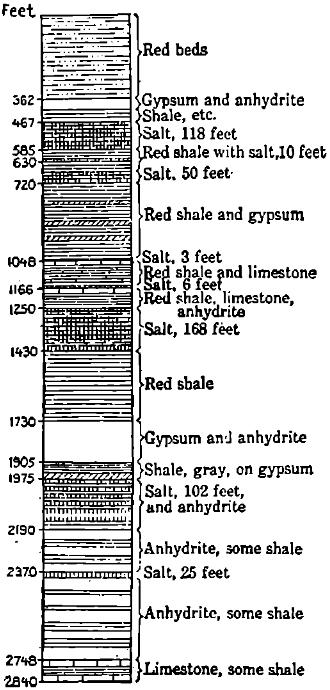

FIG. 127.—Deep boring three miles east of Gate City, Oklahoma. (*After N. H. Darton, U. S. Geol. Survey.*)

In the western arid states *playa* deposits of salt are common in the dry basins between the ranges; they are usually thin.

The most noted deposit of this kind is that of Salton, Imperial County, California, where the basin lies below the level of the sea. Before the recent flooding by the Colorado River an important production was maintained here. A large area is covered by salt crusts 10 to 20 inches in thickness. Below this lies a thin mud deposit covering another salt crust. Deeper borings encountered 22 feet of black mud containing salt and soda, and this covers 270 feet of hard clay.[2]

The desert regions of northern Africa and central Asia offer similar occurrences in abundance.

Large deposits of impure salt mixed with clay have been worked for a long time in the Alpine Triassic of Tyrol; they lie between limestone beds. Another important saline region fringes the outside of the Carpathian chain in Roumania, Transylvania, and Galicia and is contained in Miocene sands and clays. The beds are generally greatly disturbed, brecciated, and pressed. The best-known place where mining is carried on is Wieliczka, in Galicia, which is much visited by tourists on account of the picturesque and extensive workings, now about 1,000 feet deep,

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[1] M. Z. Kirk, Mineral resources of Kansas, Univ. Kansas Geol. Survey, 1898.

N. H. Darton, Permian salt deposits of the southcentral United States, *Bull.* 715, U. S. Geol. Survey, 1921, pp. 205–224; "Red Beds" and associated formations in New Mexico, *Bull.* 794, *idem*, 1928.

[2] G. E. Bailey, *Bull.* 24, California State Min. Bur., 1902, p. 126.

with elaborate carvings in solid salt. The salt beds of the Stassfurt region will be described later.

**The Salt Domes of the Gulf Coast.**[1]—The greatest salt deposits in the United States have been discovered by borings in Louisiana and the adjoining coast belt of Texas; they show many unusual features and some difficulty has been experienced in explaining their genesis. Above the low and swampy coast west of New Orleans rise a number of low mounds or knolls and below these most of the salt has been found. It does not occur in regular beds but as enormous subterranean domes, surrounded on all sides by thick and often steeply dipping beds of Quaternary and Tertiary clays and sands. At some places a thin-bedded Cretaceous limestone appears at the surface. Figure 128 gives a suggestion of the strange relations encountered. At Petite Anse, according to Harris, the drill shows 2,263 feet of almost pure salt, followed by 70 feet of foreign matter, below which the drill again enters rock salt of unknown thickness. On Côte Carline the drill entered salt at 334 feet and continued in salt without change till the drilling ceased at 2,090 feet. At Belle Isle[2] the Knapp Well No. 1 penetrated 2,000 feet of salt and, below this, 236 feet

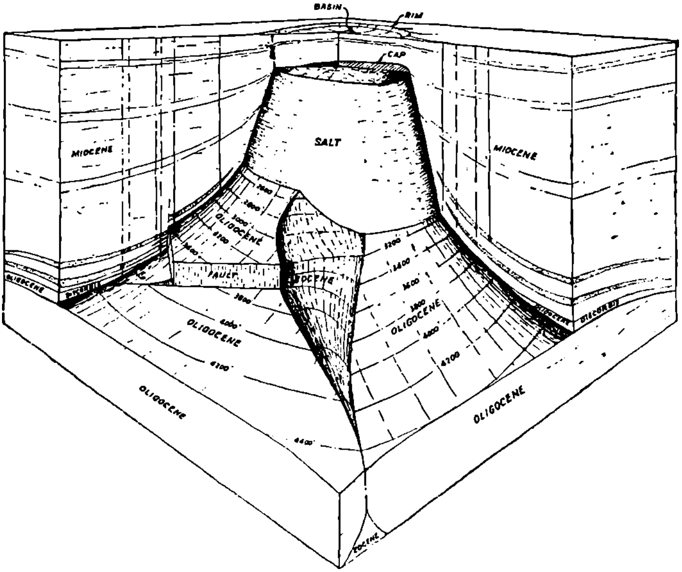

FIG. 128.—Perspective drawing of West Columbia salt dome. Depths in feet. (*After D. P. Carlton.*)

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[1] G. S. Rogers, Intrusive origin of the Gulf Coast Salt Domes, *Econ. Geol.*, 13, 1918, pp. 447–485.

E. L. DeGolyer, The theory of volcanic origin of the salt domes, *Bull.* 137, Am. Inst. Min. Eng., 1918, pp. 987–1000.

[2] A. F. Lucas, *Trans.*, Am. Inst. Min. Eng., 57, 1917, p. 1034.

of anhydrite and gypsum. Another well at Humble is said to have penetrated 5,410 feet of salt.

Oil, gas, and sulphur are often met in the drill-holes. Gypsum and anhydrite, in beds 200, 400, or even 600 feet thick, cover the salt in some places, or again the salt may be overlain (as at Spindletop, Texas) by several hundred feet of a porous limestone carrying oil. The dip of the loose strata forming the outside of the dome is steep and bore-holes only a short distance from those disclosing salt may fail to encounter it.

These enormous salt resources are as yet little utilized. In Louisiana, rock salt was mined in 1925 at Weeks Island, where the shaft is 645 feet deep, at Avery Island, at a depth of about 500 feet, and at Carline at a depth of 900 feet. The shafts are sunk in heavy, wet ground until the impermeable salt is reached. In places there is considerable danger of flooding the mine by driving into the loose strata.[1]

According to older views, these wonderful salt domes were deposited by ascending solutions; some believed that the domes are uplifts caused by laccolithic intrusions; others that they are produced by the expanding power of growing salt crystals.

None of these views is convincing. Recent literature has shown the existence of many salt domes along the coast and some quite a distance inland.[2] The same kind of salt domes are also found on the isthmus of Tehuantepec back of Puerto Mexico,[3] accompanied in places by oil and gas.

The foreign literature indicates that such salt domes also exist in northern Germany, in Transylvania, and in other countries particularly in Persia.[4]

Hopkins shows clearly that the salt dome at Palestine, Texas, is caused by a highly localized vertical uplift of quaquaversal form. Rogers and DeGolyer arrive at similar conclusions. It is probable that these domes are Carboniferous, Permian, or possibly Cretaceous salt beds forced up through the softer sediments. This is made possible by the extraordinary plasticity of rock salt, which easily yields to deformation. The nature of the force producing these uplifts remains in doubt.

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[1] F. E. Vaughn, The five islands, Louisiana, *Bull.* Am. Assoc. Petrol. Geol., 9, 1925, pp. 756–797.

[2] O. B. Hopkins, *Bull.* 616, U. S. Geol. Survey, 1917, p. 28.

[3] Burton Hartley, *Econ. Geol.*, 12, 1917, pp. 581–588.

[4] F. F. Hahn, *Econ. Geol.*, 7, 1912, pp. 120–135.

E. Fulda, Die Salzstöcke am persischen Golf, *Kali*, 25, 1931, pp. 1–5; 120–121.

J. V. Harrison, The geology of some salt plugs in Laristan, *Quart. Jour.* Geol. Soc. (London), 86, 1930, pp. 462–522.

In south Persia 107 salt domes are recognized; some of them at least are of Cambrian age. Erosion has formed crater-like depressions in places, and there are also spire-like masses of salt that rise 1,000 meters above sea level. Origin by tangential pressure on a series of sediments 25,000 feet thick is suggested.

As these salt domes on the Gulf coast are the carriers of oil, gas, and sulphur, much attention has been given to them and many new domes have been discovered by modern methods of geophysical investigation, that is, by the gravity balance or by seismometric instruments. Volume 9 of the *Bulletin* of the American Association of Petroleum Geologists contains 21 papers dealing with this subject which confirm in general the conclusion referred to above.[1]

**Composition, Production and Use.**—Salt as marketed is usually very pure. Calcium sulphate is the principal impurity and is often present to the amount of over 1 per cent. Salt from some desert lakes contains sodium carbonate and sulphate. In 1930, about 2,000,000 short tons of rock salt was mined in the United States. The total production of salt for the same year was 8,000,000 short tons. The average price was $3 per ton. New York, Kansas, Louisiana, and Michigan produced 99 per cent of the output; the same states were practically the only producers of rock salt.

The wide range of uses of salt for culinary, preservative, and industrial purposes need not be specified. Large amounts are used in the manufacture of other sodium salts, particularly the carbonate.

**Bromine and Calcium Chloride.**—In 1930, over 4,000 short tons of bromine were produced in California, Michigan, Ohio, and West Virginia from residues of pumped salt brines; the price was about 25 cents per pound. The production has increased rapidly. Experiments are in progress to recover bromine from sea water on a large scale. A large production of bromine is obtained from the German salt deposits. The output is calculated as bromine, but actually much of it is in the form of alkaline bromides.

About 116,000 short tons of calcium chloride at about $19 per ton are produced annually in the United States from pumped brines. A partial analysis of such a brine follows:[2]

ANALYSIS OF BRINE FROM POMEROY, OHIO

(Grams per Liter)

| | |
|---|---|
| Sodium chloride | 84.30 |
| Calcium chloride | 14.34 |
| Potassium chloride | 0.114 |
| Magnesium chloride | 5.50 |
| Barium chloride | 0.343 |
| Strontium chloride | 0.257 |
| Sodium iodide | 0.004 |
| Magnesium bromide | 0.155 |

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[1] E. L. DeGolyer, Origin of North American salt domes, *Bull.* Am. Assoc. Petrol. Geol., 9, 1925, pp. 831–874.

D. C. Barton, The salt domes of south Texas, *idem*, pp. 536–589.

[2] J. A. Brownocker, Ohio Geol. Survey, 8, 1906.

## THE GERMAN POTASSIUM SALTS[1]

If carried to its conclusion the process of evaporation of sea water will result in the deposition of the easily soluble chlorides and sulphates of potassium and magnesium, also chloride of calcium. Evidently this seldom takes place. Almost the only locality thus far discovered where the whole sequence of salts is present is in central Germany, north and south of the Harz Mountains, in formations of Permian age. In 1930, 40 mines yielded annually about 12,000,000 metric tons of crude potassium salts, of which 85 per cent is used as soil fertilizer and the remainder for general industrial purposes. The value of this production was about $45,000,000. The mining is done exclusively by shafts from 1,000 to 2,500 feet deep. Circular shafts lined with concrete or iron tubing are

| No. | Thickness in meters | Character of strata |
|---|---|---|
| 1 | 20 to 30 | Red clay with a little anhydrite and rock salt. |
| 2 | 50 | Rock salt. |
| 3 | 1 to 5 | Anhydrite with salt. |
| 4 | 5 to 15 | Red clay with anhydrite and rock salt. |
| 5 | 100 to 150 | Younger rock salt. |
| 6 | 40 to 90 | Main anhydrite. |
| 7 | 4 to 10 | Salt clay. |
| 8 | 30 to 40 | Carnallite zone. Carnallite ($KCl.MgCl_2+6H_2O$). |
| 9 | 20 to 40 | Kieserite zone. Kieserite ($MgSO_4+H_2O$). |
| 10 | 40 to 60 | Polyhalite zone. Polyhalite ($2CaSO_4.MgSO_4.K_2SO_4+2H_2O$). |
| 11 | 300 to 500 | Older rock salt, with narrow streaks of anhydrite, interpreted as annual deposits. |
| 12 | 70 to 100 | Older anhydrite. |
| 13 | 4 to 10 | Limestone (Zechstein Kalk). Marine deposit. |
| 14 | 0.5 to 4 | Black copper-bearing shale.<br>Conglomerate. |
|  |  | Lower Permian and Carboniferous beds. |

The older series (Nos. 8 to 11) closed with deposition of potassium salts. The younger series (Nos. 1 to 7) contains no potassium salts.

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[1] R. Ehrhardt, Die Kali-Industrie, 1907.

Beyschlag, Everding, Erdmann, Loewe, and Paxmann, Deutschlands Kalibergbau, 1907.

Summaries: F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 221–228.

H. E. Boeke and W. Eitel, Grundlagen der physikalischen chemischen Petrographie, Berlin, 1923, pp. 453–510.

R. Marc and H. Jung, Physikalische Chemie, Jena, 1930.

E. R. Lilley, Potash in Germany, *Eng. Min. Jour.*, January, 1932, pp. 33–37.

used and the greatest caution is necessary to prevent influx of water during sinking or working; if the water once breaks in, the mine will probably have to be abandoned.

Some of the products are sold for fertilizers without further chemical treatment. Carnallite is the most important of these, and next to it comes kainite; as mined, both are mixed with 30 or 40 per cent of common salt. Other chemical products are chloride and sulphate of potassium and potassium-magnesium sulphate. Kieserite is refined to magnesium sulphate.

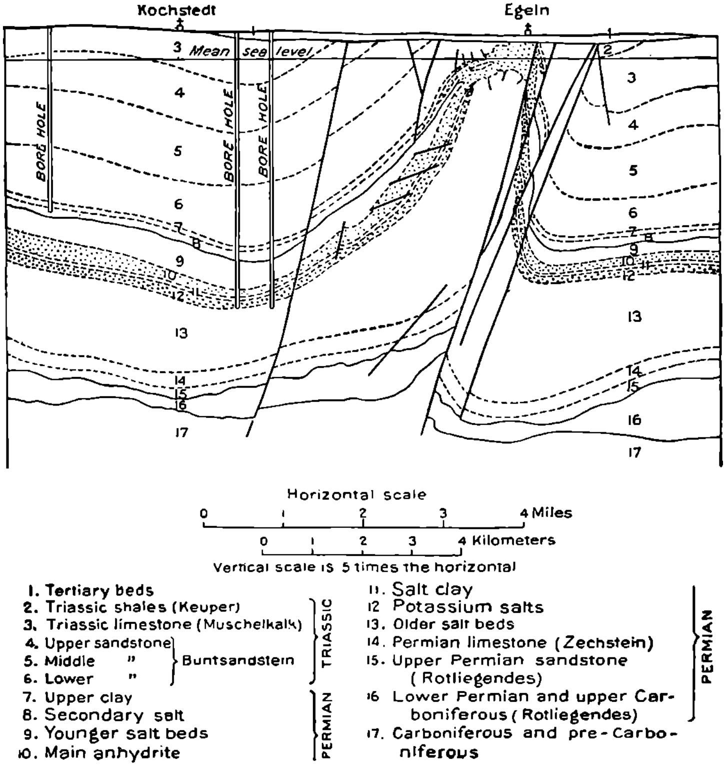

FIG. 129.—Section of the Stassfurt-Egeln anticline. (*After Everding.*)

The potassium salts lie as a relatively thin series of strata over a heavy bed of rock salt in the Permian and are in turn covered by Triassic sandstones and limestones, and finally by the Tertiary and Quaternary beds. They form a series of faulted synclines and anticlines (Fig. 129), in places approaching closely to the surface, but here generally changed by secondary leaching processes.

The general section is as shown on page 336,[1] counted from the bottom of the Triassic sandstone.

[1] H. Everding, Deutschlands Kalibergbau, 1907, p. 36.

Rock salt is really present throughout the section, for the carnallite zone, which yields the greatest quantity of crude product, contains only about 55 per cent of carnallite and includes also 25 per cent rock salt and 16 per cent kieserite. The kieserite zone yields 65 per cent rock salt and only 17 per cent kieserite. Rock salt, kieserite, sylvite (KCl), carnallite ($KCl.MgCl_2 + 6H_2O$), and kainite ($KCl.MgSO_4 + 3H_2O$) are the main products.

The German geologists have shown that extensive secondary changes have taken place in the salt beds—in part immediately after the deposition, in part much later, following the Triassic sedimentation and progressing even now. These post-Triassic changes have occurred both in the outcrops and at greater depth. The minerals just enumerated may occur in all three generations, but in addition a large number of more or less complicated secondary compounds were formed.

Kainite is in part a secondary product derived from carnallite by the leaching of $MgCl_2$. It forms along the crests of the anticlines. Under some circumstances a secondary mixture of potassium chloride, kieserite, and salt would be formed instead of kainite, and this constitutes an important product under the name of "hartsalz." Secondary deposits of the older type appear between the carnallite and the salt clay. In all these transformations the products are very complex.

Van't Hoff[1] and his associates have studied the various combinations of salts in order to ascertain their fields of stability at temperatures ranging from 25° to 83° C. Sylvite and carnallite are stable in concentrated NaCl solutions from 0° to 85° C. Many of the rarer salts (langbeinite ($2MgSO_4.K_2SO_4$), for instance) are stable under these conditions only from 37° C. upward.

Some of the temperatures required seem high; kieserite with sylvite, for instance, forming above 72° C.

However, the temperature in the middle depths of evaporating salt lakes may be very high. A. V. Kaleczinsky[2] found the temperatures of certain Hungarian salt lakes to be as much as 71° C. during the summer at a depth of 1.3 meters, while the surface and the bottom layers were much cooler, about 20° C.

The present trend of thought is, however, that many of these metamorphic changes are caused by a depression of the primary salt deposits into greater depths where a partial recrystallization took place. A later rise of the beds again brought them near the surface, and resulted, in places, in a reversal of the metamorphic reactions.[3]

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[1] Van't Hoff, Die ozeanischen Salzablagerungen, 1905 and 1909.

H. Precht and E. Cohen, Die Bildungsverhältnisse ozeanischer Salzablagerungen, Leipzig, 1912.

[2] A. V. Kaleczinsky, Ref. *Ann. Phys.* (4) 7, 1902.

[3] H. E. Boeke and W. Eitel, *op. cit.*, 1923, p. 492.

For recent papers on the transformation, see footnote.[1]

## OTHER SOURCES OF POTASSIUM SALTS

The exhausted agricultural lands of all countries need potassium salts,[2] together with phosphates and nitrogen compounds. Germany is the only country in which potassium salts in easily available form occur on a large scale. The imports of the United States in 1913 amounted to about 1,800,000 metric tons while the highly stimulated domestic production since the World War amounts to but 20 per cent of the former consumption. This brief statement indicates the acute situation. In 1930 nearly 1,000,000 short tons of potassium salts were imported.

Promising beds of potassium salts similar to those of Germany have been discovered in the Oligocene of Alsace and are now producing.[3] Other deposits occur in the Tertiary beds of Northeastern Spain,[4] and Galicia (Poland). Both of these regions now contribute to the world's supply.

Important potassium salts have been discovered by drilling at Solikamsk,[5] in the east Russian Permian and production has begun. The minerals are halite, sylvite and carnallite.

An active search for potassium salts has been carried on in the United States since 1910. Since the discovery, in 1912, by Dr. J. A. Udden, of the presence of potash in the Permian salt beds of western Texas many

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E. Jänecke, Entstehung der deutschen Kalisalzlagerstätten, 1915.

E. Jänecke, *Zeitschr. anorg. Chemie*, 100 (1917), 102 (1918), 103 (1918).

[1] K. Weber, Geologisch-petrographische Untersuchungen am Stassfurt-Egelner . . . , *Kali*, 25, Nos. 2, 3, 4, 5, 6, 7, and 8, 1931.

Older salt beds consist of anhydrite, polyhalite, and kieserite; above, potash beds carry carnallite and "hartsalz." Polyhalite and kieserite replace anhydrite; vanthoffite replaces kieserite and carnallite; kieserite is altered to sylvite. Many of the changes resulted from downward percolating KCl and $MgCl_2$. No gypsum is primary.

M. Rózsa, Entstehung und Umwandlungsvorgänge in den deutschen Kalisalzlagern, *Kali*, 25, Nos. 9 and 10, 1931. A detailed discussion of the various mineral changes in the deposits.

[2] H. S. Gale, The search for potash in the United States, *Bull.* 530, U. S. Geol. Survey, 1911. Also in *Bull.* 580, *idem*, 1914, pp. 265–317.

A. R. Schultz and Whitman Cross, Potash-bearing rocks of the Leucite Hills, Wyoming, *Bull.* 512, *idem*, 1912.

B. S. Butler, and H. S. Gale, Alunite, *Bull.* 511, *idem*, 1912.

[3] H. S. Gale, Potash deposits in Alsace, *Bull.* 715, U. S. Geol. Survey, 1921, pp. 16–56.

[4] H. S. Gale, Potash deposits in Spain, *idem*, pp. 1–16.

Keyser, . . . , Die wichtigsten Kalibergbaubezirke ausserhalb Deutschlands und Frankreichs, *Zeitschr.* Berg-Hütten-u. Salinenwesen 77(7), 1929, pp. B. 499–512.

[5] E. Kordes, *Kali*, 25, 1931, pp. 349–352.

E. Razumovskaiya, Description of the Solikamsk potash-salt deposits, *Trans.*, U. S. S. R. Geol. and Prosp. Service, 54, 1931, 43 pp. (Russian, with English summary).

wells have been drilled which prove the existence of workable beds of potash salts.[1]

The potash deposits lie in the southern part of the great Permian geosynclinal salt basin and occupy, as far as known, an area of about 40,000 square miles (Fig. 130). They are embedded in a body of Permian halite; above the salt lies Permian and Triassic red beds and Comanche Cretaceous; the underlying formations are Permian and upper Pennsylvanian. The top of the salt beds lies from 300 to 2,000 feet below the surface. The thickness of the salt series may average 1,000 feet; the thickness of the Permian beds ranges from 4,000 to 11,000 feet.

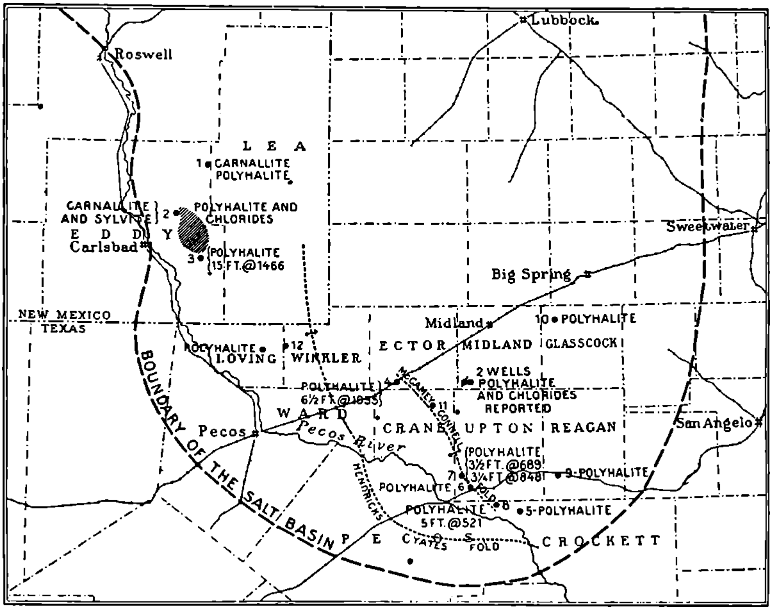

FIG. 130.—Texas-New Mexico potash area, showing location and results of core tests. (*After G. R. Mansfield and W. B. Lang.*)

Within the salt series are many beds of anhydrite and halite with which are interbedded thinner layers of potassium salts. Drilling has been done by the United States Government and by private parties. Polyhalite is the principal potash mineral, usually replacing anhydrite.

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[1] W. B. Lang, Potash investigations in 1924, *Bull.* 785, U. S. Geol. Survey, 1926, pp. 29–43. Press Bulletin, *idem*, Mar. 25, 1929.

H. W. Hoots, *Bull.* 780, *idem*, 1926, pp. 33–126 (Geology of the beds).

G. R. Mansfield and W. B. Lang, Government potash explorations in Texas and New Mexico, *Yearbook*, Am. Inst. Min. Met. Eng., 1929, pp. 241–255.

W. T. Schaller and E. P. Henderson, Mineralogy of drill cores from the potash fields of New Mexico and Texas, *Bull.* 833, U. S. Geol. Survey, 1932, 124 pp. (Detailed mineralogical descriptions of drill cores.)

In a government well in Eddy County, New Mexico, a thickness of 23 feet was found, averaging about 15 per cent $K_2O$.

The area was probably a shallow bay separated by a bar from the ocean; changes in climate or slight structural movements altered from time to time the position of the successive desiccation pans within the basin.

The earliest mineral is anhydrite, which is banded by layers of fine-grained magnesite. Polyhalite and sylvite occur in considerable quantity, but there are many rarer minerals similar to those found in the German potash beds. Many of these potash minerals originated by the reaction of pre-existing saline minerals with liquors rich in potash. The relations of halite and sylvite are difficult to decipher.

A shaft was sunk in 1930 by the United States Potash Company near Carlsbad, New Mexico, to a depth of 1,100 feet. A commercial bed of "sylvite" was met at 980 feet, and shipments began in 1931 said to contain 27 per cent $K_2O$. Output in August, 1931, was reported as 11,000 tons of ore.

Naturally it will take a long time to determine the extent of the deposits and the commercial possibilities.

**Potassium in Rocks and Minerals.**—Granites, pegmatites, some phonolites, and some leucite rocks contain a considerable amount of potassium, varying from 5 to 12 per cent. Unfortunately there is great difficulty in transforming the insoluble potassium silicates contained in the orthoclase, leucite, or glassy base in these rocks into soluble salts. If orthoclase or any potassium-bearing rock is ground to fine powder and slimed with water a certain small percentage of potash salt is converted into soluble form, probably as a potassium silicate; and it is said that such finely ground powder has some slight fertilizing power. Processes have been patented based on electrolytic treatment of the slimed rock or treatment with quicklime and calcium chloride and subsequent calcining; methods by which soluble potassium salts are said to be set free. Over 41,000 tons of leucite rocks were mined in Italy in 1930 for the manufacture of potash salts by treatment with nitric acid. It has been proposed to utilize potash-bearing shale beds in Georgia and Minnesota.

Greensand marls—for instance, the Cretaceous beds in New Jersey—contain from 3 to 6 per cent of potash besides some phosphoric acid, the former in glauconite, the latter in calcium phosphate. These marls are used in their crude state as fertilizers, and the recovery in soluble form of their potassium content has been proposed.

Another source of potassium has been sought in alunite, a mineral of inconspicuous appearance, white or pink, compact or fine granular, rarely coarse granular. ($K_2O.3Al_2O_3.4SO_3.6H_2O$; 8 to 11 per cent $K_2O$.) It is a product of rock alteration in volcanic regions probably caused by waters containing sulphuric acid; it is much more common than the soluble natural alum which sometimes appears as efflorescences. The alunite is found disseminated in the rocks or in well-defined veins. Notable western occurrences are at Goldfield, Nevada; Marysvale, Utah; and the Rosita Hills, Colorado. Among the foreign deposits which have already been utilized are those at La Tolfa, in Italy; Almeria, in Spain; and Bullah Delah, in New South Wales. The transformation of alunite into soluble potassium sulphate is easily effected by calcination; part of the sulphuric acid and all of the water are volatilized, leaving soluble potassium sulphate and an insoluble residue of alumina. The manufacture of potassium sulphate from alunite was carried on at Marysvale during the World War.

**Potassium in Brines.**—Potassium salts are easily soluble and therefore remain with calcium and magnesium chlorides in the last residues, or mother liquors, the so-called "bitterns." Many natural brines pumped from bore-holes in salt-bearing beds contain some potassium and under favorable circumstances may be used for the recovery of these salts. Some of the Michigan brines from the Marshall formation of the Lower Carboniferous (Fig. 125) contain from 3 to 5 grams per liter of potassium chloride; salt, calcium chloride, and bromine are recovered from these brines, but their potassium content appears to be too small for profitable recovery. In places certain well-defined strata yield natural brines, or residual "bitterns." One such bittern from Fairport Harbor, Ohio, on Lake Erie, contains, according to W. C. Phalen, in grams per liter, 7.4 KCl, 110.1 NaCl, 134.4 $CaCl_2$, 43.2 $MgCl_2$. Such a brine could possibly be utilized for the recovery of potassium. This stratum is almost 400 feet above the topmost salt bed from which artificial brines are pumped in Ohio. Potash brines, probably workable, are also found in the Salt Lake Desert, Utah, on the line of the Western Pacific Railroad Company.

Lakes in dry regions, especially in areas of former volcanic activity, contain appreciable quantities of potassium. The water of Owens Lake, in eastern California, yields almost 3 grams of potassium chloride per liter. No production is reported.

Evaporation in the Quaternary lakes of the Lahontan basin in Nevada and California has at many places resulted in deposits of salt of moderate thickness. Changes in drainage among these basins sometimes resulted in the residual brines, richer in potash, being drawn off into a neighboring depression, and thus it happens, as at Searles Marsh, in San Bernardino County, California, that the salt bed, which here is almost 60 feet in thickness and covers an area of at least 11 square miles, is saturated with a strong brine unusually rich in potassium. In the dissolved salts of Searles Marsh there is almost 7 per cent of $K_2O$. Plants are here in operation producing potassium salts and borax.

During the World War some potash was recovered from the brines of certain small lakes in Nebraska and Colorado.

The earliest source of potassium was, as is well known, ashes of vegetable matter. Seaweed is especially rich in this metal and also contains iodine.

At the present time potash salts are recovered in the United States, chiefly from the brines of Searles Marsh but also from other products, such as molasses, distillery waste, and the dust of cement plants and blast furnaces. The whole recovery for 1930 is 106,000 short tons of crude potash equivalent to 61,300 tons of $K_2O$; against this stands a consumption for 1930 of 1,060,244 tons crude or 390,000 tons of available $K_2O$. The future will show in how far this situation may be remedied by the working of the newly discovered potash salts in Texas and New Mexico. At the present time, France and Germany are the only large producers of potash; the former enemies have now combined to regulate the price of potash in the United States. The prices in 1930 were \$35 to \$37 a short ton for 80 per cent "muriate" and \$46 to \$48 a short ton for 90 per cent "sulphate."

# CHAPTER XXI

# MINERAL DEPOSITS RESULTING FROM PROCESSES OF ROCK DECAY AND WEATHERING[1]

## GENERAL CONDITIONS

The uplifted sedimentary beds, the lavas of the volcanoes, the granular crystalline rocks uncovered by erosion—all these, when exposed at the surface of the earth are subject to a series of changes, the sum total of which is called weathering. The agents are water, gas, heat, and vegetable and animal life. Water is essential—without it very little decomposition could take place. Oxygen is also essential, and indeed we often speak of weathering as synonymous with oxidation. Carbon dioxide dissolved in water decomposes the minerals and hastens the process of solution. Change of temperature acts mainly by promoting disintegration, most powerfully by the expansion of water when freezing in cracks and crevices, a force sufficient to break and dislodge heavy rock masses. Vegetable life furnishes carbon dioxide and disintegrates the soil by the vital energy in the roots, and bacterial life changes its composition. Animals burrow in the ground, loosening it and effecting chemical changes.

Weathering differs from alteration and metamorphism in that its ultimate result is the destruction of the rock as a unit; its chemical processes are far more radical and intense than those of the depths. Weathering effaces the texture of the rocks and segregates their chemical compounds in ways wholly different from those of the other processes mentioned. Metals closely associated in the primary rocks part company and seek new associates. Segregations of large masses of single minerals are usually a result of the process. The ordinary silicates and the carbonates of iron, magnesium, and calcium are unstable in the belt of weathering. The uppermost thin mantle of the products of weathering

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[1] G. P. Merrill, Rocks, rock-weathering, and soils, 1906.

F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 479–505.

F. K. Cameron and J. M. Bell, The mineral constituents of the soil solution, *Bull.* 30, U. S. Bur. Soils, Washington, D. C., 1905.

T. L. Lyon, and H. O. Buckman, The nature and properties of soils, New York, 1922, 588 pp.

A. Stebutt, Lehrbuch der allgemeinen Bodenkunde, Berlin, 1930, 518 pp.

H. Harrassowitz, Silicium, Aluminum, Eisen im Wechsel der Verwitterungsvorgänge, *Zeitschr. angew. Chemie*, 43, 1930, pp. 185–190.

we call the soil; in it the disintegration and chemical changes are carried to their limit, it is mixed with the products of life, and its constituents and reactions are, of course, of more interest to the agricultural chemist than to the student of ore deposits.

The depths to which weathering extends differ greatly; in some desert regions, recently glaciated areas, or areas covered by fresh lava flows it is practically absent, disintegration being the only visible effect. In regions of heavy vegetation and rainfall the weathering may extend to a depth of 100 or even 200 feet; along fractures in particularly permeable and soluble rocks like limestone oxidation may be carried to still greater depth; in mineral deposits its effects are in places felt for several hundred or in extreme cases as much as 2,000 feet. As a rule, however, weathering does not extend deeper than 50 feet, and its more intense effects are usually limited to the zone above the surface of underground water.

Disintegration and decomposition work together, but the former is likely to extend deeper than the latter. The upper layers, ordinarily colored red or brown by ferric iron, gradually change into paler-colored, more or less softened and disintegrated rock. In some areas, notably over limestone strata, there is a sharp change to the unaltered rock—so sharp, indeed, that the red clayey soil has often been taken for a different formation resting on the calcareous rock.

Erosional transportation attends disintegration, and removal of material by solution accompanies decomposition, both tending strongly to reduce the volume of the rock. On the other hand, hydration and the peculiar quality of absorption which the soils possess tend to increase the volume. On the whole weathering lessens the volume. According to G. P. Merrill, the granites of the District of Columbia may lose by weathering 13.5 per cent of their volume; T. L. Watson calculates the loss of granites of Georgia at 7 to 72 per cent. The most marked loss is the shrinking in residual clays derived from limestone; often it is more than 95 per cent. Whitney long ago arrived at the conclusion that 1 meter of residual clay in Wisconsin was derived from a thickness of 35 to 40 meters of limestone or shale.

Except in the easily soluble rocks the decomposition is never complete; for, as brought out by Cameron and Bell, even in the fine soils abundant grains of the original minerals remain unaltered. Other conditions being equal, weathering is most complete in tropical and moist countries. In the United States the most intense action of this kind has taken place in the Appalachian region south of the glaciated area, and this region contains the majority of ore deposits caused by weathering.

In the air contained in rain water both oxygen and carbon dioxide are greatly concentrated. In the soils carbon dioxide and air are absorbed; soils and clays of various kinds contain from 14 to 40 cubic centimeters of gas per 100 grams, with 14 to 34 per cent of carbon dioxide and considerably less oxygen than the air—indeed, in some soils oxygen appears to be absent.[1] Decaying vegetation still further increases the percentage of carbon dioxide. As the ground-water level is approached the oxygen decreases rapidly, as shown by the measurements made by B. Lepsius[2] in bore-holes, and below this level there is probably little left.

Naturally the processes of weathering are hastened by the presence of sulphuric acid derived from the decomposition of pyrite or exhaled from solfataric vents. W. Maxwell[3] has shown interestingly how extensive a part this acid plays in the development of soils on the slopes of volcanoes.

The processes characteristic of weathering are oxidation, hydration, and solution. In the surface waters calcium and magnesium carbonates ordinarily prevail, with a considerable amount of alkaline carbonates and relatively much soluble silica, both derived from the decomposition of the silicates. Under special conditions, as in volcanic regions or in sediments rich in salts, the surface waters may be materially different in composition, being predominantly sulphate solutions. The ground waters contain in addition small amounts of iron and manganese, carried mainly as bicarbonates, also phosphoric acid and sodium chloride.

In the weathered zone will remain the residual, almost insoluble minerals, like quartz, hydrated aluminum silicates more or less closely approaching kaolinite in composition, ferric oxides (as limonite, göthite, or hematite), and manganese dioxide, all mingled to form a red or brown clayey soil.

All these reactions involve the development of colloidal bodies like aluminum silicates and hydroxides of iron, which before their transformation into crystalline minerals are characteristic absorbents of many salts. The colloids of manganese, for instance, have a tendency to adsorb potassium and barium. The zone of weathering has indeed been called the realm of the colloids.

## DECOMPOSITION OF MINERALS

The silicates of the rocks are decomposed by water rather than dissolved, for the resulting solution does not usually contain the elements in the same proportions as the original mineral. Owing to hydrolysis the solution in most cases gives an alkaline reaction.[4]

According to the older viewpoint, the rock-forming silicates upon weathering yield the soluble carbonates of alkalies, calcium, and mag-

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[1] Cameron and Bell, *op. cit.*, p. 26.

[2] Quoted in F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, p. 477.

[3] W. Maxwell, Lavas and soils of the Hawaiian Islands, Honolulu, 1898.

N. E. A. Hinds, The weathering of Hawaiian lavas, *Am. Jour. Sci.*, 5th ser., 17, 1929, pp. 297–320.

[4] F. W. Clarke, *Bull.* 167, U. S. Geol. Survey, 1900, p. 156.

nesium, which together with much of the silica are carried away. Aluminum silicate remains with the larger part of the iron and manganese as hydroxide, but the two last metals may be in part carried off as soluble salts. The relative insolubility of alumina is an important factor in weathering.

The ferromagnesian minerals are attacked first, so that the ordinary surface waters contain more of the carbonates of calcium and magnesium than of other salts. The soda-lime feldspars come next while the alkali feldspars are more resistant. Quartz is only partially attacked. The decomposition of orthoclase is usually expressed in the following equation:

$$2KAlSi_3O_8 + 2H_2O + CO_2 = H_4Al_2Si_2O_9 + 4SiO_2 + K_2CO_3.$$

The ultimate product is kaolin, or allied colloidal bodies. Muscovite or sericite does not result from weathering, although the colloidal aluminum silicate may adsorb potassium and form amorphous compounds related in composition to the white micas.

According to a more modern viewpoint, hydrolysis, and consequently the hydrogen-ion concentration, is the dominant factor in the weathering of silicates. The equation would thus be written:

$$KAlSi_3O_8 + H.OH = KOH + HAlSi_3O_8$$

KOH combining with $CO_2$ and kaolin forming from the unstable silicate $HAlSi_3O_8$. The effect of $CO_2$ would in large part be due to its combination with the hydrolyzed bases rather than to direct solvent action. It has been shown by Schwartz and Walcker[1] that the mineral kaolinite forms only at a certain hydrogen-ion concentration. Earlier workers like Stremme had concluded that the colloidal precipitates from mixed solutions of silicic acid and aluminum salts were mixtures rather than definite compounds. Decrease in the acidity of water saturated with $CO_2$ under atmospheric pressure ($P_h$3.8 at 20° and $P_h$4.2 at 70°) would produce the optimum condition for the formation of kaolin. Both hydrolysis and attack by the hydrogen ions of bicarbonate acid ($H_2CO_3 = H^+ + HCO^-{}_3$) lead to the same result. Salt solutions such as exist in the zone of weathering may accelerate the decomposition as may also adsorption phenomena of various kinds; exchanges of bases (p. 55) of various kinds also take place.

The soil contains a certain amount of colloidal material derived from both organic and mineral sources.[2] It is assumed that the colloidal material is derived from the breaking down of the finest of the soil minerals by action of water and carbon dioxide. The average composition of this extracted soil colloid is about as shown on page 348.

[1] *Zeitschr. anorg. Chemie,* 145, 1925, pp. 304–310.

[2] Milton Whitney, The colloid chemistry of the soil in R. H. Bogue, The theory and application of colloidal behavior, New York, 1924, 2, pp. 466–487.

M. S. Anderson *et al., Bull.* 1122, U. S. Dept. Agr., 1922.

P. L. Gile *et al., Bull.* 1193, *idem,* 1923.

The composition of this soil colloid is very unlike the chemical composition of the soil. $SiO_2$ is much lower while $Al_2O_3$ and $Fe_2O_3$ are higher. $TiO_2$ and $P_2O_5$ are apparently concentrated. It does not correspond in composition to any mineral. Ordinary soils contain from 6 to 25 per cent of colloidal material, while some of the most plastic clays yield 60 per cent or more. The samples examined were collected from a maximum depth of 36 inches below the surface.

COMPOSITION OF AVERAGE SOIL COLLOID

| | | | |
|---|---|---|---|
| $SiO_2$ | 43.13 | $Na_2O$ | 0.37 |
| $Al_2O_3$ | 27.07 | $P_2O_5$ | 0.28 |
| $Fe_2O_3$ | 10.40 | $SO_3$ | 0.11 |
| $TiO_2$ | 0.64 | Cl | 0.03 |
| MnO | 0.11 | $H_2O$ | 10.40 |
| CaO | 1.02 | Organic matter | 3.40 |
| MgO | 1.63 | N | 0.26 |
| $K_2O$ | 1.42 | | |

Zeolites are undoubtedly unstable in the zone of weathering. Muscovite or sericite is slowly attacked. Cameron and Bell[1] treated 2 grams of powdered muscovite with a liter of pure water for 14 months in paraffine cylinders and obtained in the solution 10.4 parts K per million; when treating it with carbon dioxide in water they obtained 18.3 parts per million. The same quantity of orthoclase treated with pure water yielded a solution containing 1.7 parts per million of K; with water saturated with $CO_2$, 2.5 parts per million. Muscovite thus yields its potassium more easily than orthoclase. Albite treated in the same way gave 1.0 and 1.1 parts of sodium per million respectively. Earlier experiments leading to similar results have been undertaken by A. Daubrée, R. Müller, A. S. Cushman, and F. Henrich.[2]

Magnetite is soluble with difficulty, but finally yields hematite and limonite. Pyrite easily yields limonite and sulphuric acid. Apatite appears to be rather easily soluble, especially in carbonated water. Cameron and Bell[3] treated powdered chlorine apatite suspended in water at 25° C. for 7 days, passing $CO_2$ through the liquid. The solution showed 0.0378 gram CaO, 0.0096 gram $P_2O_5$, and 0.0026 gram hydrochloric acid per liter of solution. In soils and clays the phosphates are decomposed or hydrolyzed, soluble phosphates being formed; but the percolating

[1] *Op. cit.*, p. 33.

[2] A. Daubrée, Études synthétiques de géologie expérimentale, 1879.
R. Müller, *Jahrbuch*, k. k. geol. Reichsanstalt, 27, 1887.
A. S. Cushman, *Bull.* 92, Bur. Chemistry, U. S. Dept. Agr., 1905.
F. Henrich, Ueber die Einwirkung von kohlensäurehaltigem Wasser auf Gesteine, etc., *Zeitschr. prakt. Geol.*, 18, 1910, pp. 85–94.

[3] *Bull.* 41, Bur. Soils, U. S. Dept. Agr., 1907.

water contains these only in minimal quantities. It is stated[1] that humus suspended in water can adsorb calcium and a considerable amount of phosphoric acid from the calcium phosphates.

Cameron and Bell ascertained that carbon dioxide greatly aided the solution of ferric phosphate; 5 grams of which, shaken for 5 days with 100 c.c. $H_2O$ and later with 100 c.c. of saturated solutions of $CO_2$ and $CaSO_4$, yielded respectively 74, 171, and 118 milligrams of phosphoric acid. Organic matter in soils is said to be a solvent for ferric phosphate.

Zircon, pyrope garnet, tourmaline, chromite, and similar minerals are almost insoluble.

Quartz also shows great resistance and appears practically insoluble in the zone of weathering, except when exposed to the action of a stronger solution of alkaline carbonates. C. W. Hayes,[2] M. L. Fuller,[3] and C. H. Smyth[4] have observed a marked corrosion of quartz pebbles in conglomerates, but the exact nature of the reaction is uncertain. Cherty and fine-grained quartz is a little more soluble.[5] The theory of the origin of the Lake Superior iron ores, supported by Van Hise and Leith, is based on the supposed solubility of such material. It was formerly thought that certain organic acids had the power of dissolving quartz, but this is now considered very questionable.

## EXAMPLES OF THE CHEMICAL CHANGES BY WEATHERING

The following analyses[6] are quoted to show the general character of decomposition by weathering. More data are given in Clarke's Geochemistry.[7]

Merrill believes that a crystalline siliceous rock when passing into a residual soil loses about 50 per cent of the original constituents through the leaching action of meteoric waters.

The analyses generally show an apparent increase of alumina, a considerable increase of water, and losses in silica, lime, magnesia, soda, and potash. The loss in potassium is usually less than that of sodium. For purposes of comparison the alumina is assumed to be constant, which no doubt is approximately true. The action of weathering is not always

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[1] Patten and Waggaman, Absorption by soils, *Bull.* 52, Bur. Soils, U. S. Dept. Agr., 1908.

C. W. Correns, Die Bedeutung der Adsorption für die Bildung syngenetischer Erzlagerstätten, *Zeitschr. prakt. Geol.*, 1924, p. 145.

[2] *Bull.* Geol. Soc. Am., 8, 1897, p. 213.

[3] *Jour. Geol.*, 10, 1902, p. 815.

[4] *Am. Jour. Sci.*, 4th ser., 19, 1905, p. 277.

[5] G. Lunge and C. Millberg, *Zeitschr. angew. Chemie*, 1897, p. 393.

[6] G. P. Merrill, *Bull.* Geol. Soc. Am., 8, 1897, p. 160.

[7] *Bull.* 770, U. S. Geol. Survey, 1924, pp. 479–496; also C. K. Leith and W. J. Mead, Metamorphic geology, New York, 1915, pp. 3–24.

constant but in general the analyses illustrate the trend of the process.[1]

| | Micaceous Gneiss | | | Diorite | | |
|---|---|---|---|---|---|---|
| | *a* | *b* | *c* | *d* | *e* | *f* |
| $SiO_2$ | 60.69 | 45.31 | 52.45 | 46.75 | 42.44 | 37.31 |
| $Al_2O_3$ | 16.89 | 26.55 | 0.00 | 17.61 | 25.51 | 0.00 |
| $Fe_2O_3$<br>$FeO$ | 9.06 | 12.18 | 14.35 | 16.79 | 19.20 | 21.03 |
| MgO | 1.06 | 0.40 | 74.70 | 5.12 | 0.21 | 97.17 |
| CaO | 4.44 | tr | 100.00 | 9.46 | 0.37 | 97.30 |
| $Na_2O$ | 2.82 | 0.22 | 95.03 | 2.56 | 0.56 | 84.87 |
| $K_2O$ | 4.25 | 1.10 | 83.52 | 0.55 | 0.49 | 38.75 |
| Ignition | 0.62 | 13.75 | gain | 0.92 | 10.92 | gain |
| $P_2O_5$ | 0.25 | 0.47 | gain | 0.25 | 0.29 | 19.87 |
| | 100.08 | 99.98 | ....... | 100.01 | 99.99 | ....... |

*a.* Mica gneiss, Albemarle County, Virginia.

*b.* Residual soil of same rock, G. P. Merrill, *Bull.* Geol. Soc. Am., 8, 1897, p. 189.

*c.* Percentage of each constituent lost on assumption of constant alumina.

*d.* Diorite, Albemarle County, Virginia. G. P. Merrill, Rocks, Rock weathering and soils, New York, 1906, p. 224.

*e.* Same, decomposed.

*f.* Percentage of each constituent lost on assumption of constant alumina.

## CLAY[2]

**Definition.**—In a general way, clay is defined as an earthy substance containing a mixture of hydrous aluminum silicates, with residual fragments of minerals and colloidal material. The clays usually possess

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[1] For a study of the different methods used for the calculation of gains and losses, see G. Grosser, *Chemie der Erde*, 7, 1932, pp. 130–176.

[2] The literature of clays is exceedingly voluminous. For information more detailed than can be given here, consult:

H. Ries, Clays, occurrence, properties, and uses, 1908.

H. Ries, A review of the theories of origin of white residual kaolins, *Trans.*, Am. Ceramic Soc., 13, 1911.

I. E. Sproat, Refining and utilization of Georgia kaolins, *Bull.* 128, U. S. Bur. Mines, 1916.

H. O. Buckman, The chemical and physical processes involved in the formation of residual clay, *Trans.*, Am. Ceramic Soc., 13, 1911.

J. H. Watkins, White-burning clays of the southern Appalachian states, *Trans.*, Am. Inst. Min. Eng., 51, 1916, pp. 481–501.

A full bibliography of the older literature by H. Rösler is contained in *Neues Jahrbuch.*, Beil. Bd. 15, 1902, p. 231.

C. Doelter, Handbuch der Mineralchemie, 2, pt. 2, 1917, pp. 30–145, with article by Stremme.

plasticity when wet and become hard when fired to a temperature of redness (Ries).

**Transported Clays.**—The transported clays have been considered on page 215. We may distinguish marine clays, flood plain clays, lake clays, glacial clays, and eolian clays.

**Residual Clays.**—The residual clays are found in all parts of the world where rock decay has proceeded for a long time and where the products have not been swept away by strong erosion or by glacial action.

In the United States such clays occur chiefly in the southern Appalachian region. Acidic granular rocks, like granite or gneiss—particularly those rich in feldspar and poor in ferromagnesian silicates—yield the purest clays, but even these must often be purified by washing to remove quartz and limonite. At varying depths, 100 feet at most, these clays change into unaltered rocks. Pegmatite dikes sometimes yield a high-grade residual clay. Sedimentary rocks, like limestone, by long-continued solution yield residual clay; but such material is impure and rarely used for anything but brick making. Some residual clays derived from granitic rocks lack the property of plasticity. It is believed that the conditions favor the development of kaolin underneath areas covered by bogs and peat. During the oxidation of ore deposits kaolinite or allied clay minerals frequently develop, largely owing to the action on rock minerals of free sulphuric acid in the descending solutions.

**Clay of Hydrothermal Origin.**—Ascending waters, warm or tepid, may produce a clay by action on the rock minerals, under certain circumstances. It is believed, however, that below the surface sericite or white mica will be formed in preference to kaolin.

**Use and Properties.**—The ordinary varieties of residual clays are used for brick making, the purer kind for fire bricks and the finer grades for pottery. These pure white-burning clays are usually called kaolin, ball clay, paper clay, or china clay, and are also used in manufacturing paper and as fillers in paints, putty, and crayons. About 4,000,000 short tons of clays of all kinds were quarried in 1930 in the United States, the price ranging from $3 to $10 per ton; and 236,000 tons of china clay were imported at about $10 per ton, chiefly from England. Pennsylvania, Missouri, California, and Georgia have the largest clay production.

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H. Ries, W. S. Bayley *et al*, High-grade clays of eastern United States, *Bull.* 708, U. S. Geol. Survey, 1922.

R. E. Somers, *idem*, pp. 292–304.

C. S. Fox, The bauxite and aluminous occurrence of India, *Mem.*, Geol. Survey India, 49, pt. 1, 1923.

H. Harrassowitz, Laterit, Bd. 4, Hft. 14, *Fortschr. der Geol. u. Pal.*, Berlin, 1926.

M. Storz, Die sekundäre authigene Kieselsäure in ihrer petrogenetisch-geologischen Bedeutung, Monographien zur Geologie und Palaeontologie, Berlin, Teil 1, 1928, 137 pp.

Evidently, the clay industry is growing very rapidly. Most of the local china clays are residual after pegmatite and granite and are obtained in Georgia and the Carolinas. The composition of such clays before and after washing is indicated by the following analyses:

ANALYSES OF CRUDE AND WASHED KAOLIN, WEBSTER COUNTY, SOUTH CAROLINA[1]

| | Crude | Washed | Kaolinite |
|---|---|---|---|
| $SiO_2$ | 62.40 | 45.78 | 46.5 |
| $Al_2O_3$ | 26.51 | 36.46 | 39.5 |
| $Fe_2O_3$ | 1.14 | 0.28 | .............. |
| FeO | .............. | 1.08 | .............. |
| CaO | 0.57 | 0.50 | .............. |
| MgO | 0.01 | 0.04 | .............. |
| Alkalies | 0.98 | 0.25 | .............. |
| $H_2O$ | 8.80 | 13.40 | 14 |
| Moisture | 0.25 | 2.05 | .............. |
| Total | 100.66 | 99.84 | 100 |
| Clay substance | 66.14 | 93.24 | .............. |

[1] H. Ries, Economic geology, 6th ed., 1930, p. 178.

To eliminate quartz, washing in log washers or other contrivances is used. Pneumatic concentration by air currents has also been introduced. Iron in high-grade clay may be removed by treatment with acids.

Paper clays are mined in the Cretaceous beds of South Carolina and Georgia; plastic kaolins, or ball clays, are obtained from Tertiary beds in Florida and western Tennessee.

An important property of clay is plasticity, by means of which it can be molded into a desired shape which it retains when dry. Not all the residual clays are plastic, nor is the pure mineral kaolinite. The plasticity of clay depends upon the presence of colloids.[1]

The transverse strength of air-dried clays varies from 15 to over 400 pounds or more per square inch, according to Ries. The shrinkage in drying and firing is an important factor. The total shrinkage should be about 8 or 9 per cent. The fusibility varies according to the impurities present. In low grades of clay incipient fusion may occur at about 1000° C., while in refractory clays, which are low in fluxing impurities, it may not occur until 1300° or 1400° C. is reached. The melting-point of kaolin is about 1800° C.

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[1] H. E. Ashley, The colloid matter of clay and its measurement, *Bull.* 388, U. S. Geol. Survey, 1909, p. 65.

**The Clay Minerals.**[1]—The term "kaolin," often loosely used, is properly applied only to a white, residual clay that is the product of weathering or rock decomposition. The identification of the clay minerals is difficult owing to the fine distribution, the low double refraction, and the presence of amorphous, colloidal compounds. The following species may be distinguished:

1. Kaolinite (Anauxite), ($Al_2O_3.2SiO_2.2H_2O$); monoclinic, present in most clays and kaolins (sedimentary or residual), not decomposed by HCl, partly decomposed by $H_2SO_4$. The minerals nacrite and dickite are very closely related to kaolinite but occur mostly in hydrothermal ore deposits and often show definite crystals.
2. Halloysite, ($Al_2O_3.2SiO_2.nH_2O$); colloidal form of kaolinite, decomposed by $H_2SO_4$.
3. Beidellite, ($Al_2O_3.3SiO_2.nH_2O$); orthorhombic, foliated, insoluble in HCl.
4. Montmorillonite (($(MgCa)O.Al_2O_3.5SiO_2.nH_2O$); orthorhombic, foliated, insoluble in HCl.
5. Leverrierite, ($Al_2O_3.2SiO_2.nH_2O$); monoclinic, foliated, insoluble in HCl. Variety of kaolinite?
6. Allophanite, ($Al_2O_3.SiO_2.nH_2O$); colloid, decomposed by HCl.
7. Pyrophyllite, ($Al_2O_3.4SiO_2.nH_2O$); crystalline, not usually found in clays.

It seems likely that there is another colloidal form of kaolin with a formula similar to kaolinite but not decomposed by HCl.

The aluminum hydroxides, and their colloidal form, bauxite, are present in lateritic soils formed under tropical conditions.

Kaolinite is commonly present in clays though not always easy to identify. Most clays also doubtless contain a colloidal aluminum silicate as shown by the composition of the soil solution.

Beidellite and allied minerals as well as allophanite are not usually present in clays. They are more closely related to the white micas and appear as decomposition products of that or similar silicates. A colloidal form of beidellite probably also exists.

**Origin.**—There is a voluminous literature on the clay minerals, but opinions differ considerably. There is some uncertainty as to whether the products of weathering are definite silicates or mixtures of mutually precipitating colloidal silica and alumina. Both cases probably occur. Under certain conditions crystalline kaolinite will form. Under other conditions a mixture of colloids will result. The experiments noted above

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[1] E. S. Larsen and E. T. Wherry, Beidellite, a new mineral name, *Jour.*, Acad. Sci. Washington, 1925, pp. 465–468.

C. S. Ross and E. V. Shannon, *Jour.*, Am. Ceramic Soc., 9, 1926, pp. 77–96.

C. S. Ross and P. F. Kerr, The kaolin minerals, *Prof. Paper* 165-*E*, U. S. Geol. Survey, 1931.

show that colloidal substances are present in the soil (p. 348). About 1908, an animated controversy took place between Rösler,[1] who maintained that kaolin was the result of hydrothermal decomposition, and Stremme and Barnitzke,[2] who showed that in the case of Meissen, Saxony, where a high grade of chinaware is made, the deposit was decidedly a product of weathering, gradually changing in depth to fresh rocks. The fact seems to be that the clay minerals are usually products of weathering and are formed abundantly in the zone of oxidation, in both rocks and ore deposits, by the ordinary agencies of water, carbon dioxides, and sulphuric acid. Kaolinite and allied minerals may also form near the surface and in smaller amounts below this zone by the action of dilute thermal waters, but the action of thermal waters containing alkalies on rocks usually results in sericite. This has been abundantly proved in ore deposits of all kinds, but observers have often confused the kaolin formed by supergene waters with the sericite formed by the ascending alkaline waters.[3] Kaolinite, it is believed, is unstable under deep-seated processes of mineralization. G. Hickling[4] has shown that the china clays of Cornwall form a sheet covering the eroded surface of the granite and have resulted from the weathering of sericite of hydrothermal origin.

## RESIDUAL IRON ORES (LIMONITE AND HEMATITE)

**Origin.**—During the processes of weathering only a small part of the iron is carried away in solution; the greater part remains in the rock altered to limonite ($2Fe_2O_3.3H_2O$), to göthite ($Fe_2O_3.H_2O$), or to indefinite colloidal mixtures of various hydroxides of iron. Hematite may also be present. In places basic sulphates or phosphates may remain, as well as somewhat indefinite and unstable ferric silicates. Nontronite, ($H_4Fe_2Si_2O_9$), the equivalent of kaolin, is said to be present in weathered rocks. In the zone of weathering the iron shows a strong tendency to move outward and segregate in irregular or mammillary masses separated by clayey material, though much of it, of course, remains intimately mixed with clay. The same is true of manganese, some of which may be associated with the limonite, though when much manganese is present, it also tends to separate by itself.

The "centrifugal" tendency of iron hydroxide is well seen in many oxidized mineral deposits, often also in the weathering of pebbles. A fine instance was observed in the cobbles of andesite in the Tertiary river bed at Iowa Hill, California. The outside of these cobbles is hard and consists of an impure limonite; the center contains soft yellowish kaolin.

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[1] H. Rösler, *Zeitschr. prakt. Geol.*, 16, 1908, p. 251; Stremme, *idem*, p. 122.

[2] *Zeitschr. prakt. Geol.*, 17, 1909, pp. 457–472.

[3] W. Lindgren, The origin of kaolin, *Econ. Geol.*, 10, 1915, pp. 89–93.

[4] *Trans.*, Inst. Min. Eng. (London), 36, 1908–1909, p. 10.

During the concentration, the ferric hydroxides (see p. 310) were probably transported as colloids, which hardened and became crystalline, as shown by the radial structure of many concretions. The chemical character of these ores has rarely been studied in detail; probably it will be found that barite, oxidized zinc minerals, and compounds containing manganese, nickel, and cobalt are present. Many of the limonites are rather pure and are of considerable economic importance.

**Classification.**—One class of residual brown iron ores is derived from the decomposition of deposits of siderite or pyritic ores, both usually formed by ascending waters, or from the weathering of black bands or glauconite beds (Fig. 131). Such limonites in places reach considerable depths, dependent on the penetrating power of oxygenated waters. The decomposed outcrops of pyritic ores are not often used as iron ores.

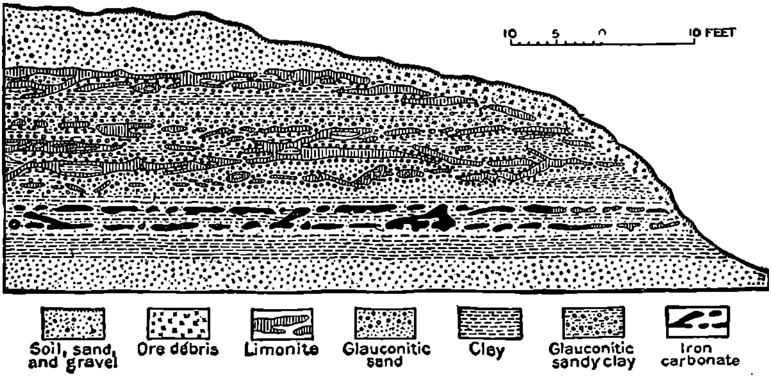

FIG. 131.—Section showing oxidation of iron carbonate to limonite in Tertiary beds, Cass County, Texas. (*After E. F. Burchard, U. S. Geol. Survey.*)

Another class consists of local segregations of limonite or allied hydroxides in the decayed rock and residual clay near the surface. These masses are particularly common in limestone area. Little or no siderite is found near the surface, but may appear in the limestone at greater depth. When oxygen is exhausted the iron is more easily transported as a bicarbonate and the metasomatic replacement of calcite by siderite may then occur. There are, however, few deposits of limonite which change in depth to large irregular replacements of siderite; so that it may be assumed that the rate of solution and downward transportation of the precipitated limonite is slow.

Finally, a third class of residual iron ores, consisting of limonites mixed with hematite, occurs as widespread sheets formed by the gradual decay of strongly ferriferous rocks.

**Brown Hematites of the Appalachian Region.**[1]—In the United States the residual iron ores are most abundant in the Appalachian region,

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[1] C. W. Hayes and E. C. Eckel, Iron ores of the Cartersville district, Georgia, *Bull.* 213, U. S. Geol. Survey, 1902, pp. 233–242.

mainly in Alabama, Georgia, Virginia, and Tennessee. The annual production of such ores is gradually diminishing. In 1930, 720,500 long tons were produced, about 1.2 per cent of the yearly output of iron ores in the United States. These so-called "brown hematites" are mined in many small deposits; their content of iron ranges from 38 to 52 per cent (limonite 59.89 per cent Fe); most of them are comparatively rich in phosphorus.

Most of the southern limonites lie in Cambro-Silurian strata and extend along the "Great Valley" between the pre-Cambrian on the east and the Paleozoic rocks on the west. They are classed as valley ores, mountain ores, and Oriskany ores.

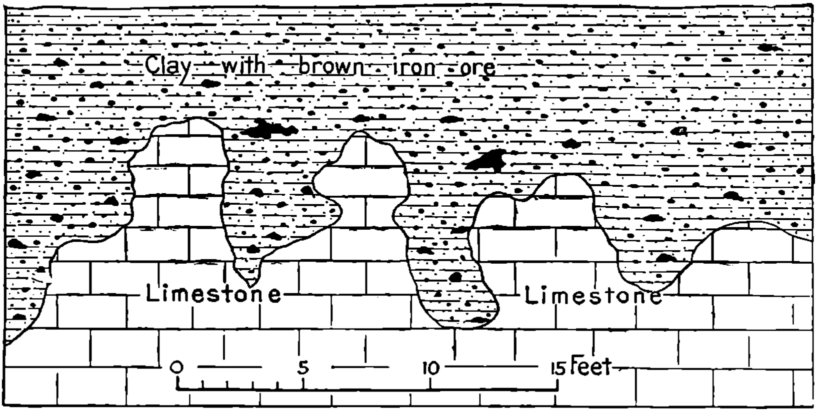

FIG. 132.—Vertical section showing structure of the valley brown ore deposits of the Rich Hill mine, Virginia. (*After E. C. Harder, U. S. Geol. Survey.*)

The *valley ores* appear as irregular deposits of shallow pockets in clay derived from the decomposition and solution of Cambro-Silurian limestone or dolomite. The ores lie as scattered lumps in the clay, not so much on the eroded surface of the limestone as higher up (Fig. 132). Each deposit is soon exhausted, and few extend below a depth of 50 feet. The ores are mixtures of limonite, göthite, and clay; the composition ranges from 40 to 56 per cent Fe, 5 to 20 per cent $SiO_2$, 0.05 to 0.5 per cent P, and 0.3 to 2.0 per cent Mn.

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E. C. Eckel, Limonite deposits of eastern New York, etc., *Bull.* 260, U. S. Geol. Survey, 1904, pp. 335–342.

R. J. Holden, in Mineral resources of Virginia, 1908.

E. C. Harder, The iron ores of the Appalachian region in Virginia, *Bull.* 380, U. S. Geol. Survey, 1908, pp. 215–254.

E. C. Harder and E. F. Burchard, *Mineral Resources*, U. S. Geol. Survey, particularly pt. 2, chapter on Iron, 1908.

E. F. Burchard, C. Butts, and E. C. Eckel, Birmingham district, *Bull.* 400, U. S. Geol. Survey, 1910, pp. 145–167.

Many of these ores were evidently concentrated under conditions different from those of to-day; most of them are probably of Tertiary age as shown particularly in the deposits south of Birmingham, Alabama. It is not unlikely that the same applies to many "mountain ores."

The *mountain ores,* according to Harder, show greater variation in occurrence and appearance. They are found as small, discontinuous pockets in residual material above the Lower Cambrian quartzite at or near the contact with the overlying formation, generally a limestone.

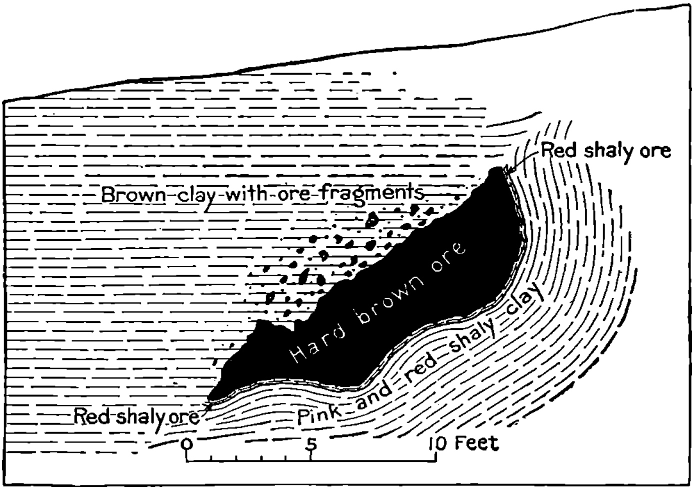

Fig. 133.—Vertical section showing the structure of mountain brown ore occurring as mammillary masses in clay. Mary Creek mine, Virginia. (*After E. C. Harder, U. S. Geol. Survey.*)

While these ores are mainly superficial, they are sometimes worked to a depth of several hundred feet. The composition ranges from 35 to 50 per cent Fe, 10 to 30 per cent $SiO_2$, 0.1 to 2 per cent P, and 0.5 to 10 per cent Mn. These limonites are often glassy and concretionary.

The occurrences are classed by Harder as follows:

1. Pocket deposits in clay, in part replacements of limestone; sometimes manganiferous (Fig. 133).
2. Small replacement deposits in shale along fractures.
3. Deposits in quartzite or sandstone, not abundant, including
   *a.* Breccia deposits accompanied by replacement.
   *b.* Vein deposits along faults.

Many limonite deposits have been worked in New York, Connecticut, and Massachusetts. Late work by R. Ruedeman and D. H. Newland (1931–1936) seems to indicate that they have been derived from Cambro-Ordovician rocks containing siderite.

The Oriskany ores[1] are mined in Virginia and form irregular replacements along local folds or fracture zones on the flanks of greater anticlines. They occur in the calcareous Oriskany sandstone which is overlain by the Romney shale (*Devonian*) and underlain by the Helderberg limestone (*Devonian*) (Fig. 134). The ore largely replaces the sandstone, sometimes also the limestone, and may be from 10 to 100 feet wide. The greatest depth reached is 600 feet; at this or lesser depth the ore grades into unaltered rock. The iron is considered by some authors to be derived from the Romney shale but is more likely derived from the sandstone itself. The ore is made up of earthy masses and rounded concretions of fibrous limonite filled with clay or sand. The ore from the Callie mine and the footwall limestone is said to contain about 0.2 per cent of zinc. At the Callie mine the ore production was about 2,700 tons per month. Probably hematite or turgite is also present for the ore does not contain enough water for limonite; it averages 43 per cent Fe, 10 to 25 per cent $SiO_2$, 0.06 to 0.5 per cent P, and 0.5 to 4 per cent manganese. Cobalt and nickel are reported to be present in traces.

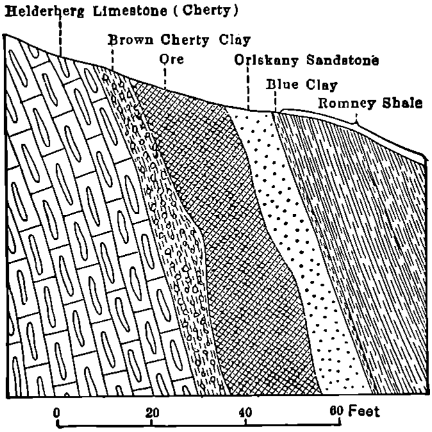

FIG. 134.—Vertical section showing the Oriskany brown ore deposit at the Callie mine, Virginia. (*After E. C. Harder, U. S. Geol. Survey.*)

The Oriskany ores, like the other "brown hematites," are subjected to a rough concentration in log washers in order to remove the clay.

**Iron Ores of Bilbao, Spain.**[2]—The great deposits of Bilbao, in northern Spain, have for many years yielded several million tons annually; the ores being exported to England and Germany.

The region shows folded and faulted Cretaceous beds; the most important from a mining standpoint is a bed of limestone, 250 feet thick. The scant igneous rocks are diabase and trachyte, of Tertiary age. The primary ore, thought by most observers to be a replacement of limestone by ascending hot waters, consists of coarse siderite and ankerite with minor amounts of pyrite and chalcopyrite and appears to continue in depth, mainly near the intersection of faults, though the quantity is

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[1] C. M. Weld, The Oriskany iron ores of Virginia, *Econ. Geol.*, 10, 1915, pp. 399–421.

[2] John, *Zeitschr. prakt. Geol.*, 19, 1911, pp. 208–212.
P. Grosch, *Geol. Rundschau*, 5, 1914–1915, pp. 392–618.
R. W. Van der Veen, *Econ. Geol.*, 17, 1922, pp. 602–618.

much smaller than the large, sheetlike superficial bodies of limonite and hematite. One of these latter was 2 miles long, 3,300 feet wide and had a thickness of 100 feet. Such oxidized ores contain from 50 to 57 per cent iron and are of Bessemer grade. They are in part residual, in part eluvial or alluvial. The primary mineralization is correlated by Van der Veen with that of the zinc and quicksilver of Santander and believed to be of Miocene age.

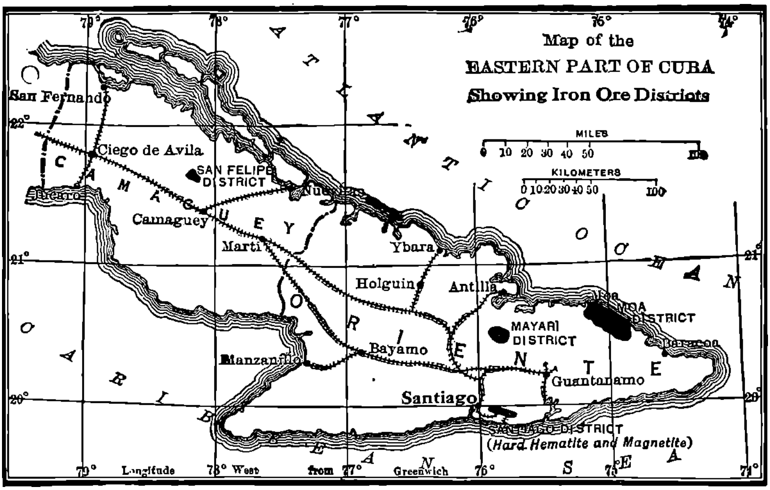

FIG. 135.—Sketch map of eastern part of Cuba. (*After W. L. Cummings, and B. L. Miller.*)

**Residual Ores of Cuba.**[1]—Iron ores have been mined for a number of years in the vicinity of Santiago, Cuba; but these ores, of contact-metamorphic origin, consist of hematite with some magnetite and contain a high percentage of sulphur. The three new districts described by Spencer and others are likewise in the eastern part of the island, but are of an entirely different type. They are the Mayari and Moa districts in Oriente Province, and the San Felipe in Camaguey (Fig. 135). The ores occur as residual mantles resulting from the weathering of serpentine and for the most part lie on plateaus at rather high elevations. They were probably formed during the Tertiary before the uplift of the present

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[1] A. C. Spencer, Three deposits of iron ore in Cuba, *Bull.* 340, U. S. Geol. Survey, 1907, pp. 318–329.

C. M. Weld, The residual iron ores of Cuba, *Trans.*, Am. Inst. Min. Eng., 40, 1909, pp. 299–312.

J. F. Kemp, The iron resources of the world, Int. Geol. Congress, Stockholm, 1910, pp. 793–795; *Trans.*, Am. Inst. Min. Eng., 51, 1916, pp. 3–30.

See also seven papers on the same subject by J. S. Cox, Jr., C. K. Leith, W. J. Mead, A. C. Spencer, C. W. Hayes, W. L. Cumings, B. L. Miller, D. E. Woodbridge, and J. E. Little, in *Trans.*, Am. Inst. Min. Eng., 42, 1911, pp. 73–152. Also C. K. Leith and W. J. Mead, *idem*, 53, 1916, pp. 75–78.

plateaus. Near the surface the material is earthy and dark red, sometimes cemented, with shot-like lumps of hematite scattered over the surface; underneath lie yellowish ores changing rather abruptly into decomposed and soft serpentine. In places a layer of cherty material is found immediately above the serpentine. In the Mayari district the average depth of the ore is about 15 feet, and it extends over an area of 10 by 4 miles. Hundreds of millions of tons are said to be available, allowing for parts of the area which are below the workable grade. The ore is removed by drag-line steam shovels.

According to analyses the ore is fairly uniform, the metallic iron varying in percentage from 40 to 50. It is remarkably free from phosphorus and evidently contains hematite, limonite, a little magnetite, and also some free aluminum hydroxide. It is, in brief, a typical iron-rich laterite (see p. 374). There is much water; according to Kemp the Moa ores yield 25 to 30 per cent hygroscopic and 10 to 12 per cent combined water; silica is low and alumina high. The presence of nickel and chromium is also notable; the latter metal is concentrated during the oxidation but is removed during the smelting; the former is favorable to the quality of the iron.

ANALYSES OF SERPENTINE AND ORE FROM THE MAYARI DISTRICT, CUBA

*(After C. K. Leith)*

| | 1 | 2 |
|---|---|---|
| $SiO_2$ | 39.80 | 2.90 |
| $Al_2O_3$ | 1.39 | 10.24 |
| $Fe_2O_3$ | 10.14 | 72.35 |
| Fe | 7.10 | 50.56 |
| MgO | 33.69 | ...... |
| Cr | 0.20 | 1.66 |
| Ni+Co | 0.97 | 0.84 |
| P | 0.001 | 0.016 |
| S | 0.06 | 0.20 |
| $H_2O+$ | 13.31 | 10.96 |
| | 99.561 | 99.166 |

1. Serpentine, at depth of 29 feet.
2. Iron ore, at depth of 6 feet.

The porosity of the ore is exceedingly great, amounting to 75 per cent of its volume, but lessens near the surface.

In considering the alteration of serpentine to ore in terms of weight it is found that the alumina has remained nearly constant. The changes in the composition of the serpentine during its alteration to ore are shown by Leith and Mead in Fig. 136, which is based on many analyses at uniform intervals. The diagram illustrates the rapid destruction of the serpentine by leaching of $SiO_2$ and MgO, and it also shows the marked relative increase of iron and alumina as oxides. Toward the surface hematite (with a little magnetite) develops from limonite, and bauxite from kaolin. In the middle part of the ore-body iron has increased in proportion to the alumina owing probably to re-deposition and oxidation of ferrous iron dissolved by the reducing action of the vegetation. Silica is lost throughout and magnesia is wholly removed.

In 100 pounds of typical serpentine there are 1.5 pounds of alumina and 10 pounds of ferrous oxide. When the magnesia and silica are removed in solution and the iron oxidized there remain approximately 11.75 pounds of limonite, 3.8 pounds of bauxite and kaolin, and, at the most, 2 pounds of minor constituents. This residual of 17.55 pounds contains 7.8 pounds, or 44.4 per cent of metallic iron, and is an iron ore.

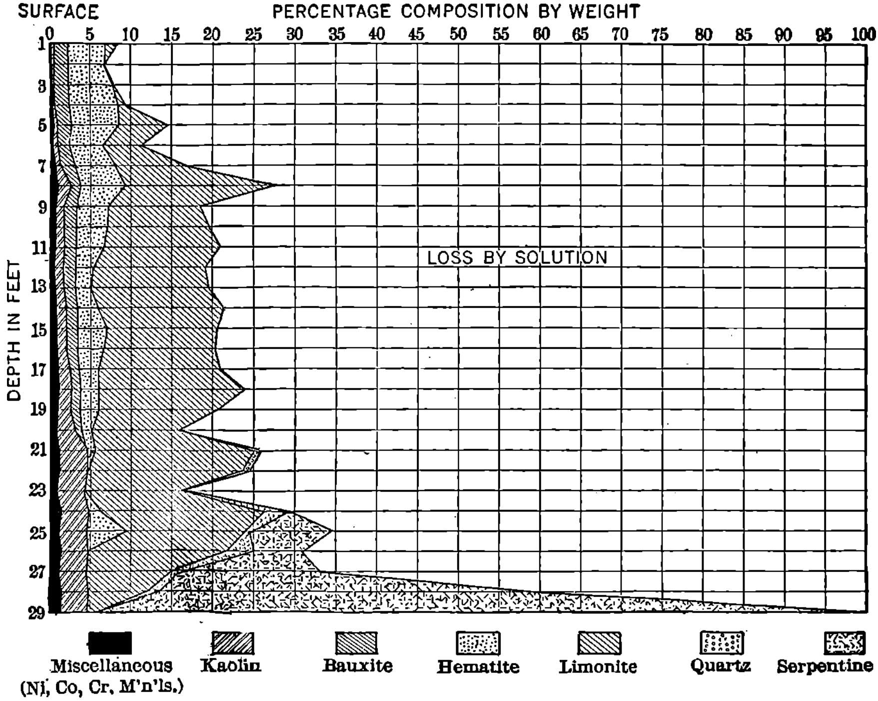

FIG. 136.—Diagram representing in terms of weight the mineralogical changes in the alteration of serpentine rock to iron ore, at Mayari, Cuba, on the assumption that alumina has remained constant. (*After C. K. Leith and W. J. Mead.*)

**Distribution and Stability of Residual Iron Ore.**—The residual iron ores are widely distributed in countries of warm climate, where secular decay has progressed without interruption for a long time. It seems, however, that great concentration has been effected only from relatively soluble rocks like limestone and serpentine. Many of the laterites of India, Africa, and other tropical countries are rich in ferric oxide and have the same characteristic concretionary pellets and shots on the surface. Limonite deposits similar to those of Cuba have been discovered in Macedonia, in Borneo and on Mindanao, in the Philippines.

Vegetation plays an important part in the origin of many of these deposits. Underneath the mat of roots and decayed vegetation the soil in tropical countries is often white or yellowish, indicating that the iron is in the ferrous state, probably as carbonate. When, as happened on the high volcanic plateau of Molokai, Hawaiian Islands,[1] the vegetation is destroyed the soil immediately turns red and hard and shows characteristic pellets of ferric oxide. In part at least the rock is thus changed directly to hematite without passing through the intermediate stage of limonite.

According to H. Wölbling,[2] the natural ferric hydroxides have great stability and cannot readily be changed to ferric oxide, at least not by exposure to air and salt solutions. The freshly precipitated hydroxides are, however, easily converted to ferric oxide; and these colloids may easily be crystallized. His experiments show that by the precipitation of ferric solutions with calcite or siderite at 100° C., $Fe_2O_3$ is easily formed, containing only 1 or 2 per cent $H_2O$, while during slow and wet oxidation of ferrosalts, ferric hydrates of iron are obtained.

It is certain, at any rate, that ferric oxide, as well as the hydrates, is very stable when once formed and is not easily altered.

No one can fail to be impressed by certain similarities of the Cuban residual ores to those of the Mesabi range (p. 304). Similar large expanses of rock, weathered under a tropical sun and covered by residual ferric oxide, undoubtedly yielded the material for the sedimentary hematite deposits.

## RESIDUAL MANGANESE ORES[3]

The minerals of the residual manganese ores consist of pyrolusite ($MnO_2$, 63.2 per cent Mn), psilomelane ($MnO_2.H_2O$, 49 to 62 per cent

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[1] W. Lindgren, The water resources of Molokai, *Water-supply Paper*, 77, U. S. Geol. Survey, 1903, p. 19.

[2] H. Wölbling, Bildung der oxydischen Eisenerzlager, *Stahl und Eisen*, 1909, p. 1248; also *Zeitschr. prakt. Geol.*, 17, 1909, p. 495.

[3] R. A. F. Penrose, Jr., Manganese, its uses, ores, and deposits, *Ann. Rept.* Arkansas Geol. Survey, 1, 1890.

E. C. Harder, Manganese deposits of the United States, *Bull.* 427, U. S. Geol. Survey, 1910, with bibliography.

E. C. Harder, Manganese ores of Russia, India, Brazil and Chile, *Trans.*, Am. Inst. Min. Met. Eng., 56, 1916, pp. 31–76.

E. C. Harder and D. F. Hewett, Recent studies of domestic manganese deposits, *idem*, 63, 1920, pp. 3–50.

Geo. A. Thiel, The manganese minerals, their identification and paragenesis, *Econ. Geol.*, 19, 1924, pp. 107–145.

J. Orcel and S. Pavlovitch, Charactères microscopiques de quelques . . . minerais . . . de manganese, *Bull.*, Soc. française de minéralogie, 54, 1931, pp. 80–99.

Mn), wad (perhaps $MnO_2.nMnO + H_2O$, varying percentage of metal), more rarely braunite ($3Mn_2O_3.MnSiO_3$ (?); 63.6 per cent Mn), manganite ($Mn_2O_3.H_2O$, 62.4 per cent Mn), and perhaps hausmannite ($Mn_3O_4$; 72.05 per cent Mn).

**Primary Sources.**—Nearly all workable manganese deposits are of secondary formation—that is, they are concentrated from manganese minerals more sparsely distributed in rocks. Manganite, pyrolusite, psilomelane, and wad are always, braunite usually, secondary, formed under the influence of weathering, even where they descend to considerable depths below the water level.

In igneous rocks manganese is always present but only in small amounts. The largest percentages (about 0.36 per cent) are found in syenite and its porphyries and in basalts. The proportion Fe:Mn is smaller in acidic than in basic rocks.

The first concentration in sedimentary rocks may contain manganese in the form of $MnO_2$ and carbonate. Nodules of $MnO_2$ occur in some deep-sea deposits.

Analyses of limestones often show a small amount of manganese. In many cherts and jaspers of the sedimentary series manganese is characteristically present as rhodonite or rhodochrosite. Important deposits of manganese may be produced by sedimentation (p. 280).

In crystalline schists, especially in those of more basic composition, manganese is found as spessartite (manganese garnet), rhodonite, piedmontite (manganese epidote), tephroite and alleghanyite.

Finally, rhodochrosite and rhodonite are rather common in ore deposits of hydrothermal or contact-metamorphic origin; and much manganese is present in some metamorphic specularite and magnetite deposits. (Cuyuna, p. 302.) Psilomelane has also been found in hot-spring deposits at the surface.

The inconspicuous, black or greenish manganese sulphide, alabandite,[1] is probably far more common than has been suspected; in hydrothermal deposits, it occurs at Bisbee and Tombstone, Arizona, and many other places in the United States, Mexico, Peru, and Hungary. According to Hewett, the normal hypogene paragenesis is rhodonite, alabandite, and rhodochrosite. Oxidation of any of these would produce manganite, psilomelane, etc.

**Origin.**—In most cases manganese has probably been transported as a bicarbonate, sometimes possibly as a sulphate. V. Vincent[2] found that in a liter of water saturated with $CO_2$, 62.5 milligrams of $MnCO_3$ is dissolved. In such water 4.6 milligrams $Mn_2O_3$ and only 1.5 milligrams

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[1] D. F. Hewett and O. N. Rove, Occurrence and relations of alabandite, *Econ. Geol.*, 25, 1930, pp. 36–56.

[2] *Comptes Rendus*, 162, 1916, pp. 259–269.

$MnO_2$ dissolved. The peroxide is thus most insoluble. Dunnington[1] showed that ferrous sulphate dissolves manganese from the carbonate, as sulphate, with the separation of ferric sulphate and limonite; from the compound solution calcium carbonate precipitates the iron but the manganese is precipitated only upon access of air.

Solutions of manganese carbonates are more stable than those of ferrous carbonate and a partial separation may thus be effected. G. A. Thiel[2] studied the precipitation of manganese salts by various rocks and found that magnesium carbonate was the most effective substance.

Of the minerals mentioned above, manganite is very often well crystallized in rhombic prisms; pyrolusite is generally believed to be a pseudomorphic form after manganite. Psilomelane is always fibrous and colloform,[3] ranging into the earthy forms of wad. Braunite and hausmannite are very often individualized in small octahedron-like crystals. Any attempt to derive a consistent order of crystallization is futile as may be seen from Thiel's paper referred to above.[4] The various minerals were evidently deposited after the oxidation of the bicarbonate solution, and almost certainly they were largely formed from colloidal solutions. Manganite in part formed from normal solutions, but we have also alternating colloform bands of manganite and psilomelane. We have hausmannite and braunite disseminated through psilomelane. The probability is that braunite is of colloidal origin; its peculiar composition and the absorbed barium indicate this almost certainly. We conclude that the residual manganese ores are largely of colloidal origin. The sol $Fe(OH)_3$ is positive while $Mn(OH)_4$ is negative.[5]

Mixing of the sols induces precipitation of manganese in complex proportions. H. Udluft[6] states that pure $Fe(OH)_3$ sol is unstable in ordinary waters containing bicarbonates, but that it may be protected by small quantities of manganese sol or humic colloids. Further investigations along these lines are desirable. In the literature the colloidal phenomena of residual manganese ores are almost wholly overlooked. These colloidal precipitates in part formed in open space, in part the gel replaced earlier rock structures; and this possibility is again generally neglected in the literature.

**Manganese Deposits in the United States.**—From the rocks above mentioned, manganese is concentrated by processes of weathering, and its ores are found in concretions embedded in residual clay or ocher and accompanied more or less closely by limonites. During this process

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[1] *Am. Jour. Sci.*, (3) 36, 1888, p. 177.

[2] *Am. Jour. Sci.*, (5) 7, 1924, p. 457.

[3] Hollandite and romanechite are said to be individualized forms of psilomelane.

[4] *Econ. Geol.*, 19, 1924, pp. 107–145.

[5] F. Behrend, *Zeitschr. prakt. Geol.*, 32, 1924, p. 81, 102.

[6] *Kolloid Zeitschr.*, 34, 1924, pp. 233–237.

some other metals, notably nickel, cobalt, zinc, lead, and barium, have a tendency to accompany the pyrolusite and psilomelane. In general such deposits are superficial or of slight depth and closely parallel the residual limonites already described.

In California small deposits of secondary manganese ores occur in areas of the radiolarian cherts or jaspers of the Franciscan formation (Jurassic ?).

In Arkansas high-grade residual ores have been mined at Batesville,[1] where they occur both in the Cason manganiferous shale, of upper Ordovician age, and in clay derived from this formation. Penrose believed that the manganese was derived from the pre-Cambrian area in southeast Missouri and deposited in the sedimentary formation. Later work of Ulrich, Miser, and others has shown that the Cason shale was probably the local source of all the manganese. This shale was a deposit on an old land surface where conditions would be favorable for concentration of manganese. Some of the manganese was transferred to the underlying Fernvale limestone and there deposited as carbonate. The final concentration took place when the region was peneplained, during the Cretaceous or Tertiary. The Batesville ores contain much hausmannite and braunite. It is possible that some manganese deposits in Arkansas and Oklahoma are derived from later veins of rhodochrosite, which break through the sediments.

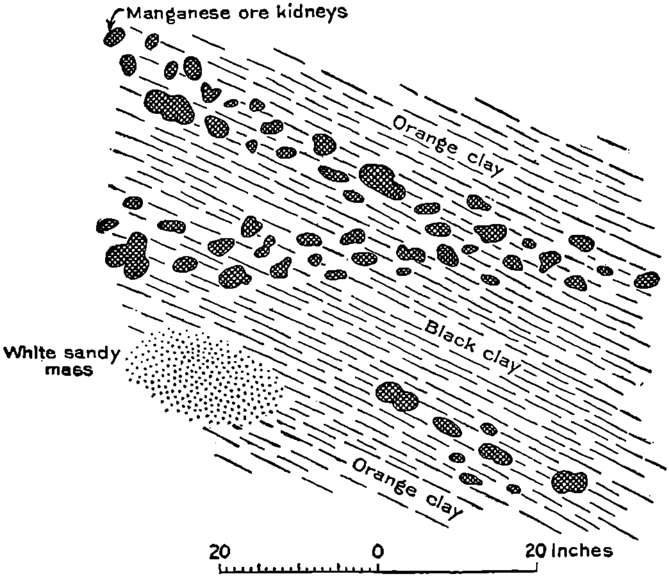

FIG. 137.—Sketch showing distribution of manganese ore lumps in clay at the Crimora mine, Virginia. (*After E. C. Harder, U. S. Geol. Survey.*)

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[1] H. D. Miser, *Bull.* 734, U. S. Geol. Survey, 1922.

In the Appalachian region small deposits occur in the granites and schists of the Piedmont region, but chiefly in the Paleozoic sediments of the Cambro-Silurian belt—that is, in the general area of the residual iron ores. At the Crimora deposit, in Virginia (Fig. 137), the ore is found as "masses of various sizes scattered through variegated clays in an elliptical basin in a canoe-shaped syncline of the Cambrian quartzite," into which the manganese penetrates as dendritic forms and crystalline coatings.[1]

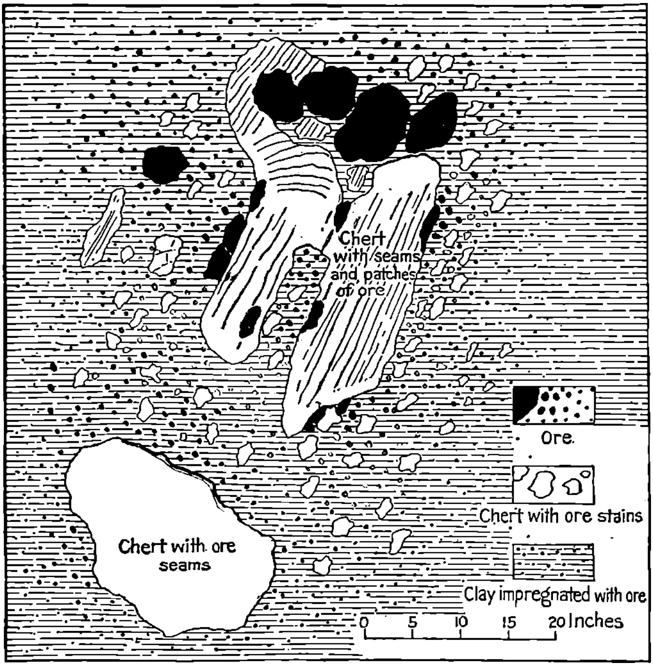

FIG. 138.—Sketch showing occurrence of manganese breccia ore at Reynolds Mountain, Virginia. (*After E. C. Harder, U. S. Geol. Survey.*)

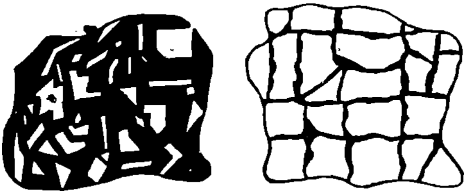

FIG. 139.—Sketch showing development of breccia ore by replacement. White areas, chert or sandstone; black, manganese ore. One-fifth natural size. (*After T. L. Watson.*)

The manganese deposits of the Appalachian region occur in a decomposed surface zone of many different rocks (Figs. 138 and 139); but most of the deposits are, according to Harder, associated with the top stratum

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[1] E. C. Harder, *Bull.* 427, U. S. Geol. Survey, 1910, p. 60.

of an impervious Cambrian quartzite overlain by limestone. Penrose held that they were laid down in local basins during the deposition of the rocks in whose residual clays they are now found. It is now believed[1] that the metal was in the first place obtained from the crystalline rocks of the Piedmont region and first concentrated in certain Paleozoic beds. The final concentration occurred in the Cretaceous and Tertiary during the several epochs of peneplanation.

ANALYSIS OF BEST QUALITY CRIMORA ORE

(*T. L. Watson, Mineral Resources of Virginia*, p. 248)

| | | | |
|---|---|---|---|
| $MnO_2$ | 81.703 | BaO | 0.829 |
| MnO | 7.281 | CaO | 0.880 |
| $Fe_2O_3$ | 0.533 | MgO | 0.630 |
| CoO | 0.354 | $P_2O_5$ | 0.171 |
| NiO | 0.096 | $(Na,K)_2O$ | 0.467 |
| ZnO | 0.623 | $H_2O$ | 3.405 |
| $Al_2O_3$ | 0.896 | $SiO_2$ | 2.132 |
| | | Total | 100.000 |
| | | Mn | 57.297 |

In central Texas,[2] in Mason, Llano, and San Saba counties, oxidized manganese ores occur as products of weathering of crystalline schists containing spessartite, piedmontite, and tephroite.

Many ore deposits contain manganese as carbonate and silicate, and in the oxidized zone the metal is often highly concentrated in the form of psilomelane, etc., mixed with limonite. These ores often contain gold and silver, but rarely much copper, lead, or zinc. Considerable quantities of such ores, used in part as flux for lead smelting and in part, if of high grade, for the manufacture of spiegeleisen, are mined at Leadville, Colorado.[3] Here the oxidized ore is apparently derived from a manganiferous siderite.

During the World War some high-grade ore was shipped from the oxidized outcrops of the Butte manganese veins; and larger bodies of such ore were found in the lodes of the Phillipsburg[4] district, Montana, here contained in limestone. Of late years rhodochrosite has been shipped from the Emma mine at Butte.

The largest part of the manganese obtained in the United States is derived from ores of the Lake Superior region, where manganese occurs as oxides associated with specularite (p. 300), and from the zinc residues of the great zinc deposit of Franklin Furnace, New Jersey. Here the manganese is contained in the franklinite ($(Fe,Zn,Mn)O.(Fe,Mn)_2O_3$)

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[1] Stose, Miser, and Katz, *Bull.* 17, Virginia Geol. Survey, 1919; *idem*, 23, 1922.

[2] Sidney Paige, *Bull.* 450, U. S. Geol. Survey, 1911.

[3] S. F. Emmons and J. D. Irving, *Bull.* 320, U. S. Geol. Survey, 1907, p. 26.

[4] J. T. Pardee, *Bull.* 725, U. S. Geol. Survey, 1922, pp. 141–244.

associated with zincite (ZnO) in a deposit of deep-seated, probably contact-metamorphic, origin.

**Brazil.**—The high-grade manganese deposits of Minas Geraes, Brazil, have been described by J. C. Branner and O. A. Derby.[1] In the main they appear to be residual ores derived from the weathering of lenses in the crystalline schists containing rhodochrosite, tephroite, and spessartite. The ores are concretions, masses, and vein-like deposits of psilomelane in the soft decomposed rock.

**India.**—Manganese ores are extensively distributed in India, and their occurrence and origin have been described in a detailed manner by L. L. Fermor.[2] These rich ores are formed by the combined replacement and decomposition of Archean rocks containing manganese silicates. In part the rocks are crystalline schists with spessartite and rhodonite, in part probably non-metamorphosed peculiar igneous rocks, one of which, for instance, consists of spessartite (spandite) and orthoclase with 3.70 per cent apatite. To a smaller extent the ores are contained in jaspery quartzites and in residual laterite.

Many deposits of the first class contain enormous masses of psilomelane, pyrolusite, and braunite. During the process of weathering almost all the silica and alumina have been removed.

Many of them extend to depths far below the water level, and Fermor believes that the oxidation may be of very ancient date, perhaps Archean. In some ways these concentrations by surface waters recall the Lake Superior iron deposits.

**West Indies.**[3]—Many residual manganese deposits occur in Cuba and in Central America. Those in Cuba are replacements in limestone and chert, perhaps also hydrothermally altered tuffs.

**Africa.**[4]—Large and important deposits of high-grade manganese ore have been discovered on the Gold Coast. They are derived from metamorphosed sediments containing spessartite and rhodonite which are similar to the rocks of India.

Another important deposit is found at Postmasburg, Griqualand, west South Africa. It carries sitaparite (braunite rich in Fe and Ca), braunite, specularite, a manganese diaspore, and barite. The original deposit was probably an oölitic sedimentary material.

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[1] Literature summarized by E. C. Harder, *Bull.* 427, U. S. Geol. Survey, 1910, p. 183.

[2] L. L. Fermor, The manganese ore deposits of India, *Mem.*, Geol. Survey India, 37, 1909.

[3] E. F. Burchard, Manganese ore deposits in Cuba, *Trans.*, Am. Inst. Min. Met. Eng., 63, 1920, pp. 51–104.

[4] A. E. Kitson and R. N. Junner, *Trans.*, Am. Inst. Min. Met. Eng., 75, 1927, pp. 372–396.

A. L. Hall, *Trans.*, Geol. Soc. South Africa, 29, 1926, pp. 17–46.

H. Schneiderhöhn, *Neues Jahrbuch*, Beil. Bd. 64-A, 1931, pp. 701–725.

**Production and Uses.**—The normal domestic output of manganese ores containing above 35 per cent Mn is small. Forced production under war conditions increased the output but not nearly sufficiently to supply the demand. Heavy imports come from Russia, the Gold Coast, Brazil, and India.

For the manufacture of spiegeleisen, an alloy of iron containing less than 20 per cent Mn, low grades of manganiferous iron ore may be used, but for other purposes the ores should contain at least 46 per cent Mn, and not more than 6 per cent Fe, 8 per cent $SiO_2$, and less than 1.5 per cent P.

The higher grades of manganese ores are used extensively for the manufacture of ferromanganese alloys, which are employed in the iron-smelting industry, particularly for hardening steel. The pure manganese dioxide ores also find an extensive chemical use, for the generation of chlorine and for the manufacture of cells for dry electric batteries. Such ores should contain not less than 65 per cent $MnO_2$ and no Cu, Ni, or Co.

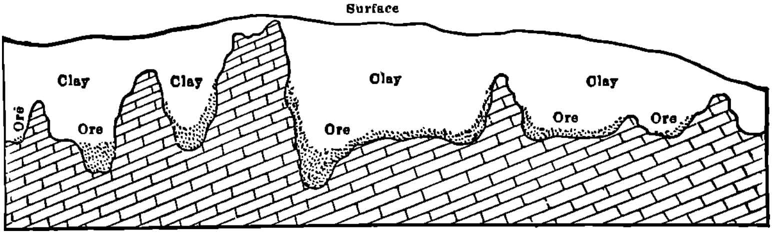

FIG. 140.—Section in open cut at the Bertha zinc mines, Virginia, showing relations of the residual ore to the limestone chimneys and the residual clay. (*After T. L. Watson.*)

In 1930, 67,000 long tons of ore with 35 + per cent Mn were mined valued at $1,438,000. The imports of the same class of ores were 586,000 long tons. Of ore containing 10 to 35 per cent Mn, 77,414 long tons were produced, and of ores with from 5 to 10 per cent, 707,973 tons.

### RESIDUAL BARITE

Barite as residual material and nodular concretions is not uncommon in the residual soils of Virginia and Georgia and in Washington County, Missouri. While it is possible that some barite is concentrated from syngenetic deposits in sedimentary rocks, the opinion is gaining ground that most of the residual barite is derived from epigenetic veins. In Missouri the barite is concentrated in the soil from veins in the Ordovician Gasconade limestone. Much of the barite produced in the United States is obtained from residual clays (p. 399).

### RESIDUAL ZINC ORE

In the Appalachian region, in western Virginia and eastern Tennessee, the Cambro-Silurian limestones contain in places sulphides of lead and zinc distributed in brecciated and crushed zones. At such localities the deep residual soil often contains calamine and smithsonite, the hydrated silicate and the carbonate of zinc, with some cerussite and galena. These ores occur next to the limestone at the bottom of the clay (Fig. 140), not scattered through it like limonite and pyrolusite.[1]

## RESIDUAL OCHERS[2]

The residual ochers are impure deep-red, yellow, or brown pulverulent materials consisting usually of predominant limonite and hematite with more or less clay and are generally used for pigments. They are no doubt colloidal precipitates. The terms Indian red, sienna, and umber, the latter two for the darker yellowish-brown and brown shades, are in use. Not all mineral pigments are natural products, for roasted pyrite, siderite, slates, and shales are also used.[3] The Clinton iron ores in New York State are also employed for these purposes.

The residual iron ore deposits of the southern states contain material which is used as ocher. Especially interesting are the Cartersville deposits,[4] in Georgia. These ochers occur only in the Weisner (Cambro-Silurian) quartzite, in the lower part of the residual zone immediately above the yet solid rock, and also in shattered zones in the quartzite itself. The quartzite contains about 90 per cent $SiO_2$, 1.5 per cent $FeS_2$, 0.5 per cent $Fe_2O_3$, and an unusual percentage of barium sulphate (4.46 per cent in the analysis given by Watson). The calculated constituents of the ocher are 66 per cent limonite, 25 per cent clay, and 9 per cent quartz; a little hematite is probably also present.

Hayes and Watson are in agreement regarding the origin of the ocher, considering it as resulting from a metasomatic replacement of the cement and quartz grains of the quartzite by limonite. The process begins by the permeation of the grains by dendritic limonite. This direct formation of the ocher is scarcely probable, but more likely it has progressed by means of an intermediate stage of siderite. The replacement

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[1] T. L. Watson, Lead and zinc deposits of Virginia, *Bull.* 1, Virginia Geol. Survey, 1905.

T. L. Watson, Lead and zinc deposits of the Virginia-Tennessee region, *Trans.*, Am. Inst. Min. Eng., 36, 1906, pp. 681–737.

[2] G. P. Merrill, Non-metallic minerals, 1910, pp. 104–111.

[3] B. L. Miller, The mineral pigments of Pennsylvania, *Rept.* 4, Topographic and Geologic Survey Commission of Pennsylvania, Harrisburg, 1911.

F. T. Agthe and J. L. Dynan, Paint-ore deposits near Lehigh Gap, Pennsylvania, *Bull.* 430, U. S. Geol. Survey, 1909, pp. 440–454.

[4] C. W. Hayes, Iron ores in the Cartersville district, Georgia, *Trans.*, Am. Inst. Min. Eng., 30, 1901, pp. 403–419.

T. L. Watson, The ocher deposits of Georgia, *Bull.* 13, Georgia Geol. Survey, 1906.

R. B. Ladoo, Non-metallic minerals, New York, 1925, pp. 368–391.

of quartz by iron carbonate is a well-known phenomenon, illustrated, for instance, in the Coeur d'Alene lead deposits of Idaho.

The annual domestic production of natural pigments amounts to about 57,000 tons. The mining is done mainly in open pits; and the material is crushed, washed in a log-washer, and allowed to settle in tanks.

## RESIDUAL PHOSPHATES

As described more fully on page 285, many sedimentary beds contain much phosphate of calcium, often in oölitic or concretionary form. When these beds are exposed to surface waters an enrichment usually takes place by solution of calcium carbonate, provided the beds are permeable to the circulating waters. Many important phosphate deposits—for instance, those of Florida, South Carolina, and Tennessee—have been thus enriched.

## DEPOSITS OF HYDRATED SILICATES OF NICKEL

The original home of nickel, cobalt, and chromium is in the peridotites and pyroxenites and in the serpentines derived from them, although traces of these metals are also frequently noted in analyses of other basic rocks. The primary condition of the nickel in the rocks is not always known; probably it occurs both as silicate and as sulphide, the latter in microscopic grains, the former as an admixture in iron-magnesium silicates. From the serpentines and peridotites the nickel is sometimes concentrated in commercially important quantities by processes of weathering, and the ores thus formed are always the green hydrated silicates of nickel. Chromite, which always occurs in these basic rocks, does not readily yield oxidized minerals in the zone of weathering. No chromium silicate analogous to garnierite exists.

Nickel silicates are uncertain in composition and generally of colloidal origin. The most important are genthite, $H_4Ni_2Mg_2(SiO_4)_3.4H_2O$; connarite, $H_4Ni_2Si_3O_{10}$; and garnierite, $(Mg,Ni)SiO_3 + nH_2O$. According to an analysis by A. Liversidge, garnierite contains 38.35 per cent $SiO_2$; 32.52 per cent NiO; 10.61 per cent MgO; 0.55 per cent $Al_2O_3$ and $Fe_2O_3$; 11.53 per cent $H_2O$ (at red heat) and 6.44 per cent $H_2O$ (at 100° C.).

Such deposits are superficial, and the oxidizing surface waters have been the carrying and concentrating agency. The ores rarely extend far below the water level and in some cases are contained in the residual clays of the completely weathered rock. These nickel ores are often accompanied by cobalt in the form of separate masses of asbolite, a rather indefinite mixture of hydrous oxides of manganese and cobalt.

These deposits do not contain sulphides, and copper is rarely present. The accompanying minerals are quartz, chalcedony, opal, and various obscure hydrous magnesium silicates, sometimes also a little magnesite.

Nickel ores of this kind are not uncommon, but have attained commercial importance only in New Caledonia.

The nickel mine at Riddles, in southern Oregon, has been described by several authors.[1] The parent rock is a peridotite containing 0.10 per cent of NiO. The olivine separated from the rock contained 0.26 per cent of NiO, and all observers agree that the nickel ores are formed from this silicate. In the finest joints of the rock, silica and nickel-magnesium silicates are deposited, and between them lies the oxidized rock converted to a limonite with some clay and chromite.

One of the two most important nickel-bearing districts of the world is in New Caledonia.[2] The island is about 250 miles long and 30 miles wide; one-third of the area is underlain by post-Eocene serpentine and peridotite. The lower slopes are covered by a deep mantle of decayed rock ("variegated clay") which really is an iron ore containing, in per cent, 18 silica, 69 ferric oxide, 0.45 alumina, 1.64 nickel oxide, and 10 water. The garnierite deposits are found at elevations of from 400 to 2,500 feet, sometimes on fairly steep slopes, or in the saddles of ridges and spurs. Underneath the "variegated clay" at depths of from 20 to 75 feet the nickel ores occur, often descending into the serpentine along fissures and accompanied by chalcedony and opal. There are many small deposits; the largest contained only 600,000 tons. The ores are worked by open cuts and carefully graded and sorted. Glasser classifies the deposits into veinlike, brecciated, impregnations, and earthy masses. In the last there is much dark brown "chocolate ore" in which the green silicate is not visible. The clayey ore averages, in per cent, 23 water, 5 to 7 nickel oxide, 10 to 12 ferric oxide, 25 magnesia, 40 silica, no lime, 1.1 chromic oxide, 0.12 cobalt, and 1.5 alumina. Most of the ore is exported. Some of it is sun dried and briquetted for local smelting, with limestone and gypsum flux, to nickel matte. Nickel metal is now also produced.

The New Caledonia deposits were discovered by the geologist Garnier in 1864; the mines were opened 10 years later, and the cheaply mined rich ores made all nickel deposits elsewhere unprofitable. In 1906 an output of 144,000 metric tons was reached. Lately, owing to the active competition of the Sudbury mines (p. 803), the output has been reduced. However, in 1931, about 147,000 metric tons of ore were

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[1] Diller and Clarke, *Bull.* 60, U. S. Geol. Survey, 1890, p. 21.

G. F. Kay, *Bull.* 315, U. S. Geol. Survey, 1907, p. 120.

[2] E. Glasser, Rapport sur les richesses minérales de la Nouvelle Calédonie, *Ann. des Mines*, (10) 5, 1904, pp. 29–154, 503–701.

G. M. Colvocoresses, *Eng. and Min. Jour.*, Sept. 21, and 28, 1907.

W. G. Miller, Nickel deposits of the world, reprinted from Report of Royal Ontario Nickel Commission, Toronto, 1917, pp. 234–264.

M. Piroutet, Étude Stratigraphique sur la Nouvelle Calédonie, Protat Frères, Macon. Also in *Bull.* Soc. géol de la France, sér. 4, 3, 1917, pp. 155–177.

produced and about 5,534 metric tons of 75 per cent nickel matte were exported to France and Belgium. The ore treated contained 5.35 to 5.67 per cent nickel. A small quantity of cobalt ore, a black, earthy asbolite, was exported for a number of years, but at present can not compete with the ore from Cobalt, Ontario. The island also produces much chromite (p. 786).

### BAUXITE[1]

**Introduction.**—Clay, as more or less impure kaolin, is the most abundant product of rock decay; but, although it carries up to 39.8 per cent alumina, its use as a source of metallic aluminum has not been found possible. Corundum is not abundant enough to be used for this purpose. Cryolite ($Na_3AlFl_6$), a mineral obtained from pegmatitic masses occurring in Greenland, was formerly an important aluminum ore and is still used, in smaller quantities, in the electrolytic processes for the extraction of aluminum and for other purposes.

In certain places the weathered zone, however, contains the hydroxides of aluminum, and of these bauxite is the most important aluminum ore. There are three aluminum hydroxides: Diaspore, $Al_2O_3.H_2O$, with 85 per cent $Al_2O_3$; gibbsite or hydrargillite, $Al_2O_3.3H_2O$, with 65.4 per cent $Al_2O_3$; and bauxite, $Al_2O_3.2H_2O$, with 74 per cent $Al_2O_3$.

The independence of bauxite as a mineral species is, however, questioned, and many authors consider it a hardened and in part crystallized hydrogel of indefinite composition. Some French and Georgian bauxites, correspond well to gibbsite. Bauxite forms compact, earthy, also very commonly pisolitic masses, the individual concretions often having a diameter of several centimeters. It is gray, cream-colored, yellowish, or brown and is usually admixed with silica and ferric oxide. The bauxites contain in places crystalline gibbsite as crusts or veinlets. Diaspore has been identified more rarely, and quite naturally, as it usually formed at higher temperature than that prevailing in residual deposits.

The bauxites always contain titanium, averaging as much as 4 per cent $TiO_2$, and some vanadium, but in this they merely share the peculiarities of residual and sedimentary clays. Some investigators state that bauxite contains residual rutile while others have failed to find any titanium mineral. In part, at least, the titanic dioxide is probably present in colloidal or metacolloidal form. Other forms of the monohydrate, called boehmite and tanatarite, have been identified from France and Russia.[2]

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[1] C. W. Hayes, Bauxite, its occurrence, geology, etc., *Sixteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1895, pp. 547–597.

T. L. Watson, *Bull.* 11, Georgia Geol. Survey, 1904 (Bibliography).

F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 496–510.

[2] J. de Lapparent, *Bull.* Soc franç. minéral., 53, 1930, pp. 255–273.

Little or no hydroxide of aluminum forms in ordinary rock weathering. Cameron and Bell[1] state that during an examination of several thousand soils from all parts of the United States, hydroxide of aluminum was observed in only one sample, which came from southern California. Bauxite, it may be concluded, is thus rarely formed in the temperate region.

In tropical countries, on the other hand, the deep residual soil very often contains aluminum hydroxide. This has been called laterite (later, brick) and is variously defined. We may say that true laterite is essentially a mixture of the hydroxides of iron and aluminum with more or less free silica, but there are all gradations toward an ordinary ferruginous clay. The laterite may be derived from any igneous or sedimentary rock but serpentine and limestone are specially favorable. The iron ore from Mayari, Cuba (p. 359) is a laterite exceptionally rich in iron. Many so-called laterites are not true residual but transported deposits. Laterites may or may not contain bauxite of economic value. They have been described from many lands and the literature is very extensive.[2]

In apparent contradiction to this, many of the worked bauxite deposits are found in temperate regions such as Georgia, Arkansas, France, Hungary, etc.; but this is explained by the fact that these are not being formed at the present time, but are of Tertiary age when a climate like that of Cuba prevailed in large parts of the temperate zone.

**Origin.**—The desilication of clay in low latitudes has been discussed extensively. The action of nitric acid, supposedly derived from rain during tropical thunderstorms, has been suggested as the cause. T. H. Holland[3] has mentioned the possibility of bacterial action and G. A. Thiel[4] has brought out experiments verifying this in part.

Clay is decomposed by sulphuric acid and by sodium hydroxide or sodium carbonate, and at some places aluminum hydroxide may have

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[1] *Bull.* 30, Bur. Soils, U. S. Dept. Agr., 1905, p. 28.

[2] A. Streng, *Zeitschr.* deutsch. geol. Gesell., 39, 1887, p. 621 (Germany).

A. Bauer, *Neues Jahrbuch*, Festband, 1907, p. 33, and 1898, Bd. 2, p. 192 (Seychelle Islands).

R. D. Oldham and T. H. Holland, *Records*, Geol. Survey India, 32, pt. 2, 1905, pp. 175–184.

L. L. Fermor, The manganese deposits of India, *Mem.*, Geol. Survey India, 37, 1909, pp. 370–380.

J. B. Harrison, Laterite in British Guiana, *Geol. Mag.*, 47, 1910, p. 440.

C. S. Fox, *Mem.*, Geol. Survey India, 49, pt. 2, 1923.

G. C. DuBois, *Tschermak's min. u. petr. Mitt.*, 22, 1903, pp. 4–18 (Surinam).

A. Lacroix, *Nouv. Arch.*, Mus. Hist. Nat. (Paris), 5th sér., 15, 1913, p. 255; reviewed by L. L. Fermor, *Geol. Mag.*, 1915, pp. 28, 77, 123 (French Guinea).

J. M. VanBemmelen, *Zeitschr. anorg. Chemie*, 66, 1910, p. 322 (General review).

J. Morrow Campbell, Laterite, its origin, structure, etc., *Min. Mag.* (London), August–November, 1917 (Tropical Africa).

[3] T. H. Holland, *Geol. Mag.*, 1903, p. 59.

[4] *Econ. Geol.*, 22, 1927, pp. 480–493.

originated in this way. W. Maxwell[1] has demonstrated this origin for some of the soils of Hawaiian volcanoes. Nevertheless it is clear that sulphuric acid does not always produce this effect, for diaspore and hydrargillite occur rarely in the oxidized portions of mineral deposits where the sericitic rocks are acted upon by sulphuric acid solutions. Bauxite also has rarely been observed. In the oxidized zone the sulphuric acid transforms sericite into kaolin, which is frequently accompanied by more or less alunite ($K_2O.3Al_2O_3.4SO_3 + 6H_2O$).

These suggestions do not suffice to explain the formation of the lateritic aluminum hydroxides. It is now generally conceded that this is caused simply by the long continued action of ordinary ground-waters under special conditions of moisture and heat.[2] W. J. Mead[3] has shown that there is a complete gradation in the case of the Arkansas deposits from the original syenite (Fig. 141), that the pisolitic structure develops in place, and that residual syenite boulders are surrounded by bauxitic material. The texture of the syenite is sometimes visible in the pisolitic bauxite. There is some evidence of downward leaching of the bauxite, for the

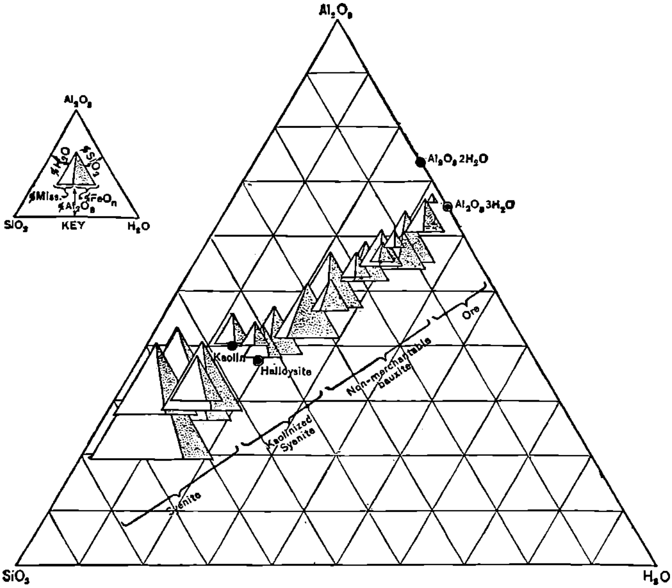

FIG. 141.—Triangular diagram showing the gradation from syenite to bauxite in terms of the principal chemical constituents. Each triangle represents an analyzed sample. (*After W. J. Mead.*)

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[1] W. Maxwell, Lavas and soils of the Hawaiian Islands, 1898.

[2] R. Schwarz, Ueber das Problem der Kaolinbildung, *Tonindustrie Zeitung*, 1924.

[3] *Econ. Geol.*, 10, 1915, pp. 28–54.

See also Leith and Mead, Metamorphic geology, New York, 1915, pp. 25–38.

top layer is usually more siliceous than the lower parts of the deposit. J. Morrow Campbell believes that bauxite forms only in the zone of percolation close to the fluctuating water level and that it never occurs far below the water level.

The sedimentary bauxites, of which numerous examples may be found in the Cretaceous and Tertiary beds of the southern states, are probably in part deposits swept out into the sea by the normal processes of erosion from bauxite rich laterites on the shore.

**Occurrences.**—The bauxite deposits of commercial importance are of several different types. In the United States they are confined to Arkansas and the southern Appalachian States.

In Arkansas[1] the mineral occurs in Pulaski and Saline counties as superficial beds over areas of various sizes up to 20 acres. The deposits range from 10 to 60 feet in thickness; both open-pit and underground methods are used; production in 1933 was 153,000 tons. They rest on nepheline syenite or on a kaolinized form of that rock; the lower part retains traces of granitic structure, while the upper part is distinctly pisolitic. Tertiary sands and clays in places cover the bed.

Other deposits of importance, described by Hayes[2] and also by Watson,[3] are found at a number of places in Georgia and Alabama. The principal occurrences are scattered between Jacksonville, Alabama and Cartersville, Georgia, along a belt about 60 miles in length, one of the typical localities being at Rock Run, Alabama. The bauxite occurs as pockets and irregular masses or curved strata of various colors, with clay and limonite, in the heavy mantle of residual clay overlying the Knox (Cambrian) dolomite, but sharply separated from it. The ore is in part pisolitic and is mined in open cuts, at some places to a depth of 50 feet or more. The bottom of the clay masses is rarely exposed; before it is reached the pockets of bauxite generally terminate in tapering points. Occasionally associated minerals are gibbsite ($Al_2O_3.3H_2O$) and halloysite, which is similar to kaolin in composition but has more water.

A suggestive fact is the occurrence of the deposits at or about the 900-foot contour, which coincides with the elevation of a probable Eocene peneplain. The ores were thus accumulated under topographic and climatic conditions different from those which prevail to-day.

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[1] C. W. Hayes, *Twenty-first Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1901, pp. 435–472.

J. C. Branner, *Jour. Geol.*, 5, 1907, pp. 263–289.

W. J. Mead, Occurrence and origin of the bauxite deposits of Arkansas, *Econ. Geol.*, 10, 1915, pp. 28–54. For recent operations see G. C. Branner, *Mining Met.*, 16, March, 1935, pp. 123–124.

[2] C. W. Hayes, The geological relations of the southern Appalachian bauxite deposits, *Trans.*, Am. Inst. Min. Eng., 24, 1895, pp. 243–254.

[3] T. L. Watson, *Bull.* 11, Georgia Geol. Survey, 1904, 169 pp.

W. A. Nelson, Appalachian bauxite deposits, *Bull.* Geol. Soc. Am., 34, 1923, pp. 525–532.

Deposits differing considerably from those already described have been found in several southern states.[1] They occur near the contact of the flat-lying sands and clays of the Tuscaloosa (Lower Cretaceous) and Claiborne (Tertiary) formations. The ore occurs either as beds resting directly upon Cretaceous clay or disseminated as nodules through it. A perfect series of transition to clay exists, as shown by analyses. Bauxite beds 10 feet in thickness have been observed; the mineral is clayey, dense, or pisolitic. The origin is still a point in controversy.

For a number of years much bauxite has been imported from British and Dutch Guiana.[2] The deposits are found along the coast and some distance inland and are mined by steam shovels, one plant producing 800 tons per day of which 70 per cent is ore. The bauxite occurs in beds 6 to 26 feet thick covered by Tertiary and recent clays and sands. The beds appear to be sedimentary, probably washed down from older deposits in the uplands. The ore contains in per cent: 59 to 61, $Al_2O_3$, 2 to 5, $SiO_2$; and 3 to 6, $Fe_2O_3$.

In France,[3] bauxite occurs as pockets and pipes in Lower Cretaceous limestone, probably formed during an interval of peneplanation and transgression in the Upper Cretaceous or perhaps between the Cretaceous and the Tertiary. In eastern Europe, bauxite is found in Istria, Italy, generally in similar deposits or in sink holes and was probably formed from clay by long-continued leaching.

Large deposits have recently been reported from the Atlas Mountains in Morocco and from the Gold Coast Colony in West Africa, 100 to 200 miles inland from Sekondi.

**Uses and Production.**—The annual production of bauxite in the United States in 1930 was 330,612 long tons, most of which was mined in Arkansas. The importations were 409,700 long tons, largely from the British and Dutch Guianas. The mines in France, also yield an increasing amount, about 608,500 metric tons in 1925. The price of bauxite is from $6 to $8 per ton. The bauxite ores of commercial grade contain

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[1] O. Veatch, *Bull.* 18, Georgia Geol. Survey, 1909, pp. 430–447.

R. F. Morse, The bauxite deposits of Mississippi, *Bull.* 19, Mississippi Geol. Survey, 1923.

E. F. Burchard, Bauxite in N. E. Mississippi, *Bull.* 750, U. S. Geol. Survey, 1925.

R. E. Rettger, The bauxite deposits of S. E. Alabama, *Econ. Geol.*, 20, 1925, pp. 671–686.

W. B. Jones, Bauxite in Alabama, *idem*, 21, 1926, pp. 792–802.

R. W. Smith, Sedimentary kaolins of the coastal plain of Georgia, *Bull.* 44, Georgia Geol. Survey, 1929, 489 pp.

[2] L. Litchfield, Bauxite mining in Dutch Guiana, *Eng. and Min. Jour.*, Aug. 17 and Sept. 21, 1929.

[3] J. de Lapparent, Les bauxites de la France méridionale, *Mém.*, Minist. Travaux Publique, Paris, 1930, 187 pp.

52 to 65 per cent $Al_2O_3$. The best ores contain 1 per cent Fe and 3 to 5 per cent $SiO_2$. Silica and iron should be less than 15 per cent. Ores with more than 4 per cent $Fe_2O_3$ are not utilized at present. The ores are mined in open cuts, often necessitating the removal of heavy overburden, washed to remove the clay, and dried. For purposes of aluminum smelting the ores must be of high grade and low in silica. In 1930, 114,518 short tons of aluminum were produced in the United States; exact data are not obtainable. The uses for the metal and its alloys are steadily increasing.

Large works for the electric smelting of aluminum are located at Niagara Falls and in Tennessee; also in Canada, Norway, and other countries. Artificial corundum (alundum) is made from the ore in the electric furnace. Bricks of bauxite for basic non-corrosive lining of furnaces are widely used. Aluminum salts, especially alum, are also manufactured from bauxite. An addition of bauxite promotes the rapid setting of cements.

ANALYSES OF BAUXITE

| | $SiO_2$ | $TiO_2$ | $Al_2O_3$ | $Fe_2O_3$ | $H_2O$ | Analyst |
|---|---|---|---|---|---|---|
| Baux, France (pisolitic). | 4.8 | 3.2 | 55.4 | 24.8 | 10.8 | Deville. |
| Jacksonville, Ala....... | 21.08 | 2.52 | 48.92 | 2.14 | 23.41 | Hillebrand. |
| Floyd County, Ga...... | 0.80 | 3.52 | 52.21 | 13.50 | 27.72 | Nichols. |
| Pulaski County, Ark.... | 2.00 | 3.50 | 62.05 | 1.66 | 30.31 | |
| Wilkinson County, Ga.. | 9.38 | 2.76 | 57.58 | 0.96 | 29.12 | E. Everhart. |

Great variations are often shown in one locality. For further analyses see G. P. Merrill, Non-metallic minerals, 1910, p. 91. The average of a long series of analyses of commercial ore from Georgia tabulated by T. L. Watson[1] gives: $SiO_2$, 4.274; $TiO_2$, 3.791; $Al_2O_3$, 58.622; $Fe_2O_3$, 1.507; and $H_2O$, 31.435; total 99.629. This corresponds to $Al_2O_3.3H_2O$.

**The "Diaspore" Clays.**[2]—During the last years bauxitic clays unusually high in alumina have been mined in Missouri. They fill sink holes in Cherokee (Pennsylvanian) sandstone approximately 250 feet in diameter and 100 feet deep. They are surrounded by a casing of non-plastic flint clay and carry from 50 to 75 per cent $Al_2O_3$. Enrichment by solution of silica is suggested.

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[1] *Bull.* 11, Georgia Geol. Survey, 1904, pp. 45–46.

[2] D. C. Wysor, Diaspore clay of Arkansas and Missouri, *Jour.* Am. Ceramic Soc., 6, 1923, pp. 501–509.

H. S. McQueen, Mining of diaspore clay, burley, and flint clay in Missouri, *idem*, 11, 1929, pp. 687–697; also 12, 1929, pp. 204–214.

# CHAPTER XXII

# DEPOSITS FORMED BY CONCENTRATION OF SUBSTANCES CONTAINED IN THE SURROUNDING ROCKS, BY MEANS OF CIRCULATING WATERS

## GENERAL STATEMENT

The water which sinks through the soil and effects the weathering of rocks becomes charged with small amounts of carbonates of calcium, sodium, magnesium, potassium, iron, and other metals, and also with soluble silica. By far the larger part of it, after a short journey through the belt of oxidation, returns to the surface as springs and seepages and is carried off in the watercourses to the sea. A smaller part of this water sinks into the ground and either joins the active circulation, descending in smaller fractures and openings to ascend on the larger fissures and other waterways, or becomes a part of the stagnant or almost stagnant and gradually diminishing ground-water of deeper levels. In comparison with the depth of the ground-water, the depth of oxidation, or rock decay, is on the whole insignificant; and that part of the dissolved substance which is carried down is also insignificant in comparison with the vast amount of underlying rocks, so that we cannot expect that the material added from the zone of weathering will produce any far-reaching changes in the composition of these rocks.

Nevertheless dissolved salts are carried down from the weathered belt and may cause deposits in open cavities or may form more or less complex replacements. In the openings silica may be deposited as chalcedony, chert, or quartz; calcium carbonate may fill fissures and replace silicates; or ferrous carbonate may be substituted for limestone. Chlorite, kaolin, and sericite may develop in igneous rocks. All these changes are, however, accompanied by renewed solution; and it is a debatable question whether the solution does not more than balance deposition. On the other hand, the water returning to the surface after a journey of varying length, more or less heavily loaded with soluble salts, deposits these by reason of decrease of temperature or by reaction with other surface waters of different composition. Finally, hydration absorbs much water, both from the active circulation and from the more stagnant ground-water; and deposits of valuable minerals may result from this simple process.

In a rough way the deposits resulting from the work of underground waters of meteoric origin may be divided into: (1) Those formed from abundant material contained in the surrounding rocks, for instance, magnesite, serpentine, sulphur (by reduction of gypsum), and certain kinds of hematite; and (2) those formed by the deposition of rarer substances dissolved by the water from the surrounding rocks or from rocks that lie deeper. As examples of the second division may be cited the vanadium and copper deposits of the "Red Beds" in the western states.

Waters of atmospheric origin doubtless have the power to dissolve many of the rarer metals contained in rocks, to carry them for considerable distances, and to concentrate them in places suitable for deposition; but unless it is aided by higher temperatures at considerable depths below the surface, this power is not often strong enough to produce important deposits of these rarer metals.

### SULPHUR[1]

**Modes of Occurrence.**—Native sulphur may be formed by various reactions. The oxidation of pyrite sometimes results in crusts of sulphur coating the cavities once occupied by the dissolved crystals. In the craters of volcanoes where sulphurous gases ascend on crevices sulphur is often found, as the result of a reaction between sulphur dioxide and hydrogen sulphide ($H_2S + 2SO_2 = H_2SO_4 + 2S$), or more probably, by incomplete oxidation of hydrogen sulphide ($2H_2S + O_2 = 2H_2O + 2S$), or by the reaction $3SO_2 + 2H_2O = 2H_2SO_4 + S$. A large deposit of this kind is worked at the Abosanobori mine, Hokkaido, Japan, and consists of clayey beds in an old crater lake. Considerable quantities have been exported from Japan to the United States. It has been proposed to utilize a similar deposit in the crater of Popocatepetl, Mexico. Other deposits are found in the volcanoes of the Chilean and Argentine Andes.

Much more commonly sulphur is found at active or extinct hot springs in the tufas or other adjoining porous rocks, like volcanic tuffs. It evidently results from the incomplete oxidation of $H_2S$, by the oxygen, or by bacterial action, or by a reaction between suspended $CaCO_3$ and $H_2S$ resulting in $Ca(SH)_2$. This, as well as soluble polysulphides and thiosulphates, is decomposed by $CO_2$ with the liberation of $H_2S$ and S. Such deposits have been observed at many places in the Western states—for instance, at Cuprite, Esmeralda County, Nevada; at the Rabbit Hole mines in Humboldt County, Nevada;[2] at Sulphur Bank, California; at the Cove Creek mine, Beaver County, Utah;[3] and at Cody and Thermopolis, in Wyoming.[4] The three last-named deposits have been worked. In Wyoming the sulphur in part replaces the limestone underlying the

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[1] O. Stutzer, Die wichtigsten Lagerstätten der Nicht-Erze, 1911, pp. 185–263.
U. S. Bur. Mines, *Information Circ.* 6329 (Sulphur), 1930.

[2] G. I. Adams, *Bull.* 225, U. S. Geol. Survey, 1904, pp. 497–500.

[3] W. T. Lee, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 485–489.

[4] E. G. Woodruff, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 451–456.
E. G. Woodruff, *Bull.* 380, *idem*, 1909, pp. 373–380.

travertine or tufa. All these deposits are superficial; and, though some are fairly productive, they play no great part in the world's output.

The greater part of the native sulphur is not connected with volcanic processes or hot springs but is obtained from sedimentary beds in close association with gypsum and limestone; calcite, aragonite, barite, celestite, opal, more rarely quartz, together with gaseous and solid hydrocarbons, are found with the sulphur. This association is constant and characteristic and recurs in almost all the great gypsum beds of the world, though the sulphur is not always present in quantities of economic importance. The sulphur is in earthy or resinous masses and forms lenticular beds, irregular masses, and veinlets in marl, limestone, and gypsum. Sulphur is widely distributed in the Miocene and Pliocene of the Mediterranean countries, everywhere accompanied by gypsum. It is mined particularly in Sicily, which is one of the great sulphur-bearing regions in the world. Another mode of occurrence is in the upper part (cap rock) of the salt domes which occur so commonly in the Gulf Coast of Louisiana and Texas. The salt domes (p. 333) are cylindrical masses of sodium chloride, usually capped with anhydrite and gypsum. They are derived from lower salt beds and have been forced up through the overlying soft rocks, in some cases almost to the surface. They are known as "diapiric" (piercing) structures. At present, by far the greater part of the world's sulphur production is obtained from these salt domes.

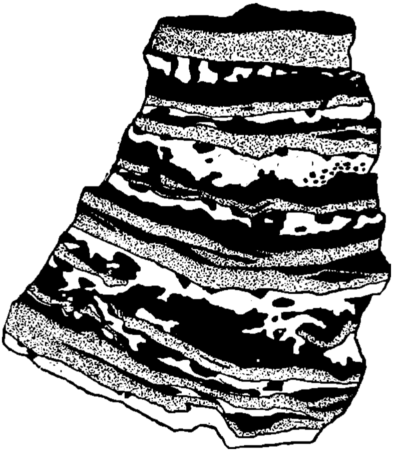

FIG. 142.—Banded sulphur rock from Sicily, one-half original size. Black, sulphur; white, calcite; stippled, limestone. (*After O. Stutzer.*)

**Sicily and Italy.**—In south-central Sicily the sulphur-bearing region extends for 160 kilometers east and west; and scattered through this territory are innumerable small mines. The largest mines lie between Agrigento and Caltanissetta in the Trabia-Tallarita folded basin. The depth attained is 800 feet. They supply about half of the production of the island. Mining and reduction are carried out by modern methods. The bed worked reaches a thickness of 70 feet in places. The sulphur is accompanied by calcite, some gypsum, and bituminous matter. It lies at the top of the Miocene and is covered by Pliocene marls. Below the sulphur bed lies heavy platy gypsum and below that the Tortonian (Upper Miocene) marls. This succession is almost constant, though the thickness of the sulphur bed varies greatly. The ore is rich, sometimes massive, much of it containing 25 per cent S (Fig. 142). Crystallized sulphur and celestite, more rarely barite, coat the walls of cavities. Sulphur deposits also occur at various places in Italy. The Altavilla-Irpina mine, 100 miles east of Naples, is actively worked with heavy production. The deposit occurs in steeply dipping Miocene beds. The sulphur occurs in kidneys and lenticular masses in clay shale, some of them 15 feet in diameter; apparently this sulphur has replaced gypsum. Above and below are alternating beds of gypsum and clay.

**Louisiana and Texas.**—The most productive sulphur deposits in the world are those which occur in the cap rock or top part of the salt domes (Figs. 128, and 143, p. 383).[1] The salt dome consists of a stock-like, round, or oblong mass of nearly pure salt extending from unknown depths into the overlying, soft, Tertiary sediments, the diameter ranging from less than ½ mile to 2 miles or more. The sides are steep, sometimes even vertical or overhanging. These diapiric or piercing structures have often a cap of thin sedimentary limestone, underneath which the sulphur ore is found, extending over the flat top and usually thickest at the edges but soon contracting along the steep sides. The cap rock of the worked sulphur deposits is from 500 to 1,500 feet below the surface, but in many domes it has not yet been found and in other cases it is still nearer to the surface.

The cap rock consists of granular calcite, sulphur and some gypsum, aragonite, and dolomite. It often contains celestite; it also contains more or less bituminous matter or oil; oil in larger quantities often occurs in the Tertiary upturned sediments surrounding the dome. Sulphur occurs in irregular masses, veins, and lenses. Below the cap rock lies a variable thickness of probably sedimentary anhydrite (also with some celestite); below the anhydrite the drill enters into the main salt mass, which rarely has been penetrated. The cap rock is essentially a recrystallized material evidently derived from gypsum and anhydrite.

None of these deposits are inexhaustible. The dome at Sulphur, Louisiana, in Calcasieu County, was exhausted in 1924 after producing 10,000,000 tons of sulphur. The largest dome known is at Boling, Texas, worked by the Texas Gulf Sulphur Company; it is reported to contain reserves of 45,000,000 tons. Other localities worked are at Long Point, and Palangana; two important deposits are Bryanmound and

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[1] For recent descriptions and map of the salt and sulphur domes see:

P. K. Kelly, The sulphur salt dome, Louisiana, *Bull.* Am. Assoc. Petr. Geol., 9, 1925, pp. 479–496.

R. F. Bacon and H. S. Davis, Recent advances in the American sulphur industry, *Proc.*, Am. Inst. Civil Eng., December, 1920.

M. I. Goldman, Petrography of salt dome cap rock, *Bull.* Am. Assoc. Petr. Geol., 9, 1925, pp. 42–98.

Levi S. Brown, Cap rock petrography, *idem*, 15, 1931, pp. 509–529.

Jack Logan, *Oil Weekly*, Oct. 3, 1930.

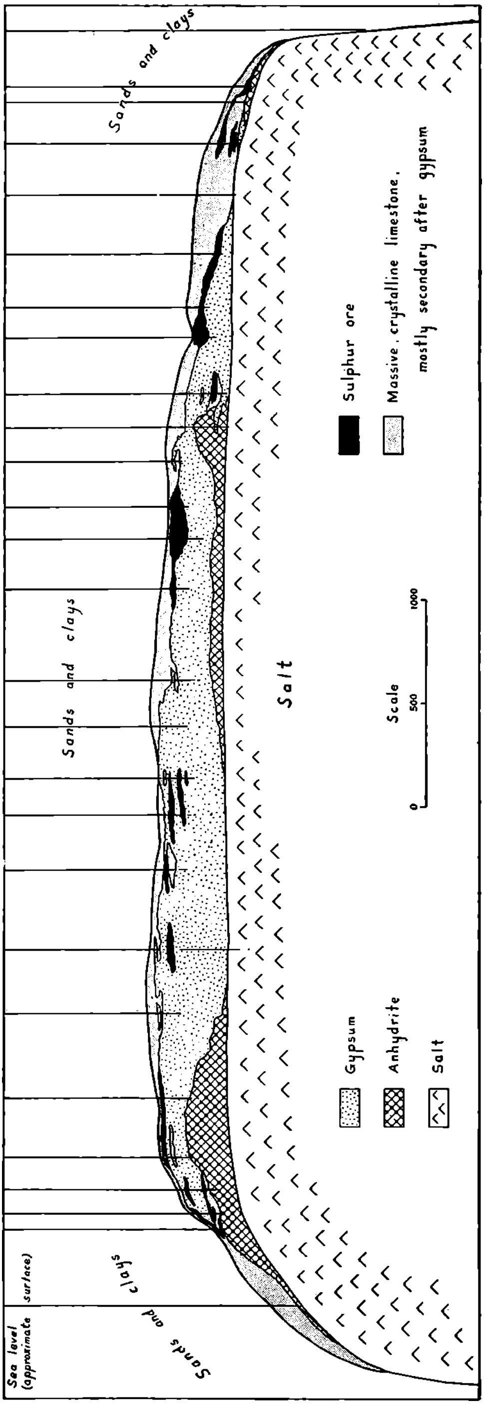

FIG. 143.—Vertical section of a sulphur dome, Texas.

Hoskins Mound, near the mouth of the Brazos River. Over 82 salt domes are known, located in part by seismic and torsion-balance methods. Probably all contain some sulphur, but few are of economic importance. Bryanmound, which so far has produced over 5,000,000 tons, occupies an area of 780 acres rising 22 feet above tidewater. The total number of wells drilled up to 1930 was 1,148. The sulphur-bearing rock is found at depths from 700 to 1,500 feet. The safest sampling is done by assaying drill cores, calculating the thickness of the sulphur in each hole, and then extending this to adjacent holes. Naturally the thickness of the sulphur-bearing formation is an important factor. The content is usually expressed in tons per acre, anything less than 5,000 tons being probably unprofitable. The extraction is, of course, never complete and probably often less than 50 per cent.

The development of these deposits was long delayed by the impossibility of sinking shafts through the quicksand. The final difficulties were overcome by the invention of the Frasch process. Through 12-inch pipes superheated water is forced down to the sulphur which is thereby melted; hot air is then supplied under pressure to aerate the molten mass and facilitate its ascent to the surface.

**Origin of Sulphur Deposits in Gypsum.**—In spite of structural differences these deposits have many similar features, such as the constant association with recrystallized calcite, presence of bitumen or oil, as well as celestite and barite. The sulphur is later than the gypsum or anhydrite. The direct reduction of gypsum by carbon compounds below 600° to 700° C., suggested by G. Bischof, appears to be ruled out by the chemists. The introduction of $H_2S$ from foreign sources (springs, mud volcanoes, etc.), advocated by V. Lasaulx, Mottura, G. Spezia, and Baldacci, does not help much, for in all these theories oxygen is needed to oxidize $H_2S$, or calcium or sodium sulphides, and in most cases the presence of much oxygen seems unlikely under the assumed conditions of deposition. For the older literature on this subject see Clarke's, "Geochemistry" and Hunt's paper quoted below.

A purely sedimentary origin of the Sicilian deposits is advocated by O. Stutzer.[1] Hydrogen sulphide is generated in large amounts in closed basins, such as the Black Sea, by the decay of organisms. Yet, no sulphur appears to have been brought up by deep dredgings in that basin, though they contain abundant iron sulphide.

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[1] O. Stutzer, Die wichtigsten Lagerstätten der Nicht-Erze, 1911, pp. 185–263.

W. F. Hunt, The origin of the sulphur deposits of Sicily, *Econ. Geol.*, 10, 1915, pp. 543–579.

Sir John Murray, The deposits of the Black Sea, *Scottish Geog. Mag.*, 16, 1900, pp. 673–702.

Stelzner and Bergeat, Erzlagerstätten, 1, 1904, p. 470. A comprehensive review by Doss is given in *Neues Jahrbuch*, 1, 1900, pp. 224–228.

It has been proved[1] that certain bacteria are able to reduce sulphates at temperatures up to 55° C. with the liberation of $H_2S$. In part these bacteria are anaërobic, and Bastin finds that they are present in deep oil-well waters. The oxidation of $H_2S$ and the action of $CO_2$ on Ca $(OH)_2$ and Ca $(SH)_2$, products formed by the reduction of $CaSO_4$, would cause a separation of sulphur and $CaCO_3$. Calcium polysulphide formed would be immediately decomposed to sulphur by $H_2S$.

The oxidation of $H_2S$ may also be performed by bacteria, the so-called sulphur bacteria (Beggiotoa) which deposit sulphur in thin cells. The following oxidation of this sulphur supports the life of the organisms, which are found in the mud of hot springs, lakes, and seas. This is the theory advanced for the sedimentary origin of the sulphur deposits of Sicily. Even if this is true, there must have been a great deal of subsequent recrystallization in the Sicilian deposits.

For the sulphur deposits in the cap rock of the salt domes this theory can not be applied; they are certainly not sedimentary; on the contrary, sulphur and allied minerals are the latest product formed in the cap while under a heavy covering of Tertiary sediments. If the theory of Bastin of the occurrence of bacteria in oil-well waters is correct, bacterial action might here too be possible, though the organisms probably must have been of the anaërobic type.

Besides the direct reduction of sulphates we have here another source of $H_2S$ from the waters ascending on the flanks of the salt domes; they are almost invariably rich in $H_2S$ and sulphates; but even if this mode of origin is admitted, there is the necessity of oxidation by free oxygen in inorganic reactions or by bacterial action. The ascending $H_2S$ waters usually meet potable sodium carbonate waters near the surface, and the mixture would probably result in sodium sulphide or polysulphides; but for these, too, some mode of oxidation must be provided.

**Production.**—The Sicilian deposits, with an annual output of about 450,000 metric tons, formerly supplied the world's demand. In 1901 the Frasch process revolutionized the trade conditions, and the production of the United States rose at once to 200,000 long tons, and the importations fell off correspondingly. The interesting trade conditions developing from these changes and the struggles of the Italian government to aid the distressed Sicilian operators have attracted much attention. In 1930, the Sicilian output was 246,814 long tons. In the same year, the production of the United States was 2,558,981 long tons. The price was about $18 per ton. Practically no sulphur has been imported since 1916, and the deposits of the United States now supply

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[1] See papers by Stutzer and Hunt cited above and E. S. Bastin, The problem of the natural reduction of sulphates, *Bull.* 10, Am. Assoc. Petr. Geol., 1925, pp. 1270–1299.

85 per cent of the world's demands. Almost the entire output is produced by the Texas Gulf Sulphur Company and the Freeport Texas Company.

**Uses.**—The manifold industrial uses of sulphur need not be specified; the larger part is used for the manufacture of sulphuric acid, for bleaching purposes by the development of sulphur dioxide, for the prevention of mildew on grapevines, for the manufacture of gunpowder, matches, etc., and for the rubber industry.

**Sulphuric Acid.**—Just before the World War, the production of sulphuric acid in the United States was about 3,500,000 short tons. In 1930, 7,625,000 tons of acid (50° Baumé) were consumed. Of the production in 1916, 40 per cent was made from Spanish (Rio Tinto) pyrite, 6 per cent from Canadian pyrite, 13 per cent from domestic pyrite, marcasite, and pyrrhotite, 22 per cent from fumes from copper and zinc smelters, leaving about 19 per cent which had to be supplied from native sulphur.[1] At the present time sulphur supplies a much larger percentage, probably over 70 per cent. The manufacture of superphosphate and petroleum refining consume the largest part of the sulphuric acid.

Many countries, particularly Spain, Norway, Portugal, France, United States, Italy, and Germany, in the order of importance stated, produce annually over 200,000 tons of pyrite.

Pyritic ores are obtained, in the United States:

(1) From pyrite deposits along the Appalachian mountains from Alabama to Vermont; (2) from pyrite deposits in California; (3) from pyrrhotite deposits in Virginia and Tennessee; (4) from marcasite as a by-product of coal mines in Illinois, Ohio, Indiana, and Pennsylvania; and (5) from marcasite as a by-product in zinc-lead mines of Wisconsin and Illinois.

The domestic production of pyrite in 1930 (including marcasite and pyrrhotite) was 347,512 long tons, of which the larger part came from Tennessee, California, and Virginia. Pyrite was also imported from Quebec, Ontario, and the Rio Tinto district in Spain (325,000 long tons in 1930). The mining and importation of pyrite have been unfavorably affected by the great increase in the production of sulphur. The production of pyrites in Spain for 1929 was about 3,000,000 metric tons.

The "pyritic deposits" comprise many types (p. 618); but aside from the minor supplies mentioned under 4 and 5, they are mainly products of high or intermediate temperature under intrusive conditions. Most of them may be considered as copper deposits of very low grade. Many among those along the Appalachian belt are of early Paleozoic age and more or less strongly dynamometamorphosed. We may mention the pyrrhotite deposits of Ducktown, Tennessee, and of the "Great Gossan

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[1] W. Y. Westervelt and A. G. White, *Bull.* 130, Am. Inst. Min. Eng., November, 1917, pp. v–xii.

Lead," Virginia, and the pyritic deposits of Louisa County, Virginia, which form long lenses in a Cambrian sericite schist and northward change into lead and zinc deposits. Other deposits of pre-Cambrian age are found in St. Lawrence County, New York. The sulphur ores contain from 30 to 50 per cent of sulphur with up to 10 per cent of insoluble. Lead, zinc, antimony, and arsenic are objectionable constituents. The residues from sulphuric acid manufacture are often used as copper and iron ores.

## THE MAGNESIAN DEPOSITS

The magnesian silicate rocks lend themselves easily to transformation and yield a number of economically valuable products, among them serpentine, magnesite, meerschaum, talc, soapstone, and asbestos. All of these result from the action of water, in most cases doubtless of atmospheric origin, on peridotites, pyroxenites, or gabbros, either near the surface or, with the cooperation of stress, at greater depths. Talc, soapstone, and asbestos belong, in part, to the latter class.

### SERPENTINE[1]

Serpentine forms by simple hydration from a rock consisting of enstatite and olivine according to the following equation:

$$\underset{\text{(Olivine)}}{Mg_2SiO_4} + \underset{\text{(Enstatite)}}{MgSiO_3} + 2H_2O = \underset{\text{(Serpentine)}}{H_4Mg_3Si_2O_9}.$$

It may also develop from olivine alone, with the removal of some magnesium as carbonate:

$$2Mg_2SiO_4 + CO_2 + 2H_2O = H_4Mg_3Si_2O_9 + MgCO_3.$$

The latter equation probably represents the usual process of serpentinization a short distance below the surface. Under oxidizing clonditions serpentine is unstable, though, of course, the change takes pace very slowly and erosion may work far ahead of decomposition.

Serpentine is, however, also formed on a large scale at greater depths, where quantities of $CO_2$ could not very well be assumed for the reason that such alteration would result in a mixture of serpentine and carbonates, whereas the large serpentine masses rarely contain admixed carbonates. The deep cañons of the Sierra Nevada, in California, show clearly that the serpentines of this range are not superficial, but descend to the depth of several thousand feet. The *modus operandi* of such extensive hydration is not fully explained. Some have held that it might have been effected by ascending waters, shortly after the intrusions.

Serpentine is generally rich in iron, for the original rocks are not of the purity indicated by the equation given above; the iron is present

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[1] H. Leitmeier in Doelter's Mineralchemie, 1, pt. 1, 1914, pp. 385–428.

as both silicate and magnetite, and also in chromite which forms a characteristic accessory. Serpentine rock that is not too much broken by joints finds fairly extensive use as building and ornamental stone. For the latter purpose the oily green translucent varieties, formed in crystalline limestone by serpentinization of the contained pyroxene, are particularly valued.

## MAGNESITE[1]

**Origin.**—Magnesite ($MgCO_3$) appears in two modifications: (1) As a cryptocrystalline (often called amorphous), earthy, hard, and compact mineral, which probably is a hardened colloidal precipitate. It is often concretionary and has a conchoidal fracture like that of unglazed porcelain. In this form it is an alteration product of serpentine or allied magnesian rocks as illustrated by the equation

$$H_4Mg_3Si_2O_9 + 3CO_2 = 3MgCO_3 + 2H_2O + 2SiO_2.$$

(2) As a crystalline mineral, isomorphous with calcite and usually holocrystalline granular. In this form it is generally a replacement of dolomite produced by magnesian solutions in connection with intrusions.[2] Brucite ($Mg\ (OH)_2$) is sometimes present.

**Occurrence.**—The amorphous magnesite is not uncommon in areas of serpentine, and occurs in fissures or crush zones or irregular masses, mixed with more or less serpentine and some opal or chalcedony. It is often pure with slight admixtures of iron, alumina, and lime and a few per cent of free silica. Magnesite occurs abundantly, but generally in small deposits, in the California Coast ranges. Until recently the pro-

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[1] G. W. Bain, Types of magnesite deposits, *Econ. Geol.*, 19, 1924, pp. 412–433.

M. Dittrich, H. Leitmeier, and K. A. Redlich in Doelter's Handbuch der Mineralchemie, 1, 1912, pp. 212–267.

F. L. Hess, The magnesite deposits of California, *Bull.* 335, U. S. Geol. Survey, 1908.

W. W. Bradley, Magnesite in California, *Bull.* 79, California State Min. Bur., 1925.

C. D. Dolman, Magnesite, etc. *Trans.*, Am. Inst. Min. Met. Eng., 63, 1920, pp. 175–187.

[2] Certain minor occurrences are of interest: Magnesite of the amorphous type is found as sedimentary beds and lenses in clays of Miocene lake beds near Bissell, San Bernardino County, California. See H. S. Gale, *Bull.* 540, U. S. Geol. Survey, 1914, p. 512.

Crystalline magnesite occurs in many crystalline schists of the Austrian Alps.

Hydromagnesite ($3MgCO_3.Mg(OH)_2 + 3H_2O$) is reported from Atlin, British Columbia, as a deposit of fine white powder several feet deep and appearing like a spring deposit. In connection with this it is recalled that H. Leitmeier found that a magnesian hydrocarbonate was deposited by the mineral waters of Rohitch in Styria. *Zeitschr. Kryst. Min.*, 47, 1909, p. 118.

duction has been small owing to distance from the eastern market. Similar deposits of great extent producing annually 134,000 tons are worked on the coast of Euboea, in Greece.[1] Other localities are found at Salem, near Madras, India, in the Transvaal, and many other lands where serpentinoid rocks abound. Magnesite does not always accompany serpentine, however; and it may be surmised that ascending springs with much $CO_2$, as are so common in California, may have some connection with its genesis.

The largest magnesite deposits in the world are at Veitsch, in Styria,[2] Austria, where in 1914 about 200,000 metric tons were mined in open quarries. This crystalline magnesite is a replacement of dolomite formed under the influences of intrusions of porphyry and other acidic and basic rocks. The Austrian magnesite, though otherwise pure, contains, like that from Greece, a few per cent of iron which makes it desirable for basic linings and bricks. In 1916 a large magnesite deposit was discovered near Chewelah, Washington, which bids fair to supply the domestic demand. Like the Styrian deposits it is a replacement of dolomite of Carboniferous age, near granite; in places the material contains a few per cent of silica, lime, and iron. Very similar to these are the deposits found some years ago in Grenville township,[3] Quebec, and which are now being worked. At this locality the magnesite contains several per cent of lime and very little iron (Fig. 144). Large deposits of magnesite have been found in Manchuria[4] in pre-Cambrian schists. The origin is still in doubt.

**Production and Use.**—The domestic production of crude magnesite in 1930 was 129,320 short tons, at $10 to $14 per ton. The imports in the same year were 842 short tons crude and 3,911 tons calcined. Much dolomite was also mined in the United States, mainly for use in refractories. Magnesite gives off its carbon dioxide at 550° C., and is, therefore, preferred to calcite in the production of this gas. After calcining, the substance is used for the manufacture of various magnesium salts, and in the paper and sugar industries. It is employed extensively with magnesium chloride for the so-called Sorel cement, used for flooring, etc. Its most important use is for basic furnace lining. Until recently the Styrian magnesite was imported for this purpose. Magnesite for bricks

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[1] H. C. Boydell, The magnesite deposits of Euboea, *Econ. Geol.*, 16, 1921, pp. 507–523.

H. S. Gale, *Bull.* 540, U. S. Geol. Survey, 1914, pp. 483–520.

[2] K. A. Redlich, Die Genese der Pinolitmagnesite, Siderite und Ankerite der Ostalpen, *Tschermak's min. u. petr. Mitt.*, 26, 1907, pp. 499–505.

K. A. Redlich, Genesis der kristallinen Magnesite, *Zeitschr. prakt. Geol.*, 21, 1913, pp. 90–101.

[3] M. E. Wilson, Magnesite deposits of Grenville districts, etc., *Mem.* 98, Canada Geol. Survey, 1917.

[4] K. Niinomy, *Econ. Geol.*, 20, 1925, pp. 25–53.

should contain a few per cent of FeO and little CaO; 8 per cent CaO being the allowable limit.[1]

Magnesite serves also as an ore for the production of metallic magnesium, which on account of its low specific gravity (1.74) is now used in alloys with aluminum and other metals. The reduction is effected by treating the chloride made from magnesite in an electric furnace. The best ore for the purpose is naturally carnallite ($KCl.MgCl_2 + 6H_2O$) from the Stassfurt salt beds (p. 337). The production of magnesium in the United States in 1930 was 559,631 pounds, the price being about $1 per pound but declining to 48 cents at the end of the year.

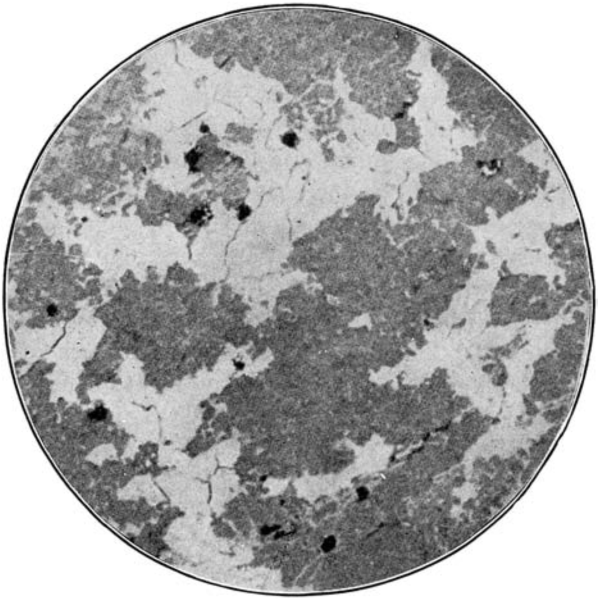

FIG. 144.—Thin section of Grenville magnesite-dolomite; dark, magnesite; light, dolomite; black spots, serpentine. Enlarged 2 diameters. (*After M. E. Wilson.*)

## MEERSCHAUM[2]

Meerschaum or sepiolite ($H_4Mg_2Si_3O_{10}$, containing $SiO_2$, 60.8 per cent; MgO, 27.1 per cent; $H_2O$, 12.1 per cent) is a hydrated silicate of magnesia of tough, compact texture, white or cream color, and smooth feel. As is well known, it finds a rather extensive use in the manufacture of pipes and cigar holders. Its analysis usually shows a little iron, alumina, and lime. It is probably derived from serpentine by slow hydration and is in most cases a colloidal precipitate. The principal occurrence is in Asia Minor at Eski-Shehr, where it is found as nodular masses near the sur-

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[1] J. S. McDowell and R. M. Horne, Basic refractories for the open hearth, *Trans.*, Am. Inst. Min. Met. Eng., 62, 1920, pp. 90–112.

[2] C. Doelter, Mineralchemie, 2, pt. 1, 1914, pp. 374–383.

face; at this and several other localities in Crimea and Bosnia serpentine rocks are found in the vicinity, although the material itself is embedded in Quaternary or Tertiary beds. A different occurrence is that discovered in New Mexico,[1] on the upper Gila River, where the substance forms veins and balls in a Paleozoic cherty limestone. Here it is probably derived from a dolomitic carbonate.

## TALC AND SOAPSTONE[2]

**General Occurrence and Origin.**—Talc ($H_2Mg_3(SiO_3)_4$, or $3MgO.4SiO_2.H_2O$; 63.5 $SiO_2$, 31.7 MgO, 4.8 $H_2O$) is a hydrated magnesium silicate, but holds much less water than serpentine. It is a soft, crystalline, foliated or compact mineral of white, gray, or pale-green color and greasy feel. The more compact, as well as some impure, varieties are usually called soapstone; and contain, besides talc, more or less serpentine, chlorite, and amphibole.

Talc often contains 1 or 2 per cent of iron and aluminum, as well as a little calcium; according to the analyses given by Merrill[3] the soapstones contain, in addition to silica and magnesia, in per cent, 5 to 11 alumina, 7 to 13 ferrous oxide, and 1 to 4 lime; some of them contain so much water that a strong admixture of serpentine must be assumed. Traces of nickel and chromium are frequent in talc. The synthetic reproduction has not been successful.

By fusion, talc is transformed into enstatite ($MgSiO_3$); and it is also evident from the formula that this compound forms the basis for talc. Talc forms pseudomorphs after a large number of minerals, such as enstatite, diopside, olivine, actinolite, hornblende, chlorite, epidote, muscovite, and calcite. It is very often, at least, a product of hydration of magnesian rocks like gabbro, pyroxenite, and peridotite, or of crystalline limestone containing such pyrometasomatic minerals as enstatite, tremolite, or actinolite. The purest talc deposits are associated with such metamorphosed limestones or dolomites.

From any magnesian amphibole or pyroxene talc may be formed if they are acted upon by $H_2O$ and $CO_2$ according to the following reaction:

$$4MgSiO_3 + CO_2 + H_2O = H_2Mg_3Si_4O_{12} + MgCO_3$$

If iron and calcium be present, corresponding carbonates will be formed. It seems that the solutions active near the contacts of intrusive rocks were most potent to effect this change and that moderately high temperature furnished the best conditions. The direct derivation from dolomite

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[1] D. B. Sterrett, *Bull.* 340, U. S. Geol. Survey, 1908.

[2] C. Doelter, *op. cit.*, 2, pt. 1, pp. 356–374.

M. E. Wilson, Talc deposits of Canada, Econ. Geol., Ser. 2, Canada Geol. Survey, 1926.

[3] G. P. Merrill, Non-metallic minerals, 1910, pp. 208–216.

under influence of siliceous magmatic emanations is advocated by Wilson for some deposits in the Grenville limestone of Ontario (Fig. 145). Talc is thus a product of the later stages of pyrometasomatism; but the magnesia is chiefly, at least, derived from the rocks in which it occurs.

Talc may probably also be formed during dynamometamorphism, and perhaps in part by the action of the deeper ground-water, but the process outlined above has produced the largest deposits. Very rarely is talc found in ordinary ore deposits.

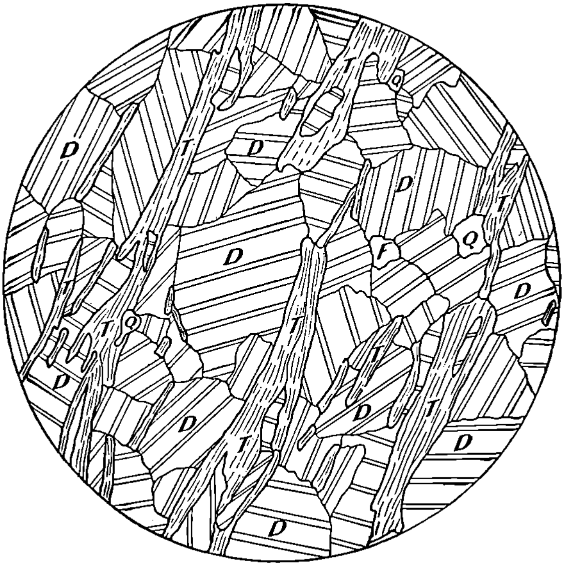

FIG. 145.—Drawing of thin section of talc-dolomite schist. *T*, talc; *D*, dolomite; *Q*, quartz; *F*, feldspar. Magnified 6.5 diameters. (*After M. E. Wilson.*)

E. Weinschenk,[1] in his description of the talc deposits of the Austrian Alps, holds that the mineral develops by replacement of schist composed of quartz, chlorite, chloritoid, and graphite along its contact with limestone and believes this transformation due to waters following the irruption of large igneous bodies. On the other hand, Heim[2] finds that certain talc deposits of Switzerland are formed from peridotites and serpentines under heavy dynamometamorphism.

**Occurrences.**—The crystalline rocks of all countries yield talc. Some occurrences are known from the Pacific coast, but the production in the United States is chiefly from the belt of ancient crystalline rocks which forms the axis of the Appalachian Mountain system from Canada to Alabama.

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[1] *Zeitschr. prakt. Geol.*, 8, 1900, pp. 41–44.

[2] A. Heim, *idem*, 26, 1918, pp. 9–10.

North Carolina is rich in talc, and one belt of Cambrian marble along the Nantahala Valley and Nottely River[1] yields many lenses as much as 200 feet long and 50 feet thick. The mineral is mined in open cuts and by shafts and tunnels.

New York and Vermont easily outrank all other States in the production of talc. The output of New York comes from a small district about 12 miles southeast of Gouverneur,[2] which has been worked for many years by underground methods. One mine at Talcville has attained a depth of 550 feet. The mineral occurs in schistose layers of enstatite, and tremolite, gradually merging into the surrounding crystalline limestone. The deposit forms a persistent layer, averaging 20 feet in width, within the enstatite-tremolite rock.

The talc deposits of Vermont[3] occur in sericitic gneiss or in serpentine. Gillson[4] finds that they are later than the serpentine and the allied basic intrusions of Paleozoic age extending from Blandford, Mass., to Quebec. The association of minerals is similar throughout. The minerals in order of deposition are: actinolite, magnesian chlorite, ferriferous dolomite and talc, and, lastly, magnetite and pyrrhotite. The deposits are regarded as replacements formed by emanations from the basic magmas, but it can hardly be considered as proved that all of the magnesia originated from the emanations.

Virginia yields most of the soapstone produced in the United States. It is derived from a belt nearly 30 miles long and less than one mile wide. The soapstone occurs as sheets or dike-like masses, 100 feet or more in thickness. Burfoot[5] concludes that the soapstone and talc are formed from pre-Cambrian pyroxenites and peridotites by the action of hot magmatic waters.

**Production and Uses.**—The production of talc in the United States was about 179,000 short tons in 1930. The larger part was sold in powdered form. The value of powdered talc is about $12 per ton.

Talc is used as a filler for paper (including wall paper), also for admixture or adulteration of pigment, as a heat insulator, lubricant, polishing powder, for toilet powders, and as an absorbent for nitroglycerine. The compact talc or soapstone is used for fire bricks, laboratory tables, gas burners, crayons, etc.[6]

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[1] Arthur Keith, *Bull.* 213, U. S. Geol. Survey, 1903, p. 443.
J. H. Pratt, North Carolina Geol. Survey, *Economic Paper* 3, 1900, p. 99.
J. H. Pratt, *Mineral Resources*, U. S. Geol. Survey, 1905, p. 1361.

[2] C. H. Smyth, Jr., *School Mines Quart.*, 17, 1896, pp. 333–341.

[3] E. C. Jacobs, Report Vermont State Geologist, 1913–1914, pp. 423–425.

[4] J. L. Gillson, Origin of the Vermont talc deposits, *Econ. Geol.*, 22, 1927, pp. 243–287

[5] J. D. Burfoot, Jr., The origin of the talc and soapstone deposits of Virginia, *Econ. Geol.*, 25, 1930, pp. 805–826.

[6] R. B. Ladoo, Talc and soapstone, *Bull.* 213, U. S. Bur. Mines, 1923.

## ASBESTOS[1]

**Amphibole Asbestos.**—The asbestos of mineralogy is a monoclinic amphibole which develops in seams and slips in normal amphibolitic rocks, especially where the rocks have been subjected to pressure and movement. Chemically it is a calcium-magnesium metasilicate. According to the series of analyses given by Merrill the silica varies from 52 to 58 per cent, the lime from 12 to 16 per cent, the magnesia from 20 to 30 per cent. Other constituents are alumina, varying from 1 to 6 per cent, and ferrous oxide, usually from 1 to 6 per cent, though in some cases considerably higher. Water is always present, the amount generally varying between 2 and 5 per cent. The water appears to be an essential constituent and the mineral is really a hydrated form of tremolite or actinolite. The extinction angle appears, however, to be that characteristic of these amphiboles, or about 18°.

Anthophyllite (rhombic, $(Mg,Fe)SiO_3$), and crocidolite (monoclinic, $NaFe(SiO_3)_2$, $FeSiO_3$), a dark-blue sodium amphibole, also yield asbestiform varieties.

Merrill has shown that the fibers are polygonal in outline and run out into needle-like points; down to a diameter of 0.002 or 0.001 millimeter the fibers retain their uniform diameter and polygonal outlines. The color of amphibole asbestos is usually white to greenish white. Only the finer kinds are utilized, but even these are more brittle than the serpentine asbestos. The mineral is classed as slip-fiber or cross-fiber, according to the position of the fibers in the veinlets. The radial or divergent structures are designated as mass-fiber.

Most of the small quantity of asbestos mined in the United States is anthophyllite.

There are many occurrences along the Appalachian Mountain system, from Vermont to Alabama. One of the most important localities worked is in Habersham County, Georgia, where, according to Hopkins,[2] the anthophyllite asbestos is derived from the alteration of a peridotite. Almost the entire domestic production is derived from Georgia, mainly of the mass-fiber variety.

Crocidolite[3] asbestos is extensively mined in Griqualand West, Cape Colony. It occurs in thin cross-fiber layers, interbedded with jaspers, slate, and ironstone in a sedimentary series. Its origin is in doubt.

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[1] G. P. Merrill, *Proc.*, U. S. Nat. Mus., 18, 1895, p. 181. *Bull.* Geol. Soc. Am., 16, 1905, p. 113. Non-metallic minerals, 1910, pp. 183–197.

F. Cirkel, Chrysotile-asbestos, Canada Dept. Mines, Mines Branch, 1910, 316 pp.

[2] O. B. Hopkins, *Bull.* 29, Georgia Geol. Survey, 1914.

[3] A. L. Hall, Asbestos in the Union of South Africa, *Mem.* 12, Geol. Surv. South Africa, 1918; also *Trans.*, Geol. Soc. South Africa, 21, 1918, pp. 1–26.

M. Peacock, Amphibole asbestos in South Africa, *Am. Mineralogist*, 13, 1928, pp. 241–286.

Like glaucophane it might be considered a product of metamorphism, though Hall expressly excepts contact metamorphism. A brown silicified variety is used as a precious stone under the name of Tiger's eye. "Amosite" is a similar amphibole, named by Hall, containing more ferrous iron and less soda and magnesia.

**Serpentine Asbestos (Chrysotile).**—Chrysotile asbestos is green or yellowish-green and is easily reduced to a white fluffy state. The fiber is short, but of very uniform diameter and great divisibility and flexibility. The decomposing effect of hydrochloric acid distinguishes it from amphibole asbestos. In composition it is practically identical with the purer kinds of serpentine. A typical analysis of the Canadian material yielded per cents as follows: 42 $SiO_2$, 42 $MgO$, 14 $H_2O$, 1 $FeO$, and 1.7 $Al_2O_3$.

This variety is found as veinlets, rarely over 6 inches thick, in serpentine or peridotite, and has almost always a cross-fiber—that is, the silky fibers lie perpendicularly to the plane of the veinlet.

The pure yellowish-green serpentine which occurs in contact-metamorphic limestone and which is an alteration product of diopside sometimes contains chrysotile of exceptionally high grade. A deposit of such material is now worked in Arizona, northeast of Globe.[1]

Chrysotile veinlets may be found in almost any serpentine area, but are rarely so abundant and large as to be of economic importance. The views regarding their origin differ.

Dresser shows that serpentinization in the Canadian deposits proceeded along irregular cracks in the peridotite, and that the chrysotile veinlets are found in the center of the serpentinized bands. These veinlets were interpreted by Pratt and Merrill as fillings of contraction cracks, but other authors consider them the result of a recrystallization of the serpentine, proceeding outward from the cracks. Along the margins definite evidence of replacement is often observed.

In conclusion, the asbestiform products are mainly formed by aqueous agencies from older magnesium minerals under conditions of fairly high temperature; and often, it would seem, their development followed soon after the intrusion of the original rocks.

S. Taber[2] believes that all cross-fiber veins are formed by a process of lateral secretion, the fibers being supplied with a supersaturated solution at their base and the growing veins pushing aside the enclosing walls. Since, however, the material in the veins is derived from the serpentine itself it is not apparent why there is any need of increase of volume.

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[1] A. M. Bateman, An Arizona asbestos deposit, *Econ. Geol.*, 18, 1923, pp. 663–683. E. D. Wilson, Asbestos deposits of Arizona, *Bull.* 126, Arizona Bur. Mines, 1928, 100 pp.

[2] S. Taber, The genesis of asbestos, *Trans.*, Am. Inst. Min. Eng., 57, 1918, pp. 62–98.

Bain and Keith,[1] in important papers, have carefully investigated the Vermont deposits. They find chrysotile veins in fresh olivine rock and conclude that they are filled torsion and crush cracks (Fig. 146). Serpentine and olivine must be removed to produce the chrysotile veins in the same intrusive body.

Since 1895 Canadian mines have supplied this country. The Canadian deposits[2] center in Asbestos Hill at Thetford, in the eastern townships of Quebec. As stated, the mineral occurs as irregular veinlets in serpentine and peridotite. These rocks are in places accompanied by somewhat later gabbro and granite, and all are intrusive into Ordovician sediments. The mineral is mainly mined in open pits, one of which, for instance, is 700 feet long, 200 feet wide, and 165 feet in greatest depth. A small percentage is obtained by hand cobbing, but the larger part—30 to 60 per cent—of the crude material quarried is crushed and screened, and the fibers separated by air currents.[3] The extraction of fiber is from 6 to 10 per cent of the rock milled.

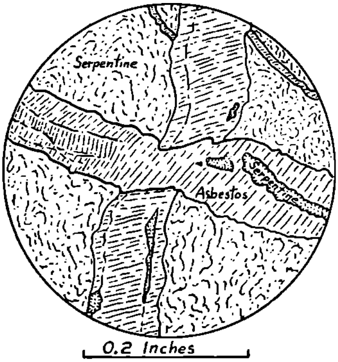

FIG. 146.—Camera lucida drawing of intersecting asbestos veins, from Moretown, Vermont. (*After S. B. Keith and G. W. Bain.*)

Of late years, the deposits in southern Rhodesia in the Bulawayo district and in the Transvaal in the Barberton district are becoming important. Other producing regions are in the Ural mountains, in Italy, and on the island of Cyprus.

In the United States[4] chrysotile of economic importance occurs in Vermont, near Caspar, Wyoming, and in Arizona. Thus far the production is small.

**Uses.**—The fundamental property of asbestos, upon which its use depends, is its flexible, fibrous structure; but coupled with this are the scarcely less important qualities of incombustibility and slow conduction of heat and electricity when the mass is fiberized and porous. The spinning and weaving of fireproof cloth form an important part of the asbestos industry carried on in the United States with Canadian raw

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[1] S. P. Keith and G. W. Bain, Chrysotile asbestos, *Econ. Geol.*, 27, 1932, pp. 169–188.

G. W. Bain, *idem*, pp. 281–296.

[2] J. A. Dresser, *Econ. Geol.*, 4, 1909, pp. 130–140.

J. A. Dresser, Preliminary report on the serpentine, etc., of southern Quebec, *Mem.* 22, Canada Geol. Survey, 1913, p. 103.

[3] W. A. Ru Keyser, Asbestos mining and milling in Quebec, *Eng. Min. Jour.-Press*, April, 15 and 22, 1922.

[4] J. S. Diller, *Bull.* 470, U. S. Geol. Survey, 1910, pp. 505–524.

material. The highest grade of the crude mineral is expensive, costing \$250 to \$400 per ton; the fines cost \$25 to \$125 per ton, while the lowest grade—a mixture of serpentine and asbestos—is sold at less than \$1 per ton. Amphibole asbestos is much cheaper, costing about \$10 to \$18 per ton. Crocidolite is more easily fusible but is more resistant than chrysotile to acids and sea water. The London price is about \$125 per ton.

The Canadian production in 1930 was 219,641 short tons; almost the whole production was exported to the United States. In the same year the output in the United States was 3,653 short tons of chrysotile asbestos, from Arizona and Vermont, valued at \$75 per ton and 589 short tons of amphibole asbestos at about \$27 per ton.

## THE ALUMINOUS DEPOSITS

**Pyrophyllite.**[1]—Pyrophyllite is a hydrous silicate of alumina ($H_2Al_2(SiO_3)_4$), containing, in per cent, 66.7 $SiO_2$, 28.3 $Al_2O_3$, and 5.0 $H_2O$. In composition and physical qualities it is similar to talc, though it does not command so high a price as the best talc. It is mined in the Deep River region, North Carolina, where it occurs in thick beds associated with schistose volcanics of pre-Cambrian age. The mineral is believed to be a late product of heated waters near intrusive rocks and formed by replacement of quartz and silicates. The alumina is derived from surrounding rocks.

**Anhydrous Aluminum Silicates.**[2]—Andalusite, sillimanite, and cyanite, all of them having the composition $Al_2SiO_5$($Al_2O_3$ 63.2 per cent), are the products of high-temperature metamorphism of sediments. A use for these minerals for refractories and spark plugs has been developed in the last few years. While frequent in disseminated form, the minerals rarely occur in large masses; but such high-grade deposits of andalusite have recently been found in the pre-Cambrian rocks of the Inyo Range, California.[3] Cyanite has been mined near Ogilby in southern California and also in Virginia and North Carolina. Large deposits occur in India,

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[1] C. H. Clapp, Alunite and pyrophyllite . . . in B. C., *Econ. Geol.*, 10, 1915, pp. 70–88.

J. L. Stuckey, On pyrophyllite . . . of North Carolina, *idem*, 20, 1925, pp. 442–463.

A. F. Buddington, Pyrophyllitization, etc., at Conception Bay, N. F., *Jour. Geol.*, 24, 1916, pp. 130–152.

[2] R. B. Ladoo, Non-metallic minerals, New York, 1925.

A. V. Petar, Sillimanite, etc., U. S. Bur. Mines, *Information Circ.* 6255, 1930.

F. H. Riddle, Sillimanite, *Eng. Min. Jour.*, March, 1932, pp. 141–142.

Dumortierite, a blue or purple borosilicate of aluminum, is sometimes used mixed with andalusite. A deposit is worked near Oreana, Nevada. It is affiliated with pegmatites.

[3] A. Knopf, *Jour.* Washington Acad. Sci., 7, 1917, pp. 549–552.

A. B. Peck, *Am. Mineralogist*, 9, 1924, pp. 123–129.

at Singhbum. Large deposits of sillimanite are reported from Assam, India.[1] Several thousand tons have been mined in the last few years.

Artificial sillimanite, mullite ($3Al_2O_3.2SiO_2$), is now manufactured in electric furnaces in the United States. The three minerals mentioned above change to mullite upon heating to 1400° C.

**Alunite.**—The occurrence of alunite ($K(AlOH)_3.(SO_4)_2.3H_2O$) has already been referred to on page 341. It is used for the manufacture of potassium and aluminum salts, and is a product of hydrothermal solutions, or descending surface waters acting on aluminous and potassic rocks at moderate temperature.

## BARITE[2]

**Modes of Occurrence.**—Barite, the sulphate of barium, also known as barytes or heavy spar, contains when pure in per cent 65.7 BaO and 34.3 $SO_3$. It fuses at 1580° C. It is usually white and coarsely crystalline with curved cleavage faces but appears also, especially in residual deposits, with granular, earthy, or even fibrous texture. Many barites contain from a fraction to several per cent of strontium sulphate. The material mined is often quite pure except for small amounts of silica, calcite, gypsum, kaolin, and iron hydroxide.

Witherite, the barium carbonate, is a much rarer mineral and is found in barite veins associated with galena. It occurs rather abundantly in such veins in Cumberland and Northumberland in England, particularly at the New Brancepeth colliery where it is secondary after barite. Allstonite and baryto-calcite are related minerals.

Barite is not a mineral of igneous rocks, nor does it occur in contact-metamorphic deposits. It is common, however, as lenses and veins in almost all kinds of rocks. It is also formed during sedimentation. In general, it is a mineral of aqueous origin.

All igneous rocks contain at least a trace of barium oxide but rarely more than 0.1 per cent. Leucite and analcite rocks from the Wyoming-Montana province are unusually rich in this metal, some analyses showing from 0.5 to 1.2 per cent of BaO, probably present in the feldspathoids as a silicate. Granites, rhyolites, andesites, and basalts are poor in barium. In more concentrated form barium is often present in many sediments though rarely in limestone. Ordovician limestones from Missouri contain, according to Steel, only 0.001 to 0.005 per cent BaO. Dickson failed to find it in many limestones. F. W. Clarke does not mention its presence in single and composite analyses of limestone in his book on Geochemistry. Some sandstones contain barite; it has also been found in shales with sedimentary manganese ores (p. 281). Sea water contains

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[1] Edwin Pascoe, *Records*, Geol. Survey India, 61, pt. 1, 1928, pp. 68–71.

[2] R. M. Santmeyers, Barite and barium products, U. S. Bur. Mines, *Information Circ.* 6221, 6223, 1930.

traces of barium and strontium and many natural waters, particularly salt brines, hold quite a little barium in solution as chloride or carbonate. Many cases of deposition of barite by hot springs or by normal meteoric waters are mentioned on pages 65–78.

Barite is soluble in water to the extent of 2.7 milligrams per liter at 18° C.; it is somewhat more soluble, probably with decomposition, in waters containing alkaline carbonates and chlorides.

Barite is a common gangue mineral in many ore deposits but is here scarcely ever of economic importance. In some cases it may have been dissolved from the rocks traversed by the ascending solutions. The barite deposits worked generally contain few other minerals and occur in sedimentary rocks of all ages as veins and lenses whose width in places may be from 10 to 50 feet. Sometimes a little quartz, fluorite, pyrite, galena, or sphalerite is associated with the barite.

The larger part of the barite mined in the United States is a residual mineral, forming lumps in clay resulting from the decay of limestone or other rocks containing barite veins.

**Deposits in the United States.**—The barite deposits now worked are mostly contained in the Paleozoic limestones in the southern Appalachian and the central states; the order of importance being Georgia, Missouri, Tennessee, Kentucky, and Virginia.

In Missouri barite often accompanies the zinc and lead deposits, but the important deposits are found in a separate area in Washington County, in southeastern Missouri,[1] not far from the great lead mines in the Bonneterre (Cambrian) dolomite (p. 439). The principal deposits are found in the Potosi, Proctor, and Gasconade dolomites (Ordovician) as filling of irregular veins and other open cavities. The order of precipitation is given by Steel as follows: A thin coating of chalcedony was first deposited; this was followed by quartz, pyrite, and marcasite, later galena and sphalerite, lastly the main barite filling. In places there is a later coating of ruby-red sphalerite on the older barite. The barite mined is derived from residual masses in the red clay covering the formations.

In Georgia barite occurs in the Cartersville district[2] as deposits from solution in fractures and cavities in the Weisner quartzite in intimate

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[1] C. W. Dickson, The concentration of barium in limestone, *School Mines Quart.*, 23, 1902, pp. 366–370.

A. A. Steel, *Trans.*, Am. Inst. Min. Eng., 40, 1910, pp. 85–117.

W. A. Tarr, *Econ. Geol.*, 14, 1919, pp. 46–67.

W. M. Weigel, The barite industry in Missouri, *Trans.*, Am. Inst. Min. Met. Eng., Yearbook, 1929, pp. 256–279.

[2] T. L. Watson and J. S. Grasty, Barite of the Appalachian States, *Trans.*, Am. Inst. Min. Eng., 51, 1916, pp. 514–559.

J. P. D. Hull, Barite deposits of Georgia, *Bull.* 36, Geol. Survey Georgia, 1920.

Albert H. Fay, Cartersville, *Eng. Min. Jour.-Press*, April 2, 1927.

association with yellow ocher, and as fragments embedded in residual clays.

The barite deposits of Virginia have been described by T. L. Watson.[1] Deep rock decay characterizes the whole region. The barite in part fills fractures and in part replaces limestone. It occurs:

1. In crystalline Cambrian or pre-Cambrian limestone as irregular, lenticular lodes, or pockets, replacing the limestone and associated with calcite and chalcopyrite (Fig. 147).

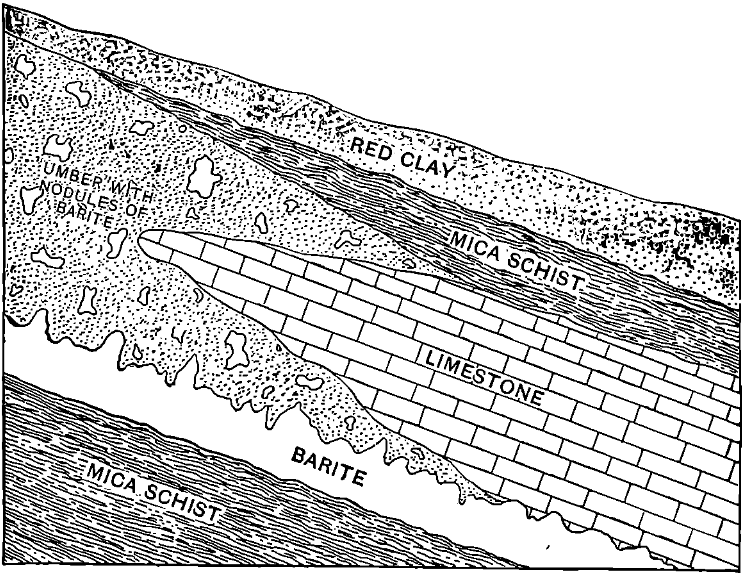

FIG. 147.—Section of the Bennett mine, Virginia, showing occurrence of barite as residual and as replacement deposit. (*After T. L. Watson.*)

2. In crystalline schists as the filling of fractures.

3. In the Shenandoah (Cambro-Ordovician) limestone as the filling of fractures or in residual soil.

4. In Triassic shales and limestone as the filling of fissures in a crushed zone.

Large and pure barite veins have been described from the Ketchikan district[2] and near Wrangell[3] in Alaska. They are contained in crystalline limestone and in schists. Witherite has been found at a few localities in the United States. At El Portal,[4] California, a witherite deposit has

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[1] *Trans.*, Am. Inst. Min. Eng., 38, 1907, pp. 953–976.

T. L. Watson and J. S. Grasty, *op. cit.*

[2] T. Chapin and G. H. Canfield, *Bull.* 642, U. S. Geol. Survey, 1916.

[3] E. F. Burchard, *Bull.* 592, U. S. Geol. Survey, 1914, pp. 109–117.

[4] A. A. Fitch, Barite and witherite from near El Portal, Mariposa County, California, *Am. Mineralogist*, 16, 1931, pp. 461–468.

For milling processes at this place see G. L. Young, *Eng. Min. Jour.-Press*, 130, 1930, p. 70.

been worked for some time. Veins up to 20 feet wide replace limestone, with development of andradite, hedenbergite, etc., near a granite intrusion and of witherite and barite farther away, the minerals replacing calcite. Here apparently witherite is a primary mineral, and in part it is replaced by barite.

**Foreign Deposits.**—Barite deposits are common in all countries. The abundant barite veins in the Northumberland region and in Derbyshire contain more or less fluorite and galena and are associated with basic sheets (Whin Sill).[1] Bodies of exceptional size and purity are found in central Germany[2] in sedimentary rocks of Permian and Triassic age, the barium being probably derived from brines from the underlying Permian salt deposits. One of the largest barite deposits of the world is located at Meggen, where it occurs as replacement of a limestone bed in folded Devonian shales (Fig. 106).

Witherite is also produced from a vein, with some sulphides, in Russian Turkestan.[3]

**Origin.**—Barite is to some extent formed during sedimentation, but the important deposits are either residual masses formed during weathering of rocks containing barite, or veins deposited by ascending waters. In the latter case the barite may be brought up in the emanations from intrusive magmas and occurs mainly in the mesothermal and epithermal veins. In other cases the evidence seems strong that pure barite veins may have been formed by leaching of rocks containing barium, or by reaction of salt brines carrying barium carbonate or chloride with other waters of a sodium sulphate or calcium sulphate type.

**Uses and Production.**—Barite is an important mineral largely on account of its inertness and weight and is used as a pigment in the manufacture of mixed paint (lithopone: barite and zinc sulphide) and to give weight to paper. It is the raw material for other barium salts, such as the nitrate which is used in pyrotechnics for green fire. For most of the purposes indicated its purity and white color are essential. The crude material is crushed and treated in log washers and jigs. After grinding the pulp is classified and the settled cream-colored mud is finally treated with sulphuric acid to remove the staining ferric hydrate. The domestic production in 1930 was about 237,505 short tons, which came from a great number of small operators in Missouri, Virginia, Kentucky, Georgia, North Carolina, and Tennessee. About 50,000 tons of barite and barium salts are imported annually, largely from England and Ger-

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[1] Maps of both districts in J. M. Sweet, Notes on British barytes, *Mineral. Mag.*, 22, 1930, pp. 257–270.

[2] R. Bartling, Die Schwerspathlagerstätten Deutschlands, Stuttgart, 1911.
A. W. Staub, *Zeitschr.* deutsch. geol. Gesell., Abhand. 80 (1), 1928, pp. 43–96.

[3] Annotated Bibliography of Economic Geology, 4, pt. 2, 1932, No. 540.

many. The average price of barite is $7.50 per ton. The carbonate (witherite) is used for paints, glass, and enamels.

## CELESTITE AND STRONTIANITE

**Occurrence.**—Strontium accompanies barium as a primary constituent of igneous rocks but is present in much smaller quantities. As celestite ($SrSO_4$) and strontianite ($SrCO_3$) it is sometimes found in fissure veins of hydrothermal origin, but the two minerals are much more commonly found as veins, nodules, and layers in sedimentary rocks, particularly limestone and anhydrite.

Celestite is found in crystals and granular masses often of bluish color; but sometimes it is dark or brownish. It usually, but not always, contains $BaSO_4$. Strontianite is crystalline, fine-grained, fibrous, or nodular and has a white, brownish, or dark color. It has often been mistaken for calcite, and always contains a few per cent of $CaCO_3$.

In geodes, veins, disseminations, and replacements celestite is found in Paleozoic dolomite and limestone of Michigan,[1] New York,[2] and Ohio. A cave at Put-in Bay, Ohio, is said to have yielded 150 tons of celestite. Deposits in the limestone quarries of northwestern Ohio and southeastern Michigan, near Toledo, also at Bagot, Ontario, Canada, are said to be of possible economic importance. Celestite also occurs in Cretaceous limestone in Texas[3] near Austin. Interesting deposits of celestite have been discovered in Tertiary lake beds in Arizona,[4] near Gila Bend, where the mineral occurs in sandstone and shale with gypsum and salt; similar beds have been described from the Awavatz Mountains near the southern end of Death Valley. Strontianite is of subordinate importance. Strontianite deposits in Miocene lake beds 10 miles north of Barstow,[5] San Bernardino County, have been discussed by A. Knopf who believes that the beds, veins, and fibrous concretions at this place are replacements of lacustrine limestone.

All these deposits are of low grade. In 1916, 250 short tons of strontium ore were mined in the United States. In 1917 the output had increased to 4,035 short tons. There has been no production of late years. About 2,000 tons of strontium ore and chemicals are imported annually. The value of crude celestite is about $8 per ton.

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[1] E. H. Kraus and W. F. Hunt, *Am. Jour. Sci.*, 4th ser., 21, 1906, p. 237.
W. A. Sherzer, *idem*, 3d ser., 50, 1895, p. 246; *Rept.*, Michigan Geol. Survey, pt. 1, 1900, p. 208.

[2] E. H. Kraus, *Am. Jour. Sci.*, 4th ser., 18, 1904, p. 30; 19, 1905, p. 286.

[3] First report, Geol. Survey Texas, 1889, p. 125.
F. L. Hess, *Eng. and Min. Jour.*, July 17, 1909, p. 117.

[4] W. C. Phalen, Celestite deposits in California and Arizona, *Bull.* 540, U. S. Geol. Survey, 1914, pp. 526–531.

[5] Strontianite deposits near Barstow, California, *Bull.* 660, U. S. Geol. Survey, 1918, pp. 257–270

The principal supply of strontium was for many years derived from the strontianite veins in Cretaceous marl and limestone of Westphalia, Germany; the mineral is here accompanied by calcite and a little pyrite.[1]

The strontium used in the United States before the World War came largely from England. Important celestite deposits are found near Bristol, where the mineral forms lenses and veins in a Triassic marl and in the underlying rocks.[2]

Celestite is also concentrated, like barite, during the formation of sulphur from gypsum and, as noted above under Sulphur, occurs in considerable quantities in the sulphur mines of Sicily; also in the anhydrite of the salt-dome cap rocks. For commercial purposes celestite should contain at least 95 per cent $SrSO_4$.

The principal use of strontium is in sugar refining, in the so-called Scheibler process, in which strontium hydroxide is used for the recovery of sugar from beet molasses. The nitrate is used in pyrotechnics for red fire.

**Origin.**—It is evident that celestite and strontianite are concentrated, probably from surrounding rocks, by meteoric waters,· this being frequently observed in sediments of various kinds. K. Dinger[3] believes that strontium carbonate was precipitated with calcium carbonate in the sea, and that later solutions carried it into gypsiferous beds where it was re-precipitated as celestite. These minerals also occur in veins, associated with sulphides and gold (*e.g.*, at Cripple Creek, Colorado), and in such cases they may well have originated in magmatic emanations.

## ORES OF COPPER, LEAD, VANADIUM, AND URANIUM IN SANDSTONE AND SHALE

**General Features.**—Ores of copper, lead, vanadium, and uranium are found disseminated in sandstones and shales far from igneous rocks. The sedimentary strata containing the ores are usually parts of thick series of terrigenous or shallow-water beds, frequently of reddish color. The ores are of low tenor and can be utilized only in exceptional cases. Nevertheless this class of deposits presents many interesting features.

The ore minerals are chalcocite, roscoelite (a vanadium mica), various copper and lead vanadates, carnotite (a vanadate of uranium, etc.). Bornite, chalcopyrite, galena, and pyrite are less common. The ores frequently carry small amounts of silver, nickel, cobalt, chromium, molybdenum, and selenium. Gangue minerals occur sparingly and are usually confined to a little barite, calcite, and gypsum. The outcrops are likely to be brilliantly colored by malachite and azurite. While

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[1] Götting, *Oesterr. Zeitschr. Berg.-u. Hüttenwesen.*, 37, 1889, p. 113.

[2] R. L. Sherlock, *Mem.*, Geol. Survey England, Special reports on mineral resources, 3, 1918, pp. 48–61.

[3] *Chemie der Erde*, 4 (2), 1929, pp. 167–177.

the deposits are confined to certain formations or members, they do not continuously follow a particular horizon and give no evidence of being of sedimentary origin. They often appear in fractured and brecciated beds or in strata rich in carbonaceous matter and plant remains. More rarely the ores follow distinct fissures in the sedimentary rocks. They do not seem to have any genetic relation with thermal springs. There is no reason why the deposits should be confined to any particular geological age; but, as a matter of fact, almost all of them are in the Upper Carboniferous, Permian, Triassic, or Jurassic.

**Origin.**—In considering the class as a whole it appears that igneous agencies had no part in the genesis. The ores are assuredly epigenetic and their universal appearance in land or shallow-water beds is significant. In all probability these ores have been concentrated by meteoric waters which leached the small quantities of metals disseminated in the strata. The sediments were rapidly accumulated, under arid conditions, from adjacent land areas and the metals were probably carried down in fine detritus and in solutions from older ore deposits.

The waters which concentrated the ores are believed to have been mainly sodium chloride and calcium sulphate solutions containing sulphates and, perhaps, chlorides of copper and lead. The mineral association and geological features indicate deposition at low temperature, probably well below 100° C., and at shallow depths, but below the zone of direct oxidation. Very likely these ores have been forming continuously since the establishment of active water circulation in the beds.

**European Occurrences.**[1]—The European occurrences are confined to the Permian and the Triassic, both, generally speaking, ages of arid climate and saline deposits.

The Russian Permian, extending far west from the Urals, consists in in its lower division of sandstones, marls (in part marine), and conglomerates. The sandstones are rich in vegetable remains. Copper ores are found over wide areas, but have not been worked extensively of late. The average tenor is said to be 0.9 per cent metallic copper. The chalcocite ores replace plant remains and tree trunks or the cement of the sandstones. The minerals mentioned from this locality are (besides secondary malachite and azurite) chalcocite, chalcopyrite, barite, vanadinite, and volborthite (vanadate of copper and calcium).

Much interest has been taken in the copper deposits of the Khirgiz Steppes,[2] between the Urals and the Altai, in the Karkaralinsk and

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[1] For an excellent review and complete index of literature, in part difficultly accessible, see Stelzner and Bergeat, Die Erzlagerstätten, 1, 1904, pp. 388–439.

[2] A. Addiassewich, A journey to central Asia, *Trans.*, Inst. Min. and Met. (London), 17, 1907–1908, pp. 498–522.

J. MacIntosh Bell, The Spassky and Atbasar copper mines, *Min. Sci. Press*, May 22 and June 5, 1920.

Akmolinsk districts. Very rich copper ores, in part oxidized and carrying pyrite, chalcopyrite, bornite, and chalcocite, have been found here in sandstones of Permo-Carboniferous age. At Nankat, west of Kokand in Turkestan, deposits of metallic copper have been discovered in sandstones and gypsiferous marls of Tertiary age, fossil wood and chalcocite are also found.[1] Some authors consider a telemagmatic genesis possible.

Over a large part of western Europe the Triassic is copper-bearing, and together with the copper more or less lead is found.

In England, at Alderley Edge and Mottram St. Andrews,[2] south of Manchester, copper ores have been mined. They occur in the cement of Triassic sandstones and conglomerates and consist of copper carbonates, galena, pyromorphite, and vanadinite, also some barite, manganese, and cobalt. The ores are said to contain at most 1.4 per cent copper. The mineral mottramite, a vanadate of copper and lead, was discovered at this place and vanadium was extracted from the ores.

In the "Buntsandstein" of the Triassic in Prussia and Lorraine, near Saarlouis and other places, a formation known as the Voltzia sandstone is particularly rich in lead and copper ores, which at times have been mined. The bed contains abundant plant remains. The minerals are cerussite, galena, chalcocite (?), and copper carbonates.

The best-known deposits of the European Triassic are those of Commern and Mechernich, not far from Aix-la-Chapelle, Rhenish Prussia. Lead ores have been mined here for several hundred years, but it is reported that the mines may soon be closed. The ores are of low grade and are mined in open cuts by removing about 130 feet of overburden. In 1903 the ores averaged 1.5 per cent lead. The ore minerals are galena and cerussite, with a little chalcopyrite and barite, the latter filling veins and veinlets in the sandstone. Small amounts of silver, nickel, and cobalt are present. The thickness of the ore-bearing sandstone is about 20 meters. The general occurrence of the galena in so-called "Knoten" or knotty concretions is very remarkable. They often enclose several sand grains and some of them are bounded by the crystal faces of the galena. The epigenetic character of the ores is beyond doubt, but whether they really belong to this division is less certain. Beyschlag regards them as impregnations from hydrothermal veins.

**American Occurrences.**[3]—On the North American continent copper ores are widely distributed in the "Red Beds" of the southwest, in Texas,

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[1] R. Beck, Lehre von den Erzlagerstätten, 2, 1909, p. 172.

[2] Phillips and Louis, Ore deposits, 1896, pp. 266–269.

E. B. Papenfus, *Econ. Geol.*, 26, 1931, p. 323.

[3] E. J. Schmitz, Copper ores in the Permian of Texas, *Trans.*, Am. Inst. Min. Eng., 26, 1896, pp. 97–108.

S. F. Emmons, Copper in the Red Beds, *Bull.* 260, U. S. Geol. Survey, 1905, pp. 221–232. W. H. Emmons, The Cashin mine, *Bull.* 285, *idem*, 1906, pp. 125–128.

E. P. Jennings, *Trans.*, Am. Inst. Min. Eng., 34, 1904, p. 839.

Oklahoma, New Mexico, Arizona, Colorado, Wyoming, Utah, and Idaho but, though conspicuous by green and blue colors, rarely prove of economic importance.

The ore occurs in arkosic sandstone, conglomerate, or clay shale and is usually associated with plant remains and fossil wood. These strata were accumulated in shallow seas or as subaerial deposits by a process of rapid degradation of adjacent land areas of the Rocky Mountain region, and have been referred to the Upper Carboniferous, Permian, Triassic, and Jurassic.

In Texas the copper-bearing beds appear over large areas in Permian sandstones and shales. They lie at several horizons, in strata rich in plant remains; covellite, chalcocite, chalcopyrite, and pyrite are the minerals mentioned. The area is said to extend from 33° to 34° in latitude and from 98° to 100° in longitude.

In Oklahoma nodules of chalcocite are found in red shales and sandstones of the same age. Fossil wood is often converted to chalcocite, sometimes with a shell of chalcopyrite; unusually high silver assays of 31 ounces per ton and traces of gold are reported.

In Colorado these ores have been recorded at several places, notably at Red Gulch, Fremont County, where Lindgren observed nodules of chalcocite with barite in black carbonaceous shale; sections show that the copper sulphide actually replaces the coal and shale. The horizon is probably the uppermost Carboniferous. North of the Colorado River they appear in the Carboniferous of the Kaibab Plateau. S. F. Emmons believed that the copper in the oxidized ore and chalcocite in the Aubrey limestone near Grandview, Arizona, was leached from the "Red Beds" and carried down into the limestone.

In southwestern Colorado copper, often accompanied by vanadium ores, is widely distributed in the Jurassic La Plata sandstone. W. H. Emmons has described the Cashin vein in this formation near Placerville. The ores are here argentiferous chalcocite, covellite, and bornite, with some calcite. No igneous rocks are present and Emmons believes that the ores were leached from the "Red Beds." There is an active circula-

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H. W. Turner, *idem*, 33, 1903, p. 678.

W. Lindgren, L. C. Graton, and C. H. Gordon, The ore deposits of New Mexico, *Prof. Paper* 68, U. S. Geol. Survey, 1910.

H. S. Gale (Idaho), *Bull.* 430, *idem*, 1909, pp. 112–121.

W. Lindgren (Colorado), *Bull.* 340, *idem*, 1907, pp. 170–174.

W. A. Tarr (Oklahoma), *Econ. Geol.*, 5, 1910, pp. 221–226. A. E. Fath (Oklahoma), *idem*, 10, 1915, pp. 140–150. L. M. Richard (Texas), *idem*, 10, 1915, pp. 634–650.

A. F. Rogers, Origin of copper ores of the "Red Bed" type, *idem*, 11, 1916, pp. 366–380.

B. S. Butler *et al.*, Ore deposits of Utah, *Prof. Paper* 111, U. S. Geol. Survey, 1920, pp. 152–158.

tion of water in the formation, and springs with salt, sulphates, and hydrogen sulphide abound. A production of about 300,000 ounces of silver and 700,000 pounds of copper is recorded from this mine.

The greatest development of the copper-bearing sandstones is in New Mexico; considerable production from picked ore has been achieved at the Nacimiento deposits, in the northern part of the State, where the Triassic "Red Beds" rest on pre-Cambrian granitic rocks which contain much older copper deposits. According to Schrader,[1] most of the copper ores occur in the basal beds and are confined within a thickness of 25 feet in a reddish-white sandstone rich in fossil wood, which is largely chalcocitized. A tree trunk 60 feet long with a basal diameter of 2½ feet is mentioned, which was almost wholly converted to chalcocite. Some barite occurs.

According to the same geologist, the copper-bearing beds of the Zuni Mountains, in northeastern New Mexico, lie at the base of the "Red Beds," resting on pre-Cambrian gneisses which contain older copper veins. The sandstones, shales, and marls for 30 to 60 feet above the base of the beds contain oxidized ores and chalcocite replacing wood.

Graton describes the ores from the Tecolote district, San Miguel County, which are partly in the "Red Beds" of the Upper Carboniferous (Abo formation), and partly at a higher horizon, the Dakota sandstone (?). The calcareous cement of the arkose is replaced by chalcocite, bornite, chalcopyrite, and pyrite, the replacement extending into the feldspar grains.

In the Oscura Range, also in New Mexico, red sandstones, probably Carboniferous, contain chalcocite, bornite, and chalcopyrite, in part as replacement of fossil wood. Turner mentions the occurrence of the Triassic *Podozamites crassifolia,* the same cycad found in the deposits at Abiquiu, first studied by Newberry.

Graton believes, contrary to Turner, that the copper ores have been introduced into the strata by hot waters ascending along a number of dislocations in the sandstone.

Finally, H. S. Gale describes copper ores from southern Idaho which occur in the Ankareh maroon shales and sandstones (Triassic or Carboniferous) (see Fig. 15).

There are then at least two main cupriferous formations in the Southwest—(1) the Upper Carboniferous "Red Beds," equivalent to the Permo-Carboniferous, or the Abo formation; (2) the undoubtedly Jurassic La Plata sandstone.

The silver deposits in the supposedly Triassic sandstones of Silver Reef,[2] in southern Utah (Harrisburg district), which created a boom

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[1] F. C. Schrader, *Prof. Paper* 68, U. S. Geol. Survey, 1910, pp. 141–149.

[2] C. M. Rolker, The silver sandstone district of Utah, *Trans.*, Am. Inst. Min. Eng., 9, 1881, pp. 21–33.

B. S. Butler, *Prof. Paper* 111, U. S. Geol. Survey, 1920, p. 582.

about 1880, are now worked only on a small scale. The ores were silver chloride above the water level, and native silver and argentite in depth; copper was also present, and selenium is reported. Plant remains were distinctly abundant. A secondary concentration from a primary argentiferous chalcocite is the probable genesis. Butler, who has written the best account of the district, points out that the ores are later than the Leeds anticline in which they occur. He rather doubts the derivation of the silver minerals from chalcocite and thinks that the genesis of the deposit is still uncertain.

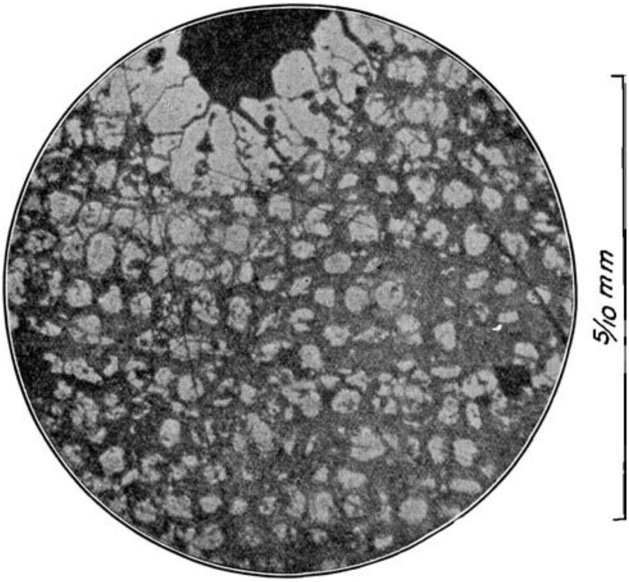

FIG. 148.—Polished section showing cross-section of wood structure, Blue Dike mine, Utah. Light areas, chalcopyrite; dark areas, chalcocite. Magnified 120 diameters. (*After B. S. Butler, U. S. Geol. Survey.*)

In Nova Scotia, Cumberland County, chalcocite nodules, with remains of pyrite, and also chalcocitized wood, are found in the Permian sandstone.[1]

**Genesis of Sedimentary Copper Ores.**—The epigenetic character of the copper deposits in sandstone is proved beyond reasonable doubt. The replacement of coal, carbonaceous shale, and calcareous or kaolinic sandstone cement by chalcocite is also proved. The gangue minerals are few and quartz is conspicuously absent. Barite in small amounts is rather common. Irregularity in dissemination is typical, though the ores often follow certain horizons rather persistently. The independence of the occurrence of igneous rocks is marked.

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[1] E. B. Papenfus, *Econ. Geol.*, 26, 1931, pp. 314–330.

The occurrences are mainly on the flanks of older continental areas containing pre-Cambrian copper deposits; the sandstones were rapidly laid down as arkoses, indicating a long epoch of rock decay, the products of which were swept away during a following arid epoch. Considering the evidence as a whole the sedimentary deposits must have contained finely divided copper ores, in part from solutions derived from the land area, in part as cupriferous detritus. When atmospheric waters charged with salt and gypsum searched these beds they must have taken this copper into solution and concentrated it at certain horizons where reducing substances like coaly vegetable matter were available. In most cases the solution probably contained the copper as sulphate, though where much salt was present it might well have been transformed into chloride.

Ascending thermal solutions of meteoric origin may have formed some deposits. Magmatic agencies seem in most cases to be definitely excluded.

Pyrite, bornite, and chalcopyrite often appear in the nodules or in the fossil wood, the chalcocite often replacing earlier sulphides (Fig. 148). Rogers and Finch have discussed the general genetic features of the deposits.[1]

Chalcocite nodules from the Nova Scotia occurrence show plainly that this mineral replaces earlier pyrite concretions; but other specimens from the same place point to a direct replacement of carbonized wood by chalcocite without an intermediate pyrite stage.

## VANADIUM AND URANIUM ORES IN SANDSTONES[2]

**Composition.**—Vanadium with some uranium and a trace of radium has been shown to be common in certain Jurassic sandstones in Colorado

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[1] A. F. Rogers, *Econ. Geol.*, 11, 1916, pp. 366–380.

J. W. Finch, *Trans.*, Am. Inst. Min. Met. Eng., 76, 1928, pp. 378–392.

[2] W. F. Hillebrand and F. L. Ransome, Carnotite, etc., in western Colorado, *Am. Jour. Sci.*, 4th ser., 10, 1900, pp. 120–144. *Bull.* 262, U. S. Geol. Survey, 1905, pp. 9–13.

R. C. Coffin *et al.*, Uranium and vanadium deposits of southwestern Colorado, *Bull.* 16, Colorado Geol. Survey, 1921.

H. S. Gale (Carnotite in Colorado), *Bull.* 340, U. S. Geol. Survey, 1908; *idem, Bull.* 315, 1906, pp. 110–117.

F. L. Hess, Vanadium deposits in Colorado, Utah, and New Mexico, *Bull.* 530, U. S. Geol. Survey, 1912.

F. L. Hess, A hypothesis for the origin of the carnotites, *Econ. Geol.*, 9, 1914, pp. 675–688.

F. L. Hess, New and known minerals from the Utah-Colorado carnotite region, *Bull.* 750, U. S. Geol. Survey, 1925, pp. 63–78.

K. L. Kithil and John A. Davis, Mining and concentration of carnotite ore, *Bull.* 103, U. S. Bur. Mines, 1917. Parsons, Moore, Lind, and Schaefer, Extraction and recovery of radium, uranium and vanadium, etc., *Bull.* 104, *idem*, 1915. R. B. Moore *et al.*, Analytical methods for certain metals, *Bull.* 212, *idem*, 1923.

and Utah. Few of the deposits in western Colorado are now worked. A reduction plant was located at Vanadium, near Placerville.

A number of unusual minerals are contained in these deposits. Most of them are fine-grained or cryptocrystalline and, in part, are probably of colloidal origin. The three most important species are carnotite, vanoxite, and roscoelite.

*Carnotite,* a potassium-uranium vanadate, usually appears as a bright-yellow crystalline powder occurring on seams and in fossil wood. Its formula is $2UO_3.V_2O_5.K_2O.xH_2O$, with about 56 per cent $UO_3$, 20 per cent $V_2O_5$, and small quantities of calcium, barium, copper, lead, and molybdenum. In tyuyamunite (named after a locality in Turkestan) calcium replaces potassium. In this, as in many other minerals of this group, the water is variable and some of it is given off at low temperatures.

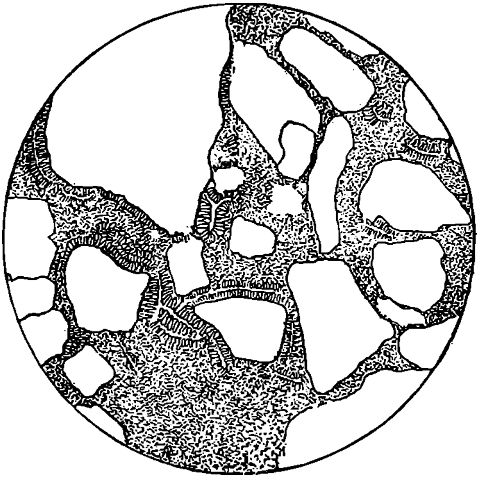

FIG. 149.—Vanadium ores in sandstone. White areas, quartz; shaded areas, roscoelite, partly radial. Magnified 25 diameters. (*After F. L. Hess, U. S. Geol. Survey.*)

*Vanoxite* is a black powdery material with the formula $2V_2O_4.V_2O_5.xH_2O$, a hydrous vanadyl-vanadate. It replaces the cement in sandstones, or occurs in gypsum veinlets or in fossil wood.

*Roscoelite*[1] is a vanadium mica of greenish color, essentially a muscovite in which aluminum is partly replaced by vanadium to the extent of 20 to 29 per cent $V_2O_5$, and occurs in minute scales replacing the cement in sandstones (Fig. 149). Some sandstones contain 20 per cent roscoelite. Among the rarer minerals are metahewettite ($CaO.3V_2O_5.xH_2O$), uranium micas of the autunite type ($CaO.2UO_3.P_2O_5.xH_2O$), and a few uranium and vanadium sulphates, also hydrous in varying degree.

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[1] Roscoelite is also known from some gold-quartz veins and is here often associated with gold and tellurides. Mariposite, a green chromium mica, is a characteristic mineral of gold-quartz veins occurring in serpentine.

Uvanite ($2UO_3.3V_2O_5.xH_2O$), a brownish-yellow uranium vanadate, was mined in Emery County, Utah. Volborthite, a hydrous vanadate of calcium, copper, and barium, occurs in places.

Chromium, in the form of mariposite, usually at a separate horizon slightly below the roscoelite sandstone, is also present in these remarkable ores. Barium, copper, lead, molybdenum, and arsenic also occur. Molybdenum is quite abundant in some places and appears as a soluble sulphate—the dark blue ilsemannite. Native selenium has been identified by Hillebrand, associated with metahewettite.

The uranium minerals, among which carnotite is the most abundant, contain a small trace of radium[1] which is recovered. Gypsum is about the only gangue mineral associated with the ores.

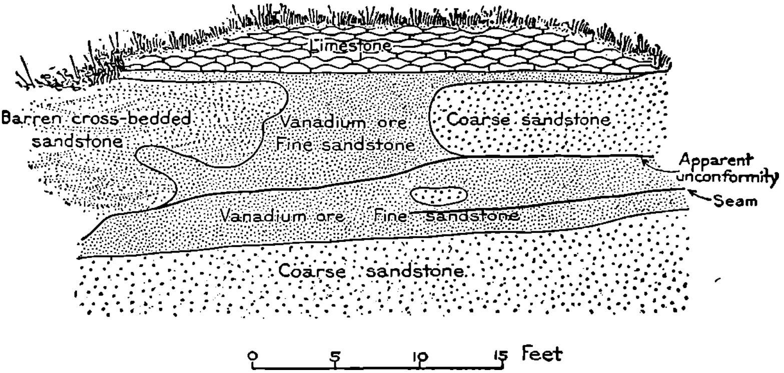

FIG. 150.—Sketch of vanadium-bearing sandstone at mine of Primos Chemical Company, on the east side of Bear Creek, Newmire, Colorado. (*After F. L. Hess, U. S. Geol. Survey.*)

The ores are not rich. The carnotite ores contain about 1.5 to 3 per cent $UO_3$ and 3 to 5 per cent $V_2O_5$. Concentration has been attempted in some cases.

The roscoelite ore at Placerville contains about 3.50 per cent $V_2O_5$ and 0.05 per cent $UO_3$.

**Occurrence.**—The ores are found in the Plateau province of horizontal or gently inclined strata in western Colorado and eastern Utah (Fig. 151). The best known localities are at Placerville, Colorado, where roscoelite ores occur in the LaPlata sandstone; also in the LaSal, Paradox, and Sindbad valleys somewhat farther west, where carnotite ores are worked in the McElmo formation of sandstone and shale covering the LaPlata sandstone. Other deposits are worked at Rifle, Colorado, and at several places in the San Rafael swell, Utah.

The ores are mainly confined to the McElmo and LaPlata formations of white, often cross-bedded, Jurassic sandstone which frequently contain much transported partly carbonized wood. They follow certain horizons

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[1] At the rate of 1 gram of radium in 3,000 kilograms of metallic uranium.

or appear in fissures or flat veins or in brecciated zones (Figs. 150 and 151) and are often associated with the fossil wood. Carnotite is always associated with fossil wood.

Some observers have thought the ores merely superficial, but it now seems certain that they may be found in depth. At Placerville the workings are said to have penetrated 2,000 feet under ground, in a horizontal direction.

Roscoelite is apparently not known in sandstones outside of the Colorado-Utah area; but carnotite and other uranium minerals, such as autunite and volborthite, are not uncommon in conglomerates, sandstones, and shales of various ages, though rarely in economic quantities.

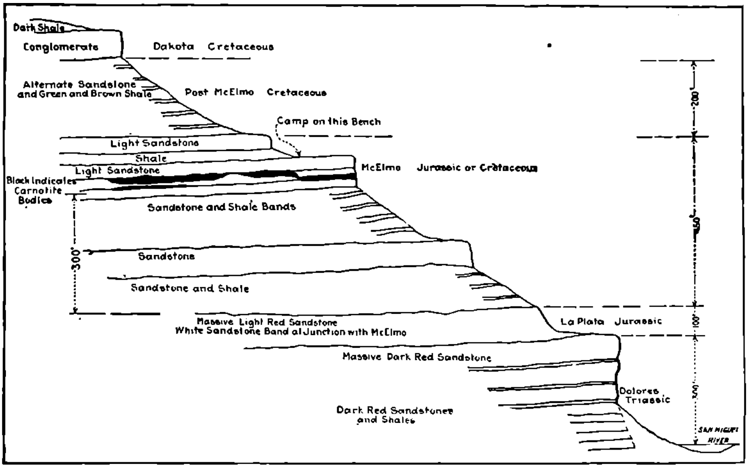

FIG. 151.—Typical geological section at Dolores, Colorado. (*After H. S. Worcester.*)

Not similar to these deposits, yet perhaps of a similar origin, are the important vanadium deposits at Mina Ragra[1] near Quisque, Province of Pasco, Peru, first described by D. F. Hewett and W. F. Hillebrand. These deposits now supply the main part of the demand for vanadium. The vanadium sulphide, patronite ($VS_4$?), occurs here on a large scale as greenish-black masses associated with a hydrocarbon and a peculiar nickel-bearing pyrite as well as with many secondary vanadium minerals. An analysis of the patronite gave in per cent 58.79 sulphur, 19.53 vanadium, 0.18 molybdenum, 1.87 nickel, and 3.47 carbon. Patronite is a fine-grained mixture of three sulphides and is probably of colloidal origin.

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[1] D. F. Hewett, Vanadium deposits of Peru, *Trans.*, Am. Inst. Min. Eng., 40, 1909, pp. 274–299.

W. F. Hillebrand, The vanadium sulphide, patronite, *Jour.*, Am. Chem. Soc., 29, 1907.

W. F. Hillebrand, *Am. Jour. Sci.*, 4th ser., 24, 1907, p. 141.

W. S. Hutchinson *et al.*, Report of the subcommittee on vanadium, *Bull.* Mining Met. Soc. Am., 175, 1925.

The deposit is intersected by a porphyry dike, and dips with the beds. It appears to be a lens, probably conformably enclosed, in gently dipping Cretaceous shales which contain much gypsum. It extends for 400 feet north and south, is from 50 to 100 feet wide, and has been followed to a depth of 300 feet. The larger part of the world's production of vanadium has been obtained from this rich deposit; the ores contain from 10 to 13 per cent vanadium.

**Genesis.**—The intimate relation of vanadium with arsenic and phosphorus is shown by its position in the periodic table of elements. In its occurrence it is closely allied to titanium and iron. While the igneous rocks of the crust average 0.026 per cent $V_2O_3$ it is not present as any definite vanadium mineral, but augite, biotite, rutile, magnetite, and ilmenite often contain notable amounts of it in isomorphous mixtures. During the decomposition of these minerals vanadium remains with the iron, largely at least, and so it is present in limonite and limonitic soils.

During sedimentation it seems to be concentrated in fine and clayey sediments probably in colloidal form. Many vanadium minerals show a marked tendency to colloform development. Lastly, vanadium plays an important, as yet scarcely understood part in biochemical reactions. Replacing phosphorus it is a constituent of the blood of certain holothurians and is doubtless absorbed from sea water. And as it is a common constituent in the ashes of bitumens, asphalt, coals, and petroleum, the inference is justified that it enters into vegetable and low animal growth. Frequently it is contained in carbonaceous shales; such Cretaceous shales from Peru contain in places 0.1 to 0.4 per cent $V_2O_3$, and the vanadium phosphate sincosite appears in them. It is not surprising then to find vanadium with coaly materials as in Colorado and Utah, or as at Mina Ragra in Peru. At the latter place the probable genesis is by leaching of vanadiferous sediments by, possibly warm, meteoric waters.

In the Colorado-Utah field the carnotite, vanoxite, and hewettite ores are perhaps secondary enrichments of an unknown vanadium mineral; but the roscoelite, with which little uranium is associated, appears more like an original product of the meteoric circulation; and these waters may well have been tepid. Carnotite is strongly attacked by sulphuric acid and by sulphate solutions.[1]

It is more difficult to explain why the vanadium minerals are mainly confined to certain horizons, as in Colorado. Hess believes that the ores were concentrated by meteoric waters and that they were chiefly carried down to the flood plains during an epoch of deep disintegration of veins and pegmatite dikes of adjacent crystalline, pre-Cambrian rocks. He thinks both uranium and vanadium were mainly transported as

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[1] F. B. Notestein, Some chemical experiments on uranium-vanadium, *Econ. Geol.*, 13, 1918, pp. 50–64.

sulphates. At any rate the concentration must surely have been effected by meteoric waters, and they were probably of the sulphate type. In fact gypsum is the common associate of the ores.

The salic extracts (the pegmatites) contain few, if any, vanadium minerals. The later volcanic emanations contain little vanadium for, in the deposits formed by hot ascending waters, roscoelite is about the only hypogene vanadium mineral and that very scarce. It is true that especially in lead deposits, lead and copper vanadates, such as vanadinite and descloizite, are not uncommon in the oxidized zone; but in this case the vanadium was probably derived from the country rock. It is possible, however, that vanadium is present, to a greater extent than is now suspected, in the sericitic minerals and in the sulphides of ore deposits.

Uranium while associated with vanadium is not chemically related to it. The home of the uranium minerals is in the pegmatites (p. 761) and in certain mesothermal veins like those of Joachimsthal, Bohemia, and Gilpin County, Colorado (p. 596). The primary uranium minerals yield many products of decomposition among which carnotite and the uranium micas (autunite group) are prominent. It is easily transported in surface waters as shown by the occurrence of the uranium micas in sandstone and in alluvial peats (Madagascar, Lacroix).

Uranium is also found in ashes of certain bitumens and coals, particularly in the Swedish "Kohm" of Cambrian age. It is probable that it enters into biochemical reactions of vegetable life; it forms many colloidal compounds.

In conclusion, it seems fair to assume that, in the sediments here discussed, it has been concentrated with vanadium, by meteoric waters, which derived the metals from terrigenous sediments resulting from the disintegration of pre-Cambrian igneous rocks and pegmatites.

The constant association with nickel, molybdenum, chromium, copper, and selenium is very suggestive.

**Production and Use.**—In 1915, 47,000 tons of ore were mined in Colorado containing, according to U. S. Geological Survey, 19.9 tons of metallic uranium, 627 tons of vanadium (mainly from roscoelite ores) and 6.1 grams of radium, the value of the metals being about $700,000. Some years ago most of the uranium ores were exported and radium extracted abroad. In 1914, the ores mined contained 87.2 tons of uranium and 22.3 grams of radium. The pure vanadium ores of Placerville were roasted with sodium chloride, the resulting sodium vanadate extracted with water and precipitated with ferrous sulphate as iron vanadate which was shipped east for reduction to ferrovanadium. The price of ferrovanadium, with about 35 per cent vanadium, is about $3.25 per pound of contained vanadium. About 1 per cent vanadium added to steel increases its toughness and resistance to torsion and high temperature. It is, however, less essential to the steel industry than tungsten. The principal supply is obtained from the Peruvian patronite mine, from which 6,000 tons of 12 per cent ore were imported in 1930. Large amounts of vanadate of lead and copper are mined from oxidized deposits in Southwest Africa and in northern Rhodesia. Minor amounts of vanadium salts are used as mordants for dyeing and cloth printing and for other chemical purposes.

Uranium salts have a limited use for a yellowish-green glass and for pottery; also as a mordant in dyeing. Ferro-uranium is at present not used in the steel industry. Radium[1] is separated from uranium by a complicated process explained in *Bulletin* 104, United States Bureau of Mines. It is produced as a chloride or bromide and its principal use is in medical science, various diseases yielding to its emanations.

Radium in corresponding quantities is also contained in uraninite (crystalline) and pitchblende (amorphous), both essentially $UO_2$ with 80 ± per cent uranium oxides. These minerals are found in pegmatite dikes and in hydrothermal veins.

In recent years[2] the output of the vanadium-uranium deposits in Colorado and Utah has been small, owing to the domination of the radium market by the Katanga deposits lately discovered in the Belgian Congo, where, in 1930, 1,296 metric tons of uranium-radium ore were mined. This ore is rich in uranium and as each per cent of uranium means 2.57 milligrams of radium per ton, the cost of reduction is said to be only a few dollars per milligram, whereas formerly the cost of extraction was reported to be $37 per milligram. It is reported that, in 1930, 60 grams of radium salts were produced of which 17 grams were imported into the United States. Price quoted $70 per milligram.

About 115 tons of uranium salts, chiefly sodium uranate, were imported at $1.50 per pound. Of these a large part was produced at Joachimsthal, Czechoslovakia. A few grams of radium were produced at this place. In 1931 a possibly important new source of radium was discovered, for much pitchblende occurs in the silver-cobalt veins of Great Bear Lake in the Northwest Territories, Canada.

## THE COPPER-BEARING SHALES OF MANSFELD[3]

It seems proper to consider at this place the celebrated cupriferous shale (Kupferschiefer) of Mansfeld, in central Germany, for, though not identical with the deposits described in this chapter, it presents most interesting analogies to them.

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[1] P. M. Taylor, Radium, U. S. Bur. Mines, *Information Circ.* 6312, 1930.

[2] F. L. Hess, Vanadium, uranium, and radium, *Mineral Resources*, U. S. Bur. Mines, 1930. J. Thoreau and R. de Terdonck, Le gîte d'uranium de Katanga (Shinkolobwe), 16, Int. Geol. Congress, Washington, 1933 (abstract). Related to the Katanga copper deposits. Of magmatic origin.

[3] Best description with literature in Stelzner and Bergeat, Die Erzlagerstätten, 1, 1904, pp. 391–417.

A flourishing mining industry, which began in A.D. 1150, is still based on the Kupferschiefer, the annual production being approximately 700,000 metric tons of ore containing between 2 and 3 per cent of copper and yielding 24,000 metric tons of copper and 125 tons of silver.

The first stratum deposited in the subsiding basin of the upper Permian in central Germany was a marine conglomerate of slight thickness. "Above it extends like a black shroud the thin bed of cupriferous shale, one of the most remarkable products of the geologic ages. Characterized by its fauna as a shallow-sea deposit, full of plant remains carried in from adjacent coasts, the formation bears the stamp of an organic mud deposit admixed with inorganic precipitates."[1]

Above the cupriferous shale (*k*), which is less than 1 meter thick, extends a marine limestone, Zechstein (*z*), 6 to 10 meters thick, and above that lie the great gypsum and salt masses (*s*) of the uppermost Permian, again covered by Triassic sandstone (*b*). Folding and faulting have since affected the beds, and the mining now follows the inclined strata to a depth of 500 meters. The cupriferous bed averages 50 centimeters in thickness, but only the lower 20 to 30 centimeters is utilized as ore (Fig. 152). The Permian "Rothliegendes" (*r*) underlies the copper bed.

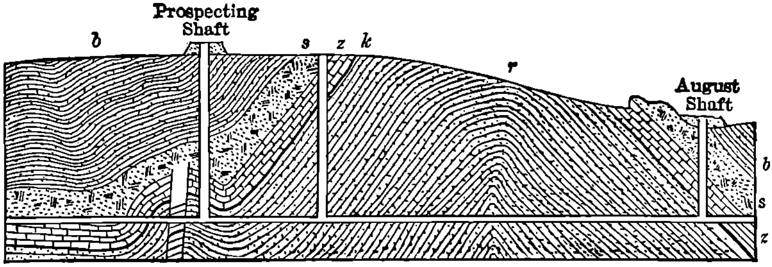

FIG. 152.—Section of a part of the copper-bearing shale bed at Mansfeld, Germany. (*After Schrader.*)

The ores are sulphides, in minute distribution through the shale giving it a bronzy appearance. Bornite predominates, but there are also chalcocite, and rarely galena, sphalerite, and tetrahedrite. Pyrite and chalcopyrite, when present, are later products. Small quantities of nickel, cobalt, selenium, vanadium, and molybdenum have been recognized; there is also about 0.015 per cent of silver. Zinc is present, and

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[1] F. Beyschlag, in Deutschlands Kalibergbau, 1907, p. 4.

P. Krusch, *Zeitschr. prakt. Geol.*, 27, 1919, pp. 76–84.

H. Schneiderhöhn, *Neues Jahrbuch.*, Beil. Bd. 47, 1923, with literature (microscopic study of ores).

P. D. Trask, *Econ. Geol.*, 20, 1925, pp. 746–761 (review of theories).

J. W. Gregory, The copper shale of Mansfeld, *Bull.* 315, Inst. Min. and Met. (London), 1930, 28 pp.

A. Cissarz, *Metall u. Erz.* 27, June 2, 1930, pp. 316–319 (spectroscopic analysis of ore).

in the upper part of the bed, not mined, there is more zinc than copper. An average analysis quoted by Stelzner and Bergeat is as follows:

AVERAGE ANALYSIS OF CUPRIFEROUS SHALE

(*Dr. Haase, Analyst*)

| | Per cent | | Per cent |
|---|---|---|---|
| $SiO_2$ | 33.15 | Ag | 0.014 |
| $Al_2O_3$ | 17.3 | Ni | 0.018 |
| CaO | 10.4 | S | 2.31 |
| MgO | 1.0 | $CO_2$ | 9.24 |
| Fe | 2.6 | $H_2O$ | 1.7 |
| Zn | 1.276 | Bitumen | 9.06 |
| Cu | 2.75 | | |

There is about 3 per cent $K_2O$ and 1 per cent $Na_2O$; lead amounts to 1.50 per cent, manganese to about 0.25 per cent.

There is practically no gangue, except veinlets of gypsum and barite. The bed is cut by faults, along which in places occur barite, anhydrite, calcite, niccolite, pyrite, and chalcopyrite, and near these breaks (called "Rücken") the metal content is subject to enrichment, impoverishment, or removal upward in adjacent beds. Bergeat declares that these changes take place on secondary fissures and cracks.

There has been much controversy about the Mansfeld deposits. The majority of geologists regard them as sedimentary and syngenetic: von Groddeck, Stelzner, Freiesleben, and von Cotta held this view, and it is shared by Bergeat. Posepny and Beck, as well as Krusch, believe them epigenetic and think that the metals were probably introduced in the shale from the "Rücken."

The Kupferschiefer is certainly not an ordinary marine deposit precipitated from the sea water.[1] It was laid down in a shallow sea which was full of decaying vegetable and animal remains and into which were probably discharged cupriferous waters from the surrounding littoral, most likely, sulphate solutions derived from the eruptives and the ore deposits of the early Permian epochs. The arguments for a syngenetic origin are very strong.

The presence of nickel, cobalt, vanadium, molybdenum, and selenium recalls the epigenetic deposits in sandstone so abundant around the shores of the Permian sea, in Bohemia, and Russia, for instance. The Mansfeld basin was simply, then, the final collecting place of the solutions derived from adjacent desert shores. Schneiderhöhn concludes that the copper sulphides were formed syngenetically as mixed iron-copper sulphide gels under the influence of bacteria of the sulphur cycle.

## COPPER SULPHIDE VEINS IN BASIC LAVAS

**General Features.**—All basic lavas contain copper, but in many cases conditions were evidently unfavorable for the concentration of copper

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[1] Sea water contains a trace of copper, but the amount present seems utterly insufficient to account for the Mansfeld deposits.

immediately after the eruption, and the rocks retained their copper until later opportunities for ore formation were offered. The existence of vast masses of such basic lavas near the surface, without any indication of copper concentration (*e.g.*, the Columbia River lava or the basalts of the Hawaiian volcanoes), shows plainly that the ordinary surface waters at slight depth are not competent to dissolve and concentrate accessory metals contained in these rocks. A depth of perhaps a few thousand feet seems to be necessary, under the most favorable conditions, for waters of meteoric origin to extract the copper; though it is, of course, possible that such waters, when ascending in suitable channels, may deposit the dissolved copper at higher horizons. In some of the veins here discussed epidote is present, but more frequently it is absent, and the veins then assume the well-known type of chalcopyrite in a quartz-calcite-siderite gangue. Such veins, deposited by ascending surface waters of the deeper circulation, may not be easy to distinguish from those whose development is a phase of intrusive after-effects. Nor can it be denied that in these veins may be concentrated some gold and silver from the igneous rock; in general, however, they will be found much poorer in gold and silver than deposits connected with the intrusive processes.

Whether native copper, bornite, or chalcopyrite will form may be dependent upon the quantity of sulphur which the lavas were able to retain at their eruption.

**The Kennecott Mine, Alaska.**—The copper deposits in the Nikolai greenstones of the Copper River region, described by F. C. Schrader, W. C. Mendenhall, A. C. Spencer, and again by F. H. Moffit,[1] are of special interest. Flows of Triassic or Carboniferous basalts about 4,000 feet in thickness are covered by 2,000 feet of Triassic dolomitic limestone, which in turn is overlain by a thick series of Jurassic strata. The latter are cut by monzonitic intrusives, which are accompanied by a different kind of metallization, characterized by gold deposits.

The Nikolai greenstones are amygdaloid flows of basalt; the amygdules contain chlorite, chalcedony, quartz, epidote, and some zeolites and calcite, with specks of native copper, chalcocite, bornite, and chalcopyrite. Copper sulphides also occur in slips, brecciated zones, and faults. Even the apparently fresh greenstone contains copper.

The ore-bodies of the Bonanza and Jumbo mines,[2] which have yielded annually for many years from 10,000 to 20,000 tons of copper, are located at an elevation of about 6,000 feet in the Kennecott Range far above the

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[1] F. H. Moffit and S. R. Capps, *Bull.* 448, U. S. Geol. Survey, 1911.

F. H. Moffit, *Bull.* 662, U. S. Geol. Survey, 1917, pp. 155–182.

[2] A. M. Bateman and D. H. McLaughlin, Geology of the ore deposits of Kennecott, Alaska, *Econ. Geol.*, 15, 1920, pp. 1–80; also *idem*, 27, 1932, pp. 297–306.

S. G. Lasky, A colloidal origin of some of the Kennecott ore minerals, *Econ. Geol.*, 25, 1930, pp. 737–757.

snowline and in a region of enormous glaciers heading in the Wrangel Range at elevations of about 16,000 feet (Fig. 153). The ground-water is frozen so that no oxidation is proceeding at the present time. The Bonanza ore-body was exposed in the cliffs above the glacier and there was a great quantity of almost solid chalcocite in the talus masses below the outcrop. The ores occur only in the limestone or dolomite and extend up into the carbonate rock from near the greenstone contact (Fig. 154). The chalcocite forms large wedge-shaped masses, one, for instance, traced for 400 feet with a greatest width of 25 feet; but eventually these masses, which follow joints transverse to the greenstone contact, contract to small dimensions. Along the greenstone contact there is a strike fault which follows the lowest shaly beds of the limestone. There is some

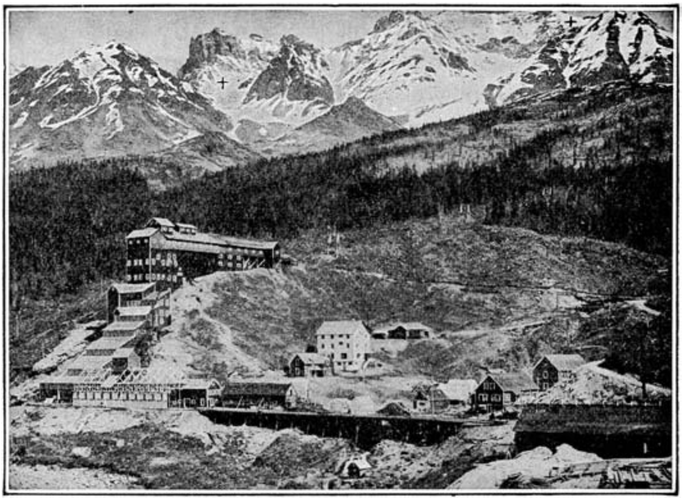

Fig. 153.—View of Kennecott mill and mine, Alaska. Mine near summit of Snowy Peaks. + indicates position of mines. (*After A. M. Bateman and D. H. McLaughlin.*)

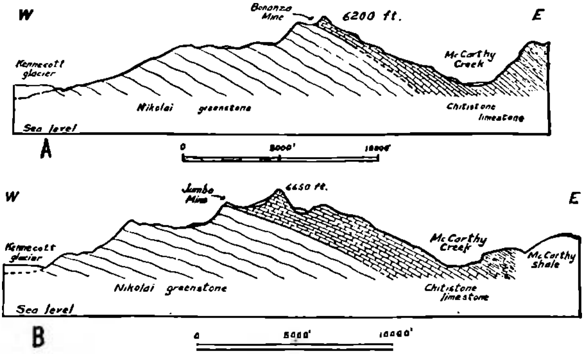

Fig. 154.—Vertical cross-section showing position of (*A*) Bonanza mine; (*B*) Jumbo mine. No vertical exaggeration. (*After A. M. Bateman and D. H. McLaughlin.*)

oxidation but it evidently antedates the present glacial epoch. The ore consists of chalcocite with very small amounts of bornite, covellite, enargite, luzonite, tennantite, pyrite, galena, and sphalerite. The limestone shows no alteration near the ore. The ore as mined contains from 7 to 10 per cent of copper, and the earlier shipments were much richer. The chalcocite is in part of an apparently colloidal variety, in part orthorhombic. Some of it is granular showing cleavage, which Bateman and McLaughlin consider to indicate the isometric variety, and contains 9 per cent covellite in solid solution, which would prevent reversal to the rhombic variety. Graton believes the cleavage to be inherited from bornite. Concentric or colloform structures occur in the ore. Covellite is demonstrated to be hypogene and generally earlier than chalcocite. Succession is said to be chalcopyrite, bornite, tennantite, enargite, covellite, chalcocite.

The genesis of this most remarkable deposit is not fully clear. Bateman and McLaughlin believe that before the present erosion epoch and shortly after the intrusion of monzonite porphyry (which is 1½ miles distant and apparently not directly related to the deposit) meteoric warm waters searched the underlying greenstone, whose copper content antedates the main deposits, and carried copper solutions up into the limestone where it was precipitated as chalcocite with minor amounts of other minerals. Evidently the temperature during deposition was low; in the case of rhombic chalcocite it must have been below 90° C. The absence of alteration of the country rock points to comparatively low temperature.

Naturally the greenstone mineralization recalls Michigan deposits, and the chalcocite deposition might be regarded as its last phase; at any rate the magmatic origin of the solutions can not be regarded as proved conclusively.

## COPPER SULPHIDE VEINS IN INTRUSIVE BASIC ROCKS

Veins containing pyrite and chalcopyrite, occasionally with other sulphides, in a gangue of quartz, calcite, dolomite, and siderite, more commonly quartz alone, occur abundantly in intrusive basic rocks, such as diabase or gabbro. Here, however, the distinction between the results effected by water of atmospheric origin and those effected by magmatic solutions becomes increasingly difficult.

F. E. Wright[1] has pointed out the fact that the intrusive Keweenawan gabbro of Mount Bohemia contains veins with chalcopyrite, bornite, chalcocite, calcite, and quartz, while in the surface lavas of the same series native copper is the principal ore mineral. This seems an excellent illustration of the retention of volatile sulphur by intrusives, contrasted with

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[1] F. E. Wright, The intrusive rocks of Mount Bohemia, Michigan, *Seventh Ann. Rept.*, Geol. Survey Michigan, 1908.

its escape from the extrusive flows. The origin of the water which was the vehicle of deposition in these veins may remain an open question.

Along the foothills of the Sierra Nevada of California extends a belt of andesitic rocks of Jurassic age collectively called "greenstones." They consist of massive and schistose rocks, including surface flows, tuffs, and intrusions mixed. Within this belt, for instance in Yuba and Nevada counties, short and irregular quartz veins with pyrite and chalcopyrite are common. Probably these veins derived their copper from the greenstones, and undoubtedly they were formed at a time when the rocks now exposed were covered by several thousand feet of overlying and now eroded igneous rocks.

Other deposits in Placer County and at Campo Seco in Calaveras County are, according to Knopf, replacement deposits in shear zones in amphibolite and greenstone. The minerals are pyrite and chalcopyrite with a little galena and sphalerite. The sulphides contain a little silver and a trace of gold. Whether these deposits were formed by meteoric waters is, however, very uncertain.

Other and much larger copper deposits are found in the same region to the north in Shasta County, but in this case the evidence points unmistakably to an origin by solutions derived directly from the magma (p. 621).

### OTHER VEINS DEPOSITED BY WATERS OF THE UPPER CIRCULATION

In the region of the saline Paleozoic and Mesozoic beds of central Germany, a certain amount of vein deposition by saline meteoric waters appears to have taken place. Veins of this origin are likely to contain an abundant gangue of calcite, dolomite, or barite, with some quartz and a scant amount of sulphides.[1]

Similar examples of ore deposition by saline waters also exist in the western part of the United States; one, the Cashin mine of Colorado, is mentioned above (p. 406). The prevailing influence of igneous intrusions on ore deposition is, however, so strong that it is difficult to establish the proofs of the less conspicuous deposition by purely meteoric water.

That the ordinary surface waters are in most cases quite incompetent to effect concentration is plainly shown by the lack of important mineralization in fissures and joints cutting the rocks of mining districts. In the Globe district, Arizona, for instance, the Paleozoic rocks are intersected by a network of dislocations which would offer excellent paths for these waters; and yet the important deposits are in no way connected with these fractures. Similar post-mineral fissures traverse lead and zinc veins in the Coeur d' Alene district, Idaho, but generally show no trace of mineralization.

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[1] P. Krusch, Ueber die Zusammensetzung der Westfälischen Spaltenwässer, *Zeitschr. prakt. Geol.*, 12, 1904, p. 252.

The Cordilleran region contains many great ranges of pre-Cambrian rocks capped in places by Paleozoic and Mesozoic strata. Among them may be mentioned the Front ranges of the Rocky Mountains in Colorado, the Wind River Range in Wyoming, and the Mission Range in Montana. Uplift, folding, and faulting have in each of these ranges intensified the circulation of meteoric waters, but in spite of this the ranges are remarkably poor in mineral deposits, which appear only in the vicinity of intrusive masses. These relations show very plainly the slight concentrating power of ordinary cool surface waters and even of the waters of atmospheric origin that have become a part of the deeper circulation.

## CHAPTER XXIII

## LEAD AND ZINC DEPOSITS IN SEDIMENTARY ROCKS; ORIGIN INDEPENDENT OF IGNEOUS ACTIVITY

**Characteristic Features.**—The lead and zinc deposits which form the subject of this chapter represent a type of world-wide distribution and, in spite of local variations, of remarkably constant characteristics. They appear to be independent of igneous rocks and occur in limestones, dolomites, cherts (derived from limestone), or calcareous shales. In the United States this type is represented by the ores of the Mississippi Valley, chiefly in Oklahoma and Missouri. In Europe characteristic occurrences are in Belgium and in Upper Silesia.

The mineral composition is simple, and the ore minerals few. Galena and sphalerite are essential constituents, with their train of oxidized minerals near the surface[1] (sulphates, carbonates, and silicates); there is more or less pyrite, almost always marcasite, occasionally a little chalcopyrite. Gold, antimony, arsenic, and molybdenum are almost wholly absent; in some districts the galena contains a little silver, but on the whole the deposits are non-argentiferous. Cadmium is often contained in the sphalerite, which is mainly red, light brown, or yellow and carries little iron. Cadmium sulphide, greenockite, occurs as a secondary mineral. Nickel and cobalt are often present in small quantities. Among the gangue minerals dolomite is the most characteristic; quartz in crystals is not common, but a secondary chert with bitumen is typical of many districts; barite is found, but is not characteristic.

The ores commonly fill open spaces in zones of local brecciation or in crevices (gash veins) or joints which have been enlarged by solution. Less commonly they occupy fault fissures; sometimes they are purely metasomatic, the minerals occurring disseminated in limestone or dolomite and closely following certain sedimentary horizons. Even in this case they are not spread over irregular areas, but tend to follow certain lines in the plane of stratification (so-called "runs"). In regions of slightly disturbed strata, many observers have noted the tendency of the ore to follow pitching troughs. The ores usually lie within a few hundred feet of the surface and are oxidized in the vicinity of the water level. Frequently they are found below impervious shale beds. The

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[1] The oxidation of lead and zinc sulphides is treated in Chap. XXXII. The principal oxidized zinc minerals are smithsonite and calamine, while hydrozincite and willemite are rare; goslarite, the soluble sulphate, is frequently found as efflorescences.

mineral association and the frequent colloform texture indicate deposition at low temperature.

**Origin.**—Simple as the deposits of this type are, the views as to their origin diverge strongly. The earliest interpretation of them, as marine deposits, is generally abandoned; it is recognized that even if the metals are derived from primary ocean sediments the finely divided sulphides must have been concentrated and redeposited. Their epigenetic nature is clear. Some geologists hold the ores to be deposited by ascending waters; others see in them the work of descending surface waters.

In either case many American geologists believe that atmospheric waters have effected the concentration of the lead and zinc from sedimentary Paleozoic rocks and that igneous agencies have had nothing to do with the deposition. This opinion is not unanimous for a number of investigators have suggested an origin by thermal waters ascending from great depths. Some of these consider that the metals were extracted by the hot waters from the underlying pre-Cambrian rocks, while others believe they can see a relationship of the deposits with deep intrusions and magmatic sources. Beyschlag, Krusch, and Vogt in their handbook on ore deposits uphold the theory of origin by thermal waters. By authors holding this view, the deposits are often referred to as "telemagmatic" (tele = distant).

No one can, however, deny that galena and sphalerite are of widespread occurrence in many limestones and dolomites far from regions of deep fissuring and igneous action.[1] Before appealing to igneous agencies it will be advisable to examine into the competency of waters of atmospheric origin to effect the concentration of these metals.

In the first place, the mineral association indicates a shallow deposition at temperatures and pressures not very different from those prevailing at the surface. The deposits contain no characteristic substances carried by thermal waters of volcanic origin, and no primary silicate minerals. The marcasite suggests strongly deposition near the surface.

Regarding nickel and cobalt, it has already been pointed out that minerals of these metals are not uncommon in sedimentary strata, as is shown, for example, by their occurrence with the marine siderites and limonitic oölites, or by the occasional discovery of millerite in limestone. This granted, it remains to account for the two principal metals, lead and zinc. The majority of geologists who have studied these deposits believe

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[1] *Cf.* for instance the repeated finds of galena and other minerals in deep borings on the Gulf Coast. See A. C. Veatch and G. D. Harris, *Bull.* 7, Louisiana Geol. Survey, 1908, p. 25; also, G. D. Harris, *Bull.* 429, U. S. Geol. Survey, 1910, p. 45. Regarding zinc in sedimentary iron ore, see p. 274.

Gilbert Van Ingen has pointed out the frequent occurrence of grains of galena and sphalerite in fossils where perhaps decaying organic matter might have brought about precipitation. *Bull.* Geol. Soc. Am., 26, 1915, p. 85.

that the lead and zinc originally were contained as silicates or sulphides in the older crystalline rocks from which the limestones and other sedimentary rocks were derived. A number of analyses have been made showing the presence of copper, lead, and zinc. Still more convincing is the analysis by George Steiger of a composite sample of 329 igneous rocks which have been analyzed in the laboratory of the United States Geological Survey. This gave in per cent 0.00513 zinc, 0.00075 lead, 0.00932 copper, 0.00515 nickel, and 0.00048 arsenic.[1] Another series of analyses of igneous rocks from England by A. M. Finlayson[2] gave an average of 0.0032 per cent lead and 0.0016 per cent zinc.

According to J. B. Weems and J. D. Robertson, the Cambrian and Ordovician limestones of Missouri average in per cent 0.00425 zinc, 0.00096 lead, and 0.00126 copper.[3] Considered in conjunction with the composite analyses of silts from the Mississippi River delta (p. 253) these figures in part support the opinion referred to regarding decreasing metal content in successive sedimentations.

No analyses of shales are included among those given above. They can not be safely excluded, however; and it is probable that they will average higher in metal content than the limestones as suggested by partial analyses quoted by E. R. Buckley[4] and G. H. Cox.[5]

It is assumed by Siebenthal that the metal contents of the igneous rocks are gradually dissipated in successive sedimentations. This may be offset, however, by the fact that limestones are far more easily leached by waters than the crystalline rocks.

In the publication referred to Siebenthal has compiled all available analyses of foreign and domestic waters and has shown that zinc particularly, but also copper and lead, is contained in many samples of the deeper circulation of meteoric waters. Out of 392 waters from Kentucky analyzed by A. M. Peters 89 contained zinc; of these waters 36 also contained $H_2S$ or $Na_2S$. Most of these waters were obtained from Silurian or Ordovician formations. Similar, though less extensive data are shown from Missouri waters. The zinc is carried by sulphureted salt waters and by alkaline-earth carbonate waters, the latter usually containing $H_2S$ or $CO_2$ or both. That acid waters derived from pyritic shales also contain zinc, copper, lead, and nickel is shown by the analyses given on pages 51 and 52, and these certainly demonstrate that the metals may be extracted from sedimentary silt deposits. Siebenthal finally found that reservoir deposits from fifteen deep wells of alkaline or saline type in Missouri, Kansas, and Oklahoma contained much iron sulphide as

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[1] C. E. Siebenthal, *Bull.* 606, U. S. Geol. Survey, 1915, p. 67.

[2] *Quart. Jour.*, Geol. Soc. London, 66, 1916, p. 301.

[3] E. C. Siebenthal, *op. cit.*, p. 79.

[4] Missouri Bur. Geol. and Mines, 9, 1909, p. 221.

[5] *Econ. Geol.*, 6, 1916, p. 587.

well as zinc, lead, and copper, all of them probably also present as sulphides. The dried deposits contained a maximum of 0.6 per cent zinc, 0.2 per cent lead, and 0.1 per cent copper. Zinc was present in thirteen samples, lead in eleven, and copper in nine. The waters themselves commonly yielded a trace of zinc, the greatest amount found being 0.6 part per million.

According to these investigations[1] which represent the most detailed evidence offered by those who advocate an origin from meteoric waters of the deposits under discussion, the zinc and lead existed as finely disseminated sulphides in the older Paleozoic limestone. Waters containing carbon dioxide decomposed the sulphides with the formation of bicarbonates and hydrogen sulphide. In the presence of carbon dioxide, $H_2S$ is not an effective precipitating agent, but when the moving solutions become stagnant in places suitable for deposition, $CO_2$ would escape and the remaining $H_2S$ precipitated the metals as sulphides.

The chlorides of lead and zinc are far more soluble than the bicarbonates, and strong brines of sodium chloride are undoubtedly effective in the transportation of the metals. R. C. Wells found that weak salt solutions decomposed but little zinc sulphide.[2] Stronger solutions might be more active. The theory explained is then based on the leaching of lead and zinc occurring as minutely disseminated sulphides in limestone and shale.

**Moresnet.**[3]—The Moresnet district in Belgium, Luxembourg, and Rhenish Prussia, is situated in a region of folded Devonian and Carboniferous limestones and slates cut by several large faults and covered unconformably by Cretaceous beds. In the main the ore follows these dislocations, in part as filled veins, in part as large replacement deposits in limestone at the slate contacts or at the intersection of faults. Dolomitization of the limestone is often mentioned. The ore contains sphalerite, galena, iron sulphides, and calcite; and the galena and sphalerite are often intimately intergrown. Nickel is occasionally present. Masses of calamine appeared near the surface and extended to depths of 160 feet; in some cases, notably at Vieille Montagne, they were of enormous size and reached a depth of 330 feet; the sulphides appeared at depths of 170 to 330 feet, much below the water level.[4] Concentric intergrowths, of colloidal origin, of wurtzite and sphalerite (schalenblende),[5] are very common.

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[1] E. C. Siebenthal, *op. cit.*, pp. 42–66.

[2] *Bull.* 606, U. S. Geol. Survey, 1915, p. 58.

[3] Ch. Timmerhans, Les gîtes métallifères de la région de Moresnet, Liège, 1905, p. 28.

F. Klockmann, Die Erzlagerstätten der Gegend von Aachen, Berlin, 1910.

[4] The oxidation of this deposit may be of pre-Cretaceous age.

[5] H. Ehrenberg, Der Aufbau der Schalenblenden, *Neues Jahrbuch*, Beil. Bd. 64, 1931, pp. 397–422.

The quantity of ore is said to diminish in depth, and large amounts of water are found. A considerable part of the world's production of zinc has been obtained from these deposits.

**Silesia.**[1]—Silesia, until 1919 a province of Prussia, remains one of the world's most important zinc-producing regions. The deposits are mainly in the territory transferred to Poland. The ore occurs in Triassic sandstone and limestone, which lie in flat synclines (Fig. 155). In the lower part of the "Muschelkalk" extensive dolomitization has taken place, mainly along fissures, and the dolomite is underlain by an impermeable "Sohlenstein" or clay rock. Between the two lies a thin layer of decomposed limestone (vitriol-letten), and the ore lies immediately above the latter. The ground-water circulation, dolomitization, and mineralization are all apparently closely connected. The replacement ore occurs along two horizons—the lower in a bed of dolomite several meters thick, at most 400 feet below the surface, carrying galena, sphalerite, and abundant marcasite; and the upper in a bed of smithsonite (zinc carbonate) of considerable thickness. The galena contains a little silver (0.02 to 0.03 per cent); manganese, as psilomelane, is sometimes present. The pyrite contains a little arsenic and a trace of nickel. Jordanite, a lead sulpharsenide, occurs at the Bleischarley mine. The succession is marcasite (oldest), sphalerite, and galena.[2] The sphalerite and marcasite show fine colloform structures. The ore yields about 17 per cent zinc and 5 per cent lead. In 1930 the Polish reduction works produced 174,400 metric tons of zinc.

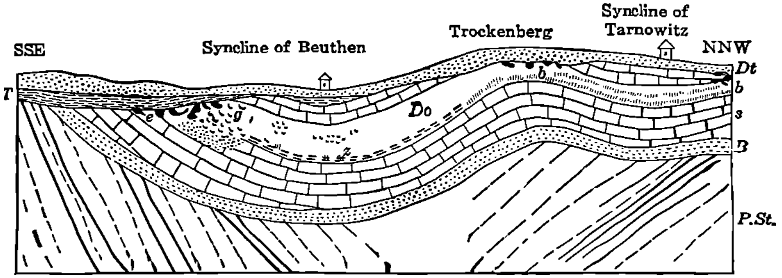

FIG. 155.—Section through the synclines of Tarnowitz and Beuthen, Silesia. *P.St.*, Carboniferous; *B*, Triassic sandstone; *s*, Triassic limestone; *Do*, dolomite; *b*, galena deposits; *z*, sphalerite deposits; *g*, oxidized zinc ores; *e*, limonite; *T*, Tertiary beds; *Dt*, Quarternary beds. (*After Gürich.*)

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[1] G. Gürich, Zur Genesis der oberschlesischen Erzlagerstätten, *Zeitschr. prakt. Geol.*, 11, 1903, pp. 202–205.

A. Sachs, Die Bildung der schlesischen Erzlagerstätten, *Centralblatt f. Min.*, 1904, pp. 40–49; *Zeitschr.* deutsch. geol. Gesell., 56, 1904, pp. 269–272.

[2] E. Schultz, *Geol. Rundschau*, 4, 1913, pp. 126–136.

There has been much discussion concerning the origin of these ores. Beyschlag and Michael[1] have shown that some ore-bearing fissures descend into the Carboniferous, and hence believe that ascending waters did the work; others, like Sachs, believe that the ores resulted from descending waters and that organic matter caused the precipitation.

In the recent literature on these ores, quoted below, there is general agreement as to the low-temperature origin of the ore, with frequent transitions from pronounced colloidal to crystalline structures. While some authors do not commit themselves as to a possible telemagmatic origin, Stappenbeck agrees in substance with Siebenthal's theory of leaching and deposition by ascending waters. Bernicke gives the paragenesis as: sphalerite (oldest), wurtzite and sphalerite, galena, iron sulphide, main galena deposition.

**Alpine Trias.**—The Alpine Trias in Austria contains a number of deposits of this type. At Bleiberg, in Carinthia, the ores occupy filled flats and gash veins. They consist of light-colored sphalerite and marcasite, with calcite and barite gangue, and a little anhydrite and fluorite, but no quartz. No silver, antimony, copper, or arsenic is present.

At Raibl, made famous by Posepny's investigations,[2] the ores form fillings and replacements along three dislocations. The minerals are sphalerite, occasionally with wurtzite and galena, with a little marcasite and chalcopyrite; and their deposition was accompanied by extensive dolomitization. Posepny describes stalactites of galena, pyrite, and sphalerite, but such occurrences are exceptional.

**Other European Localities.**—The great deposits of Santander, Spain, are contained in Carboniferous limestone and are said to be replacements connected with dislocations. The light-yellow sphalerite from these deposits is famous. Some cinnabar is present.

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[1] Beyschlag, *Zeitschr. prakt. Geol.*, 10, 1902, p. 143.

Michael, *Zeitschr.* deutsch. geol. Gesell., 56, 1904, Protocol, pp. 127–139.

R. Stappenbeck, Ausbildung, etc., der oberschlesischen Pb-Zn Lagerstätten, *Archiv Lagerstättenforsch.*, Preuss. Geol. Landesanstalt, 41, 1928, 143 pp.

P. Krusch, Ueber kolloidale Vorgänge bei der Entstehung der oberschlesischen Pb-Zn Lagerstätten, *Zeitschr. Oberschles. Berg-u. Hüttenm. Vereins*, June, July, 1929, pp. 284–290; 344–349.

Duwensee, Die oberschlesischen Pb-Zn Lagerstätten, *Metall u. Erz*, 26, 1929, pp. 481–492.

H. Schneiderhöhn, Mineralchem., etc., der Deutsch-Bleischarley-Grube, Oberschlesien, *Chemie der Erde*, 5, 1930, pp. 385–395.

F. Bernicke, Die primären Erzmineralien der Deutsch-Bleischarley-Grube, bei Beuthen, O. S., *Archiv. Lagerstättenforsch.*, Preuss. Geol. Landesanstalt, 53, 1931, 124 pp.

[2] F. Posepny, *Jahrb.* k. k. geol. Reichsanstalt, 23, 1873, pp. 315–420.

The genesis of ore deposits, *Trans.*, Am. Inst. Min. Eng., 23, 1894, pp. 197–369.

An elaborate series of illustrations of this deposit has been published by the Department of Agriculture of Austria.

At Monteponi,[1] Sardinia, large "stocks" of galena with sphalerite and pyrite are contained in Paleozoic limestones. There is much dolomitization, and a little quartz and barite occur. Cinnabar is reported and the ores contain silver in part. Igneous rocks are represented only by a diabase.

The genetic relations of the Austrian, Spanish, and Sardinian deposits are as yet uncertain. Probably they are formed by ascending waters, and may have affiliation with Tertiary igneous rocks.

**The Lead-zinc Ores of the Mississippi Valley.**—One of the most remarkable metallogenetic provinces characterized by lead and zinc ores extends over the valley of the Mississippi in the generally flat-lying limestones of the Paleozoic, ranging from the Ordovician to the lower Carboniferous (Mississippian) inclusive. These ores are found in Arkansas, Missouri, Oklahoma, Kansas, Illinois, Wisconsin, and Iowa and reach eastward as far as western Virginia and Tennessee. The ores are mined on a large scale in comparatively few regions. Small deposits of lead and zinc are widely spread and are even found in Pennsylvania, New York, and Ontario.[2] Igneous rocks are absent. There are, however, a few small deposits in southern Arkansas, Kentucky, and southern Illinois in which gold, silver, antimony, mercury, or fluorite is present and which appear to be genetically related to local intrusions of igneous rocks. The main characteristics of the predominating type are sufficiently described in the introduction to this chapter. In details they differ considerably.

In point of production the deposits in the "Tri-state field," including parts of Missouri, Oklahoma, and Kansas (Fig. 156) easily predominate. The zinc-mining industry in these areas, particularly in Oklahoma, in 1930, yielded 216,961 short tons of spelter, of a value of about $20,828,256, making about 44 per cent of the production of the United States. The lead mining in the southeastern part of Missouri in the same year produced 198,622 short tons of lead, having a total value of $19,862,200. This is 35 per cent of the lead production of the United States. The ore mined is generally referred to as "dirt"; the concentrates are spoken of as "ore." The total quantity of crude lead and zinc ore raised annually in Missouri, Oklahoma, and Kansas is now about 21,000,000 tons, consequently it is of low grade. Practically all of it is treated in concentrating works, to yield high-grade material suitable for the reduction plants.

*The Tri-state Region.*[3]—The Joplin region includes adjacent parts of Kansas and Oklahoma (Fig. 156). The principal camps are at Aurora,

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[1] R. Beck, Lehre von den Erzlagerstätten, 2, 1909, p. 257.

[2] Similar deposits have been found in Devonian beds at Great Slave Lake, Canada. J. Mackintosh Bell, *Econ. Geol.* 26, 1931, pp. 611–624.

[3] H. F. Bain (with C. R. Van Hise), Preliminary report on the lead and zinc deposits of the Ozark region, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901.

Granby, Webb City, Joplin, Galena, Baxter Springs, and Picher. In the early days lead was the only metal won, but since 1870 zinc ores have been mined and now predominate entirely. From the districts in Missouri, near Joplin, the quantity of lead recovered is about one-fourth that of the zinc. The yield of lead and zinc from the crude ore averages about 3.0 per cent zinc and 0.7 per cent lead. The concentrates average, respectively, 59 per cent zinc, and 79 per cent lead. They are smelted in Kansas and Oklahoma.

The districts are situated on the flanks of the Ozark uplift (Fig. 157). The ore occurs in the Boone formation and in rocks of Kinderhook age, both belonging to the lower Carboniferous (Mississippian). The beds form a very flat anticline pitching gradually northwest and are displaced slightly by the Seneca fault in Missouri and Oklahoma as well as by the Miami fault in Oklahoma and Kansas. All three structural features appear to be of importance in the ore deposition. The Boone formation contains much light-colored chert, especially in the Grand Falls chert member, which contains the so-called "sheet ground" deposits. The surface of the Boone contained numerous sink-holes and caves, perhaps also drainage channels, and over this "Karst" topography were deposited

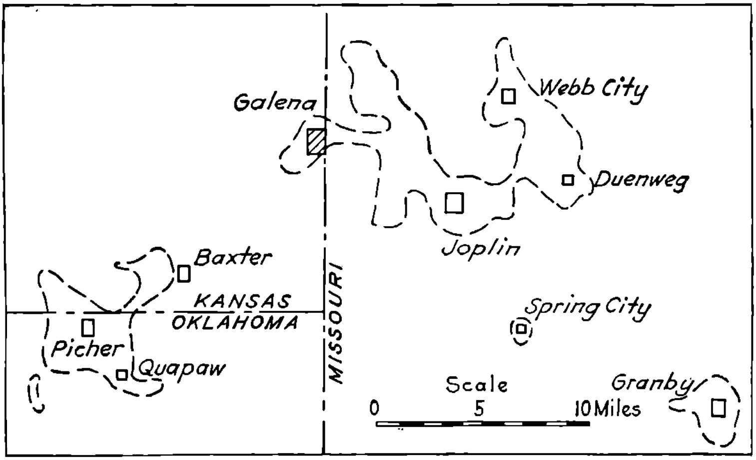

FIG. 156.—Sketch showing position of known mineralized area in the Tri-State field.

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W. S. T. Smith and C. E. Siebenthal, *U. S. Geol. Atlas*, Folio 148 (Joplin), 1914.

E. R. Buckley and H. A. Buehler, The geology of the Granby area, Missouri Bur. Geol. and Mines, 4, 1909.

C. E. Siebenthal, Contour map of the surface of the beds underlying the Cherokee shale (Picher district), U. S. Geol. Survey, 1925.

R. V. Ageton, *idem*, Quapaw district, 1931.

G. M. Fowler and J. P. Lyden, The ore deposits of the Tri-state district, *Tech. Publ.* 446, Am. Inst. Min. Met. Eng., 1931, 46 pp.

S. Weidman, The Miami-Picher zinc-lead district, *Bull.* 56, Oklahoma Geol. Survey, 1932, 177 pp.

the sandstones and shales, in part carbonaceous, of the Coal Measures (Pennsylvanian). There is, then, an unconformity by erosion. Post-Carboniferous erosion has now removed much of these rocks, but near Joplin the Pennsylvanian shale still remains in many of the old depressions (Fig. 158) and covers the whole district in Oklahoma.

The succession of the ore minerals is given by Smith and Siebenthal as follows: Dolomite (oldest), chalcopyrite, galena, sphalerite, galena, chalcopyrite, marcasite, pyrite, calcite, barite, and marcasite, the whole series being seldom found in one locality. All the minerals are frequently well crystallized. There is general agreement among the investigators that the mineralization began by dolomitization, and Bain sees in it a result of the more rapid diffusion of the magnesia in the ore-forming solutions than of the zinc. The sphalerite occurs as crystals and grains in the secondary chert which forms the gangue of the ore, the primary chert containing no metasomatic sphalerite (Fig. 159). This secondary chert, largely made up of cryptocrystalline or microcrystalline quartz, contains much organic matter with minutely disseminated sulphides,[1] and is darker than the primary chert, which antedated ore deposition. The chert is probably in part a replacement of limestone, in part, where cementing breccias, a silicified mud. It is apparently a colloidal, recrystallized deposit.

A composite sample of the zinc concentrate representing 3,800 lots has the composition shown on page 432.[2]

The galena contains only a trace of silver.

The ores are found as irregular deposits in the "broken ground" near the surface and as flat "blanket deposits," or "sheet ground," in a chert member of the Boone formation at depths of 150 to 300 feet. Below this horizon there are, as yet, un-

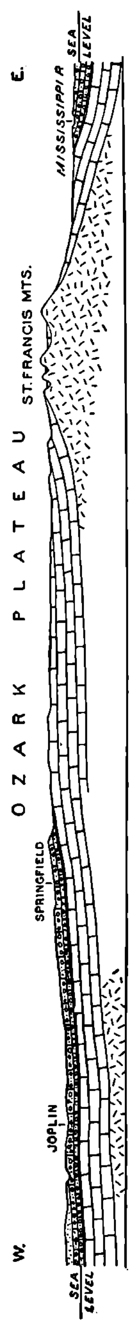

FIG. 157.—General east-west section through Joplin, Missouri, and the St. Francis Mountains, showing the flat-topped dome of Cambrian and Ordovician limestones and pre-Cambrian crystalline rocks overlain by thin Devonian shale and Carboniferous sediments. (*After Smith and Siebenthal, U. S. Geol. Survey.*)

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[1] Cox, Dean, and Gottschalk, Studies on the origin of Missouri cherts and zinc ores, *Bull.* Univ. Missouri School of Mines., Tech. Ser. 3, No. 2, 1916, 34 pp.

[2] W. G. Waring, The zinc ores of the Joplin district, *Trans.*, Am. Inst. Min. Eng., 57, 1918, pp. 657–670. Waring has also found thallium, indium, gallium, and germanium in the flue dust and in the zinc metal.

worked deposits of disseminated ore of doubtful value. The ore occurs mainly as fillings of cavities, the fillings of distinct veins or crevices being subordinate. The ore minerals, with secondary chert, fill spaces of brecciation or solution cavities along the stratification, perhaps also spaces of discission in limestone caused by stretching and adjustments.

ANALYSIS OF ZINC CONCENTRATE FROM THE JOPLIN REGION

| | | | |
|---|---|---|---|
| Zinc | 58.260 | Sulphur | 30.720 |
| Cadmium | 0.304 | Calcium carbonate | 1.880 |
| Lead | 0.700 | Magnesium carbonate | 0.850 |
| Iron | 2.230 | Barium sulphate | 0.820 |
| Manganese | 0.010 | Silica | 3.950 |
| Copper | 0.049 | | |
| | | | 99.773 |

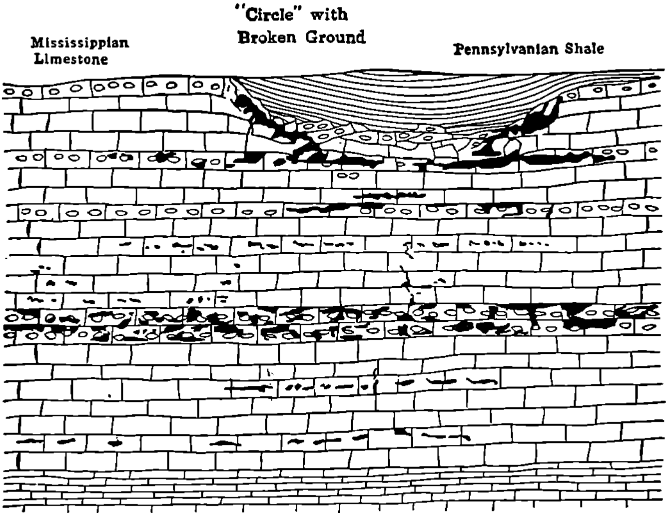

FIG. 158.—Diagram of zinc-lead deposits at Joplin showing "broken ground" around "circle" near surface and "sheet ground" deposit in Grand Falls chert member below. Black areas represent ore. Scale 100 feet-1 inch.

In the "broken ground," which extends for 100 or 150 feet below the surface, the ores occur in clayey chert breccias in old sink-holes filled with Pennsylvanian sediments, or along the outside of such sink-holes, forming "circles" where the slipping and settling provided open ground (Fig. 158). In these occurrences the galena predominates, partly because of solution and oxidation of sphalerite, and partly because the deposition of galena prevailed at these upper levels, below the shale. Large masses of galena are found here, in contrast to the conditions in the sheet ground.

Both at Joplin and at Aurora (Fig. 160), as well as in camps in Oklahoma, the "runs" are also a characteristic form of the upper deposits; these sometimes extend for 1 or 2 miles, following the same horizon at depths less than 250 feet—usually much less. The width of the run is rarely more than 50 to 150 feet, the height ranges from a few feet up towards 80 feet. These runs appear to be determined by solution cavities along joint systems in the rocks.

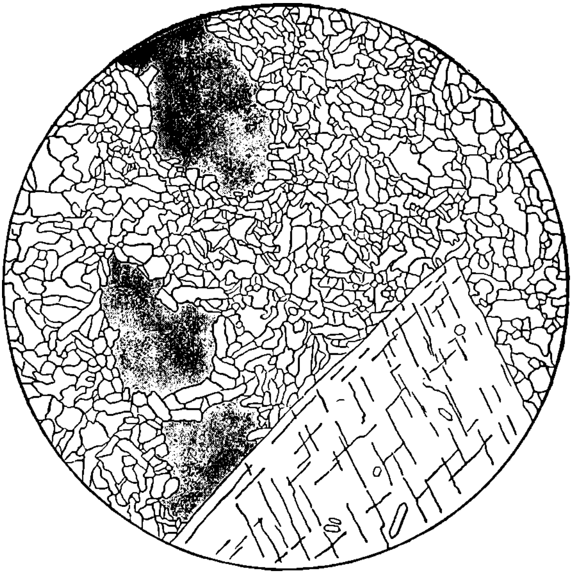

FIG. 159.—Thin section of "black chert" showing matrix of fine-grained quartz with grains of sphalerite (shaded) and crystals of dolomite. Note quartz crystals developing in dolomite. Magnified 53 diameters. (*After Smith and Siebenthal, U. S. Geol. Survey.*)

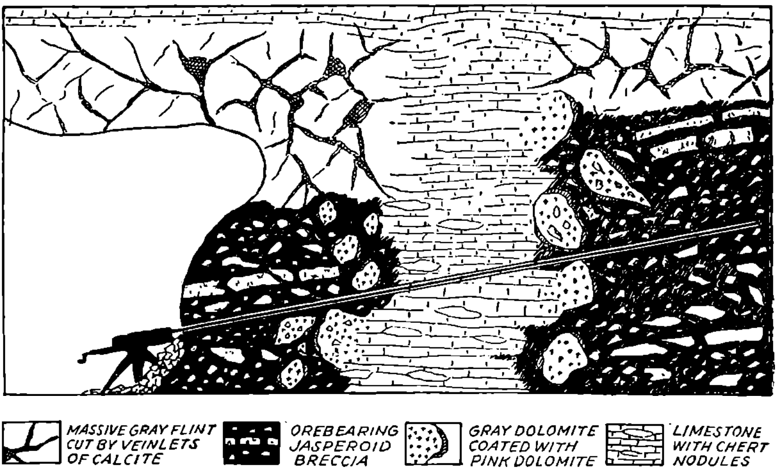

FIG. 160.—Detail of section through limestone bar between runs, showing relation of dolomite to ore-bodies. (*After N. F. Netzeband.*)

While in places the brecciation and mineralization continue down to the blanket deposits of the sheet ground, the latter extends in the main independently of the old pre-Pennsylvanian surface. In this sheet ground, which is from 6 to 15 feet thick, the galena and sphalerite occur in dark chert, filling brecciated old chert, and in elongated, narrow solution cavities due to dissolved streaks of limestone in the prevailing chert. The sheet ground is poor and is not the most important source of ore (Fig. 162).

The deposits in northeastern Oklahoma, at Picher[1] (Fig. 161), are now by far the most productive. Within an area of about 30 square miles there were (1925) 132 mines, which in that year, yielded 272,000 tons of zinc and 78,000 tons of lead. Since 1908 this district is said to have produced zinc and lead to the astonishing value of $800,000,000. Few mines are deeper than 250 feet. An individual mine or lease covering say 40 acres is soon worked out. There are, however, possibilities for an extension of the field. There are 100 mills equipped for wet concentration supplemented by flotation. The mills are as ephemeral as the mines.

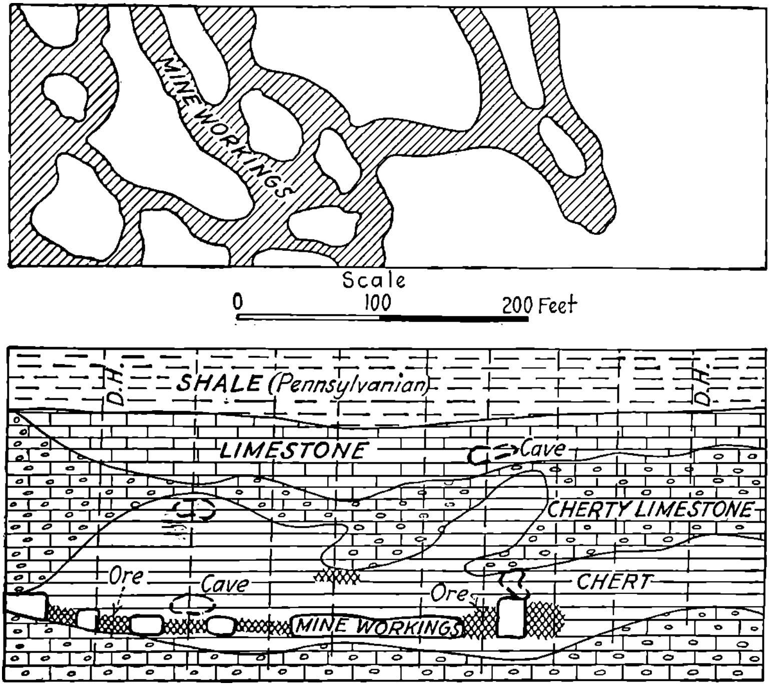

FIG. 161.—Plan and section showing workings in typical lead-zinc mine at Picher, Oklahoma.

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[1] C. E. Siebenthal, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 187–228.

Maps by H. A. Buehler in "War minerals of the Joplin district," Am. Inst. Min. Eng., Joplin-Miami meeting, October, 1917.

The production for 1930 was 4,100,000 tons of crude ore averaging 3.08 per cent zinc and 0.71 per cent lead.

The ores are similar to those of Joplin. Chert, dolomite, and calcite are the gangue minerals; sphalerite and galena the principal ore minerals, with a little chalcopyrite and locally marcasite. An analysis of a large composite sample of concentrate shows in per cent 58.40 Zn, 1.80 Fe, 0.92 Pb, and 0.42 Cd. There is no evidence of colloidal deposition or colloform structures, the ore minerals being well developed and often beautifully crystallized. The ore lies in the Short Creek oölite horizon and about 200 to 250 feet below the surface. It is closely connected with the extensive chert replacement, and where there is no chert the limestone contains little ore. The runs, while very irregular, have a tendency to extend northeast and follow irregular zones of fracture.

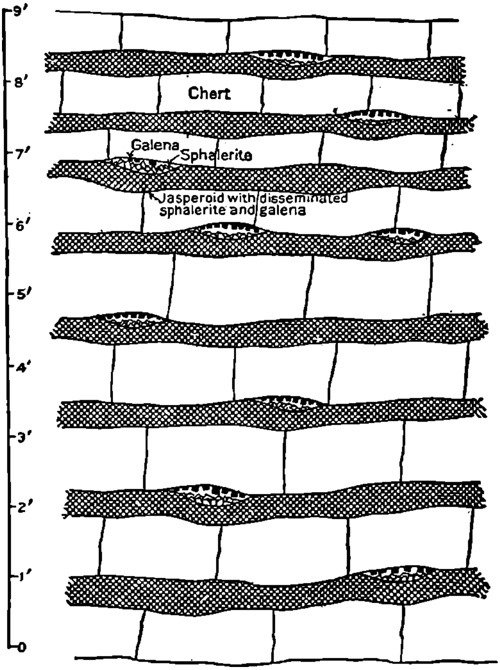

FIG. 162.—Sketch illustrating the occurrence of galena and sphalerite in cavities in the sheet ground, Joplin, Missouri. (*After C. E. Siebenthal, U. S. Geol. Survey.*)

In few places does the ore reach up to the base of the shale; generally it lies about 100 feet below this erosion surface. The shale is from 100 to 150 feet thick and contains no ore. The runs are up to 50 or 70 feet wide, usually, less, and have a "face," or height of from a few feet up to 80 feet. In the case of large runs the tonnage becomes very large. The ore forms brecciated masses, fragments of white (early) chert being cemented by "black chert," dolomite, calcite, and sulphide (Fig. 160).

In part there is "open ground"; in many places the ore occurs along the obscure stratification, in lenticular openings or solution cavities. Often there are "runs" on two or more levels, say within 100 feet, and the upper and lower runs may connect.

There is evidence of an ascending circulation of water in places heavily charged with hydrogen sulphide. Bitumen is abundant in places. The whole occurrence supports strongly the theory of origin by ascending waters. Possibly a lower ore horizon of "sheet ground" at the level of the Grand Falls chert, may be developed. A few mines are operating in this horizon, but it is not believed to contain deposits as large as those now worked.

**Arkansas.**—In northern Arkansas,[1] a short distance southeast from the Joplin region, the sphalerite generally without galena, occurs in fissures or crevices, in fault breccias, and in solution breccias, accompanied by secondary chert or dolomite, sometimes also by crystallized quartz. The ores are found in Ordovician limestone and also in the Boone (Mississippian) formation.

**Upper Mississippi Valley.**[2]—The districts of the upper valley lie in Wisconsin, Iowa, and Illinois. The important districts are in Wisconsin and Illinois and yielded in 1930 about 1,800 short tons of lead and 12,600 short tons of zinc. The ore deposits are found in Ordovician strata of almost horizontal position. The following formations are recognized:

| | Feet |
|---|---|
| Cincinnati or Maquoketa shale | 160 |
| Galena dolomite | 240 |
| Platteville limestone (Trenton) | 55 |
| St. Peter sandstone | 100 |
| Lower magnesian limestone | 350 |

Below the magnesian limestone is 700 feet of the Cambrian Potsdam sandstone. The so-called "oil rock," a thin bed of bituminous shale, is found at the base of the Galena formation or at the top of the Trenton. The rocks dip gently southwest and are flexed into very shallow troughs.

The ores are confined to the Galena limestone and the upper part of the Platteville limestone; the minerals consist of marcasite, sphalerite,

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[1] G. I. Adams, *Prof. Paper* 24, U. S. Geol. Survey, 1904.

J. C. Branner, Arkansas Geol. Survey, 5, 1900.

E. T. McKnight, Lead-zinc deposits of northern Arkansas, *Press Bull.*, U. S. Geol. Survey, Nov. 28, 1932. Ore said to contain a little enargite.

[2] T. C. Chamberlin, Geology of Wisconsin, 4, 1882.

H. F. Bain, *Bull.* 294, U. S. Geol. Survey, 1906.

U. S. Grant, *Bull.* 14, Wisconsin Geol. Survey, 1906.

G. H. Cox, *Econ. Geol.*, 6, 1911, pp. 427–448, 582–603.

H. C. George, *Trans.*, Am. Inst. Min. Eng., 59, 1918, pp. 117–150.

and galena, deposited in the order noted. The gangue is crystallized calcite, rarely barite. Cadmium is absent, but a trace of silver is found. The abundance of marcasite causes metallurgical difficulties and necessitates treatment of the concentrates in electrostatic or magnetic separators, in the latter case preceded by partial roasting. The ores occur as fillings of open spaces, vertical crevices of varying strike, or "gash veins" connected with "pitches" or "flats," all probably due to solution along joint planes and tension cracks (Fig. 163). Stalactites of sulphides are sometimes found indicating that the spaces were not always filled by solutions. In part there are also flat bodies of disseminated ores. The galena predominates near the surface, probably largely because the sphalerite has been dissolved as sulphate and transformed to silicate or carbonate in the lower levels. In depth sphalerite with a little galena is the principal ore. Mining operations extend to a depth of at most 200 feet. The distribution of the oil shale, according to Bain, seems to coincide with the extent of the deposits. Cox, however, holds that the metal was derived from the overlying Maquoketa shale and carried down to be concentrated in the Galena limestone.

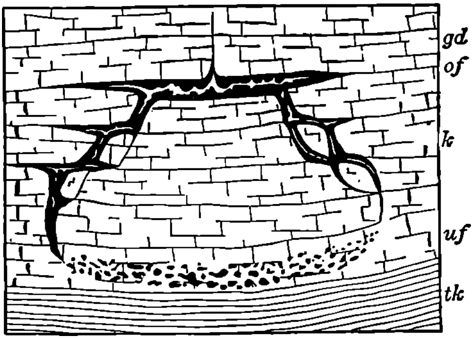

FIG. 163.—Section showing occurrence of lead and zinc in vertical crevices, flats, and pitches; also of disseminated ores in the rocks. *gd*, Galena dolomite; *tk*, Trenton limestone; *of*, upper flat; *uf*, lower flat; *k*, connecting flats, pitches, and verticals. (*After T. C. Chamberlin.*)

**Virginia and Tennessee.**[1]—In western Virginia, and near Knoxville, Tennessee, lead and zinc ores occur in the Cambro-Ordovician (Shenandoah) limestones, mostly where the rocks are faulted or brecciated. The

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[1] T. L. Watson, Lead and zinc deposits of Virginia, *Bull.* 1, Geol. Survey Virginia, 1905.

Frank L. Nason, Characteristics of zinc deposits in North America, *Trans.*, Am. Inst. Min. Eng., 57, 1918, pp. 830–855.

H. A. Coy and J. A. Noble, Mining methods at Mascot, Tenn., *idem*, 72, 1924, pp. 55–76.

M. H. Secrist, Zinc deposits of East Tennessee, *Bull.* 31, Tennessee Geol. Survey, 1924.

gangue consists of calcite, dolomite, and rarely barite. There is little quartz or pyrite; sphalerite predominates and is of very light color.

**Southeastern Missouri.**[1]—In eastern Missouri not far from the Mississippi River and south of St. Louis, lead mining has been carried on more or less extensively since the early part of the eighteenth century; but in the last years the industry has assumed very large proportions, the St. Joseph Lead Company being the principal producer. In 1930 the yield of lead from this region was 278,000 short tons, worth $16,559,000. The crude ores, which yield on the average 3.4 per cent of lead, are treated at the rate of 20,000 tons per day in large concentrating works and a part of the galena is smelted in the district. Practically no zinc is contained in the ore.

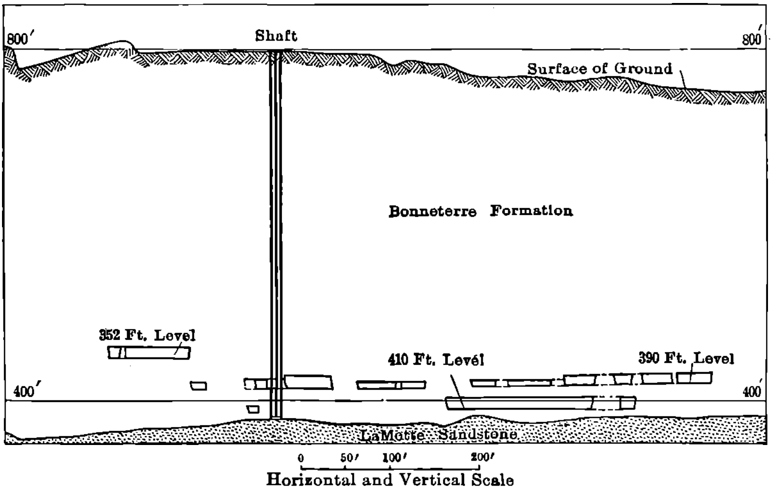

FIG. 164.—Vertical section showing workings in mine No. 4, Federal Lead Company, southeastern Missouri. Horizontal bodies of disseminated ore, following bedding of shaly dolomite of Bonneterre formation. (*After E. R. Buckley.*)

The geological position of the deposits is in the Cambro-Ordovician, and therefore lower than those of the other Mississippi Valley ores. On an irregular surface of pre-Cambrian granite and porphyry rests the basal La Motte sandstone, about 200 feet thick (Fig. 164). Above this lies the arenaceous dolomite of the Bonneterre formation, often chloritic, with beds of shale, having in all a thickness of 300 to 400 feet. Covering the Bonneterre are the Davis, Derby, Doe Run, and Potosi formations, mainly dolomites and shales.

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[1] A. Winslow, Missouri Geol. Survey, 6 and 7, 1894.

C. R. Keyes, *idem*, 9, 1896.

E. R. Buckley, Geology of the disseminated lead deposits, Missouri Bur. Geol. and Mines, 9, pts. 1 and 2, 1909.

A. P. Watt, Concentration practice in southeastern Missouri, *Trans.*, Am. Inst. Min. Eng., 57, 1918, pp. 322–419.

The principal ore horizon is in the lower part of the Bonneterre dolomite, though some galena occurs throughout that formation. A second, less important ore horizon is in the Potosi dolomite, where the galena is accompanied by barite. The strata are horizontal or have very gentle dip.

The ore minerals are mainly galena accompanied by calcite, a little pyrite, and sometimes chalcopyrite. In places—for instance, at Mine La Motte and Fredericktown—the ores contain nickel and cobalt, as linnæite $(Co,Ni)_3S_4$; some of the ores have been worked for these metals. Watt quotes a representative analysis of the crude ore of the southeastern district.

ANALYSIS OF DISSEMINATED ORE FROM SOUTHEASTERN MISSOURI

| | | | |
|---|---|---|---|
| Pb | 4.32 | $Al_2O_3$ | 1.16 |
| Cu | 0.03 | CaO | 30.80 |
| Zn | 0.50 | MgO | 17.96 |
| S | 0.97 | $CO_2$ | 32.79 |
| $SiO_2$ | 4.83 | | |
| $Fe_2O_3$ | 6.64 | | 100.00 |

Silver 0.12 oz. per ton. Trace Ni, Co, Mn.

Watt states that the silver follows the zinc. Concentrates of sphalerite contain up to 10 ounces of silver per ton.

The ores are often called disseminated, for the galena usually occurs as grains or crystals disseminated in the greenish-gray dolomite (Fig. 165); sometimes these crystals are several centimeters in diameter. According to Buckley the ores of the lower part of the Bonneterre occur as follows:

1. As horizontal sheets along bedding planes, generally along the upper side of thin shale beds.
2. Disseminated in dolomite.
3. Filling or lining joints.
4. In cavities or vugs.

The galena is persistently associated with dark dolomite and black shale.

FIG. 165.—Crystals of galena developing in shaly dolomite. Black, galena; shaded and stippled, shaly dolomite; white, quartz. Magnified about 10 diameters. (*After E. R. Buckley.*)

The ores are mined from vertical shafts, 100 to 550 feet deep. The ore does not extend in all directions like a coal bed, but the flat shoots or "runs" follow rather persistently one or two directions, undoubtedly controlled by joints and small faults. Some of these runs have been followed for miles and may be several hundred feet wide. Some of the mine workings in the Bonneterre district are 100 feet high.

**Genesis of the Mississippi Valley Deposits.**—An unusually extensive literature, full of controversy and divergent views, covers the question of the genesis of these ores. A majority of the authors agree that the source of the ores was in the Paleozoic sedimentary beds, also that the deposition was effected by atmospheric waters. At present there are three strongly contrasting opinions regarding the Missouri deposits. The descensionists are represented by Whitney, Chamberlin, Blake, Robertson, Winslow, Buckley and Buehler and the ascensionists by Van Hise, Bain, W. S. T. Smith and Siebenthal. Buckley and Buehler hold that the source of the lead and zinc was in the Pennsylvanian sediments, which, however, contain no important deposits and only in places small amounts of galena and sphalerite. The finely distributed sulphides were dissolved as sulphates and carried downward in acid solutions which finally mingled with neutral or alkaline solutions from the unoxidized parts of Pennsylvanian sediments. These mingled waters deposited galena and sphalerite in the sink-holes and drainage channels of the underlying Mississippian limestone and chert. The obvious difficulties in their theory are that it requires the waters to have descended through an impervious shale cover and that the shale seems utterly incompetent to furnish the enormous amounts of metals required. Decidedly, the ascensionists have the best arguments.

A similar theory was advocated by Buckley for the lead deposits of southeastern Missouri.

The ascensionists believe that the source of the ores is in the various formations below the Pennsylvanian, particularly in the Cambro-Ordovician, and that atmospheric waters penetrating these rocks were carried up towards the impervious beds of the Pennsylvanian and here deposited the ores in the pre-Pennsylvanian breccias and sink-holes. Smith and Siebenthal hold that the ores were formed much later than the Pennsylvanian, after the Ozark uplift (Fig. 157) had established an artesian circulation. The surface waters entered the older Paleozoic outcrops to the south and east of the Joplin region. After following these beds they passed upward through the jointed and brecciated Mississippian limestone until they reached the vicinity of the impermeable shales. There is, indeed, in the deep wells of Joplin as well as at Picher and in southeastern Wisconsin, good evidence of the existence of artesian pressure. Siebenthal's strong arguments in favor of the artesian theory have already been reviewed in the general part of this chapter.

An interesting hypothesis bearing on this subject has been advanced by E. S. Bastin.[1] Certain deep waters in the oil fields contain anaërobic bacteria capable of reducing sulphates and generating hydrogen sulphide. If now solutions carrying metals in balance with the sulphate, chloride, or

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[1] A hypothesis of bacterial influence in the genesis of certain sulphide deposits, *Jour. Geol.*, 34, 1926, pp. 773–792.

bicarbonate radicles come in contact with relatively stagnant $H_2S$ bearing water, precipitation as sulphides must occur. The importance of this hypothesis is clear, and it is particularly applicable to the Oklahoma area.

For the deposits of Wisconsin and Illinois, Van Hise and Bain assume that the metals were minutely disseminated as sulphides through the Galena dolomite and concentrated, probably in late Tertiary time, by the action of surface waters descending in shallow troughs through the fractured Galena limestone; and that the reduction was effected by the organic matter of the oil rock. In the Galena limestone the solutions were practically confined between two beds of almost impermeable shale.

According to Cox, the metals were derived from the overlying Maquoketa shale, in which he finds some evidence of the presence of sulphides, particularly sphalerite. He points out, with good reason, that the shales are shore formations in which the metallic substances derived from adjacent continents would most easily be deposited as detritus or precipitated from solutions.

There is a third opinion on the Mississippi Valley deposits which was early advocated by Jenney and by Nason. According to these men, thermal waters ascended from regions below the Paleozoic beds, and the source of the ore is deep-seated. More recently W. A. Tarr, who studied the barite-galena veins of Washington County, Missouri (p. 369), held that the ores are of magmatic origin. In 1915, L. V. Pirsson[1] expressed the view that the zinc-lead deposits of the Mississippi-valley type might well result from "the quiet upward movement of volatile magmatic material." Still more recently, J. E. Spurr expressed the same opinion. Beginning with the galena-fluorite veins of the Kentucky-Illinois field (p. 634), he later expanded his views to the Joplin, Wisconsin, and southeastern Missouri fields,[2] and explained the ore deposition as the result of a "magmatic surgence" of igneous material which rarely reached the surface, though the "ore magmas" or solutions (both words being used) derived from these hypothetical igneous rocks deposited their content in favorable places. One of the strongest arguments against this view is that the ores have never been traced in depth; in other words the avenues of approach can not be found. On the other hand the enormous quantity of ore at these places, particularly at Picher, seems difficult to reconcile with a leaching of limestone and still more with the views of the descensionists.

In 1932, Fowler and Lyden[3] presented a paper containing a renewed study of the Tri-state district. A detailed subdivision was given of the

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[1] Origin of certain ore deposits, *Econ. Geol.*, 10, 1915, pp. 180–186.

[2] *Eng. Min. Jour.-Press*, Feb. 9, 1924; Dec. 18, 1926; Jan. 29, 1927.

[3] The ore deposits of the Tri-state district, *Tech. Publ.* 446, Am. Inst. Min. Met. Eng., 1932, 46 pp.

Boone formation; the ore deposition was connected with a series of shear and tension fissures, which is more emphasized than in the earlier literature. The presence of an intrusive granite in the Paleozoic of Kansas[1] was pointed out and the ore deposits were considered as of probable magmatic origin.

In the same year S. Weidman published an account of the Oklahoma zinc-lead deposits and came to a similar conclusion. He mentions a drill hole at the Bird-Dog mine which encountered granite at a depth of 1,246 feet. This granite, he states, is intrusive and probably of Mesozoic age.

Later C. K. Leith[2] held that there is only slight evidence of deep-reaching regional stresses, and, in agreement with the views of Buehler, that most of the structures resulted from settling during the consolidation of the strata on an uneven basement. The Grand Falls chert is held to represent silicification on an old erosion surface.

Buehler points out that the deposits follow the depressions in the pre-Cambrian bed rock; none occur on the ridges (determined by magnetometric observations).

Still more recently, W. H. Newhouse[3] showed that the galena from the Tri-state field contained many large inclusions of a strong NaCl solution and considered these as part of the original mineralizing fluid. This, taken in connection with Siebenthal's data and particularly with the lead-zinc sediments in the saline sulphur springs of Arkansas,[4] seems to prove definitely that sodium chloride solutions with $H_2S$ and probably $CO_2$ were the cause of the ore deposition. Naturally, this does not solve the problem of the origin of the water. The very large volumes of water needed can scarcely have been of magmatic origin. If the metals ascended from the supposed intrusions in depth, the solutions were no doubt very liberally mixed with waters of meteoric origin.

W. H. Emmons[5] emphasizes that all the deposits occur in Paleozoic rocks and in some relation to fault lines; also that they are of Late Paleozoic age and are formed by ascending magmatic solutions related to underlying ("cryptobatholitic") intrusions. They are thus telemagmatic deposits. He believes there is evidence of horizontal zoning as follows: (1) Iron ores with specularite, pyrite, and quartz; (2) iron ores with pyrite, chalcopyrite, and in places a little lead and zinc; (3) zinc-

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[1] W. H. Twenhofel, Intrusive granite of the Rose dome, Kansas, *Bull.* Geol. Soc. Am., 37, 1926, pp. 403–412.

[2] Structures of the Wisconsin and Tri-state lead and zinc deposits, *Econ. Geol.*, 27, 1932, pp. 405–418.

[3] *Econ. Geol.* 27, 1932, pp. 419–436.

[4] *Econ. Geol.*, 9, 1914, pp. 758–767.

[5] William H. Emmons, The origin of the deposits of sulphide ores of the Mississippi Valley, Econ. Geol., 24, 221–271, 1929.

lead ores with chert, dolomite, and barite; (4) manganese ores (Batesville, Arkansas). The paper is a very valuable contribution, though not entirely convincing. Many geologists believe that the deposits are of Cretaceous age.

The problem is not yet solved, but looking beyond these controversies, we cannot deny that in many countries transition types appear which seem to connect these apparently distinct non-igneous deposits with deposits of igneous affiliations.

**Northern Rhodesia.**—Important deposits of zinc, lead, and vanadium have been opened by the Rhodesia Broken Hill Company on the railroad 100 miles south of the Congo border.[1] From the scant information available these ores occur in the same dolomite which farther north carries the Katanga copper deposits. The ores are said to occur as replacements along "more or less vertical bedding planes and fissures," and extend over a considerable area. To a depth of 100 feet or more oxidized ores prevail with many beautifully crystallized species of calamine, cerussite, and smithsonite, and a host of rare phosphates, vanadates and arsenates of zinc and lead. The deeper ores, developed to 350 feet, contain sphalerite and galena with a little pyrite. The vanadium plant was completed in 1930 and is expected to produce 570 tons of ferrovanadium per annum. If this is realized, it will make the deposit the largest source of vanadium next to Mina Ragra. There are no igneous rocks present, "within miles."

There is much uncertainty about the genetic position, but it is clearly analogous to the vanadium ores near the Tsumeb mine, Southwest Africa.

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[1] *Mining Mag.*, October, 1919, Mar. and Apr. 1927. See also *Guidebook* 22, 15th Int. Geol. Congress, 1930.

# CHAPTER XXIV

# EPITHERMAL DEPOSITS

## METALLIFEROUS DEPOSITS FORMED NEAR THE SURFACE BY ASCENDING THERMAL WATERS AND IN GENETIC CONNECTION WITH IGNEOUS ROCKS

### CHARACTER AND ORIGIN

**General Features.**—The deposits at the orifices of hot ascending springs have been described in Chapter VI. It has been shown that they consist of opal, chalcedony, quartz, calcite, aragonite, barite, and fluorite, with a number of other gangue minerals; and that they also contain in places metallic gold and certain sulphides, such as cinnabar, stibnite, and pyrite, but not the other common ore minerals such as chalcopyrite, galena, sphalerite, and arsenopyrite. The ore deposits described in the present chapter present some striking analogies to those products of the hot springs.

In regions of comparatively recent volcanic activity where the measure of erosion since the eruptions ceased is in hundreds rather than in thousands of feet, we find a group of important ore deposits, usually in the form of fissure veins. They generally occur in igneous flow rocks[1] and most commonly in andesite, latite, trachyte, and rhyolite. They are rarely found in basalts. But they also cut the underlying or adjacent formations. They constitute the source of a large part of the world's production of gold, silver, and quicksilver, and contain the spectacular bonanzas of the Cordilleran region, of which examples are found at Tuscarora, Virginia City, Goldfield, Cripple Creek, Pachuca, Guanajuato, and many other districts. Following the Tertiary outbursts of effusive rocks, these deposits accompany the "circle of fire" that encompasses the Pacific Ocean. We find them in Japan, in the East Indian Islands, and in New Zealand. They are characteristically developed in that classical mining region of the Old World, in Hungary and Transylvania,

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[1] We are accustomed to consider as intrusive rocks those which have congealed with granular texture far below the surface. Intrusions are, however, not confined to any particular depth or texture. Intrusive bodies may be found in any series of rocks even near the surface and may then have fine-grained, trachytic or even glassy texture. The distinction between flows and intrusions may in such cases become difficult; and, as shown in case of the Tonopah, Waihi, and other districts, the relations may have far-reaching bearing upon the richness and continuation of the deposits contained in such a series of rocks.

where one of the elements—tellurium—which so often accompanies them, was first found.

Though most of these ore-deposits are found in the Tertiary flow rocks they are not confined to rocks of this period. There is good reason to believe that veins are developing now in some regions of recent volcanism, and that similar veins have been formed during pre-Tertiary outbreaks, although erosion has removed most of the older representatives of this type. These deposits have certain well-marked characteristics which are partly of a mechanical, partly of a chemical origin.

Because the fissuring of the rocks took place near the surface, under slight load, open cavities were abundant, and filling, crustification, and comb structure are conspicuous. The walls are likely to be irregular, and the vein matter is often "frozen" to the walls. Splitting, chambering, and brecciation are features of the veins. While metasomatic processes have been active in the surrounding rocks, the ore is usually confined to the open fissures. Short and irregular veins are more frequent than the regularly developed conjugated fractures resulting from strong compressive stress. Divergent systems of fractures or several parallel systems with little apparent relationship are thought to be due to the gravitative settling of volcanic piles.

Banding caused by crustification is common, as illustrated in Figs. 180, 187, 193, 197, and 204. It is much more delicate and frequent than in deposits formed at greater depth and higher temperature.

The occurrence of the ore in "stockworks," or in pipes, or below impervious beds is often observed. In superimposed lava flows of different kinds, some are usually better adapted to the deposition of ore than others, and this difference may result in the development of ore-shoots which are approximately horizontal.

Among the metals contained in these deposits gold and silver are by far the most important. Base metals are present, plentifully enough in places, but the mines are rarely worked for these. Large bodies of galena and sphalerite occur in some places, but it is decidedly rare to find important copper deposits. The "pyritic" deposits are not represented; they are confined to the deeper zones or to those of higher temperatures. Arsenic and antimony, bismuth, tellurium, and selenium are common but are rarely of economic importance; quicksilver is present in some deposits and indeed the typical quicksilver deposits belong to this class. Cobalt and nickel, tungsten, and molybdenum are not unknown, but are entirely subordinate. Their home is in the deeper deposits.

The pure gold deposits are relatively scarce. Those carrying silver only are common in certain regions, like Mexico. The usual metals are gold and silver occurring together in varying proportions.

Among the ore minerals, native gold should be mentioned first. It contains silver, as a rule, and is of pale yellow color; a proportion sometimes occurring is ounce for ounce when the mineral is of very pale grayish-yellow color (electrum). Deep yellow gold is not unknown, however. The gold is often present in very fine mechanical distribution, being sometimes so closely intergrown with ore minerals and gangue that no colors can be obtained by panning. When derived by oxidation of tellurides it is of dull brown color and is difficult to recognize even in rich specimens. The whole series of tellurides is present. As the gold generally occurs in minute particles rich placers below the outcrops of these deposits are rather unusual.

Native silver is ordinarily a product of oxidation. The primary and most abundant silver mineral is argentite; complex silver sulphantimonides and sulpharsenides are also characteristic; it is often difficult to say which are secondary (supergene) and which are primary (hypogene). Among them are proustite, pyrargyrite, miargyrite, stephanite, polybasite, tetrahedrite, and, more rarely, enargite.

Stibnite is plentiful in deposits of certain types. Among the base minerals pyrite is always present, but in small quantity and fine distribution; marcasite, a mineral typical of deposition near the surface, is not unusual; often it is secondary. There are also galena, sphalerite, chalcopyrite, and sometimes alabandite; rarely arsenopyrite; never pyrrhotite or magnetite.

Of gangue minerals quartz is the most abundant, and crystals of it are plentiful but rarely large; an amethyst color is often noticeable. The quartz aggregates are not glassy or milky but usually fine-grained (hornstone) and often chalcedonic, with rapidly changing grain. In many cases the quartz is a crystallized colloid and the original, delicate colloform banding is still visible.[1] Calcite, dolomite, barite, and fluorite are locally the dominant gangue minerals, while siderite is rare. Manganese minerals, like rhodochrosite and sometimes rhodonite, are typical of certain groups. Kaolin accompanies the veins, sometimes in large amounts, but is probably in most cases a product of secondary changes by descending waters. The alteration of the country rock is propylitic with chlorite, carbonates, pyrite, less abundantly sericite as secondary minerals. In limestones and in siliceous rocks there may be silicification. Zeolites occur in a few deposits.

One of the most widespread and characteristic gangue minerals and the most difficult to explain is adularia (or valencianite).[2] Discovered by Breithaupt in specimens from the Valenciana mine at Guanajuato,

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[1] Such quartz from many districts is described by S. F. Adams, A microscopic study of vein quartz, *Econ. Geol.*, 15, 1920, pp. 623–663.

[2] A. Breithaupt, Ueber die Felsite und einige neue Specien ihres Geschlechts, *Schweigg. Jour.*, Bd. 60, 1830, p. 322.

W. Lindgren, Orthoclase as a gangue mineral in fissure veins, *Am. Jour. Sci.*, 4th ser., 5, 1898, p. 418. *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, p. 167.

Mexico, this mineral has since been found in numerous other places, mainly in the Cordilleran region, as part of the filling, and as a metasomatic product in the country rock. Among the places where this feldspar plays an important part may be mentioned Silver City (Idaho), Tuscarora, Tonopah, and Rawhide (Nevada), Gold Road (Arizona), Republic (Washington), and Cripple Creek and Creede (Colorado). It does not occur at Goldfield, Nevada, where solutions of acid reaction appear to have deposited the ore. The mineral is usually a pure potassium feldspar, although varieties with several per cent of sodium have been found at Waihi, New Zealand, and in the Gold Spring district, Utah.[1] Sometimes the adularia replaces orthoclase, biotite, and other rock minerals (Fig. 166); it is also found in the form of well-developed crystals of prism and dome intergrown with vein quartz (Fig. 167). The cross-sections of the adularia crystals are usually of rhombic shape. The mineral also occurs abundantly in some veins that had originally a calcite gangue, now replaced by an intimate intergrowth of adularia and quartz.

Fig. 166.—Adularia (*ad*) replacing soda-lime feldspar (*An*) in andesite from Tonopah, Nevada. Magnified 17 diameters. (*After J. E. Spurr, U. S. Geol. Survey.*)

The high-temperature minerals, such as augite, amphibole, olivine, biotite, tourmaline, topaz, garnet, magnetite, ilmenite, and chromite, are conspicuously absent.

**Successive Phases of Mineralization.**—Veins formed near the surface in volcanic regions are sometimes subject to peculiar changes, which are rarely observed in deposits of more deep-seated origin. An earlier

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[1] B. S. Butler, *Prof. Paper* 111, U. S. Geol. Survey, 1920, p. 565.

gangue mineral, such as calcite or barite, may be wholly wiped out and replaced by a new gangue of quartz and adularia. This alteration has nothing to do with surface waters, though the latter by the solution of calcite may sometimes produce a similar cellular or lamellar structure; it is plainly caused by a change in the composition of the ascending currents. Indications of this process may be seen even where it has not been

FIG. 167.—Intergrowth of quartz (*q*) and adularia (*a*), Fraction vein, Tonopah, Nevada. Magnified 38 diameters. (*After J. E. Spurr, U. S. Geol. Survey.*)

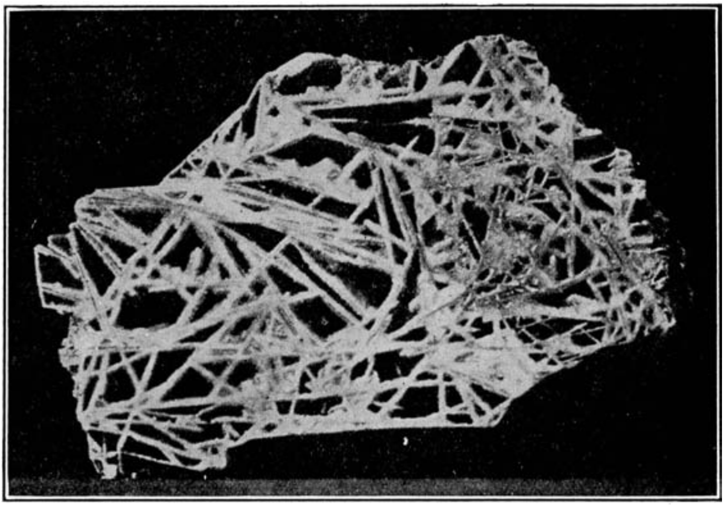

FIG. 168.—Section of lamellar ore, De Lamar, Idaho. Natural size.

carried to completion. In many veins at Cripple Creek deposition began by the growth of slender crystals of celestite from the walls, and these crystals were subsequently replaced by quartz, in which the pseudomorphs are now embedded.

In many instances the vein was completely filled by calcite, each grain separated by a slender partition of quartz; at the beginning of the second stage this calcite was dissolved, leaving a skeleton of thin silica walls; secondary quartz and often also adularia were deposited upon these walls, giving them more strength, but the ore remains a delicate aggregate of "hackly" or lamellar quartz, such as is characteristic of some mining districts. At De Lamar, Idaho, this ore consists only of quartz (Fig. 168). In the veins at Oatman, Arizona, the original gangue material consisted of calcite and fluorite and the "pseudomorphic" ore consists of quartz and large amounts of adularia.

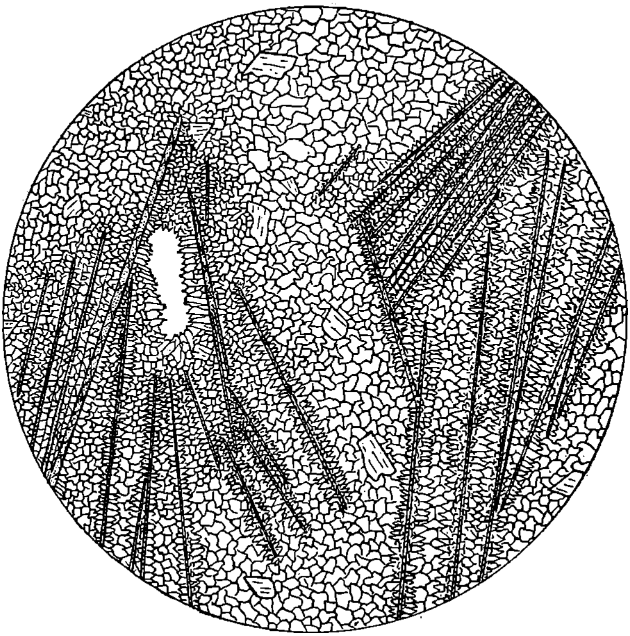

FIG. 169.—Lamellar aggregate of quartz, calcite, and adularia, Tom Reed mine, Arizona. Drawn from thin section. The thin plates of calcite are fringed with bristling quartz crystals and separated by granular quartz (clear) with some adularia (light-line shading). One small cavity lined with quartz crystals. Magnified 15 diameters. (*After F. L. Ransome, U. S. Geol. Survey.*)

As plainly shown by A. Knopf,[1] F. L. Ransome,[2] and E. V. Shannon,[3] much of the earlier calcite was the lamellar variety called "argentine"

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[1] *Bull.* 527, U. S. Geol. Survey, 1913.

[2] *Bull.* 743, *idem*, 1923.

[3] Minerals of Idaho, *Bull.* 131, U. S. Nat. Mus., 1926.

developed parallel to the basal plane, and was replaced by quartz producing a thin platy, microcrystalline quartz (Figs. 169 and 170) with adularia. Similar ore may be seen in the Mount Baldy district, southern Utah, at Jarbidge, Nevada, and at many other places. This important development of adularia, involving transportation of alumina by siliceous solutions, remains without full explanation. The composition of the ore may be similar to that of a pegmatite dike, but the structure is wholly different, and the evidence in favor of aqueous deposition is incontrovertible.

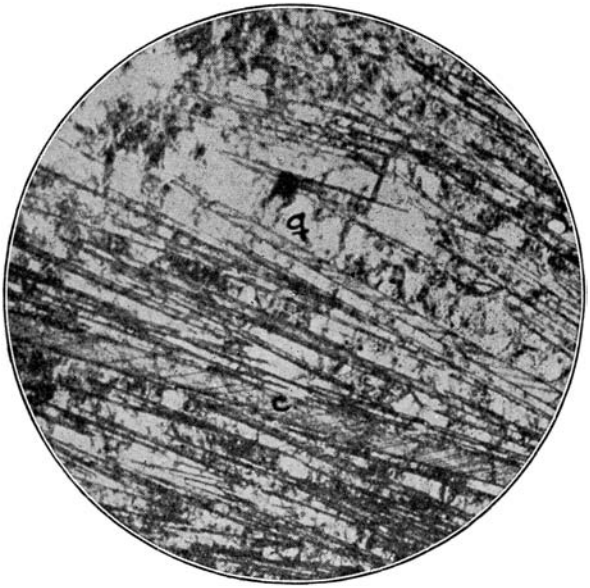

FIG. 170.—Lamellar calcite ore, Belmont mine, Marysville, Montana. Thin section. *c*, calcite; *q*, quartz. Magnified 30 diameters. (*After A. Knopf, U. S. Geol. Survey.*)

There is reason to believe that this "pseudomorphism" is accompanied by a change in the metal content of the vein. At least it seems as if the original filling of barite, calcite, and fluorite carried more silver and as if the silicification and feldspathization were accompanied by a concentration of the gold.

**Paragenesis.**—In general the paragenesis of the ore minerals conforms with that observed in mesothermal veins (p. 543), though the scarcity of metallic minerals makes the investigation more difficult. In the Smuggler mine, Colorado,[1] the hypogene succession is as follows: (1) Quartz; (2) pyrite, arsenopyrite, and rhodochrosite; (3) chalcopyrite; (4) sphalerite and chalcopyrite; (5) galena; (6) tetrahedrite, polybasite, and chalcopyrite; (7) gold; (8) calcite.

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[1] M. E. Hurst, *Econ. Geol.*, 17, 1922, p. 695.

**Zeolitic Replacement.**—Zeolites are foreign to veins of the deeper zones; in the veins formed near the surface they are occasionally found, but they are rare. At a few places zeolites are reported in the altered country rock (Tonopah, the Comstock, and Waihi). At Guanajuato, apophyllite, laumontite, and stilbite are found in the filling of the veins, but here they always belong to the latest phases of vein formation. Few of these occurrences in the vein filling have been carefully studied. In the Southern Republic mine at Republic, Washington, laumontite, associated with calcite, occurs on a fairly large scale.[1] At this place the ordinary fine-grained banded quartz filling had evidently been dissolved and the laumontite-calcite aggregate was deposited in its place. The ore in this zeolitic zone, or shoot, contains mainly silver, whereas elsewhere in the mine gold predominates in the quartz gangue. It seems to be worth while to draw attention to this occurrence with a view to ascertaining whether the development of zeolites is not favorable to silver enrichment. Many facts noted in veins of other classes, like those of Kongsberg, Norway, and Andreasberg, Germany, point in this direction.

**Primary Ore-shoots, Oxidation, and Sulphide Enrichment.**—Rich oxidized ores are often encountered in these deposits at the surface and down to the water level. Whether the primary ore is greatly enriched in this zone depends really more on the texture and composition of the ore than on its original tenor. In veins of hard fine-grained quartz oxidation often fails to produce an ore of higher grade. There are many districts (Cripple Creek and Tonopah, for instance) in which the oxidized ores are little, if any, richer than those below the oxidized zone.

The largest and richest masses of ore are often found just below the oxidized zone and, in general, contain sulphides, sulphantimonides, and sulpharsenides. It will suffice to call attention to the great silver bonanzas of Guanajuato and Pachuca, Mexico; to the Comstock, Nevada, where in one month silver-gold ores valued at $6,000,000 were extracted; to Tonopah, Nevada, where in three months ore yielding over $3,000,000 was extracted; to the Caledonian mine at Thames, New Zealand, which in one year from a small ore-shoot produced $6,000,000 in gold; to Cripple Creek, Colorado, where in one year from a small area but a considerable number of mines $18,000,000 in gold was produced; to Goldfield, Nevada, where during a single year over $10,000,000 in gold was recovered from one property, the ore averaging $38.50 per ton, and where, of the phenomenally rich ore shipped in 1907, one carload of 47 tons yielded $600,000 in gold.

Most of these high yields proved ephemeral. The bonanzas were extracted, poorer ore was found in depth, and the mine was abandoned or continued in feeble existence.

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[1] W. Lindgren, *Trans.*, Canadian Min. Inst., 15, 1912, pp. 187–191.

These great bonanzas were in part formed by primary (hypogene) deposition—in large degree probably by the later, reconcentrating phases of primary deposition; in part they owe their origin to sulphide concentration by descending (supergene) waters charged with precious metals from the upper parts of the veins. It should not be overlooked that ore-shoots of primary origin are common enough. Take, for instance, the Cripple Creek gold ores, in which evidence of enrichment is conspicuously lacking; these primary shoots are usually of a markedly irregular form; many of the smaller ore-bodies are likely to follow intersections of fissures. In depth the rich shoots show a tendency to contract, to feather out, or simply to become impoverished to a scant mineralization of pyrite and sphalerite.

On the other hand sulphide enrichment is conspicuously marked, especially in silver veins. In gold-bearing veins the enrichment in gold is likely to be localized in the lower part of the zone of oxidation. The secondary (supergene) silver minerals are cerargyrite, native silver, argentite, stromeyerite, stephanite, polybasite, and pearceite. It is often exceedingly difficult to distinguish primary ore and enriched ore, and when in addition to this we recognize that the later effects of primary mineralization may sometimes closely simulate the products of descending surface waters, the difficulties of correct interpretation will be fully realized.[1] The distinction is made easier when the secondary sulphides form a distinct zone immediately below the oxidized part of the lode.

The conditions for the deposition of gold and silver are apparently much more favorable near the surface than at greater depths; on the other hand, deposition took place rapidly and the gold and silver contents of the solutions were doubtless exhausted before they reached the actual surface.

**Types of Deposits.**—The merging of the various types makes it difficult to establish a rigid classification.

One type, namely, the zeolitic copper ores in amygdaloid rocks presents certain controversial features, and will be described in Chapter XXV. The deposits are classified as follows:

1. *Cinnabar Deposits.*—Cinnabar, marcasite, stibnite, hydrocarbons, quartz, opal, calcite.
2. *Stibnite Deposits.*—Stibnite, pyrite, and some other sulphides; also quartz.
3. *Base Metal Deposits.*—Chalcopyrite, galena, sphalerite, tetrahedrite, in an abundant gangue of quartz, carbonate, fluorite, or barite. Principal values usually in gold and silver.
4. *Gold Deposits.*—Native gold, alloyed with silver. Subordinate argentite, ruby silver, etc. Quartz.

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[1] E. S. Bastin and F. B. Laney, The genesis of the ores at Tonopah, Nev., *Prof. Paper* 104, U. S. Geol. Survey, 1918.

5. *Argentite-gold Deposits.*—Argentite, silver selenides, ruby silver, tetrahedrite, etc.; native gold, quartz, calcite.

6. *Argentite Deposits.*—Argentite, ruby silver, tetrahedrite, etc.; quartz or calcite, barite, and fluorite.

7. *Gold-telluride Deposits.*—Gold tellurides, quartz, or quartz and fluorite.

8. *Gold-telluride Deposits with Alunite.*—Gold tellurides, gold pyrite, alunite, kaolin.

9. *Gold-selenide Deposits.*—Gold selenides, pyrite, quartz, calcite.

**Older Representatives of This Class.**—The types just enumerated almost wholly represent veins or similar deposits in Tertiary lavas of the Cordilleran, Pacific, or Hungarian regions, but a close examination will easily discover examples of similar deposits of a greater geological age. Beck[1] described relatively unimportant deposits in the Mesozoic melaphyres and quartz porphyries at Imsbach, in the German Palatinate, that carry chalcopyrite, galena, and tetrahedrite in a gangue of calcite, barite, and rhodochrosite and are probably ancient representatives of this class. The celebrated veins of Freiberg, or at least three types of them, namely, the "noble quartz formation," the "noble carbonate formation," and especially the "barytic lead formation," should be mentioned in this connection. There seems to be good evidence that these are Carboniferous representatives of veins formed near the surface, although the lavas in which they probably reached the surface are now eroded. Their structure and composition point clearly to a shallow deposition, and were the physiographic conditions in the Erzgebirge fully analyzed, the results would probably confirm this view. The "barytic lead formation," for instance, carries barite, fluorite, quartz, and dense quartz as gangue minerals with beautiful crustification, while the ore minerals are argentiferous galena, tetrahedrite, bournonite, and chalcopyrite.

Other occurrences comprise the veins in the Triassic diabase flows at Bergen Hill, New Jersey, which contain pyrite and galena in a gangue quartz and adularia, with secondary zeolites; also the copper veins at Bristol, Connecticut.

**Proof of Depth below Surface.**—Physiography furnishes the data on the original surface during deposition. We may be able to trace the old surface of the volcanic plateau and ascertain the relation of the outcrops to the uppermost flow, or in dissected volcanic piles it may be possible to reconstruct approximately the old surface of the volcanic cone. Of this latter possibility Cripple Creek is an instance (Fig. 188); the present surface was probably less than 1,500 feet below the original surface of the volcanic cone. Ransome estimates that at Goldfield, Nevada, the sur-

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[1] R. Beck, Lehre von den Erzlagerstätten, 1, 1909, p. 334.

face has been degraded but a few hundred feet below the original contours of the flows. A fine example showing the connection of deposits formed near the surface with those of more deep-seated type is offered by the San Juan region, in Colorado, where erosion has not only intersected the flows but laid bare the intrusive masses forced into them—all within a vertical interval of 6,000 feet.

**Proof of Temperature.**—The similarity to hot-spring deposits is least marked in deep-seated veins, but becomes striking in the veins here under consideration. The fine-grained chalcedonic and banded quartz of spring deposits (Fig. 5, p. 67) is entirely similar to the often delicate and beautifully banded and crustified portions of these veins. Both kinds are colloidal deposits. The evidence indicates beyond any doubt deposition by hot waters that held in solution large quantities of substances and which at the surface could not have had a temperature of more than 100° C. The minerals present are those which we have reason to believe were developed at a temperature less than 200° C. At Republic, Washington, the veins now exposed were probably not more than 1,000 feet below the original surface. They carry much adularia. If it be conceded that they were formed by aqueous deposition the temperature at the levels now exposed could hardly have exceeded 150° C., assuming that the waters reached the surface at boiling point. Doelter believes that adularia may form at a temperature of 100° C.

Considering the case of the Camp Bird mine, Colorado, Boydell[1] finds that the temperature of the ascending water would not be materially influenced by the heat of a single lava flow, and that the assumption of 150° C. at the place of deposition would be entirely reasonable.

An interesting sidelight on the temperature is obtained from the gold deposits at Manhattan, Nevada.[2] Here we have normal quartz-adularia veins in volcanic rocks and also replacements in the Cambrian White Caps limestone. In the latter the gold occurs in very fine distribution; adularia is present. The limestone is recrystallized but the temperature was not high enough to produce lime silicates in the rock. There are also stibnite and realgar, which may have been added by still later hydrothermal action.

The hot springs are volcanic "after effects" and usually ascend through the cooled lavas. In some cases the waters rise through bodies of hot rocks, and then the temperature may become so high that the solutions issue at the surface as gases and form "fumaroles" and "soffioni" which sometimes, at their orifices, have a temperature of as much as 600° C. In these rarer instances the high-temperature deposits, marked for instance by tourmaline or cassiterite, may develop close to the surface.

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[1] H. C. Boydell, Temperature of formation of an epithermal ore deposit, *Bull.* 331, Inst. Min. and Met. (London), paper read April, 1932, 43 pp.

[2] H. G. Ferguson, *Bull.* 723, U. S. Geol. Survey, 1924.

**Genesis.**—There is strong evidence connecting the class of deposits here discussed with igneous action and ascending hot waters. The best proof that the ores were not formed by the ordinary circulation of surface waters is the fact that deposition has not proceeded uniformly, but that the vein-forming epochs were of brief duration and followed closely after each considerable eruption. At Jarbidge, Nevada, the veins are contained in the early rhyolite, while the later rhyolite is barren. At Waihi, New Zealand, the rich veins are sharply truncated by erosion and capped by a rhyolite of later age.

The occurrence of these deposits in lavas really counts for but little; there are vast areas of lava flows absolutely barren of mineral deposits. Several of the Hungarian authors have pointed out that the veins are confined mainly to the vicinity of volcanic necks or centers of eruption. Exactly the same conclusions have been reached in the United States.

These veins do not disappear in depth. The gangue minerals continue as far as followed, but the ore becomes low grade, therefore the exploitation stops. The deposits have nothing to do with the superficial volcanic phenomena, though some authors seem to think so. The solutions were truly of deepseated origin but their load was precipitated within about 3,000 feet of the surface.

The question naturally arises: Why should not the veins gradually assume the aspect of the deeper zones? Indeed, there is some scant evidence from the San Juan region in Colorado that they do so occasionally, but this is not a final answer. It would seem logical to suggest that the epithermal veins were formed from hot solutions which had their origin in the deep reservoirs where the magma was differentiated into the various types of flow rocks, and not, like most mesothermal and hypothermal deposits, in batholithic intrusions nearer to the surface. Why the ores should not have been precipitated before they reached the epithermal region still remains an open question.[1] J. E. Spurr has introduced the apt term of "telescoped" veins, which I should interpret to mean: (1) That the ascending cooling solutions encountered precipitating influences near the surface causing rapid deposition of their content within a short vertical distance; or (2) that the same happened in a vein previously formed at a higher temperature; or (3) that several ore zones were deposited rapidly covering each other owing to the solutions gradually becoming hotter, an exceptional condition which, however, might locally obtain.

## METASOMATIC PROCESSES

**Extent of Alteration.**—At considerable depths, the ore-forming solutions move in the paths prescribed by fissuring and brecciation; they rarely penetrate great masses of rocks. Near the surface, especially in

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[1] J. E. Spurr, Ore magmas, New York, 1923, p. 300.

the great volcanic piles, different conditions prevail. There are here thick beds of tuffs and agglomerates with great porosity, and the stresses may irregularly shatter large volumes of rock. The solutions and gases—of meteoric or telluric origin—move far more freely and alteration is effected on the largest scale. Here, too, we find most emphasized the peculiar effects of the mingling of ascending and descending solutions.

Any one who has visited an active or recently extinct volcano has undoubtedly observed the areas of discolored reddish, brown, and yellow rocks which indicate alteration. Erosion of older volcanoes discloses similar zones of alteration on a large scale and exposes metalliferous deposits formed in their interior. Types of alteration are described below.

**Silicification.**—Gases given off by ascending lavas penetrate the volcanic flows, often reaching the surface as fumaroles. They are admixed with steam in part or wholly of magmatic origin. The gases are ordinarily acid and contain carbon dioxide, hydrogen sulphide, hydrochloric, hydrofluoric, and boric acids, and many other minor constituents. The rocks are decomposed and generally silicified with the formation of kaolin, alunite, jarosite, and diaspore. Small deposits of magnetite and sulphides may form.

The fumarolic gases are condensed, by surface waters, with development of heat and the mixtures forming ascending springs penetrate the volcanic flows. These springs contain a moderate amount of sulphates, a little free sulphuric acid, and a large amount of silica (up to several hundred milligrams per liter). They produce the same alteration in the rocks as above indicated.[1]

In volcanic regions extensive areas of strong silicification are found in places, as have been described, for instance, from the Rosita, Silverton, and Bonanza districts, Colorado. The type of alteration has been well described by W. S. Burbank[2] from Bonanza. This barren silicification resulted in the (probably colloidal) replacement of the igneous rocks by fine-grained quartz and chalcedony. The silica was in part derived from the original silicates, but in part new material was carried in by the waters. In addition, kaolin minerals, alunite, diaspore, rutile, and zunyite [$8Al_2O_3.6SiO_2.9H_2O(O,F_2,Cl_2)$] develop; more rarely, barite, hematite, and pyrite.

The silicification described by Ransome[3] from Red Mountain, near Silverton, is of the same type. This rock contains, in per cent: 60.9 quartz, 26.3 kaolinite, 3.8 diaspore, 5.6 pyrite, 0.6 rutile, and 0.6 apatite. This silicification was apparently not a superficial process, but proceeded

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[1] A. L. Day and E. T. Allen, The volcanic activity and hot springs of Larsen Peak, *Publ.* 360, pt. 2, Carnegie Inst. Washington, 1925.

[2] *Prof. Paper* 169, U. S. Geol. Survey, 1932, pp. 71–75.

[3] *Bull.* 182, U. S. Geol. Survey, 1901.

at considerable depths and penetrated large rock masses. It represents an early stage of alteration which was later followed by the deposition of economically important ores with propylitization and sericitization.

This extreme mode of alteration, by acid solutions, results in the almost total elimination of calcium, magnesium, alkali metals, and carbon dioxide; its operation is somewhat similar to that of weathering by descending acid surface waters.

**The Development of Kaolin.**—Kaolin in mineral deposits results chiefly from the leaching by surface waters containing free sulphuric acid. This mode of alteration is, then, superimposed on the products of chloritization and sericitization formed by ascending alkaline waters.

Below the water level and the influence of free oxygen, sulphides may be deposited by supergene enrichment. Such sulphides are marcasite, chalcocite, covellite, argentite, ruby silver, etc. (see Chap. XXXII). In places such supergene kaolin may contain finely divided gold, undoubtedly concentrated by secondary reactions. This supergene kaolin is often reported in the literature to be of hypogene origin.

Hypogene kaolin doubtless exists, and occurs mainly, in the silicified areas described above, as an earlier type of alteration, as, for instance, at Bonanza, and the Red Mountain and Silverton districts. It is often well crystallized and belongs to the variety of kaolin which has recently been named dickite.[1] The author fully agrees with the conclusions of A. L. Day, and E. T. Allen.[2] They say that "kaolin is the product of acid waters, hot or cold; where sericite is formed, the waters are presumably alkaline." Which one of the clay minerals is formed, depends on temperature and pH value; weakly acid waters would probably be most favorable for kaolin deposition.

**Propylitization.**—After the epoch of fumaroles and hot acid waters has passed and the rocks have cooled somewhat, the epoch of metallization proper begins; and the ore deposits are formed by rising hot solutions which now contain carbon dioxide, silica, and hydrogen sulphide and are of alkaline rather than acid reaction. Some of these waters move slowly, percolating through great masses of rocks; others move rapidly in prescribed channels and effect extensive changes in the immediately adjoining rock.

One of the most common types of alteration is that resulting in the *propylitic facies.*[3] It affects mainly andesites and basalts, more rarely

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[1] C. S. Ross and P. F. Kerr, The kaolin minerals, *Prof. Paper* 165, U. S. Geol. Survey, 1931, p. 158.

[2] Volcanic activity and hot springs of Larsen Peak, *Publ.* 360, Carnegie Inst. Washington, 1925, p. 141.

[3] Propylite, named by von Richthofen, was formerly supposed to be a primary andesitic rock, but G. F. Becker showed at the Comstock mines that it was really a highly altered volcanic rock.

rhyolites, often spreading over wide areas in mineralized districts. Its mineralogical characteristics consist in the abundant development of chlorite and pyrite, sometimes also epidote; in places it is accompanied by the development of carbonates and a little sericite. The rock, which assumes a dull green color, usually directly adjoins the ore-bearing vein. The chemical changes consist of a moderate leaching of both potassium and sodium; the silica is usually decreased, as are also calcium and magnesium, except when carbonates of these metals are formed. The composition of the rock changes but little and the additions consist only of sulphur and some water of hydration. Takeo Kato[1] finds that in some propylitized rocks of Japan the soda-lime feldspars are altered to albite or potash-albite. Propylitization generally precedes the metallization.

Still another type of alteration, seen mostly in siliceous rocks like rhyolite, consists in a general silicification of the ground mass and phenocrysts, with aureoles of quartz developing around quartz phenocrysts.

Where limestone adjoins epithermal veins, a fine-grained silicification without lime silicates may be observed. Sandstones and shales are usually little altered.

**Sericitization and Vein Quartz.**—Near the veins, the sericitization may become more intense, although here, too, simply chloritized rock may often adjoin the fissure filling. In sericitization sodium is almost entirely carried away and potassium is accumulated in a marked degree in sericite and adularia; the latter mineral has a wide distribution, both in the altered country rock and in the fissure filling. Unless carbonates are formed, calcium and, to a less degree, magnesium are carried away; much pyrite is introduced which usually contains at least traces of precious metals. The percentage of silica is reduced. Close to the vein, silicification often assumes the ascendancy and quartz, adularia, and sericite, with more or less sulphides, develop and may form part of the ore. In rare cases hydrargillite and zeolites may appear in the altered rock. Rutile appears to be the only stable titanium mineral.

**Alunitization.**—Nearer the surface another potassium-aluminum mineral—alunite—appears in considerable quantities. This hydrous sulphate is characteristic of large altered areas in volcanic regions,[2] but, being inconspicuous, is easily overlooked. Its development in this phase is probably confined to the zone where the descending waters carrying free sulphuric acid meet the ascending currents of alkaline waters.

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[1] The last stage of magmatic differentiation as represented by Tertiary gold-silver veins, 15th Internat. Geol. Congress, Pretoria, *C. R.* 2: 1930, pp. 9–12.

[2] B. S. Butler and H. S. Gale, Alunite, *Bull.* 511, U. S. Geol. Survey, 1912.

R. J. Leonard, *Econ. Geol.*, 22, 1927, pp. 14–17. Leonard experimented with the action of various alkaline salts and acids on feldspar. He obtained alunite readily at 100° C. and 200° C. but no kaolinite or sericite.

As soon as the mineralized rocks become exposed to the air, oxidation begins and sulphuric acid is generated by the action of oxygen on sulphides. This sulphuric acid descends with the surface water and converts the sericitized rocks into kaolin mixed with alunite and other oxidation products. Where waters exceptionally rich in sulphuric acid have acted on the rocks, almost everything but quartz is carried away and the final result is a loose quartz aggregate. Descending still farther, these sulphuric acid solutions soon lose their oxygen, and, under certain circumstances, secondary sulphides, with alunite and sericite, may again develop. This indicates the complication of the series of reactions and how the same minerals may form at different steps of the process.

ANALYSES OF FRESH AND ALTERED ROCKS IN THE THAMES DISTRICT

| | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| $SiO_2$ | 57.42 | 52.69 | 57.99 | 55.38 | 58.98 |
| $TiO_2$ | 0.68 | 0.53 | 0.51 | 0.24 | 0.11 |
| $Al_2O_3$ | 17.61 | 18.33 | 17.59 | 15.63 | 11.21 |
| $Fe_2O_3$ | 2.34 | 2.32 | 1.56 | 1.88 | 1.45 |
| $FeO$ | 3.77 | 2.98 | 2.37 | 2.95 | 2.42 |
| $MnO$ | 0.43 | 0.25 | 0.21 | 0.23 | 0.11 |
| $MgO$ | 2.19 | 3.09 | 2.01 | 1.88 | 1.43 |
| $CaO$ | 5.69 | 7.87 | 5.45 | 6.01 | 8.11 |
| $Na_2O$ | 3.22 | 2.62 | 1.98 | 0.83 | 0.61 |
| $K_2O$ | 1.94 | 0.98 | 1.65 | 3.28 | 3.93 |
| $H_2O-$ | 0.85 | 0.73 | 1.56 | 2.41 | 2.54 |
| $H_2O+$ | 2.62 | 3.71 | 1.89 | 1.92 | 1.15 |
| $CO_2$ | 0.95 | 3.59 | 3.89 | 4.58 | 4.69 |
| $P_2O_5$ | 0.31 | 0.42 | 0.35 | 0.11 | 0.06 |
| $FeS_2$ | .......... | .......... | 1.42 | 2.35 | 3.13 |
| Total | 100.02 | 100.11 | 100.43 | 99.68 | 99.93 |

1. Fresh hornblende andesite, Thames.
2. Chloritized hornblende andesite, Halcyon mine.
3. Altered andesite, 14 feet from Ophir vein, Halcyon mine.
4. Altered andesite, 5 feet from Ophir vein, Halcyon mine.
5. Altered andesite, adjoining Ophir vein, Halcyon mine.

(3–5) Sericite-pyrite-carbonate rock from the 386-foot level.

**Metasomatic Processes at Thames and Waihi.**—Extensive propylitization has been described by several authors from observations in the Hauraki Peninsula, in the northern island of New Zealand,[1] where rich

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[1] A. M. Finlayson, Problems in the geology of the Hauraki gold fields, *Econ. Geol.*, 4, 1909, pp. 632–645.

gold-bearing veins appear in volcanic rocks, like pyroxene andesite or dacite. The extreme phase, close to the veins, is a grayish-white rock, but a more widespread type is a chloritized andesite which corresponds to the propylitic facies as defined on a previous page. In this second type the ferric minerals are chloritized, the pyroxene often passing first through a uralitic stage, while the plagioclase remains comparatively fresh, but contains some calcite and sericite.

Finlayson has presented two extremely valuable series of analyses, which are given on pages 459 and 460. They represent rocks from Thames and Waihi, the two most important fields in the peninsula. The first column in each table gives the composition of the fresh rock, the second that of the propylitic or chloritic facies; and the rest are analyses of the more altered forms in which sericite and adularia are the predominating metasomatic products.

ANALYSES OF FRESH AND ALTERED ROCKS AT THE WAIHI MINE

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| $SiO_2$ | 63.45 | 58.39 | 61.78 | 69.35 | 76.61 | 85.65 |
| $TiO_2$ | 0.75 | 0.68 | 0.69 | 0.43 | 0.28 | tr. |
| $Al_2O_3$ | 15.26 | 16.51 | 14.89 | 11.66 | 8.31 | 1.35 |
| $Fe_2O_3$ | 2.28 | 2.46 | 2.08 | 1.53 | 1.08 | 0.43 |
| FeO | 3.01 | 2.98 | 2.51 | 1.66 | 0.59 | 0.21 |
| MnO | 0.36 | 0.32 | 0.28 | 0.11 | 0.11 | 0.12 |
| MgO | 1.29 | 1.66 | 1.08 | 0.46 | 0.51 | 0.31 |
| CaO | 3.44 | 4.08 | 3.16 | 2.09 | 3.61 | 2.56 |
| $Na_2O$ | 2.21 | 2.08 | 2.18 | 1.06 | 0.29 | 0.28 |
| $K_2O$ | 1.78 | 2.89 | 3.68 | 3.31 | 1.98 | 1.41 |
| $H_2O-$ | 1.10 | 2.41 | 1.89 | 1.61 | 0.43 | 0.24 |
| $H_2O+$ | 2.90 | 2.87 | 3.05 | 2.12 | 1.08 | 1.33 |
| $CO_2$ | 1.08 | 1.56 | 2.01 | 2.24 | 1.87 | 2.04 |
| $P_2O_5$ | 0.29 | 0.31 | 0.30 | 0.26 | 0.11 | tr. |
| $FeS_2$ | ........ | ........ | 0.65 | 1.88 | 3.59 | 4.69 |
| Total | 99.20 | 99.20 | 100.23 | 99.77 | 100.45 | 100.62 |

1. Fresh hornblende dacite, Waihi.
2. Chloritized hornblende dacite, 45 feet from Empire vein.
3. Altered dacite, 30 feet from Empire vein.
4. Altered dacite, 15 feet from Empire vein. } 850-foot level.
5. Altered dacite, adjoining Empire vein.
6. Replacement ore, Empire vein.

The hydration is caused by the development of chlorite. The percentage of alkali metals decreases, but only to a moderate degree. If sericite or adularia has formed, the potassium may be higher in the altered rock. The two excellent series of analyses quoted above show a slightly differing trend. At Thames the altered rocks contain 10 or 11 per cent of carbonates, while at Waihi the carbonates form only one-half of that amount. As in the California gold-quartz veins, this development of carbonates results, at Thames, in the fixing of calcium, while magnesium shows slight changes. At Waihi there is little change in calcium, while the magnesium has been somewhat reduced. In both places there is strong leaching of sodium and progressive accumulation of potassium, except that at Waihi the potassium finally diminishes in the highly quartzose vein material. Iron, in ferric and ferrous states, is converted to pyrite, but the total iron is not much increased. At Thames, where carbonates are abundant, the silica tends to decrease; at Waihi the opposite is true. In both places there is an apparent decrease in alumina, and a remarkable and unmistakable leaching of titanium, phosphorus, and manganese, as has also been noted by Spurr at Tonopah.

Mineralogically the alteration close to the vein results in sericite, calcite, siderite, pyrite, quartz, and adularia, the last mineral in places forming pseudomorphs after soda-lime feldspars, while also occurring in small fissures. The adularia (valencianite) from Waihi was analyzed by Finlayson and found to contain 11.25 per cent $K_2O$ and 4.11 per cent $Na_2O$.

Stilbite and laumontite have been identified in the altered rocks of Waihi,[1] and analyses 4 (Waihi) and 5 (Thames) suggest the possibility of their presence.

Finlayson does not accept Spurr's view that the vein-forming waters, filtered through rock masses, caused propylitization, but thinks that this alteration is due to solutions or gases rich in $CO_2$, which permeated the rocks immediately after solidification. The sericite-pyrite carbonate rock along the veins, according to Finlayson, is caused by ascending solutions of a later type.

**Metasomatic Processes at Tonopah.**—During the alteration of the trachyte near the veins at Tonopah, Nevada,[2] biotite and hornblende have usually been completely destroyed; their outlines are marked by aggregates of sericite, quartz, pyrite, and siderite, the latter two often crystallizing together. The primary andesine-oligoclase has changed to quartz and sericite or to adularia; the latter two are not often associated in the same specimens. The microlitic groundmass is largely altered to fine-grained quartz with fibers of sericite; pyrite and siderite are disseminated. Apatite and zircon are residual minerals. Kaolin, when present, is believed to result from the alteration of sericite by descending solutions.

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[1] P. G. Morgan, *Trans.*, Australian Inst. Min. Eng., 8, 1902, p. 186.

[2] J. E. Spurr, *Prof. Paper* 42, U. S. Geol. Survey, 1905; *Econ. Geol.*, 10, 1915, pp. 713–769.

Within short distances from the larger veins, a propylitic type of alteration appears, in which calcite and chlorite, together with pyrite and siderite, are the important minerals. The feldspars are altered to calcite with a little quartz; epidote is not abundant. There are transitions between the propylitic and the sericitic alteration; and, according to Spurr, they were produced by the same waters. Near the veins these waters introduced silica, potassium, and metallic sulphides; as they penetrated farther from these channels their metal contents were exhausted, while silica and potassium were still introduced; finally only carbon dioxide and hydrogen sulphide were left in the cooling waters, which, therefore, had little to precipitate and small power of abstracting. The wall rock acted as a screen for the traversing solutions.

ANALYSES OF FRESH AND ALTERED ROCKS, TONOPAH, NEVADA

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| $SiO_2$ | 67.69 | 55.60 | 72.98 | 91.40 |
| $TiO_2$ | .......... | 0.72 | 0.44 | 0.07 |
| $Al_2O_3$ | 17.67 | 16.70 | 14.66 | 4.31 |
| $Fe_2O_3$ | 2.43 | 2.23 | 1.01 | 0.77 |
| FeO | 0.80 | 3.51 | 0.16 | 0.11 |
| MgO | 0.88 | 2.60 | 0.33 | 0.18 |
| CaO | 0.45 | 4.27 | 0.18 | none |
| BaO | .......... | 0.12 | .......... | 0.02 |
| $Na_2O$ | 2.54 | 4.08 | none | 0.06 |
| $K_2O$ | 5.11 | 3.17 | 6.03 | 1.68 |
| $H_2O-$ | .......... | 0.88 | 0.97 | 0.46 |
| $H_2O+$ | .......... | 3.06 | 2.95 | 0.98 |
| $CO_2$ | .......... | 2.76 | none | none |
| $P_2O_5$ | .......... | 0.28 | 0.16 | 0.04 |
| | 97.57 | 99.98 | 99.87 | [1]100.08 |

1. Partial analysis of relatively fresh "Mizpah" trachyte. Booth, Garrett, and Blair, analyst.
2. Altered andesite, Siebert shaft. Propylitic alteration to quartz, calcite, chlorite, and sericite. George Steiger, analyst.
3. Altered trachyte, Mizpah mine. No original minerals remaining. Sericitic alteration. George Steiger, analyst.
4. Ore material of Mizpah vein. Dense quartzose rock mixed with kaolin-like material. Silicified trachyte. George Steiger, analyst.

[1] Also 0.12 $ZrO_2$ and 0.06 MnO

As noted above, these views are not entirely accepted by Finlayson.

The most prominent features of the alteration are the almost complete removal, adjacent to the veins, of ferrous iron, calcium, magnesium, and sodium and the partial removal of ferric oxide. Even the resistant apatite and rutile seem to have been dissolved to some degree. On the other hand, there is a decided increase of silica, and the potassium has increased. There is a moderate hydration, but no carbonates appear.

**Marysville, Montana.**—In a description of certain epithermal veins, including the Drumlummon lode, A. Knopf[1] gives comparative analyses which show a decided sericitization of the quartz monzonite wall rock. The altered rock contains only a few per cent of carbonates, but 27 per cent of sericite. He states that the sericitization has been selective, the orthoclase being practically exempt from alteration. A net decrease of mass amounting to 5 per cent is calculated for the altered rock. This lode is credited with a production of $15,000,000 in gold and silver.

**Summary.**—We may say that propylitization is the dominant process of rock alteration of epithermal veins. Close to the vein, silicification with adularia and sericite may appear; and nearer the surface, or under the influence of descending sulphuric acid solutions, we find kaolin, or, where sulphuric acid is present in abundance, alunite.

## QUICKSILVER DEPOSITS

**The Ores and Their General Occurrence.**—The principal quicksilver ore is cinnabar ($HgS$), which contains 86.2 per cent mercury. It is usually massive, rarely well crystallized. A black modification of this mineral, called metacinnabarite, rarely occurs in large amounts and is probably a secondary sulphide deposited by descending waters.[2] Native quicksilver, calomel ($HgCl$), montroydite ($HgO$), and several oxychlorides are evidently secondary minerals, developing from the sulphide (p. 869). Primary but rare minerals are the black telluride of quicksilver, coloradoite; the selenide, tiemannite; the sulphoselenide, onofrite;[3] and other still rarer combinations of the selenides of copper, lead, and quicksilver. Mercurial tetrahedrite is not uncommon, and some varieties contain as much as 17 per cent quicksilver, although the percentage is usually much smaller. In smaller quantities this metal is also sometimes present in other minerals, for instance, in the native silver of Kongsberg, Norway, and in the dyscrasite of the silver-bearing veins of Cobalt, Ontario (p. 608).

The occurrence of quicksilver minerals is by no means confined to any certain kind of deposits or to any given age or epoch of metallization. However, such minerals are not known to occur in deposits of distinctly

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[1] *Bull.* 527, U. S. Geol. Survey, 1913.

[2] E. T. Allen and J. L. Crenshaw, The sulphides of zinc, cadmium, and mercury, *Am. Jour. Sci.*, 4th ser., 34, 1912, pp. 367–383.

[3] Tiemannite and onofrite occurred as fissure filling in limestone at Marysvale, Utah, and the ore was worked for some time. Volcanic rocks are abundant. Selenides are reported from Guadalcazar, S. L. P., Mexico; livingstonite ($HgSb_4S_7$) at Huitzuco, Guerrero.

igneous origin, or in pegmatite dikes, or in veins of the deepest zone. High temperature is evidently unfavorable for their development. However, there is coloradoite in the gold telluride veins of Western Australia, which contain, among other minerals, magnetite and tourmaline, indicating deposition at fairly high temperature. In gold-bearing quartz veins of the ordinary type, believed to have been formed at a considerable depth, but at lower temperature and pressure than pegmatite dikes, cinnabar is not an uncommon mineral. It occurs in several of these veins in California,[1] as well as in the similar veins of central Idaho, and is frequently found in the placers derived from the erosion of these veins, as at Stanley Basin and Warren, Idaho. In northeastern Oregon, the gold-quartz veins contain mercurial tetrahedrite, as well as secondary cinnabar formed from that mineral.[2] In the placers below the veins of Susanville, in the same region, are pebbles showing masses of cinnabar inclosed in massive white vein quartz. In small quantities cinnabar occurs in the lead and zinc deposits of Monteponi, Sardinia, and at Santander, in Spain.

In few of these deposits are the mercurial minerals abundant enough to constitute an ore, and in the majority of the deposits formed at a considerable depth the metal is apparently entirely absent. The commercial quicksilver ores are practically confined to a small and well-defined group of deposits, which are of particular interest because their genesis can be fairly accurately ascertained.

A scant association of ore minerals characterizes these deposits. Besides cinnabar and a few secondary minerals, they contain almost invariably pyrite or marcasite. Stibnite is common. Chalcopyrite and millerite are rare. Pyrargyrite occurred in a deposit near Calistoga, California. Among gangue minerals we have predominatingly opal, chalcedony, and quartz, also calcite and dolomite, more rarely barite and alunite, very seldom fluorite; zeolites are of exceptional occurrence. Replacements of adjoining country rock by dolomitic carbonates or by opal are common, but the cinnabar nearly always occurs in these or in unaltered rock directly associated with quartz or chalcedonic silica. Hydrocarbons are frequent. The ores should contain at least 0.5 per cent Hg. The richest ores are those of Almaden, Spain, said to average 6 to 8 per cent.

As to form and structure, the ores occur in irregular and "chambered" veins and brecciated zones, often also as "stockworks" of minute seams, or as disseminations and replacements in more or less porous rocks. The irregularity and brecciated character of the deposits suggest their develop-

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[1] H. W. Turner, *Am. Jour. Sci.*, 3d ser., 47, 1894, p. 467.

H. D. McCaskey, *Min. Resources*, U. S. Geol. Survey, pt. 1, 1910, p. 905.

[2] W. Lindgren, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 604, 708.

ment near the surface. Schuette emphasizes the almost universal presence of impervious roofs which have the tendency to localize the ore-shoots.

The deposits occur in rocks of any kind and any age, but almost always in close connection with effusive rocks or in regions of volcanic activity. Hot springs are frequent in many quicksilver districts, and many springs in volcanic regions are depositing cinnabar at the present time.

Although most of the quicksilver deposits have been formed at a relatively late time and in connection with the eruption of Tertiary and recent lavas, it does not necessarily follow that their development has been confined to late geological time. Older surface eruptions were undoubtedly accompanied in places by the formation of quicksilver deposits, but as these were near the surface, they were quickly eroded.

The quicksilver deposits generally become impoverished at a depth of less than 1,000 feet. Very few deposits have been profitable to a depth of 1,500 feet.

**Distribution, Production, and Use.**—Quicksilver deposits are widely distributed, although the main production of the metal comes from a few occurrences. The ores follow the belts of Tertiary and Quaternary eruptions, especially along the important fracture lines of the globe.

The Coast Range of California, in which the orogenic movements are largely post-Miocene, contains a belt of quicksilver deposits several hundred miles in length, from which a large production has been obtained. Scattered deposits are found in Oregon, Washington, and even in British Columbia. A second belt, less well defined and certainly less productive, extends from north to south over a similar length in western Nevada. There are a few occurrences in Arizona.[1]

The Mexican area, which in spite of comprising many deposits, yields only a slight production, begins in western Texas, in the Terlingua district, and may be considered to end in the State of San Luis Potosi, Mexico.[2]

Farther south, in Colombia and Peru,[3] quicksilver deposits appear again. The Yauli and Huancavelica districts, Peru, are best known; the latter at one time was an important producer.

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[1] E. D. Gordon and C. Lausen, Quicksilver resources of Arizona, *Bull.* 122, Univ. Arizona, 1927.

[2] J. D. Villarello, Génesis de los yacimientos mercuriales de Palomas y Huitzuco, *Mem.*, Soc. Cient. Ant. Alzate, 19, 1903, pp. 95–136.

J. D. Villarello, Descripción de los criaderos de mercurio de Chiquilistan, Jalisco, *idem*, 20, 1904, pp. 389–397.

P. A. Babb, Dulces Nombres quicksilver deposit, Mexico, *Eng. and Min. Jour.*, Oct. 2, 1909.

F. J. H. Merrill, The mercury deposits of Mexico, *Mining World*, 24, 1906, p. 244.

[3] A. F. Umlauff, El cinabrio de Huancavelica, *Bol.* 17, Cuerpo de Ingenieros de Minas, Lima, 1904, p. 61.

The ores occur in Jurassic beds or in Tertiary volcanic rocks.

In Europe, an extensive region in Italy, Austria, and adjacent countries contains quicksilver deposits. This area includes the deposits of Tuscany, Vallalta-Sagron, Idria and vicinity, Avala,[1] in Serbia, and less important occurrences in Bosnia.

Isolated yet highly productive deposits occur at Almaden, in Spain.

On the northern slope of the Altai Mountains, Ferghana, Russian Turkestan,[2] a long belt of deposits occurs in limestone and schist. They contain cinnabar and stibnite with calcite and fluorite, and were extensively worked by the Chinese a thousand years ago.

Some deposits have been found on the western side of the Pacific, mainly in Japan, China,[3] Borneo, Australia, and New Zealand.

In 1930 the world's production was 107,000 flasks, of 76 pounds, of which 56,069 came from Italy (Idria), 46,244 from Spain (Almaden), and 21,553 from the United States. California produced 11,451 flasks. The average price was about $115 per flask.

The principal use of quicksilver is for gold amalgamation, drugs, paints, and mercurial fulminate ($Hg(ONC)_2$), an explosive used for priming shells.

**The Adriatic Region.**—In the Adriatic region of Europe the ores occur in rocks of many kinds: In Tuscany,[4] Mesozoic and Tertiary limestones and sandstones with trachyte are the enclosing rocks; hot springs are present; at Idria, in the Alpine Triassic; at Avala, in serpentine and probably Cretaceous limestone cut by trachytic dikes. De Launay[5] has shown that these Adriatic ores coincide in their extension with Tertiary eruptives, and that, in all probability, even where these eruptives are locally absent, as at Idria, the deposits owe their origin to the aftereffects of this igneous activity in the form of ascending springs.

At Idria,[6] in territory now annexed by Italy, is located one of the great quicksilver mines of the world. The ores are contained in Triassic beds

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[1] H. Fischer, Die Quecksilberlagerstätten am Avala-Berge in Serbien, *Zeitschr. prakt. Geol.*, 1906, 14, p. 245.

[2] T. B. Borowskaja, Zwetnye metally, 12, Moscow, 1930, pp. 1720–1731.

[3] W. H. Wong, The mineral resources of China, *Mem.*, ser. B.,.1, China Geol. Survey, 1919.

F. R. Tegengren, The quicksilver deposits of China, *idem*, *Bull.* 2, 1920.

[4] B. Lotti, Il campo cinabrifero dell' Abbadia, etc., *Rass. Min.*, 7, 1898; *Zeitschr. prakt. Geol.*, 1898, p. 258. See also *Rass. Min.*, 17, 1902.

A. Rzehak, Die Zinnoberlagerstätte von Vallalta-Sagron, *Zeitschr. prakt. Geol.*, 1905, pp. 325–330.

H. Troegel and F. Ahlfeld, Die Zinnober-vorkommen in der südlichen Toscana, *Zeitschr. prakt. Geol.*, 1920, pp. 21–28; 37–46.

[5] L. de Launay, La métallogénie de l'Italie, Tenth Internat. Geol. Congress, Mexico, 1906.

[6] F. Kossmat, Ueber die geologischen Verhältnisse des Bergbaugebietes von Idria, *Jahrb. K. k. geol. Reichsanstalt*, 49, 1899, pp. 259–286.

Geologisch-bergmännische Karten, etc. von Idria. Text by Plaminek. Published by the Agricultural Department, Vienna, 1893 (Literature).

of shale, marl, and dolomite, and are connected with and in part occur in great overthrusts and faults. The ore-bodies, which apparently do not extend below a depth of 1,000 feet, in places follow the stratification. The ores are usually designated as "impregnations" in shale or marls, but small veins and stockworks are also found, especially in the dolomite. The age of deposition is certainly post-Cretaceous, probably Tertiary.

**Almaden.**[1]—Almaden is the richest and most productive quicksilver mine in the world. The value of its metallic product during the period 1564 to 1927 is estimated at about $250,000,000. The mine is 1,150 feet deep; the three ore-shoots have a length of about 1,000 feet and an average width of about 12 feet. The ores occur in three fractured beds of steeply dipping Silurian quartzite separated by bituminous slates. Beck has shown that cinnabar replaces the sandstone grains (Fig. 171). Ransome proved the same, still more convincingly. Sericite, pyrite, and a zeolite (natrolite?) occur in the ore. Van der Veen showed that dikes of diabase occur in the mine. He correlates this diabase with the

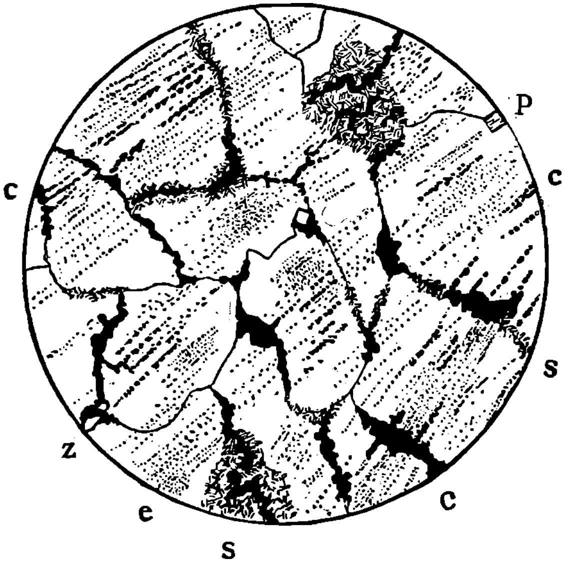

Fig. 171.—Rich ore, Almaden, Spain, showing cinnabar between grains of quartzite and formed by replacement in quartz. *c*, cinnabar; *s*, sericite; *p*, pyrite; *z*, zircon. Magnified 70 diameters. (*After R. Beck.*)

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[1] Courtney de Kalb, The Almaden quicksilver mine, *Econ. Geol.*, 16, 1921, pp. 301–312.

F. L. Ransome, *idem*, pp. 313–321.

R. W. Van der Veen, *idem*, 19, 1924, pp. 146–156.

Minas de Almaden, Guidebook, Fourteenth Internat. Geol. Congress, Madrid, 1926.

Tertiary diabase intrusions so common in many parts of Spain and suggests that the ores are probably of a similar age.

**California.**[1]—The California region offers types of almost all the various structural developments. In the region north of San Francisco, near Clear Lake, serpentine, radiolarian cherts, altered Franciscan sandstone (Jurassic?), and Cretaceous sandstones prevail. The rocks are

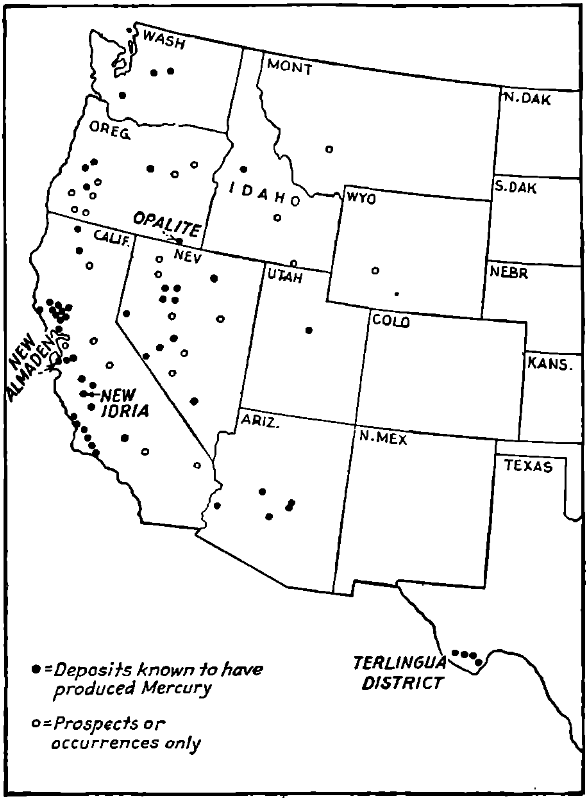

FIG. 172.—Map showing the location of the mercury deposits of the United States. (*After F. L. Ransome, U. S. Mineral Resources.*)

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[1] G. F. Becker, The quicksilver resources of the Pacific slope, *Mon.* 13, U. S. Geol. Survey, 1888.

W. Forstner, The quicksilver resources of California, *Bull.* 27, California State Min. Bur., 1903.

H. D. McCaskey, F. L. Ransome, C. P. Ross, and J. W. Furness, *Mineral Resources,* U. S. Geol. Survey, particularly in issue of pt. 1, 1917, pp. 367–456, with literature by I. P. Evans; pt. 1, 1918, pp. 143–182.

W. W. Bradley, Quicksilver resources of California, *Bull.* 78, California State Min. Bur., 1918, pp. 389.

C. N. Schuette, Occurrence of quicksilver ore-bodies, *Trans.,* Am. Inst. Min. Met. Eng., General Volume, 1931, pp. 403–488. An excellent paper.

greatly shattered and late Tertiary to Quaternary andesites and basalts break through them. The occurrences of cinnabar are numerous, and some of them have yielded a large production, but the ore-bodies generally become impoverished at a depth of a few hundred feet. The deposits form fissure veins largely filled with attrition material, and this is impregnated with cinnabar, pyrite, opaline silica, and calcite. Masses of ore often extend into the country rock from these fissures. Or again, as in the Great Eastern mine, the ore forms tabular masses between serpentine and sandstone, or pipes in opaline or chalcedonic rocks between the same formations, or finally, it may be developed on the contact of basalt and sandstone. The Redington mine was operated on a large chambered deposit at the surface which was continued below by more regular and narrow veins. Throughout this region hot springs are found in and around the ore deposits.

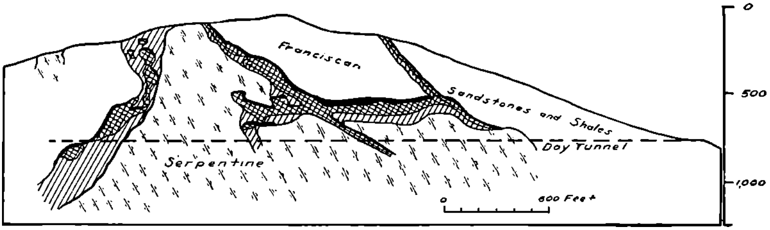

FIG. 173.—Generalized section through New Almaden mine, California, along the Day tunnel. Shaded areas indicate the fractured zone; cross-hatched areas indicate ore; "alta" shown in black. (*After C. N. Schuette.*)

The great mine of New Almaden, south of San Francisco, is credited with a production valued at $75,000,000. It has been in operation for about 100 years. The ore-bodies lie in vein-like fissure zones underneath the "alta," a slicken-sided clay. The footwall is serpentine; above the "alta" is sandstone and clay of the Franciscan series (Jurassic?) (Fig. 173). On the whole, the ore, then, lies underneath an impervious capping and practically on the contact with the intrusive serpentine. Continuous ore-bodies extended to the 1,600-foot level. The greatest depth attained is 2,100 feet, but no work has been done lately in the lower levels. There are no hot springs or lava flows near by.

The New Idria mine, in San Benito County, has been in operation since 1850. Similarly to New Almaden, the ore lies along a clayey streak separating sandstones of the Franciscan series (unconformably covered by tilted Chico (Cretaceous) and Tejon (Eocene) sandstones) from the serpentinized rocks of the hanging wall. The ores appear in three forms—as normal veins, as irregular stockworks, and as impregnations in sandstone. The mine is opened by tunnels, the lowest level being at a vertical depth of 1,400 feet below the surface. There are no effusive rocks in the immediate vicinity. The ores in places contain much meta-cinnabarite.

**Other Districts.**—In the Terlingua district,[1] Texas, near the Mexican boundary, the ores are found in the Upper Cretaceous shales and the Lower Cretaceous limestone over a considerable area. Volcanic rocks are represented by sheets, dikes, and flows of andesite, rhyolite, and basalt. In the lower limestones the ores are mainly in nearly vertical calcite veins, or in lodes of friction breccia (Fig. 174). The other associated minerals are chalcedony, gypsum, aragonite, and pyrite; also many rare oxidation products of cinnabar. In part, the ore-bodies occur below the impervious shale beds. The Chisos mine is 900 feet deep.

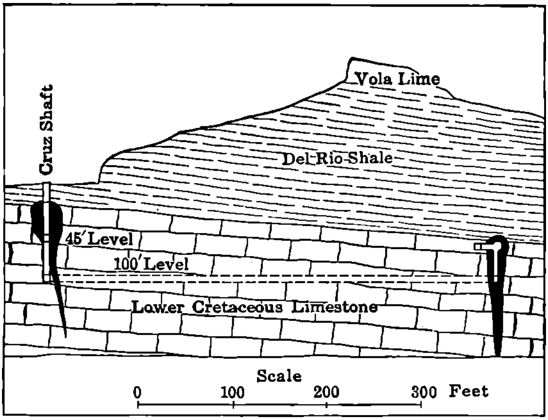

FIG. 174.—Vertical cross-section of California Hill, Terlingua, Texas, showing cinnabar veins with large ore-bodies below impervious shale. (*After H. W. Turner.*)

The quicksilver belt of western Nevada has been described by H. D. McCaskey[2] and A. Knopf.[3] It is found in Humboldt, Pershing, Esmeralda, and Nye counties and has so far yielded no great production. The ores occur in Paleozoic limestone and in Tertiary rhyolite, and show the usual association, except that at Beatty, the rhyolites are completely silicified and alunitized over large areas. Cinnabar with gold and stibnite also occurs near National, Nevada.

A new cinnabar deposit, said to be of some promise, is found in central Idaho in the Yellow Pine district.[4] The cinnabar occurs here in brecciated zones in a Paleozoic (?) limestone included in an intrusive quartz

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[1] W. B. Phillips, The quicksilver deposits of Brewster County, Texas, *Econ. Geol.*, 1, 1905, pp. 155–162.

H. W. Turner, The Terlingua quicksilver deposits, *idem*, 1, 1905, pp. 265–281.

J. A. Udden, The anticlinal theory applied to some quicksilver deposits, *Bull.* 1822, Texas Univ., 1918, 30 pp.

[2] *Mineral Resources*, U. S. Geol. Survey, 1911, pt. 1, pp. 906–919.

[3] Some cinnabar deposits of western Nevada, *Bull.* 620, U. S. Geol. Survey, 1916, pp. 59–110.

[4] D. C. Livingston and E. S. Larsen, *Bull.* 715, U. S. Geol. Survey, 1920, pp. 73–83.

monzonite. Thick beds of Tertiary effusives occur in the vicinity. Cinnabar deposits have recently been found in Carboniferous sandstone in central Arkansas.[1] Associated minerals are quartz, dickite (kaolinite), and stibnite.

Many minor occurrences of cinnabar are believed to be secondary and to result from the decomposition of mercurial tetrahedrite.[2]

**Genesis.**—The uniform character of the quicksilver deposits points to a common genesis for all of them. The earlier belief, that the ores were products of sublimation, is generally abandoned, for the usual mode of occurrence, with minerals of aqueous origin, such as calcite, opal, chalcedony, and often barite, is decidedly opposed to such a view. Becker has pointed out that, as the character of the enclosing rocks has little influence on the deposits they are most probably derived from a common, deep-seated source. Their structure indicates deposition near the surface, as does also the physiographic evidence at many places—for instance, where the ore appears in the crevices of Quaternary and little-eroded lava flows.

When it is noted that hot springs and volcanic surface flows are present in almost all regions of importance (except at Almaden and Idria), and that cinnabar in considerable quantities is associated with hot spring deposits, or is actually found deposited by hot springs, the argument becomes very strong indeed that such solutions have formed the majority of the deposits. For the few deposits that have no such clear connection with volcanic rocks the characteristic mineral association still holds good, and we are forced to the hypothesis that volcanism and hot-spring action are the causes of these also, though the products of the igneous activity may have failed to reach the surface and the hot springs may have subsided.

The evidence relating to cinnabar deposited by hot springs is summarized in the following paragraphs.

At Steamboat Springs, Nevada, near the California boundary, cinnabar is contained in the hot ascending sodium chloride waters together with antimony, arsenic, and sulphur, and is actually being deposited in the sinter.[3] Close by, but at a higher level, is a low-grade quicksilver deposit in decomposed granite, and this in all probability was also formed by the same springs when issuing at a higher level. Underneath the sinters of the present springs the gravels contain crystallized stibnite and pyrite.

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[1] J. C. Branner, Cinnabar in southwestern Arkansas, *Information Circ.* 2, Arkansas Geol. Survey, 1932, 51 pp.

[2] K. Hummel, *Zeitschr. prakt. Geol.*, 1925, pp. 154–167, describes an interesting case where cinnabar with chalcocite has been formed by supergene enrichment of pre-Variscan age.

[3] G. F. Becker, *Mon.* 13, U. S. Geol. Survey, 1888, Chap. XI.

At Sulphur Bank,[1] in the California quicksilver belt, Le Conte, Christy, Rising, Becker, and Posepny have studied the deposition of cinnabar and sulphur by ascending hot sodium carbonate and borate waters and have all arrived at the conclusion that such deposition, together with that of pyrite and opal, is actually taking place. The Cretaceous sandstones and associated Franciscan metamorphic rocks are here overlain by flows of basalt and by cinder cones, pointing to very recent eruption. The hot springs have altered and bleached the basalt. Sulphur is deposited at the surface. Below the superficial deposit of sulphur, cinnabar is found in the basalt, as well as in the underlying shales and sandstones; it occurs mostly in veinlets and joints together with the pyrite and opal above mentioned (*cf.* p. 77).

The Rabbit Hole sulphur deposit, in Humboldt County, Nevada, described by G. I. Adams,[2] is evidently a product of hot springs, and near it are considerable areas of rhyolite. The rocks are silicified, and opal, alunite, gypsum, and some cinnabar are present.

In the Hauraki Peninsula of New Zealand, near Omapere Lake,[3] where basalts overlie Mesozoic shales and sandstones, mercury and cinnabar have been found in the deposits of the hot springs at several places.

A careful investigation would doubtless disclose the presence of cinnabar in many other spring deposits.[4] If it is found in more than traces the best way to test such material, as well as ores, is by the miner's pan, in which the bright-red grains of cinnabar show conspicuously.

Quicksilver is apparently contained in hot-spring waters carrying sodium carbonate, sodium chloride, or sodium borate; sometimes all three salts as well as carbon dioxide and some hydrogen sulphide are present.

Regarding the mode in which mercury is carried in solution, Becker's views,[5] based on the laboratory experiments of W. H. Melville, still appear to furnish the best explanation. While the solubility of mercuric

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[1] J. LeConte and W. B. Rising, The phenomena of metalliferous vein formation now in progress at Sulphur Bank, Calif., *Am. Jour. Sci.*, 3d ser., 24, 1882, pp. 23–33.

G. F. Becker, *op. cit.*, Chap. VII.

F. Posepny, The genesis of ore deposits, 2d ed., 1902, pp. 32–36.

[2] *Bull.* 225, U. S. Geol. Survey, 1904, pp. 497–502.

[3] J. M. Bell and E. de C. Clarke, *Bull.* 8, New Zealand Geol. Survey, 1909, p. 87.

J. Park, *Trans.*, N. Z. Inst. Min. Eng., 88, 1904, p. 27.

Andre P. Griffiths, The Ohaeawai quicksilver deposits, *Trans.*, N. Z. Inst. Min. Eng., 2, 1898, p. 48.

[4] Occurrences which merit close examination are those described by T. Warner from Coso Range, Inyo County, California (Mining in California, 26, January, 1930, pp. 59–63, California Dept. Nat. Research Div. Mines), and that at Opalite, southeastern Oregon.

[5] G. F. Becker, *Mon.* 13, U. S. Geol. Survey, 1888, Chap. XV.

sulphide in alkaline compounds containing sulphur had long been recognized, the evidence was to some degree conflicting.

Becker showed that mercuric sulphide is freely soluble in solutions of sodium sulphide, as well as in a mixture of $Na_2S$ and NaOH, and also in warm sodium sulphydrate (NaHS). When neutral sodium carbonate is treated with hydrogen sulphide, sodium sulphydrate and probably also sodium sulphide will form; these dissolve mercurial sulphide, and double salts of the general formula $HgS.nNa_2S$ doubtless form. Incidentally it was found that the same reagents would dissolve metallic gold, pyrite, sphalerite, and cupric sulphide. The solubility of the sulphides of arsenic and antimony under similar conditions is well known.

It is, therefore, easy to perceive that a spring water containing sodium carbonate and hydrogen sulphide would form a suitable solvent for the compounds mentioned. The precipitation would be easily effected by oxidation of the water and the development of free acids, by dilution, by cooling, or by the presence of organic or ammoniacal compounds. That the latter two agents are active in many cases there is little doubt.

**Relation to Other Ore Deposits.**—Although the cinnabar deposits form a well-defined group, they are not to be separated entirely from other classes of ore deposits. Some of them contain other metallic minerals, and there are many that show a transition to the stibnite and arsenical veins. The Nevada belt, especially, furnishes many instances of a close relationship to gold and silver bearing veins on one hand and to stibnite veins on the other hand. It is true, however, that no cinnabar deposit has yet been found to change gradually into ores of different character as depth is attained. No deposits have been worked below a depth of 2,000 feet vertically beneath the outcrops. There is little doubt that these ores were formed from the most volatile parts of the magmas, carried in solution by ascending waters until they came close to the surface. But the irregular distribution of the deposits and their entire absence from large areas of volcanism form problems that are yet unsolved.

## STIBNITE DEPOSITS

**Mineralogy, Production and Uses.**—Stibnite ($Sb_2S_3$ with 71.4 per cent Sb) is the principal ore mineral of antimony. Its oxidation near the surface results in various oxides (p. 876) of yellowish or white color such as senarmontite ($Sb_2O_3$), cervantite ($SbO_2$), and stibiconite ($H_2Sb_2O_5$). While stibnite occurs in many deposits in small quantities, especially in quicksilver ores, it is the characteristic and dominant mineral in the stibnite veins where it is accompanied by quartz gangue and a scant amount of other sulphides, such as arsenopyrite, realgar, pyrite, and sphalerite, more rarely jamesonite and similar sulphantimonides. Such ores often also carry gold, and the association of the stibnite with some gold-quartz veins has often been noted. For the purpose of making pure antimony the presence of arsenic and copper is undesirable.

In past years the production of antimony has not been great owing to its low price and limited usefulness. During the World War the price increased to 28 cents per pound for the best grade; in 1929 and 1930 the price had dropped to about 8 cents. The world's production of antimony metal may be estimated at 30,000 short tons. The supply is mainly obtained from China, but smaller amounts come from central France, from the State of Queretaro, Mexico, as well as from Italy, Jugoslavia, Asia Minor, and Algeria. In the United States little pure antimony is produced. Antimony is used for bearing metals, type alloys, shrapnel bullets, and its salts find a varied use in the industries. The sulphide is used in pyrotechnics.

Another source of antimony is in the replacement deposits and veins containing mainly galena but associated with tetrahedrite and more rarely with jamesonite, bournonite, boulangerite, and other lead sulphantimonides. As a rule these are related to intrusive action; and the small amount of antimony contained is recovered as "hard lead" or antimonial lead in the smelting operations. About 25,000 tons of this alloy were produced annually in the United States but in 1930 the output had dropped to 13,700 tons. It contains about 27 per cent Sb.

No antimonial mineral is known to occur in magmatic deposits; they are certainly rare in the contact-metamorphic and other high temperature deposits though in these jamesonite, tetrahedrite, and stibnite have been occasionally reported.

**Occurrence.**—The stibnite veins have wide distribution but are rarely rich. They are in part formed near the surface (epithermal), but many deposits are of more deepseated origin and occur in or near intrusive rocks. To the former type belong the stibnite veins with a gangue of fine-grained and drusy quartz which intersect flows of rhyolite and basalt in western Nevada. The antimony sulphide is as a rule beautifully crystallized in acicular and prismatic forms; it is often accompanied by a little pyrite, sphalerite, and arsenopyrite, sometimes also by tetrahedrite and cinnabar. Such veins carry a little silver and less gold. The intimate relationship of these veins with the gold and silver veins proper is, however, shown by the occurrence in one of them, at National, Nevada,[1] of a shoot of remarkably coarse gold of the electrum variety.

Stibnite veins of uncertain affiliations are found in the Carboniferous of central Arkansas. Stibnite shows a marked tendency to form replacements in limestone and shale. Such deposits in Eocene shale below a thick series of andesites have been described from southern Utah.[2]

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[1] W. Lindgren, *Bull.* 601, U. S. Geol. Survey, 1915.

[2] G. B. Richardson, *idem*, pp. 253–256.

They are undoubtedly hot-spring deposits. Of such nature are also the deposits at Pereta, in Tuscany, where the mineral is associated with realgar and cinnabar and occupies veins, seams, and irregular pockets in Tertiary calcareous and detrital rocks. The country rock is in part silicified, in part altered to gypsum or alunite, and exhalations of hydrogen sulphide testify to the recent age of the deposit.

Beck[1] describes important deposits of stibnite at Kostainik, in Serbia, where the mineral occurs in nests and veins in trachyte or in Triassic clay shales, and also as metasomatic bodies replacing the beds along the contact of limestone and shale. The gangue is a drusy fine-grained quartz.

The stibnite veins of Japan, renowned for their beautiful crystals, are found in Mesozoic and Paleozoic rocks near intrusions of quartz porphyry.

From 60 to 90 per cent of the world's production of antimony metal comes from China,[2] mainly from the Hunan Province. The deposits appear to be normal epithermal veins; but they occur in part in Paleozoic rocks.

The deposit of Djebel Hamimat,[3] in Algeria, containing senarmontite, and that of Altar,[4] Sonora, from which stibnite is reported as the principal ore mineral, appear both to be replacement deposits in limestone.

Stibnite veins affiliated with intrusive rocks differ but little from the deposits described above. They are known from central France, where narrow veins intersect granite and surrounding schist. Similar deposits occur, for instance, in Kern County, California, where the quartz veins also contain gold, and in the Coeur d' Alene district, Idaho. The pre-Cambrian gold-quartz veins of Rhodesia often contain stibnite.

Stibnite is very common in Alaska and generally is found in gold-quartz veins. A. H. Brooks[5] who enumerates 67 occurrences suggests that the stibnite may have been introduced in older gold-quartz veins during a later and Tertiary mineralization.

## GOLD-QUARTZ VEINS IN ANDESITE

**Transylvania.**[6]—In northwestern Hungary and in adjoining parts of Transylvania gold-bearing veins of Tertiary age have been formed after eruptions of andesites and dacites. A mining industry, begun centuries

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[1] R. Beck (after W. von Fircks), *Zeitschr. prakt. Geol.*, 1900, pp. 33–36.

[2] F. R. Tegengren, The Hsi-Ku'ang-Shan antimony mining fields, Hsin-Hua district, Hunan, *Bull.* 3, China Geol. Survey, 1921, pp. 1–21.

[3] L. de Launay, Gîtes minéraux, 1, 1913, p. 772.

[4] E. T. Cox, *Am. Jour. Sci.*, 3d ser., 20, 1880, pp. 421–423.

[5] Antimony deposits of Alaska, *Bull.* 649, U. S. Geol. Survey, 1916.

[6] Bela von Inkey, Nagyag und seine Erzlagerstätten, Buda-Pest, 1885.

Bela von Inkey, De la relation entre l'état propylitique des roches andésitiques et leur filons minéraux, 10th Internat. Geol. Congress, Mexico, 1906.

M. v. Pálfy, Das Goldvorkommen im siebenbürgischen Erzgebirge, etc., *Zeitschr.*

ago, still flourishes in this region. The literature is very extensive, and only a few deposits can be mentioned here as examples. The annual production is still about $1,500,000. The mines are now mainly in Roumanian territory.

The geological formations in the western part of the gold mining region of Transylvania consist of Mesozoic melaphyres, Cretaceous shales and sandstones, and Miocene sediments, all penetrated by late Tertiary eruptions of andesites and dacites. The igneous rocks appear as lava flows, tuffs, and volcanic necks. The veins near Brád, at present the most productive district, fill well-defined, steeply dipping, in places branching fissures which generally intersect volcanic rocks or Cretaceous sediments. The simple veins are as much as 1 meter in thickness; the lodes attain a thickness of 10 to 20 meters. The deposits have been worked to a depth of about 300 meters. The fissures are tectonic, not contraction joints. They are of Miocene age.

The surrounding rocks, particularly the andesite, have suffered extensive propylitization, the femic minerals being extensively decomposed, while feldspars remain fresh. Pyrite is not common except near the veins. Calcite is abundant. Schumacher does not believe that propylitization is caused by "intensive penetration by gases from the not yet wholly solidified intrusions," an opinion expressed by Stelzner and Bergeat. He nevertheless considers the process distinctly earlier than the veins and independent of them. The alteration continues to the greatest depths attained. "Kaolinization" near the veins is a wholly different process, according to Schumacher, and is superimposed upon propylitization. The alteration of the wall rock in a vein 0.5 meter thick extends only about 10 centimeters from the vein, but many smaller veins have proportionately wider zones of alteration.

An older set of veins is composed of clastic material of shale and igneous rocks ("Glauch veins"). They are interpreted as having been filled by ascending liquid muds. Similar veins in the Silverton district, Colorado, have been described by Ransome.

The gangue of the ore-bearing veins is composed of quartz (rarely chalcedony), calcite, rhodochrosite, and barite, a frequently recurring association in veins of this class. The quartz is usually fine-grained, sometimes drusy, cellular, or honeycombed, but not amethystine. Pyrite in small crystals is abundant; the concentrates contain 10 grams of gold and 69 grams of silver per ton, while the pyrite in the country rock contains 7 to 15 grams of gold and 10 to 15 grams of silver per ton;

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*prakt. Geol.*, 1907, pp. 144–148.

C. Semper, Beiträge zur Kenntniss der Golderzlagerstätten des siebenbürgischen Erzgebirges, *Abh.* K. Preuss. geol. Landesanstalt, Neue Folge, Fasc. 33, 1900.

F. Schumacher, Die Golderzlagerstätten der Rudaer Zwölf-Apostel Gewerkschaft zu Brád in Siebenbürgen, *Zeitschr. prakt. Geol.*, 1912, pp. 1–85.

both kinds are therefore poor in gold. Sphalerite is associated with pyrite and is poor in gold but contains more silver than the pyrite. Chalcopyrite and galena, where present, are poor in gold but contain several hundred grams of silver per ton.

The principal ore mineral is native gold, which occurs commonly in crushed quartz or in little fissures (Fig. 175), or as sheets or wires between the quartz combs of veinlets. Some of it is found in coarse quartz and is apparently older than the quartz or of contemporaneous origin. It occurs also in sheets along the cleavage planes of calcite and in lumps or nodules in clay. In part it is therefore of relatively late origin. The gold contains 28 per cent silver, and the ores average 10 grams of gold per ton. Tellurides and rich silver minerals are rare.

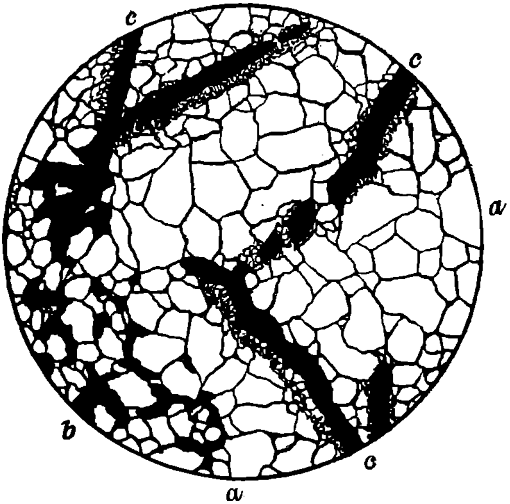

FIG. 175.—Rich gold-bearing quartz, Brád, Transylvania. *a*, granular quartz; *b*, gold between grains; *c*, plates of gold accompanied by crushed quartz. Magnified. (*After F. Schumacher.*)

The structure of the veins is irregularly massive, though in places crusted, banded, and drusy or brecciated. Small prismatic and rectangular pseudomorphs of quartz are considered as replacements of gypsum, but strongly resemble the similar casts of celestite at Cripple Creek, Colorado.

The ore-shoots extend irregularly with steep dip for 100 or 200 feet vertically. Shoots often occur at junctions and intersections (Fig. 176). At a junction of two veins with a narrow pyritic seam was found a pocket from which in one day gold weighing 55 kilograms was extracted. Near the surface the veins were poor. The richest zone extended from about 100 meters below the surface down to a depth of 320 meters. The remarkable dependence of the rich ore on narrow seams of pyrite is evident and recalls analogous conditions in the Thames district, New Zealand, the "indicators" of Victorian quartz mines and many other gold deposits (Fig. 177).

Schumacher believes that the gold was deposited by ascending hot waters in the vicinity of necks of intrusive rocks. Beyond the intrusive necks the veins persist but contain only gangue minerals. M. Dittrich examined fresh and propylitic andesite, using the cyanide process, but found gold in neither.

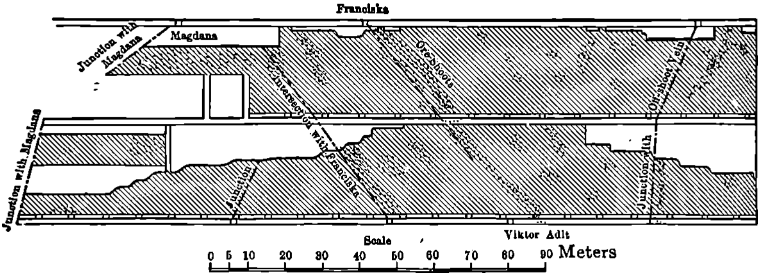

FIG. 176.—Section of stoped area in vein at Brád, Transylvania, showing pockets of rich ore; also rich shoots following intersections with barren veins. (*After F. Schumacher.*)

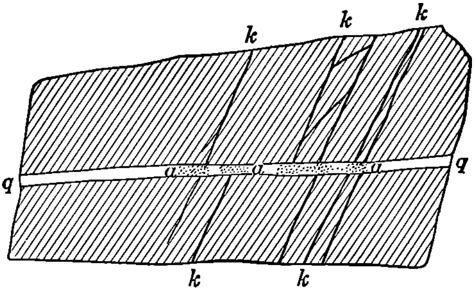

FIG. 177.—Pockets of native gold (*a*) in quartz vein (*q*) along intersections with pyritic seams (*k*). (*After F. Schumacher.*)

**Hauraki Peninsula, New Zealand.**[1]—The Hauraki region, in the northern island of New Zealand, is richly mineralized in several districts.

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[1] James Park, Geology of Hauraki gold field, *Trans.*, N. Z. Inst. Min. Eng., 1, 1897, p. 3.

P. G. Morgan, Geology, etc., of Waihi, *Bull.* 26, New Zealand Geol. Surv. 1924; *Econ. Geol.*, 20, 1925, p. 203.

J. M. Bell and C. Fraser, The great Waihi gold mine, *Bull.* 15, New Zealand Geol. Survey, 1912.

A. M. Finlayson, *Econ. Geol.*, 4, 1909, pp. 632–645 (with literature).

Arthur Jarman, The geology of the Waihi-Grand Junction mine, *Trans.*, Inst. Min. and Met. (London), 25, 1916, pp. 3–40, with discussion.

E. G. Banks, Mining and metallurgical methods at Waihi, *Bull.* 250, Inst. Min. and Met. (London), 1925.

Its rocks consist mainly of andesite and dacite flows of Eocene or Miocene age covered by Pliocene rhyolites.

A production of about $30,000,000 is recorded from the Thames district, though but little gold is now obtained there. The veins are contained in broad belts of soft, propylitic andesite and dip 40° or more. Great masses of low-grade quartz occur, but the gold is derived mainly from rich pockets occurring down to a depth of 400 to 600 feet below the surface. One of these pockets in the Caledonia mine, about 1871, yielded 9 tons of gold in 15 months. The veins have been followed from a height of 1,500 feet above the sea to 640 feet below it, but owing to intervening faults the real vertical extent is only 1,200 feet. Park states that the veins do not continue into the underlying Jurassic shale and that they are thus limited to the thickness of the lava flows in which they occur. The rich shoots occur mainly where the veins are intersected by small stringers or "leaders." Opinions differ widely as to whether this concentration in pockets is due to descending waters or not. In all probability, however, it was one of the latest phases of the primary mineralization. The principal ore mineral is gold alloyed with 30 to 40 per cent silver, but some pyrite, chalcopyrite, sphalerite, galena, stibnite, and pyrargyrite also occur. Arsenopyrite and native arsenic occur at Coromandel. The gangue minerals, besides quartz, are dolomite and, occasionally, rhodonite.

The Karangahake deposits, 40 miles south of the Thames district, are also in propylitized andesite and dacite, but differ somewhat from the type described and consist in brief of calcite and quartz with more or less sulphides. The best known deposits are at Waihi. The Waihi lodes are conspicuous and were discovered in 1878; in part the outcrops are covered by rhyolite, and the development of the deposit therefore falls between the two eruptions. The ore proved difficult to amalgamate and the mines achieved importance only after the introduction of the cyanide process. To the end of 1922 the total production amounted to about $80,000,000. In 1917 the ore averaged $8 in gold and 1 ounce of silver per ton. An annual output of about $2,000,000 has been maintained up to 1931 inclusive. The country rock is a green propylitic dacite with some pyrite, calcite, and seams of quartz and adularia. This rock often adjoins the veins without further alteration, but transitions to the quartz filling by silicification are said to exist.

The vein system is complex, and sixteen steeply dipping and interconnecting veins are known. Of most importance is the Martha lode (Fig. 178), a wide fissure vein with brecciated walls; the quartz is formed largely by filling, in part by silicification. On the 500-foot level the lode is in some places 80 feet wide; for half of this width it is barren, but the other half is said to average $15 to $20 per ton. The proportion of gold to silver by weight is 1:3 or 1:4, and this average was maintained, from the surface down. The water level stood within 200 or 300 feet of the surface. The lode is said to contain ore for a horizontal distance of 1,700 feet. The developments in the deepest levels are said to be disappointing as to the quantity of ore, but the lode itself maintains its strength. The deepest shaft reaches 1,788 feet vertically, and 2,000 gallons of water are handled per minute.

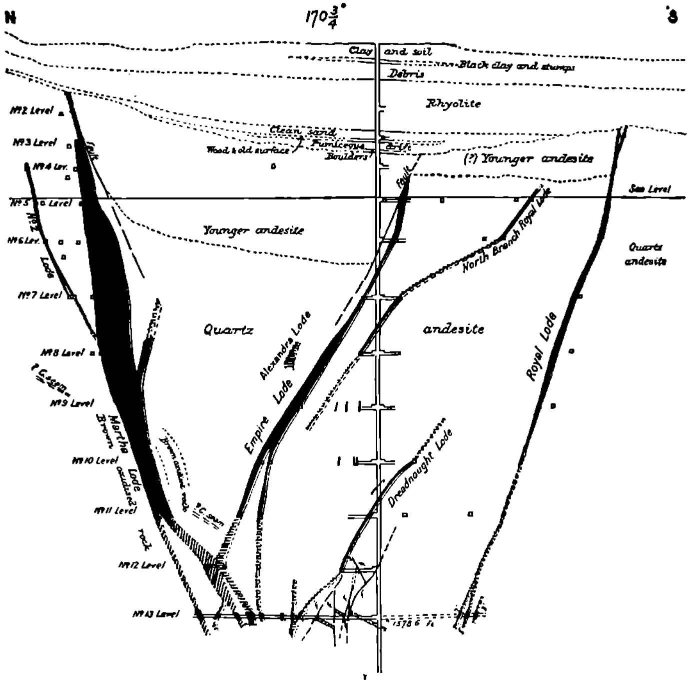

FIG. 178.—Cross-section of Waihi mine, New Zealand, showing lode system in dacite covered by later volcanic flows. (*After P. G. Morgan.*)

A lively controversy has lately developed in regard to near-surface intrusions. Bell and Frazer consider the dacite intrusive in andesite flows, the inference being that ore may only be expected in that rock. Jarman believes there are no intrusives but only a series of flows.

A little pyrite was found in the first level in the Martha lode; on the second level the sulphide ore on the foot-wall was a few feet wide; on the 500-foot level 20 feet of sulphide ore was exposed on the foot-wall, while the remainder of the vein, at this place 40 feet wide, was thoroughly oxidized, with much black manganese oxide. This sulphide ore is of nearly the same value as the oxidized ore, containing perhaps a little more gold and a little less silver.

The ore consists of quartz and calcite, with 3 per cent of pyrite, sphalerite, galena, and argentite. The sulphides are often banded and the gold values are mainly in the pyrite but there is no visible gold; the bullion contains some selenium. Throughout the oxidized zone the calcite is dissolved, leaving a lamellar quartz ore stained by manganese; but this change is produced mainly by descending waters. In other mines of the district there are indications of a pseudomorphic deposition of silica, similar to that of the De Lamar mine, Idaho (p. 448), by a late phase of ascending solutions. In at least some mines in the Karangahake district the ore becomes poor when the zone of the calcite, unchanged by descending waters, is reached. Much of the quartz shows evidence of colloidal deposition.

The depth of the oxidation in the Waihi mine below water level is noteworthy and probably indicates a dry, intervolcanic epoch.

**El Oro, Mexico.**[1]—As there are few important gold deposits in Mexico, the occurrence at El Oro, 70 miles northwest of the federal capital, is of special interest. The district is situated on the volcanic high plateau at an elevation of about 10,000 feet. The barren and unaltered andesites of this plateau overlie the ore-bearing formation which consists of a thick flat dipping series of well stratified black bituminous shale with some sandstone; in places these Jurassic sediments are covered by earlier andesites, which near the vein contain pyrite and chlorite and in other places they are intruded by sills of similar andesitic rocks. These earlier andesites are held to be of Miocene age.

The lodes, about ten in number, outcrop only at one or two places and have been opened by cross cut tunnels and shafts. Almost all of the important work has been done since 1904. There are two principal master lodes, the San Rafael and Dos Estrellas striking N.N.W. and dipping steeply W.S.W. The production from the San Rafael alone since 1904 is approximately $40,000,000 from not less than 5,000,000 metric tons of ore. The Dos Estrellas lode for many years yielded about $5,000,000 per annum. The mines are approaching exhaustion.

The lodes occupy fault fissures, which in their upper parts at least were open and much of the ore has been deposited by filling of open space. In depth and especially in andesite rock much of the quartz is formed by replacement. In 1913, the greatest depth attained was 2,000 feet; 200 feet of which was in barren cap andesite. The filling consists of fine-grained quartz intergrown with much coarse-grained calcite. There is only a small percentage of pyrite and sphalerite. The gold is never visible and even close panning often fails to bring a color from rich ore.

In the upper levels some stopes are from 60 to 100 feet wide but in depth the lode contracts to widths of 3 to 15 feet.

The "branch veins" are an interesting feature of the large lodes at El Oro. They are steep and persistent stringers caused by the settling

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[1] T. Flores, El Oro y Tlalpujahua, *Bull.* 37, Inst. Geol. Mexico, 1920, 85 pp.

of the hanging wall (Fig. 179) and are usually rich, containing from $15 to $35 in gold and from 5 to 20 ounces of silver to the ton. The ore from the main lode contains only $5 to $15 in gold with 2 or 3 ounces of silver per ton.

The ore-shoots in the main lodes are of primary origin and extend horizontally, while reaching to a depth of 500 to 700 feet below the capping andesite. The San Rafael has been stoped continuously for one and one-half miles through three properties. These horizontal shoots are probably caused by an upper, impermeable barrier, now eroded, of andesite or clayey rock.

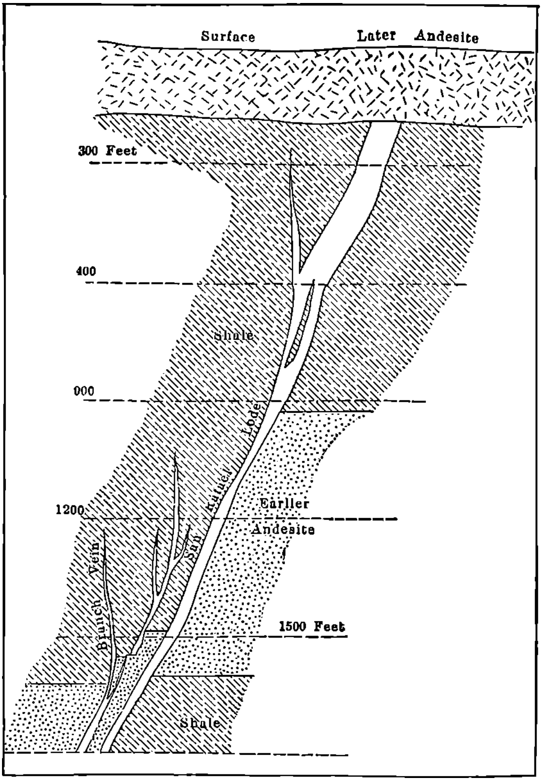

FIG. 179.—Cross-section of the San Rafael lode, El Oro, Mexico, showing branch veins.

The ore of the branch veins is usually well banded by crustification, probably indicating colloidal deposition (Fig. 180), and is much richer in sulphides, mainly sphalerite and pyrite, than the main lode.

Oxidation of the main lode antedates the capping of younger andesite and is practically complete to a depth of 800 feet. The calcite is dissolved and the quartz remains as a porous, cellular mass. The gold being in the quartz there is a considerable apparent enrichment of gold in the oxidized zone. Some silver has probably been leached, but no evidence was found of transportation of gold. There is little manganese in the ore.

**Black Mountains, Arizona.**[1]—Not far east of Needles, on the Colorado River, rises the Black Range, which is built up of a series of tilted volcanic flows of andesite-latite rocks with minor intrusive masses of granite-porphyry and quartz-monzonite porphyry. The flows rest on pre-Cambrian rocks and have a great thickness (Fig. 181). The gold-bearing veins are formed along fault fissures with normal throw, and some of them have been very productive though the shoots are irregular. Most of the ore occurs in the Oatman andesite (the chloritic andesite of Schrader). The Tom Reed and the Gold Road veins have been the most productive. The United Eastern mine has been working a shoot with an area of about 800 by 800 feet which did not outcrop and which has yielded $8,000,000. The greatest depth reached is about 1,300 feet. The veins are stringer lodes often beautifully crustified and "chalcedonic," and the filling consists of calcite, quartz, adularia, and fluorite. The earlier calcite has been replaced by the latter two minerals. There is also a later generation of calcite. The gold, generally finely disseminated, is about the only metallic mineral and occurs chiefly in a peculiar honey-

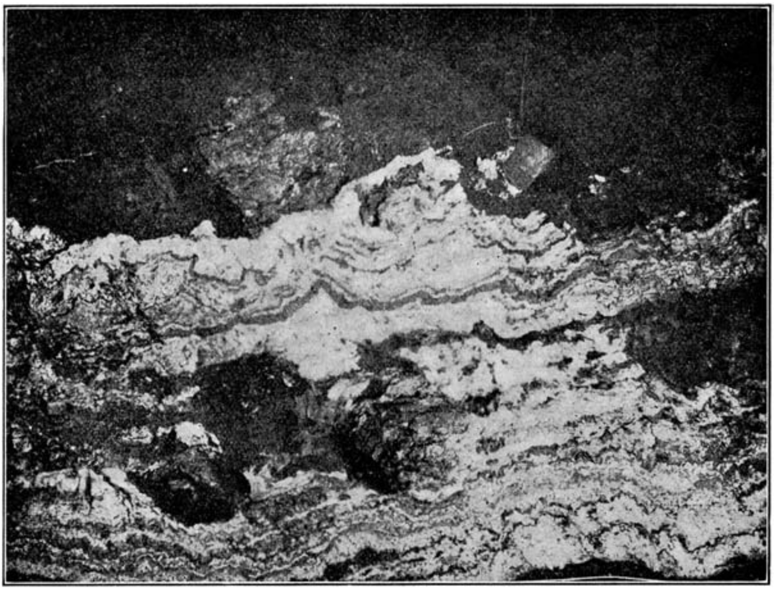

FIG. 180.—Photograph of branch vein on 1,000-foot level, El Oro Mining and Railway Company, showing pronounced crustification by sulphides; vein 3 feet wide contains 9½ ounces gold and 165 ounces silver per ton. Open cavity in middle.

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[1] F. C. Schrader, *Bull.* 397, U. S. Geol. Survey, 1909.
F. L. Ransome, *Bull.* 743, *idem,* 1923.

yellow quartz layer (Fig. 182). The rock alteration is chloritic an sericitic.

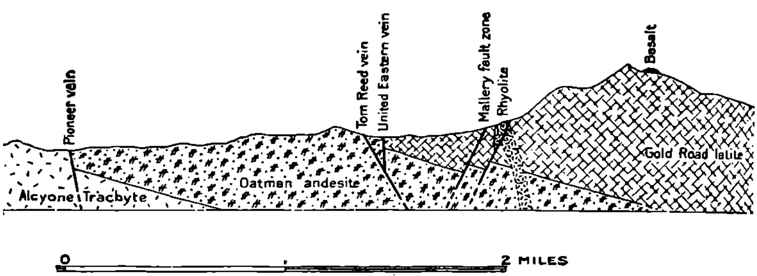

FIG. 181.—Section across the Black Mountains, Arizona, showing position of fault veins. (*After F. L. Ransome, U. S. Geol. Survey.*)

## GOLD-QUARTZ VEINS IN RHYOLITE

The Tertiary rhyolites in the Cordilleran region often contain gold-bearing veins. These veins are poor in ore minerals other than gold but usually contain some argentite, pyrargyrite, and pyrite. The gold is frequently coarse and accompanied by more or less silver. Among the

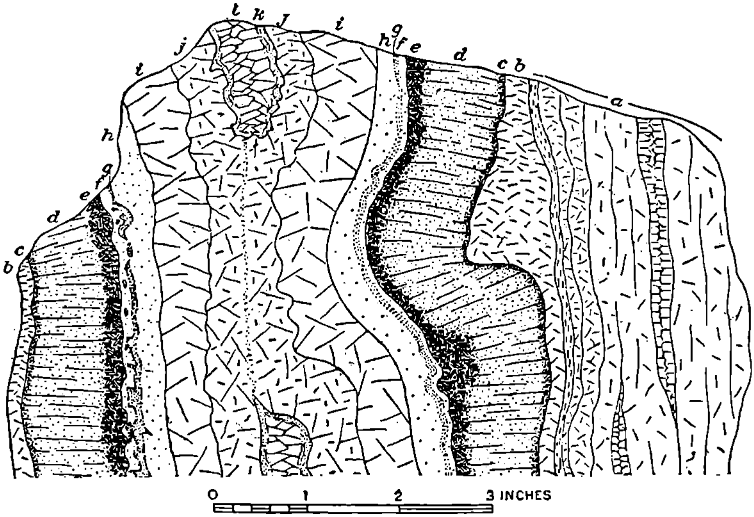

FIG. 182.—Typical banded ore from United Eastern mine. *a*, brown quartz; *b*, white milky quartz; *c*, thin layer of honey-yellow quartz; *d*, clear quartz; *e*, layer of dark calcite, quartz and adularia; *f*, fine-grained quartz, *g*, honey-yellow quartz; *h*, like *f*; *i*, milk-white quartz; *k*, honey-yellow quartz; *l*, clear comb quartz. (*After F. L. Ransome, U. S. Geol. Survey.*)

gangue minerals quartz prevails, but in most cases it is associated with much adularia, probably derived from the surrounding potassic rock. Calcite and fluorite are also common, but barite is rare. The veins are almost always characteristically pseudomorphic, with bladed and cellular quartz and adularia, which replace calcite and fluorite. Both veins and sheeted zones occur; in the latter there is little gangue and the gold, as in the Jumbo mine at Hart, California, may be embedded in apparently fresh rhyolite.

There is no real propylitic alteration of the country rock but often extensive silicification and much finely disseminated pyrite. The silicification is attended by concentration of potassium as adularia or sericite. The decomposed upper zone of the veins contains clay seams that may be extremely rich in gold and secondary silver minerals, as at De Lamar, Idaho, and Rawhide, Nevada.

At the De Lamar mine[1] a series of parallel, gently dipping veins of the kind described abut against a fissure filled with tough clay (the "iron dike") near which the best ore is found (Fig. 183). Below a vertical depth of about 800 feet the values are low, although the veins persist. Free gold is rarely seen. The proportion of gold to silver, by weight, is 1:20. Argentite and naumannite[2] ($Ag_2Se$) occur, the latter in large specimens. The veins are ordinarily 1 to 6 feet thick and distinctly filled, though transitions by silicification were also noted. The filling is wholly quartz, pseudomorphic after calcite, and forms a cellular mass of thin plates covered by minute crystals (Fig. 168). The value of the ore ordinarily ranges from $10 to $20 per ton. The De Lamar mine has yielded gold and silver to the value of about $7,000,000, but is now closed.

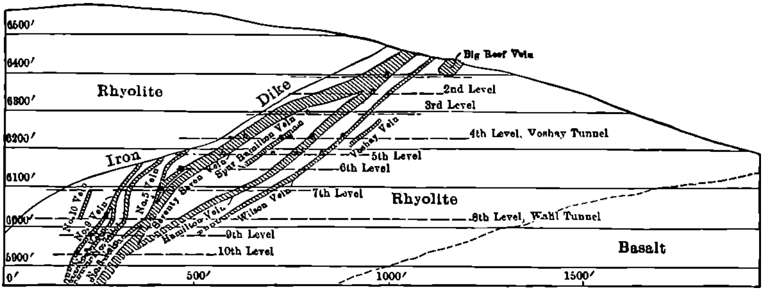

FIG. 183.—Vertical section of the vein system at De Lamar, Idaho.

The gold veins of Jarbidge[3] in northern Nevada are very similar to those of the Black Mountains, Arizona. Gold-bearing veins intersect an older rhyolite flow, which is covered by later barren rhyolite. The fine-grained gold is accompanied by a little pyrite, chalcopyrite, argentite, and naumannite. The gangue is pseudomorphic quartz, with unusually

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[1] W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, p. 122. A. M. Piper and F. B. Laney, *Bull.* 11, Idaho Bur. Mines, 1926.

[2] E. V. Shannon, The minerals of Idaho, *Bull.* 131, U. S. Nat. Mus., 1926.

[3] F. C. Schrader, *Bull.* 741, U. S. Geol. Survey, 1923.

large amounts of adularia sometimes making up 50 per cent of the materials. The gold appears to be the latest mineral introduced.

Entirely similar veins are found at Rawhide,[1] Gold Circle,[2] Round Mountain,[3] Beatty,[4] and many other places in Nevada[5] and Utah.[6]

The extremely rich ore of the National[7] vein in northwestern Nevada (Fig. 184) has yielded about $3,000,000 from a narrow shoot of coarse gold followed to a depth of 800 feet on the vein. The veins of that district, except for this occurrence, are of the stibnite type. The native gold contains 50 per cent silver and is more properly called electrum.

Fig. 184.—Dendritic gold (black) in extremely fine-grained quartz of probable colloidal deposition. White area is coarsely crystalline comb quartz, National, Nevada. Magnified 15 diameters.

## ARGENTITE-GOLD-QUARTZ VEINS

**Tonopah, Nevada.**[8]—The Tonopah district, discovered in 1900, is situated in a group of desert hills in western Nevada and was for many years the most important of the western silver- and gold-producing

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[1] A. F. Rogers, *Econ. Geol.*, 6, 1911, p. 790.

[2] W. H. Emmons, *Bull.* 408, U. S. Geol. Survey, 1910.

[3] F. L. Ransome, *Bull.* 380, *idem*, 1909, pp. 44–47.

[4] F. L. Ransome, W. H. Emmons, and G. H. Garrey, *Bull.* 407, *idem*, 1910.

[5] S. H. Ball, *Bull.* 308, *idem*, 1907, p. 46.

[6] B. S. Butler *et al.*, *Prof. Paper*, 111, *idem*, 1920, p. 179.

[7] W. Lindgren, *Bull.* 601, *idem*, 1915.

[8] J. E. Spurr, Geology of the Tonopah mining district, *Prof. Paper* 42, *idem*, 1905; also, *Min. and Sci. Press*, Apr. 22, 1911.

J. A. Burgess, Geology of the producing part of the Tonopah district, *Econ. Geol.*, 4, 1909, pp. 681–712.

A. Locke, The geology of the Tonopah mining district, *Trans.*, Am. Inst. Min. Eng., 43, 1913, pp. 157–166.

J. E. Spurr, Geology and ore deposition at Tonopah, Nevada, *Econ. Geol.*, 10, localities. In 1916, the production amounted to nearly $2,000,000 in gold and 8,700,000 ounces of silver. In 1930, this had diminished to $406,000 in gold and 1,900,000 ounces of silver, obtained from 114,500 tons of ore. The ore is treated by concentration and cyaniding. The total output, to 1930 inclusive, amounted to $35,371,000 in gold and 161,600,000 ounces of silver. Naturally, the decline in silver prices has affected the district unfavorably.

The veins of Tonopah are typically epithermal and intersect a complex Tertiary volcanic series in which the succession has been variously interpreted. Opinions also differ as to the proper nomenclature for the rock. Spurr considers some of the igneous bodies as intrusive, while Burgess and Locke hold that the various formations originated as surface flows. The importance of the volcanic succession is readily seen, for, if any apparently underlying rock is really intrusive and later than the vein formation, no continuation of the bonanza veins can be expected in it. Spurr held there were several epochs of vein formation, whereas Burgess, Locke, and Nolan are inclined to recognize only one epoch.

Much of the later work at Tonopah has been carried on in the West End of the district and has disclosed valuable information on succession and mineralization.

Nolan recognizes an early "Tonopah formation" probably of Miocene age, consisting of rhyolite flows, tuffs, breccias, and well-laminated silts. Some dark lavas (Sandgrass andesite) are interbedded. The Mizpah trachyte overlies the Tonopah formation conformably, but in places the West End rhyolite is intruded between them. Later formations, younger than the ore, are the Midway andesite and various dikes of the Oddie rhyolite.

In the West End, the flows and beds dip rather steeply to the west. There are much fracturing and faulting in detail. In the West End, the tilted formations are cut by a remarkable compound fault, flat and convex upward. This has played an important part in the protracted litigations concerning apex rights. The ore-bodies thus far mined are localized either in this fault or in others that split from it or radiate from its crest. The strike of the productive fissures is generally east-west.

The main faulting occurred after the intrusion of the West End rhyolite, and the mineralization followed soon afterward. The silver ores are considered as somewhat later than the early barren quartz. The

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1915, pp. 713–769.

E. S. Bastin and F. B. Laney, The genesis of the ores at Tonopah, *Prof. Paper* 104, U. S. Geol. Survey, 1918.

G. J. Young, *Eng. and Min. Jour.-Press*, 120, 1925, p. 824 (present conditions).

T. B. Nolan, The underground geology of the western part of the Tonopah district, *Bull.* 24 (4), Univ. Nevada, 1930, 35 pp.

deepest workings extend to about 2,000 feet. Among the principal mines are; the Tonopah, the Tonopah Belmont, Tonopah Extension, and the West End. While the silver content of the ore decreases in depth, it is held that further search may disclose important ore-bodies.

The mineralization is similar in the Tonopah formation, in the Mizpah trachyte, and in the West End rhyolite. Quartz, pyrite, sericite, and adularia, often also rhodochrosite, are the common gangue minerals. The quartz is of fine, though variable, grain with banded structure and "chalcedonic" appearance containing narrow bands of finely divided sulphides. Argentite and polybasite are the principal ore minerals, with small amounts of pyrite, chalcopyrite, galena, and sphalerite. Selenium is present probably as a silver selenide. The concentrates are very rich in silver containing up to 25 per cent Ag and 2 per cent Se. Arsenic and antimony are not abundant. Small amounts of scheelite and hübnerite have been found. The ore worked contains about 25 ounces of silver and 0.125 ounce of gold per ton. The costs of mining and reduction are about $15.00 per ton, hence declining silver prices make operations difficult. The occurrence of the ore-shoots are dependent upon the permeability and intersections. Many of the fissures have been sealed by the earlier barren quartz.

For supergene enrichment see Chapter XXXII, page 866.

**The Comstock Lode.**[1]—Discovered in 1859, the celebrated Comstock Lode has yielded, to the end of 1911, a total production of $381,400,000 in silver and gold, of which the gold amounted to $153,000,000. The bonanza period fell in the seventies of the last century and the production since then has declined greatly. Present work is confined to upper levels and repeated attempts to mine a large tonnage of low grade ore have been unsuccessful. In 1924, the lode yielded $2,200,000 in gold and silver from 544,000 tons of ore; in 1930, $60,000 represented the value of the output. The Comstock Lode, situated near the summit of the Virginia Range, east of the Sierra Nevada, in Nevada, is a fault fissure of great throw (Fig. 185), traceable two and one-half miles and in places several hundred feet wide, the vein matter of the lode spreading into the hanging wall. Great bonanzas of crushed quartz, in part exceedingly rich in silver minerals, were found at intervals along the lode, especially in chambers or vertical fissures probably produced by normal faulting of the hanging wall. The greatest vertical depth attained below the outcrop is about 3,000 feet. Mining has been greatly hampered by enormous quantities of hot water containing mainly calcium sulphate. None of the great bonanzas was found below a depth of 2,000 feet.

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[1] G. F. Becker, *Mon.* 3, U. S. Geol. Survey, 1882.

J. A. Reid, *Bull.* 4, California Univ., Dept. Geol., 1905, pp. 177–199.

E. S. Bastin, Bonanza ores of the Comstock lode, *Bull.* 735, U. S. Geol. Survey, 1923, pp. 41–63.

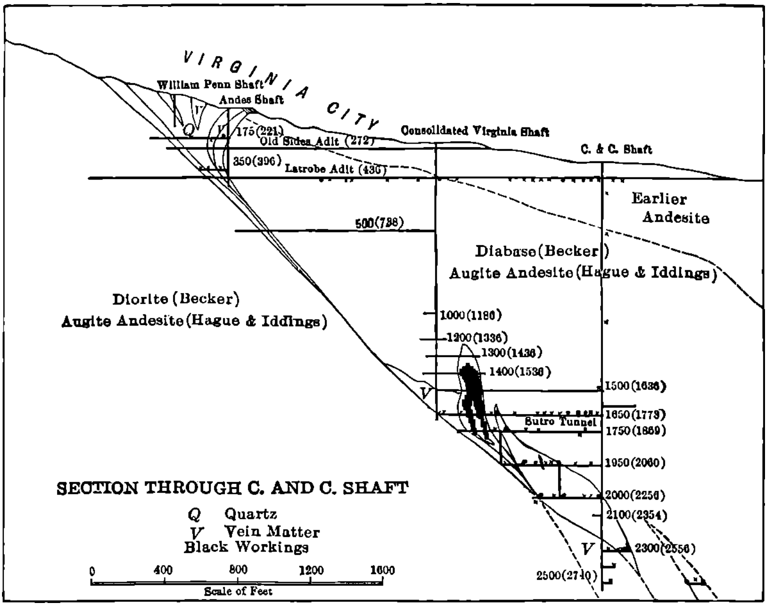

FIG. 185.—Vertical cross-section through the Comstock lode, showing chambered outcrop and bonanza in vertical vein in hanging wall. (*After G. F. Becker, U. S. Geol. Survey.*)

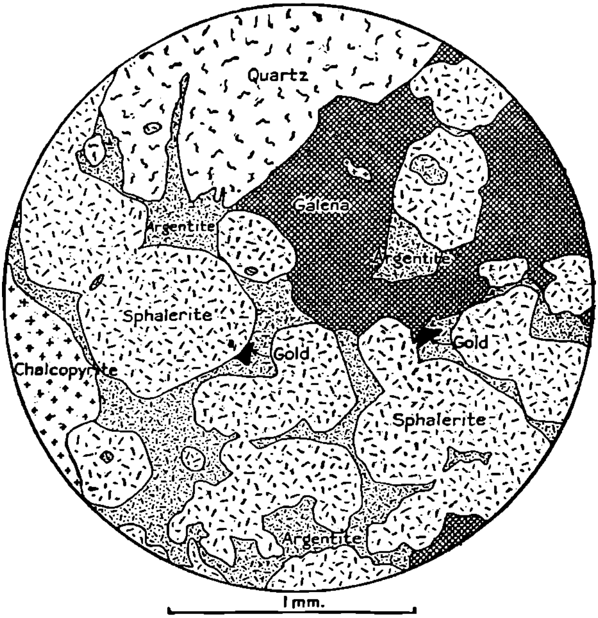

FIG. 186.—Bonanza ore from C. & C. mine Comstock lode, East vein, between 2,000- and 2,050-foot levels, carrying abundant primary argentite associated with galena, sphalerite, chalcopyrite, quartz, and gold. Camera lucida drawing. (*After E. S. Bastin.*)

The lode intersects igneous rocks of deep-seated type, showing transitions and variously classified as diorite, diabase, and augite andesite.[1] These are covered by andesite flows of Tertiary age, which are also mineralized. Both classes of rocks have suffered propylitic alteration, and analyses of the clay gouge near the veins show that sericitic alteration has set in along the principal channels which the solutions followed. The ores consist of quartz and some calcite, in places banded with pyrite, galena, chalcopyrite, and sphalerite. The valuable minerals are mainly pale gold, argentite, and polybasite. They appear to replace quartz, sphalerite and galena (Fig. 186).

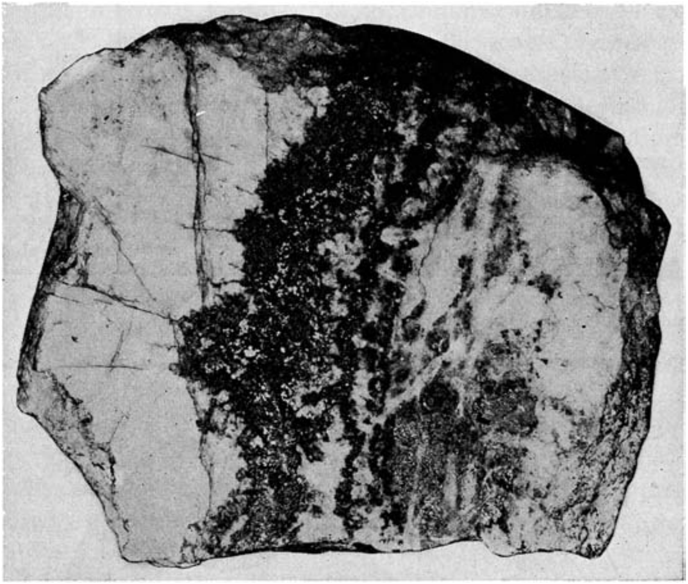

Fig. 187.—Rich ore, Ophir mine, Comstock lode. Right side, altered country rock with pyrite and quartz veinlets. Left side, three generations of galena, sphalerite, chalcopyrite and argentite (dark), and quartz (white). Natural size.

There are in places two generations of quartz, as shown in Fig. 187, the older quartz containing principally pyrite. The quartz is fine-grained and in great part probably replaces crushed rock. There is no lamellar or hackly quartz. According to Reid, the descending waters, rich in sulphates, contained notable amounts of gold and silver, and small quantities of these metals were also present in the ascending hot sulphate and carbonate waters. Bastin recognizes as supergene minerals only silver, with some argentite and polybasite above the 500-foot level.

## ARGENTITE VEINS

The argentite veins have numerous representatives in Mexico, as at Pachuca, Real del Monte, and Guanajuato. In general they intersect

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[1] A. Hague and J. P. Iddings, *Bull.* 17, U. S. Geol. Survey, 1885.

andesitic rocks of supposedly Miocene age but also cut adjacent or underlying Mesozoic sediments.

The rich and long-worked veins of Pachuca[1] have come into renewed prominence by the successful application of the cyanide process to their ores. A complicated vein system intersects andesite flows covering Cretaceous sediments. The andesite is extensively propylitized; and this altered rock also adjoins the veins, near which, however, a silicification is often superimposed upon the chloritization. The veins, 8 to 14 feet wide, are filled fissures, crustified or brecciated, with quartz, sometimes amethystine, as the principal constituent of the gangue; there are also rhodochrosite, rhodonite, adularia, and calcite, the last named being the most recent. The ore-shoots are more or less horizontal.

The ore minerals consist of argentite, stephanite, polybasite, pyrite, galena, and sphalerite. The veins have been followed to a depth of 2,000 feet where the veins are said to become poor, and here contain argentite, pyrite, and sphalerite. The ores average about 16 ounces of silver to the ton with very little gold. The cost of mining and milling is about $6 per ton. With economy the veins can be worked with silver at 30 cents per ounce. The oxidation is marked by the zones of the "colorados" and the "negros," the first of which contains limonite with silver haloid salts and the second much manganese.

The veins of the old and famous silver-mining district of Guanajuato[2] are similar to those of Pachuca.

## GOLD TELLURIDE VEINS

**Cripple Creek.**[3]—The veins of Cripple Creek, situated in an otherwise barren part of Colorado, have since 1891 annually yielded a large amount of gold, which in 1900 reached a maximum of $18,000,000. In 1916, the production of gold was valued at $11,800,000; in 1925, at about $5,000,000; in 1931, at $2,400,000. The silver recovered was about 9,000 ounces in 1930. The total output of the district to the end of 1931 was nearly $360,000,000. Individual mines have produced from $10,000,000 to $30,000,000 each. The district lies on a granitic plateau a few miles

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[1] J. Aguilera and E. Ordóñez, El mineral de Pachuca, *Boletín*, Inst. geol. Mexico, Nos. 7, 8, 9, 1879.

H. V. Winchell, *Trans.*, Am. Inst. Min. Met. Eng., 66, 1922, pp. 27–36.

M. Santillán, *Bol. Minero*, February, 1931.

[2] A. Wandke and J. Martinez, *Econ. Geol.*, 23, 1928, pp. 1–44.

[3] W. Cross and R. A. F. Penrose, Jr., *Sixteenth Ann. Report*, U. S. Geol. Survey, pt. 2, 1896, pp. 1–209.

W. Lindgren and F. L. Ransome, *Prof. Paper* 54, *idem*, 1906.

Horace B. Patton, The Cresson Bonanza at Cripple Creek, *Min. and Sci. Press*, Sept. 15, 1917.

G. F. Loughlin, Ore at deep levels in the Cripple Creek district, *Trans.*, Am. Inst. Min. Met. Eng., 75, 1927, pp. 42–73.

G. F. Loughlin and A. H. Koshmann, *Proc.*, Colo. Sci. Soc. 13, 1935, pp. 217–435.

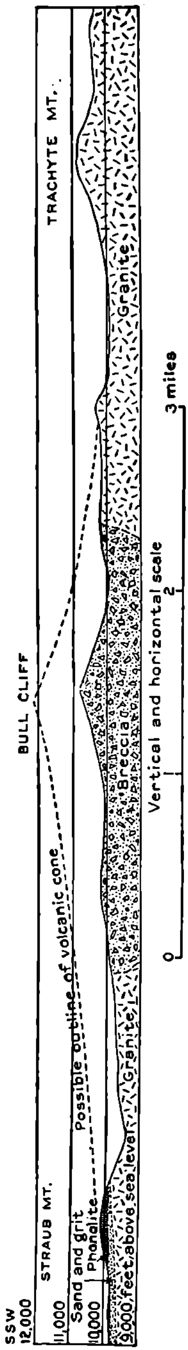

FIG. 188.—Section showing probable outline of the Cripple Creek volcanic cone at the close of the volcanic epoch. (*After Lindgren and Ransome, U. S. Geol. Survey.*)

southwest of Pikes Peak, at elevations of 9,000 to 11,000 feet. Within a few square miles are a large number of producing mines; 64 reported in 1916 the production of 975,270 short tons of ore, averaging $12 per ton. In earlier years the average value was $30 to $40 per ton; but during recent years an increasing quantity of low-grade ore from dumps, etc., has been treated, some of it containing only $3 or $4 per ton. In 1927 the most important mines were the Golden Cycle, the Portland, and the Cresson. Most of the ore is reduced in large mills at Colorado Springs, by roasting and cyaniding.

The mining operations have always suffered from a large quantity of mine waters. The greatest depth attained is about 3,000 feet in the Portland mine. The Roosevelt tunnel, nearly 3 miles long, now drains the mines to an elevation of 8,020 feet. This tunnel, in 1916, discharged about 10,000 gallons per minute and has now reached the Portland mine at an elevation of 8,112 feet and 164 feet below the 1,900 foot level.

The rocks constituting the plateau are pre-Cambrian red granite, with some gneiss and fibrolite schist. Breaking through this basement is a mass of Tertiary volcanic rocks, the area having a diameter of 2 or 3 miles. As shown by the mining operations, the contact surface between the granite and the volcanic mass is steep, or even vertical or overhanging; and there is little doubt that this "plug" of volcanic material represents the core of a Tertiary volcano which formerly rose above the plateau, as tentatively indicated in Fig. 188. The bulk of the remaining core is composed of tuffs and breccias of latite-phonolite which are cut by dikes and intrusive masses of phonolite and syenite. The latest manifestations of volcanism were basic dikes of monchiquite and vogesite; and the veins were apparently formed soon after these dikes had been intruded. Many of the rocks contain a notable amount of combined water. The deposits are veins, in many places complex and numerous, which followed a system of roughly radiating, steep fissures (Fig. 39), believed to have resulted from compressive stress developed in a settling volcanic mass. The physiographic history of the district indicates that the surface at the time of vein formation was only from a few hundred to 1,000 feet above the present surface.

In the granite adjoining the contact are also found irregular bodies of ore, formed by replacement. Most of the veins are closely spaced sheeted zones (Fig. 48) a few feet wide, though some attain a width of 20 to 40 feet. The ore deposition has generally taken place by filling along the narrow fissures and only to a smaller extent by replacement of the intervening or adjoining rock. Low-grade ores are formed by mineralization of narrow seams in the country rock.

Fig. 189.—Vein filling, Portland mine, Cripple Creek (purple quartz). *f*, fluorite; *q*, quartz of coarser and finer grain; *p*, pyrite. Magnified 50 diameters.

Fig. 190.—Filled space of dissolution in granite, Independence mine, Cripple Creek (granite ore). *o*, orthoclase of granite; *m*, biotite replaced by adularia and pyrite; *v*, adularia (secondary orthoclase), showing crustification; *q*, quartz; *p*, pyrite. Magnified 14 diameters.

There is little or no native gold, except in the oxidized zone (p. 859). The principal ore mineral is calaverite ($AuTe_2$, with but little silver), of yellowish white color and often well crystallized. Associated with this are small quantities of pyrite, sphalerite tetrahedrite, stibnite, cinnabar, and molybdenite, rarely hübnerite. The gangue consists of quartz and fluorite, with some dolomitic carbonate, in places also roscoelite, a vanadium mica, which often accompanies rich ore. The fluorite, quartz, and calaverite are often intergrown, forming a fine-grained rock which has the purple color of the fluorite (Fig. 189). The vein structure is drusy, and the calaverite was among the latest minerals formed. Specimen ore may contain about 60 per cent quartz, 20 per cent dolomite, 20 per cent fluorite, 0.1 per cent gold, and 0.2 per cent tellurium, with iron, copper, zinc, and molybdenum present in fractions of 1 per cent. Extremely rich shoots of calaverite ore are sometimes found. The Cresson bonanza forms a large pipe or chimney of basaltic breccia, subsequently fissured and mineralized, with a diameter of many hundred feet. It is still being mined.[1] From 1903 to 1930, inclusive, the Cresson pipe produced 1,800,000 tons of ore valued at $32,000,000.

The replacement ore consists of the ordinary red granite, often drusy and partly replaced by adularia, fluorite, and calaverite (Fig. 190). Loughlin recognizes three stages of mineralization at Cripple Creek, all three beginning with quartz and pyrite, the tellurides appearing principally in the second stage.

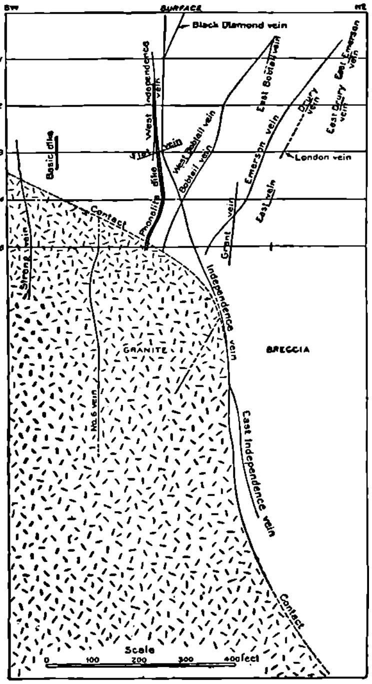

FIG. 191.—Section through Stratton's Independence mine, Cripple Creek, showing relation of veins to granite-breccia contact. (*After Lindgren and Ransome, U. S. Geol, Survey.*)

In the upper levels of the mines there are a great number of short veins all of which carry more or less ore. In the aggregate these veins contain an enormous amount of ore, some of which is of exceedingly high grade. The veins are less abundant in the lower levels and of lower grade. A few veins continue, however, to the lowest levels. The vents from which the ore solutions rose were located near the Portland, Vindicator, and Golden Cycle mines.

As shown in the report cited, the ore-shoots are to a marked degree influenced by intersections with other veins or dikes, but many of the largest and richest shoots have no such relation. The greatest horizontal extension of a shoot is 1,300 feet. Many shoots terminate in depth, while others have continued to the greatest depths attained. The relations at Stratton's Independence and Portland mines are illustrated in Figs. 191 and 192 and the latest available data, published by G. F. Loughlin, show the whole shoot of the Portland Mine down to the 3,000-foot level. The lower levels, below the drainage tunnel, were abandoned in 1930.

The tuffs and breccias are generally altered and contain some fine-grained pyrite, which has little value; the dark silicates alter to carbon-

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[1] It is illustrated in publication by G. F. Loughlin, *op. cit.* The contact between granite and breccia near the Portland mine is also mapped in detail.

ates, fluorite, and pyrite and the feldspars to sericite and adularia. Cross and Penrose thought this propylitic alteration earlier than the veins, while Lindgren and Ransome consider it to be caused by the same kinds of solutions that filled the fissures. Similar differences of opinion have been expressed in relation to the propylitization at other places (p. 461).

There is no evidence that there has ever been an active circulation of surface water in the district. The porous breccias and in general the whole volcanic plug are filled with stagnant water, while there is little water in the surrounding granite. The general conclusion of Lindgren and Ransome is that the vein-forming epoch was brief and that the remarkable and abundant telluride ores were formed by alkaline solutions emanating from deeper igneous masses, the last effects of these emanations being the exhalations of carbon dioxide and nitrogen, which have not yet subsided. The waters ascended rapidly in the deeper parts of the volcanic plug, but near the surface they spread out in more numerous fissures and precipitation, followed by cooling or mixture with descending solutions.

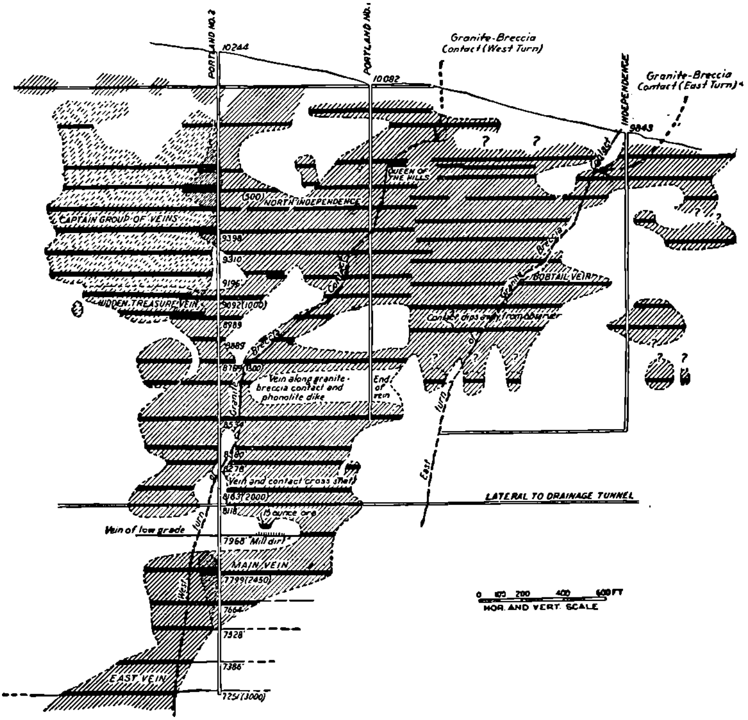

FIG. 192.—Longitudinal section showing principal stopes on veins in Portland and Independence mines, Cripple Creek. (*After G. F. Loughlin.*)

## GOLD SELENIDE VEINS

**Occurrence of Selenides.**[1]—In minute quantities selenium is present in many deposits, particularly in the pyritic copper deposits and the ores of the Mansfeld district, Germany; it is recovered during the electrolytic refining of copper. As distinct minerals the selenides are found occasionally in small veins in pre-Cambrian rocks as at Falun and Skrikerum in Sweden, and in Paleozoic rocks in The Harz, Germany. But their real home is in the epithermal veins; they are not usually associated with tellurides. Their presence in some rare quicksilver deposits has already been mentioned. In the silver veins of Mexico near Guanajuato selenides of silver and lead have been found; and in some silver-gold veins, like those at Tonopah and De Lamar, or gold-silver veins, like those at Waihi, they are important constituents. At both Tonopah and Waihi other minerals are present in quantities. In larger amounts lead and copper selenides are present in veins in limestone in Sierra de Umango,[2] Argentina, (umangite, $Cu_3Se_2$; tiemannite, $HgSe$) and in "trachyte" in Cerro de Cacheuta,[3] Argentina (naumannite, $Ag_2PbSe$; clausthalite, $PbSe$; zorgite, $(PbCu)Se$). Little is known concerning these occurrences. In 1930, about 278,300 pounds of selenium were produced in the United States, mainly from the residues of copper refineries. Selenium is valuable for coloring and decolorization of glass; for electric apparatus on account of the variation of conductivity, and it could be used to advantage in the rubber industry if a cheaper supply could be obtained. The price is about $2 per pound. There is little demand for tellurium, which commands the same price.

The type of veins described in these paragraphs is remarkably free from ore minerals other than native gold and selenides, and is rare, only two examples being known, that of Republic, in Washington, and that of Radjang Lebong, in Sumatra. In some respects, however, the Tonopah veins are allied to this type.

In both the places mentioned there is a predominating gangue of very fine-grained quartz, beautifully banded by crustification, but not markedly drusy; it has a "chalcedonic" appearance, although there is really but little chalcedony present, and resembles strongly, both in hand specimen and in thin section, some kinds of sinter deposited at the orifices of hot springs. The gold is present in very fine distribution and the gold selenide has not yet been positively identified. These veins have probably been deposited close to the surface as colloids, and subsequently crystallized as fine-grained quartz.

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[1] For best description of selenium minerals see Doelter's Handbuch der Mineralchemie, 4 (1), 1926, pp. 804-832.

[2] F. Klockmann, *Zeitschr. Kryst.*, 19, 1891, p. 267.

[3] J. Domeyko, *Compt. Rend.*, 63, 1864, p. 866.

**Republic, Washington.**—At Republic,[1] a series of Miocene andesite and latite flows filling an old valley have been intruded by a mass of latite porphyry belonging to the same general period of eruption. Over a considerable area the andesite and porphyry have suffered normal propylitization, chlorite, earthy carbonates, and pyrite being the principal minerals formed. A series of parallel fractures dipping from 38° to 80° have been opened in the volcanic rocks and are occupied by sharply defined veins averaging 3 or 4 feet in width. Against these the propylitic rock borders, usually with little further alteration. The banded filling (Fig. 193) consists of quartz and calcite and also includes dark masses of jasperoid of uncertain derivation. There is adularia in the filling and in some places can be seen the beginning of a replacement of calcite by fine-grained quartz and adularia. In one mine the quartz filling has been replaced by laumontite containing much silver.

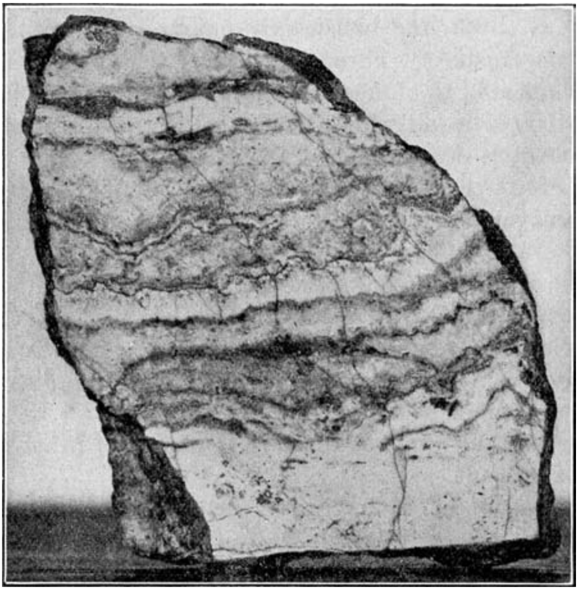

FIG. 193.—Typical ore, Republic mine, Washington. Fine-grained colloform quartz, with narrow streaks of sulphides and selenides (?). One-half natural size.

Free gold is rarely visible, but the valuable portions of the veins lie along narrow dark bands that are parallel to the crustification and are believed to represent finely divided gold selenide. Local crusts are rich in free gold, tetrahedrite, and chalcopyrite and this material contains about 2 per cent of selenium, which, according to experiments by Dr. Chase Palmer, of the United States Geological Survey, is probably combined with gold.

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[1] Howland Bancroft and W. Lindgren, *Bull.* 550, U. S. Geol. Survey, 1914.

The ores of Republic have proved difficult to treat by the cyanide process. Their grade varies greatly, averaging perhaps $11 per ton. The proportion of gold to silver by weight is about 1:3. Oxidation has resulted in small changes but has set some silver free. Since 1897 the district has yielded about $9,400,000. The greatest depth attained is 950 feet.

**Sumatra.**—The Radjang-Lebong field,[1] in southern Sumatra, has yielded much gold in recent years. The annual production is about 50,000 ounces of gold and 300,000 ounces of silver. Andesite is the country rock and the principal vein, which has a width of 17 feet, is divided into five well-defined seams separated by silicified andesite. According to R. Beck, the bluish-gray quartz is beautifully banded in thin concentric crusts of "fibrous quartz." The rich ore, like that of Republic, is indicated by thin dark dendritic crusts similar in appearance to the quicksilver selenide from Lerbach, and the silver selenide at Tilkerode (both localities in the Harz Mountains).

The ore, of which only a small part is amenable to amalgamation, contains on an average 41 grams of gold and 318 grams of silver per metric ton. There are a little pyrite and chalcopyrite. The gold bullion, according to Truscott,[2] contains 4.35 per cent selenium and small amounts of copper, lead, zinc, and iron.

Closely related to this is the silver vein Gedang Ilir,[3] which contains much argentite and a little chalcopyrite, also scattered grains of magnetite in a gangue of quartz, adularia, and inesite ($2MnSiO_3.H_2O$). Most of the rich ore is oxidized and contains large amounts of $MnO_2$ which makes the treatment difficult.

Argentite-gold veins have been described from Omui, Hokkaido,[4] Japan in which apparently these minerals have crystallized from a colloidal precipitate. The country rock is rhyolite. Possibly they belong to this class.

## THE BASE-METAL VEINS

Among the deposits formed relatively near the surface by ascending thermal waters in genetic connection with igneous rocks, ores rich in the baser metals and worked principally for these metals are rather exceptional. Heavy deposits of pyrite and chalcopyrite ore are very seldom

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[1] R. Beck, Lehre von den Erzlagerstätten, 1, 1909, p. 488.

N. Wing-Easton, Die wichtigsten Edelmetall Lagerstätten in Sumatra, Preuss. geol. Landesanstalt, *Archiv für Lagerstätt., Forsch.* 35, 1926.

[2] M. Maclaren, Gold, London, 1909, p. 298.

S. J. Truscott, *Trans.*, Inst. Min. and Met. (London), 10, 1902, p. 53.

[3] G. Clevenger and M. H. Caron, The treatment of manganese silver ores, *Bull.* 226, U. S. Bur. Mines, 1925, p. 61.

[4] C. Iwasaki, Kotô Commemoration Volume, Tokyo, 1925.

found; one instance is furnished by the Nacozari district, in Mexico, where, at the Los Pilares mine,[1] a large body of low-grade pyrite and chalcopyrite is worked. The ores occur mainly in the interstices of a brecciated rhyolite in a series of fractures arranged in almost circular form. (Fig. 40.) The pyrite is often crystallized; the gangue is quartz; the ore is poor in gold and silver.

Somewhat similar are the veins of Ashio, Japan, which contain chalcopyrite and quartz. They cut Tertiary rhyolites.

The veins and stocks of the San Juan region, Colorado, described in more detail below, are in places rich in lead, zinc, and copper, but yield principally gold and silver. Both galena and sphalerite occur, but the copper is derived mainly from tetrahedrite or enargite.

The Schemnitz deposits, in Hungary, consist of a strong vein system intersecting rhyolite and andesite above Triassic slates and Eocene strata. The gangue minerals are fine-grained quartz and amethyst, together with later calcite, ankerite, rhodochrosite, rhodonite, siderite, and barite, with much pyrite, galena, chalcopyrite, and sphalerite. Among the rarer minerals are adularia, fluorite, and diaspore. The proportion of gold to silver by weight is 1:23.

The Bull-Domingo and Bassick deposits,[2] at Silver Cliff, Colorado, yielded principally gold and silver. The ore consisted, however, largely of sulphides and tellurides, which in the Bassick mine occurred in what is considered a volcanic neck. The Bassick deposit was mined to a depth of 800 feet and yielded rich ore. The cross-section covered about 25 by 100 feet and the ores encrusted the fragments of volcanic rocks filling the shoot.

When the solutions depositing veins in volcanic rocks leave the flows and enter into the surrounding limestones and other sedimentary rocks, deposition by selective precipitation comes into play and ores rich in sulphides, particularly galena, may be formed. Examples of this are found in several deposits of the Ouray district, Colorado, described by J. D. Irving (*Bull.* 260, U. S. Geol. Survey, 1905). Fine-grained silica here replaces the limestone showing that the temperature was lower than that necessary for the development of lime silicates.

## THE SAN JUAN REGION, COLORADO[3]

**General Features.**—One of the most interesting metallogenetic provinces is that of the rugged San Juan region, in southwest Colorado,

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[1] S. F. Emmons, *Econ. Geol.*, 1, 1906, pp. 629–643.

[2] S. F. Emmons, The mines of Custer County, Colorado, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, pp. 430–447.

[3] W. Cross and C. W. Purington, *U. S. Folio* 57, Telluride; Cross, Spencer, and Purington, *U. S. Folio* 60, La Plata; Cross, Howe, and Ransome, *U. S. Folio* 120, Silverton; Cross and Ransome, *U. S. Folio* 130, Rico; Cross, Howe, Irving, and

including the mining districts of Telluride, Ouray, Silverton, Bonanza, Lake City, Rico, Needle Mountains, La Plata, and Creede (Fig. 194).

The San Juan Mountains consist chiefly of volcanic rocks, poured out over a basement of Paleozoic and Mesozoic sediments; pre-Cambrian rocks are also exposed in places (Fig. 195). The volcanic flows, occupying over 3,000 square miles, have a total thickness of many thousand feet

FIG. 194.—Map showing approximate distribution of the principal mining regions in Colorado. (*After J. E. Spurr, U. S. Geol. Survey.*)

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W. H. Emmons, *U. S. Folio* 131, Needle Mountains; Cross and Hole, *U. S. Folio* 171, Engineer Mountain; Cross, Howe, and Irving, *U. S. Folio* 153, Ouray.

T. A. Rickard, Across the San Juan Mountains, New York, 1903.

J. E. Spurr, summary in *Prof. Paper* 63, U. S. Geol. Survey, 1908, pp. 111–168.

E. S. Bastin, Silver enrichment in the San Juan mountains, *Bull.* 735, U. S. Geol. Survey, 1922, pp. 65–129.

B. S. Butler, Relation of the ore deposits of the southern Rocky Mountain region to the Colorado plateau, *Proc.* Colorado Sci. Soc., 12, 1929, pp. 23–36.

and were erupted at intervals during the whole Tertiary period. The lowest formation, called the San Juan tuff, consists of about 3,000 feet of andesitic material. Covering this are andesites, latites, and rhyolites, called the Silverton series, 4,000 feet thick; this in turn is overlain by the Potosi series, consisting of rhyolite and latite. The eruptions were separated by epochs of erosion and were closed by the effusion of the Hinsdale series of rhyolites, andesites, and basalts.

Deep erosion has laid bare these flows to a depth of several thousand feet and exposed a number of smaller intrusive stocks and sheets of dioritic or monzonitic character and granular or porphyritic texture. The very latest intrusions are represented by dark dikes of lamprophyric character.

This vast uplifted and eroded dome of volcanic rocks is traversed by numerous systems of strong fissure veins, many of them continuous for several miles. Their development followed closely after the latest epoch of volcanic activity and they intersect several of the intrusive masses. While the vein systems bear the marks of deposition within a moderate distance of the original surface, there are some features that tend to connect them with deposits formed at greater depth and in more direct genetic connection with igneous intrusions.[1] Many types are represented, including normal veins, stocks, replacement deposits in sedimentary formations, and some small contact-metamorphic deposits near the contacts of the intrusive masses. Some of the veins are exposed over vertical distances of several thousand feet. The Virginius vein, above Ouray, for instance, has been mined for a vertical distance of 3,300 feet; the Revenue tunnel is 2,400 feet below the outcrops.

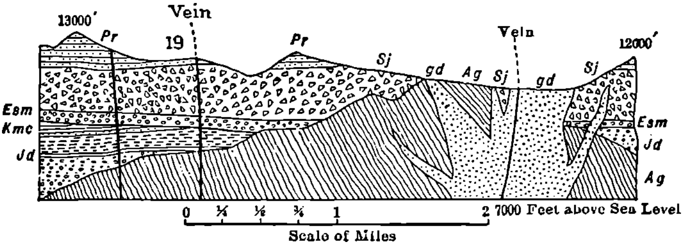

FIG. 195.—Section at Telluride, Colorado. *Ag*, Algonkian quartzite; *Jd*, Dolores formation (Jurassic); *Kmc*, Mancos shale (Cretaceous); *Esm*, San Miguel conglomerate (Eocene); *Sj*, San Juan tuff; *Pr*, Potosi rhyolite; *gd*, gabbro-diorite; *Sj*, *Pr*, *gd*, Eocene and Miocene. (*After Whitman Cross, U. S. Geol. Survey.*)

The predominating values are in gold and silver, but in some districts much lead, copper, and zinc are also present, mainly as galena, tetrahedrite, and sphalerite. The gangue is mainly quartz, but rhodo-

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[1] W. S. Burbank, Revision of geologic structure and stratigraphy in the Ouray district, Colorado, and its bearing on ore deposition, *Proc.* Colorado Sci. Soc., 12, 1930, pp. 152–232.

chrosite, rhodonite, barite, and fluorite are often present. Propylitic alteration predominates near the veins.

**Telluride District.**[1]—In the Telluride district the annual production was for many years about $1,000,000 in gold and 1,500,000 ounces of silver, with some copper and lead, but has now declined. Strong veins have been worked by the Liberty Bell, Smuggler-Union, and Tomboy mines. All of these are now closed.[2] The veins are filled fissures, averaging 3 or 4 feet in width, with crustification-banding and drusy structure. Sericitization is marked close to the vein; beyond this there is extensive propylitization. The larger lodes appear often as a number of parallel plates of filled veins separated by sheets of altered rock. The Smuggler vein, for instance, cuts through the San Juan tuff, the andesite, and the rhyolite, a vertical distance of 2,300 feet (Fig. 195). Only in one place has a vein been followed into the underlying sedimentary beds. The vein continued strong but was poor. In the Smuggler mine the ore-shoot has been mined continuously for 8,000 feet, probably a record length, and through a vertical distance of 2,300 feet. The ore is contained only in the San Juan breccia; in the overlying rhyolite the vein is poor (Fig. 196).

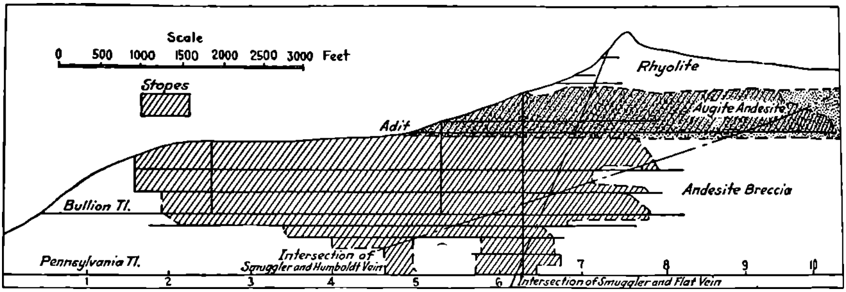

FIG. 196.—Longitudinal section of the Smuggler vein, Colorado. (*Courtesy Smuggler Union Mining Co., R. Livermore, General Manager.*)

The gangue minerals are quartz, calcite, siderite, rhodochrosite, adularia, barite, and fluorite, the last being abundant in the Tomboy

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[1] C. W. Purington, *Eighteenth Ann. Rept.*, U. S. Geol. Survey, 3, 1897, pp. 745–850.

C. W. Purington, Ore horizons in the San Juan Mts., *Econ. Geol.*, 1, 1905, pp. 129–133.

A. Winslow, The Liberty Bell Mine, *Trans.*, Am. Inst. Min. Eng., 29, 1900, pp. 285–307.

C. A. Chase, *idem*, 42, 1911, pp. 694–741.

M. E. Hurst, Rock alteration and ore deposition at Telluride, *Econ. Geol.*, 17, 1922, pp. 675–702.

C. N. Bell, Mining methods of the Telluride district, *Trans.*, Am. Inst. Min. Met. Eng., 72, 1925, pp. 550–565.

[2] Total productions: Smuggler-Union, $26,000,000; Tomboy, $26,000,000; Liberty Bell, $16,000,000.

mine (Fig. 197). Native gold, pyrite, galena, sphalerite, chalcopyrite, tetrahedrite, and polybasite are the principal metallic minerals. The ores contain 2 or 3 per cent sulphides and yield about $6 per ton. The treatment consists of a combination of amalgamation, concentration, and cyaniding.

**Silverton District.**[1]—The Silverton district is rich in veins, which contain more base metals than those of the Telluride district. The annual production has varied greatly. In 1929, in San Juan County, the production had a value of nearly $4,000,000 in lead, zinc, gold, silver, and copper. In large part this came from the Sunnyside mine, which, since 1912, has been working a low-grade vein in which the paragenesis is quartz, pyrite, sphalerite, galena, chalcopyrite, and rhodochrosite.[2] The deposits are simple veins or lodes averaging about 3 feet in thickness. The structure is commonly massive but sometimes banded (Fig. 198)

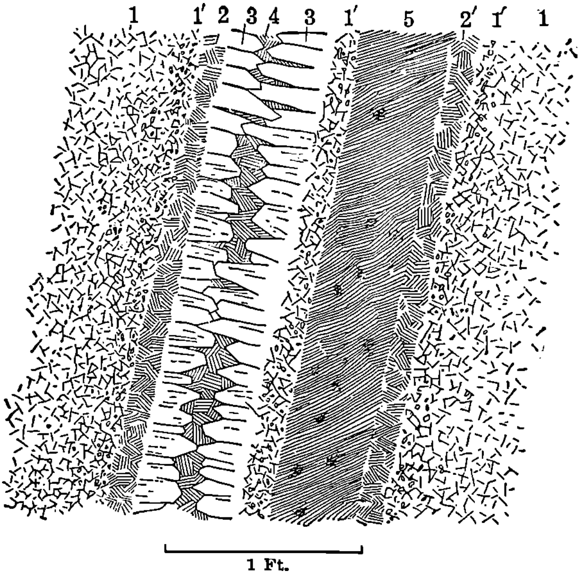

Fig. 197.—Section showing succession of ore minerals in Mendota workings, Smuggler vein, Telluride, Colorado. 1, country rock; 1′, sericitized and impregnated country rock; 2, sphalerite with calcite; 2′, sphalerite with galena; 3, white quartz; 4, rhodochrosite; 5, blue quartz with finely disseminated sulphides. (*After C. W. Purington, U. S. Geol. Survey.*)

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[1] F. L. Ransome, Economic geology of the Silverton quadrangle, *Bull.* 182, U. S. Geol. Survey, 1901. C. D. Hulin, *Econ. Geol.*, 24, 1929, p. 32.

C. W. Purington, The Camp Bird mine, *Trans.*, Am. Inst. Min. Eng., 33, 1904, pp. 499–528.

J. E. Spurr, The Camp Bird compound vein dike, *Econ. Geol.*, 20, 1925, pp. 115–152. H. C. Boydell, Discussion, Camp Bird mine, *idem*, 21, 1926, pp. 185–190.

[2] High temperature manganese minerals, like tephroite, helvite, and alleghanyite, are said to appear at a depth of 2,200 feet.

and drusy; on the whole it resembles that of deeper-seated veins. Many of the veins are rich in sulphides.

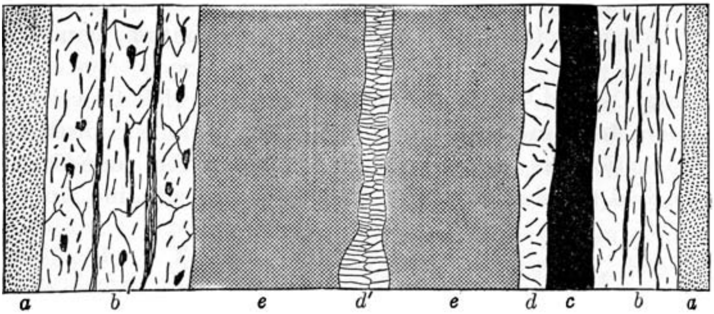

Fig. 198.—Cross-section of banded vein near London shaft, Silverton, Colorado. *a*, country rock; *b*, quartz and chalcopyrite; *c*, tetrahedrite; *d*, *d'*, quartz; *e*, galena. Vein 6 inches wide. (*After F. L. Ransome, U. S. Geol. Survey.*)

The prevailing gangue mineral is quartz and this is of coarser-grained texture than is common in the veins deposited near the surface (Fig. 199).

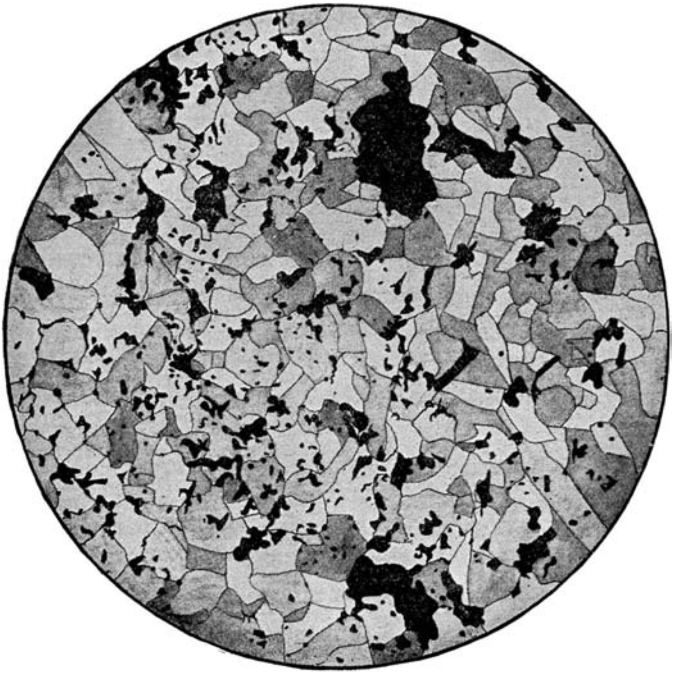

Fig. 199.—Thin section of ore from Ridgeway mine, Silverton, Colorado. Large black areas, pyrite; small black areas, argentite, with a little galena and sphalerite; shaded grains, quartz. Magnified 17 diameters. (*After F. L. Ransome, U. S. Geol. Survey.*)

The gangue also includes much calcite, dolomite, rhodochrosite, rhodonite, barite, and fluorite. The ore minerals are pyrite, galena, chalcopyrite, sphalerite, tetrahedrite, enargite, argentite, and native gold, more rarely hübnerite, molybdenite, and various sulphantimonides and bismuthides. Tetrahedrite and galena are very abundant. Tellurides are rare. Rock alteration in this district is discussed in some detail on pages 456 to 457.

The Camp Bird lode probably represents the easterly continuation of one of the Telluride lodes. The Camp Bird mine yielded $27,000,000 from 1903 to 1916, principally in gold, though some silver, copper and lead were recovered. The ore contained about $22 in gold per ton. Since 1916, lessees have been extracting ore from the upper levels. The lode intersects San Juan tuff and andesite and is described as a sheeted zone 4 or 5 feet thick made up of alternating fissure filling and altered rock; filling was, however, the predominating process. The gangue is quartz, often beautifully crustified, with rhodochrosite, calcite, and fluorite. The metallic minerals are very fine native gold, with a few per cent of galena, pyrite, and sphalerite, also some finely distributed tellurides.

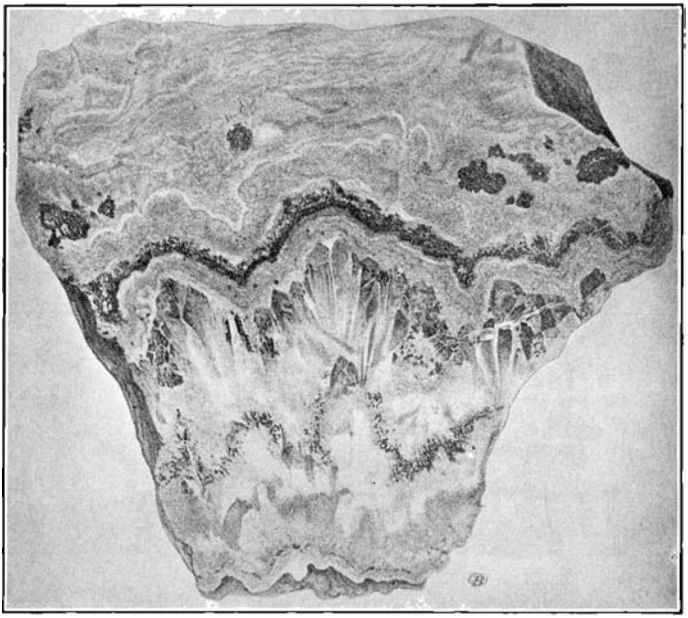

FIG. 200.—Rich ore from Camp Bird mine, Colorado, showing colloform deposition. At base chalcedony and carbonates surrounded by quartz crystals, in which is contained a dark curve containing native gold. Above quartz crystals six distinct layers of cryptocrystalline quartz, calcite, and fluorite. Above this comes conspicuous dark banding containing free gold. Above this colloform deposition of quartz, fluorite, and rhodochrosite. (*After F. L. Ransome, U. S. Geol. Survey.*)

Spurr distinguishes justly between an early and poor pyrite-galena-sphalerite stage, in part with specularite, and a later mineralization when the rich rhodochrosite ore was introduced. Ransome and Purington show the crustified and colloform deposition (Fig. 200) of this later ore for which Spurr's explanation by an injection of "ore magma" which solidified in place would seem entirely untenable. The extent and dependence of the rich shoot on the enclosing rocks are excellently shown in Fig. 201.

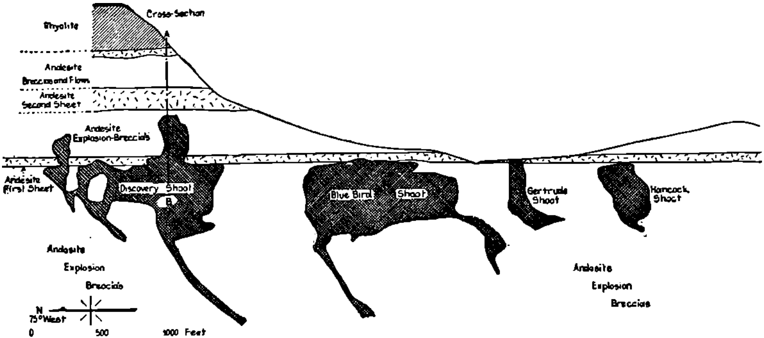

FIG. 201.—Projection of stopes on plane of Camp Bird vein. (*After J. E. Spurr, "Economic Geology."*)

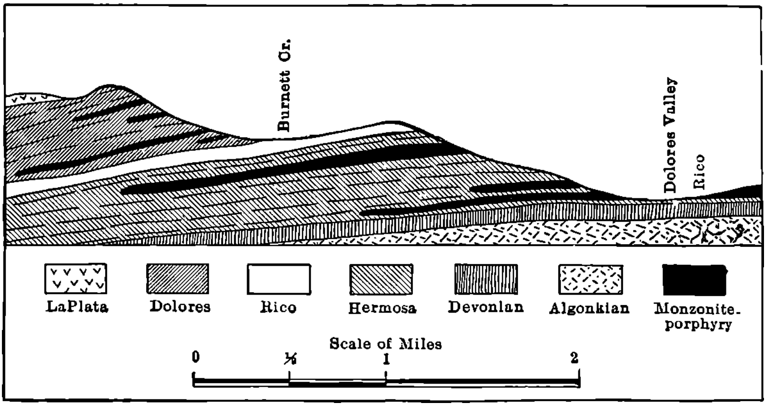

FIG. 202.—Geological section through part of the Rico dome, Colorado. (*After Cross and Spencer, U. S. Geol. Survey.*)

**Rico District.**[1]—The Rico Mountains are a domelike uplift of sedimentary rocks ranging from Algonkian to Jurassic in age, intruded by stocks, sheets, and sills of monzonite, or monzonite porphyry (Fig. 202). The ores occupy minor fissures while the structural faults are barren of

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[1] T. A. Rickard, The Enterprise mine, *Trans.*, Am. Inst. Min. Eng. 26, 1897, pp. 906–980.

F. L. Ransome, Ore deposits of the Rico Mountains, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, 1901, pp. 229–300.

G. E. Collins, Localization of orebodies at Rico and Red Mountains, Colorado, *Proc.*, Colorado Sci. Soc., 12, 1931, pp. 407–424.

metals. Burbank connects the veins with the Eocene intrusions which antedate the volcanic flows. The deposits form lodes, bed-veins (blankets), and replacements. The blankets often lie parallel to the sheets of intruded rocks or below impervious shales (Fig. 85). The abundant ore minerals consist of pyrite, galena, sphalerite, silver-bearing tennantite and tetrahedrite, with later, but probably hypogene, argentite and pearceite in a gangue of quartz, rhodochrosite, calcite, and fluorite. The filling is often beautifully banded (Fig. 203). A very limited vertical range is characteristic of the deposits. The greater part of the production has come from the blankets, a short distance below which the veins have become impoverished in the Hermosa (Pennsylvanian) formation of sandstone, shale, and limestone. Ascending warm waters of the sodium carbonate type and exhalations of $CO_2$ are common.

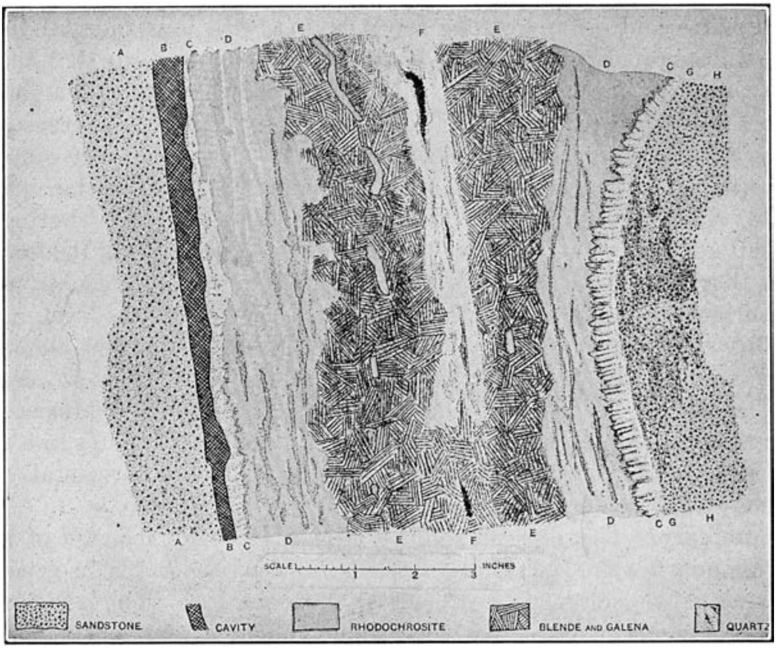

FIG. 203.—Banded ore, Rico, Colorado. (*After T. A. Rickard.*)

**Lake City District.**[1]—The deposits in the Lake City district range from simple veins to more complicated lodes, but filling with banding was the predominating process in their development. The width ranges from a few inches to 20 feet. Many of the fissures are short and were found to pinch out at relatively slight depth. The alteration of the wall rock is slight but is marked by some silicification and sericitization.

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[1] J. D. Irving and H. Bancroft, Geology and ore deposits near Lake City, Colorado, *Bull.* 478, U. S. Geol. Survey, 1911.

Pyritization of the surrounding rhyolite or andesite extends to considerable distances. The veins intersect all rocks below the Potosi series and also cut the intrusions of monzonite porphyry. Silver predominates with quartz, rhodochrosite, and barite as gangue. The following types of veins are recognized: (1) Tetrahedrite-rhodochrosite veins, (2) quartz-galena-sphalerite veins, and (3) telluride veins.

Similar telluride veins are found in the La Plata, Durango, and Needle Mountain quadrangles to the southwest of Lake City.

**Creede District.**[1]—The ores appear in a thick volcanic complex mainly of rhyolites and occupy strong fault fissures with large normal throw. The veins are markedly banded and contain some pyrite but mostly argentiferous galena, sphalerite, with a little chalcopyrite, and argentite. The ores are oxidized and enriched in part. In 1910, they averaged 0.09 ounce of gold and 12.29 ounces of silver per ton, with 6.5 per cent lead and 1.9 per cent zinc. The gangue minerals are amethystine quartz, barite, rhodochrosite, also an unusual amount of chlorite (thuringite). The adjacent rhyolite is somewhat silicified and sericitized; it also contains thuringite. The ore zone is horizontal and did not generally reach the surface; it apparently does not extend below the tunnel level.

The total output since 1890 includes about 45,000,000 ounces of silver and 100,000 tons of lead. There is little production at present.

The silver-gold veins of the Mogollon[2] district, New Mexico, are somewhat similar in that they occupy important fault fissures in a complex series of Miocene flows. The ore-shoots are mainly horizontal. The ore shows fine colloform texture.

**Summary.**—The deposits of the San Juan region consist of gold-bearing quartz veins, gold-telluride veins, and base-metal veins generally carrying galena, tennantite, argentite, and pearceite. The descriptions show a merging of the types and certain minerals common to most of them, such as rhodochrosite, and barite. The ores are often banded and crustified.

The ores are clearly independent of the character of the country rock, as has been already noted by J. D. Irving. They also intersect the intrusive monzonites, where these are exposed by erosion. They formed both in the lavas and in the cool sedimentary rocks (Rico). Colloidal deposition is indicated in part. Limestone when adjacent was not silicated but silicified.

Spurr believes that fissuring and mineralization were caused by and followed the dome-like uplifts and that these were due to deep-seated intrusions, not yet exposed by erosion. The metals in the deposits would then have been derived from the ascending emanations of these deep-

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[1] W. H. Emmons and E. S. Larsen, *Bull.* 718, U. S. Geol. Survey, 1923.
E. S. Larsen, *Bull.* 811, *idem*, 1930, pp. 89–112.

[2] H. G. Ferguson, Bull. 787, *idem*, 1927.

seated intrusives. The region shows, as few others do, the relation of the upper and deeper vein zones.

At Rico, Ouray, and likewise at other places, W. S. Burbank recognizes two periods of ore deposition, both polymetallic. The earlier was related to laccolithic intrusions (late Cretaceous or early Tertiary) and yielded lead-zinc veins and associated replacement deposits with high-grade silver ores, low-grade lead-zinc replacement deposits and pyritic gold deposits. The later period yielded gold and silver veins with base metals and was related to the Tertiary volcanic activity (see footnote, p. 501).

**Casapalca, Peru.**[1]—The epithermal silver veins of Casapalca are very similar to some of the San Juan deposits. They intersect Tertiary sediments and covering andesite flows. The gangue minerals are quartz, calcite, rhodochrosite, and rhodonite, also barite. The sulphides are, in order of deposition, pyrite, sphalerite, galena, tetrahedrite, and bournonite; there is a still later development of carbonates and quartz, with realgar and stibnite.

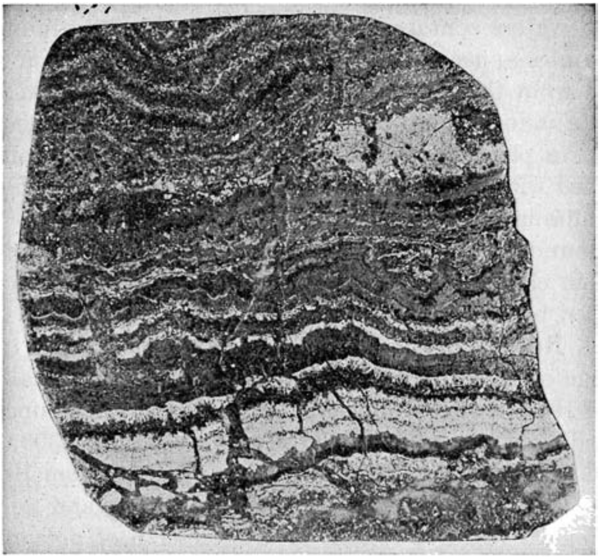

FIG. 204.—Polished section of crustified ore from Parral, Mexico. Dark gray, sphalerite; white, pyrite and marcasite. Probably colloform texture. Magnified 2 diameters. (*After Robert K. Doten.*)

**Parral, Mexico.**[2]—There are many epithermal veins in Mexico which contain abundant sulphides. One of the best examples is found at Parral

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[1] H. E. McKinstry and J. A. Noble, *Econ. Geol.* 27, 1932, pp. 501–522.

J. Orcel, *Bull.* Soc. minéralog. de France, 51, 1928, pp. 213–255.

[2] Harrison Schmitt, *Trans.*, Am. Inst. Min. Met. Eng., General Volume, 1931, pp. 268–290.

R. K. Doten, M. S. Thesis, Massachusetts Institute of Technology, 1929.

where veins carrying quartz, fluorite, pyrite, marcasite, sphalerite, galena and argentite intersect Jurassic rocks. Figure 204 shows excellent crustification with frequent repetition of pyrite, marcasite, and sphalerite, strongly suggesting colloidal deposition.

## GOLD-ALUNITE DEPOSITS

**General Features.**—In volcanic regions it is not uncommon to find considerable areas of bleached and altered lavas which contain more or less alunite ($K_2O.3Al_2O_3.4SO_3.6H_2O$), an earthy or compact, rarely coarsely crystalline mineral of inconspicuous appearance (p. 341).[1] Occasionally diaspore or gibbsite is associated with alunite. In most cases this basic sulphate, which is insoluble in water, is probably formed by the action of waters containing free sulphuric acid on aluminous rocks. Pyrite sometimes appears as a primary constituent, its iron being probably derived from the ferromagnesian silicates of the rock. Less commonly alunite is found in altered wall rocks of mesothermal veins. It is also found in places in the oxidized zones of veins containing pyrite, here associated with alunogen, jarosite, halotrichite, and other sulphates of iron and aluminum.

Rich gold and silver ores are rarely associated with areas of alunitization. Of such character is, however, the remarkable bonanza of Goldfield, described by F. L. Ransome.[2]

**Goldfield, Nevada.**—The Goldfield district, discovered in 1902, lies in a low range of desert hills in western Nevada. The total production to the end of 1917 was $80,700,000 in gold and 1,250,000 ounces of silver. In 1911, the district yielded gold and silver valued at $10,300,000; in 1917 only $2,000,000; in 1929, $269,018. Since then, the small production has been derived from leasing operations, and dumps and tailings.

The geological features consist of a succession of volcanic flows, of which 15 members are recognized, resting on a basement of granite and Cambrian shale. The age of these lava flows and intercalated lacustrine beds probably ranges from Eocene to the latest Pliocene. On the basement rest several flows of rhyolite; then come heavy masses of andesite or latite. In the andesite is intruded a thick sheet of dacite in which most of the ore-bodies are found. The andesite is overlain by 1,000 feet of

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[1] Whitman Cross, Geology of Silver Cliff and the Rosita Hills, *Seventeenth Ann. Report*, U. S. Geol. Survey, pt. 2, 1896, pp. 263–403.

E. S. Larsen, Alunite in the San Cristobal quadrangle, Colorado, *Bull.* 530, U. S. Geol. Survey, 1913, pp. 179–183.

R. T. Hill, Camp Alunite, Nevada, *Eng. and Min. Jour.*, 86, 1908, pp. 1203–1206.

[2] *Prof. Paper* 66, U. S. Geol. Survey, 1909.

*Econ. Geol.*, 5, 1910, pp. 301–311, 438–470.

Alunitized areas have been described from Vancouver island, in Jurassic and Triassic lavas. These altered rocks contain a little gold. See C. H. Clapp, *Econ. Geol.*, 10, 1915, pp. 70–88.

lacustrine beds, and still later than these are thin flows of basalt and rhyolite. The dacite sheet, almost 700 feet thick, consists of a rock of intermediate composition with glassy groundmass. Silica amounts to 60 per cent, and the potash about equals the soda. The pre-lacustrine rocks are considered Eocene. The lacustrine conditions ended by deformation and a low dome-like uplift resulted.

The principal producing area is small; most of the deposits are in the dacite, though a few are contained in rhyolite, andesite, or latite. The deposits are probably younger than the lake beds but older than the latest rhyolite and basalt; the age is considered late Miocene or early Pliocene. They were thus formed in intervolcanic time, like so many others of the ore deposits connected with effusive rocks. The deposits are contained in an area of highly altered rock and are irregular, silicified and fractured rock masses to which the term "ledges" is applied.

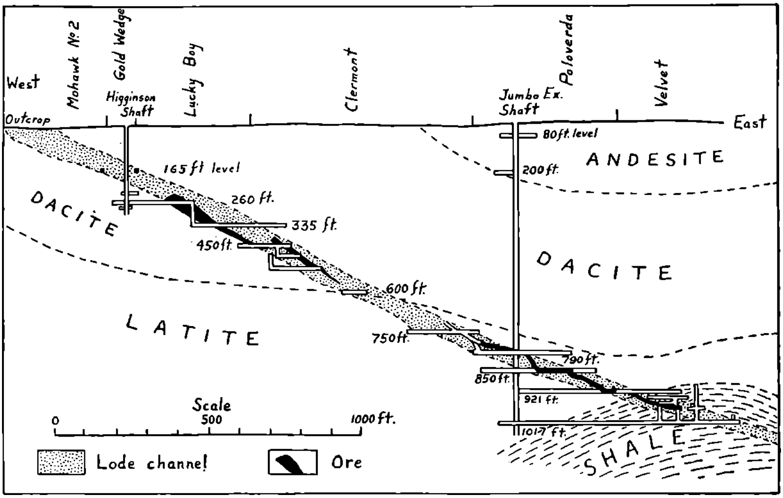

FIG. 205.—Vertical section of the Mohawk-Jumbo lode, Goldfield, Nevada. (*After J. K. Turner.*)

The ore-shoots form irregular bodies in the ledges and have no visible limits other than those indicated by the assay, but the ore is usually marked by more distinct brecciation and seaming than the surrounding ledge rock. During the earlier years little of the ore contained less than $30 per ton, but in 1916 ores yielding only $8 per ton were treated. Some shoots are roughly equidimensional masses; others are tabular.

The developments on the lower levels[1] have shown that at a depth of 1,000 feet the deposits pass into the Cambrian shale, as indeed predicted by Ransome, and that the general shape of the "lode channel" is flat,

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[1] J. K. Turner, *Min. and Sci. Press*, June 26, 1915.

with a dip of about 30°, and that it follows, in general, the contact planes between andesite and dacite, between dacite and latite, and between the latite and the shale, continuing for a greater distance along the shale contact (Fig. 205). Ores of varying, generally low, tenor are said to occur on this contact and contain alunite, much pyrite, chalcopyrite, and famatinite but with little free gold and tellurides. Explorations have been continued in the southern part of the district to the shale contact at 1,750 feet and later to 2,400 feet, but to judge from the production no important ore-bodies have been found.

The four minerals characteristic of this type of alteration are quartz, kaolinite, alunite, and pyrite. The altered dacite contains approximately 49 per cent quartz, 24 per cent kaolinite, 16 per cent alunite, and 7 per cent pyrite; and the replacement has been attended by losses of much CaO, MgO, and $Na_2O$, and of some $SiO_2$, $Al_2O_3$, and $K_2O$. Water, sulphuric acid, and sulphur have been added. Under the assumption that no change in volume has taken place, the altered rock has lost 5.3 per cent in weight and now has a porosity of 10 per cent.

While the shoots of the upper levels are not large compared to those of many other mines, the ore has been extremely rich. Much of it has averaged $419 per ton, containing 20 ounces of gold and 3 ounces of silver to the ton. One shipment of 14½ tons is said to have brought $45,783.

The unoxidized ore contains fine-grained pyrite and marcasite, bismuthinite, goldfieldite ($5Cu_2S.(Sb, Bi, As)_2(S, Te)_3$), an arsenical famatinite ($Cu_3SbS_4$), and native gold, also tellurides,[1] all in a dark-gray flinty quartz gangue. Concentric shells of ore minerals, with much finely divided yellowish-brown native gold, about greatly altered fragments of rock are rather characteristic of the rich ore. Galena and sphalerite are rare at Goldfield.

An analysis of rich ore from the Mohawk mine showed 2 per cent gold (541 ounces per ton), 0.25 silver, 2.42 tellurium, 0.35 bismuth, 2.08 copper, 3.83 iron, trace of zinc, 66.30 silica, 9.09 alumina, and 6.30 sulphur.

While the deposition of pyrite, alunite, and kaolinite proceeded during the whole epoch of ledge formation, the richest ores were deposited somewhat later in brecciated and shattered parts. The water level stands 100 to 150 feet below the surface; above this the ores are oxidized and contain some gypsum but do not differ greatly in tenor from the fresh ores below water level.

These remarkable deposits are believed to have been formed by ascending alkaline waters, containing hydrogen sulphide, which derived their load of rare metals from deep-seated magmas. Through oxidation near the surface sulphuric acid was generated from the hydrogen sulphide

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[1] F. L. Ransome, *op. cit.*, pp. 115–116.

W. J. Sharwood, Gold tellurides, *Min. and Sci. Press*, 94, 1907, p. 731.

and this acid attacked the rocks, causing the alunitic alteration.[1] The sulphuric acid, descending mingled with surface waters, acted as a precipitant for the gold solution; and this combination of ascending alkaline and descending acid waters has, according to Ransome, probably resulted in the development of this unusual type.

The surface at the time of ore deposition was probably only a few hundred feet above the present surface.

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[1] The facts shown indicate hydrothermal deposition by waters ascending on fissures after the effusion of the lavas. They indicate rapid precipitation of the gold near the surface and probably a very deep source of the rising solutions. Whether or not the sulphates were derived from the oxidation of hydrogen sulphide is a debatable question. See C. F. Tolman and J. W. Ambrose, The rich ores of Goldfield, *Nevada Econ. Geol.* 29 (4), 1934, pp. 255–279.

# CHAPTER XXV

# DEPOSITS OF NATIVE COPPER

## GENERAL STATEMENT

Deposits of native copper, with subordinate chalcocite and bornite, are often found in flows of basic rocks, particularly in basalts; or in near-surface intrusives. They are associated with epidote, calcite, prehnite, datolite, quartz, chlorite, and sometimes adularia. Minerals of the zeolite group, such as laumontite, analcite, natrolite, stilbite, and apophyllite are usually present, sometimes in large amounts. These gangue minerals, together with the copper minerals, fill vacuoles or blowholes in the basic rocks or replace the rock. Pyrite and sulphides of metals other than copper rarely occur.

The deposits have been formed under hydrothermal conditions. The mineral association indicates that the deposition proceeded at a temperature lower than 250° C. They should probably be classed with the epithermal ores although differing markedly from the types described in the last chapter.

Instances of native copper occurring in this manner are plentiful. Among the numerous localities the following may be mentioned: The Faeroer,[1] north of Scotland; Stirling,[2] in Scotland; Oberstein a. d. Nahe, Germany; São Paulo,[3] Brazil; the Kristiania region,[4] Norway; the Triassic "traps" of New Jersey[5] and Connecticut; New Guinea;[6] the Transbaikalian provinces[7] on the Dochida River; the Bay of Fundy,[8] Nova Scotia; Novaja Semlja;[9] the Komodor Islands, west of Kamtschatka.[10] Few of these occurrences are of economic importance.

[1] F. Cornu, *Zeitschr. prakt. Geol.*, 15, 1907, p. 321.

[2] Carl Hintze, Handbuch der Mineralogie, 1898.

[3] E. Hussak, *Centralblatt f. Min.*, 1906, p. 333. (No zeolites; copper between the peripheral covering of the amygdules, consisting of an iron silicate and the filling of chalcedony.)

[4] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 7, 1899, p. 12.

[5] Volney Lewis, *Ann. Rept.*, Geol. Survey New Jersey, 1907, pp. 157 and 165.

[6] R. Beck, Lehre von den Erzlagerstätten, 1, 1909, p. 345.

[7] *Idem*, p. 346; also *Zeitschr. prakt. Geol.*, 9, 1901, p. 391. (With opal, chalcedony, calcite, epidote, and prehnite.)

[8] R. W. Ells, Mineral Resourcs Canada, Copper in the provinces of Nova Scotia, New Brunswick, and Quebec, Canada Geol. Survey, 1904, 58 pp.

[9] P. W. Voit, *Zeitschr. prakt. Geol.*, 21, 1913, p. 42.

[10] J. Morozevicz, *Mém.* 72, com. géol., St. Petersburg, 1912; *Neues Jahrbuch*, 1, 1914, p. 250.

(Fresh basalt contains 0.04 to 0.07 per cent Cu said to be present in solid solution in magnetite, with calcite and zeolites.)

A pre-Cambrian series of basaltic rocks in the Lake Superior region contains the most prominent example of this class of deposits in the world, and will be described in more detail below.

In eastern Oregon,[1] about 20 miles east of Baker City, and along the Snake River cañon, opposite the Seven Devils Mountains in Idaho, are extensive areas covered by a basaltic amygdaloid flow of Permian or Mesozoic age. This rock contains native copper and chalcocite, sparsely disseminated or along obscure fracture zones, in association with epidote, chlorite, calcite, and zeolites. The ores are of low grade and have not yet been worked with profit.

Another occurrence of interest, in the White River region in Alaska, has been described by Adolph Knopf.[2] The basaltic amygdaloids probably in part submarine, with tuffs and breccias, are interbedded with sediments of Carboniferous age. Placer copper is common in the creeks, and some large masses have been found. The minerals are chalcocite, chalcopyrite, and native metal in stringers and seams, with prehnite, laumontite, thomsonite, and calcite; also native copper with zeolites filling blowholes in reddish, highly amygdaloidal lava.

These statements will serve to show that these deposits in basaltic rocks represent a type of world-wide distribution.

In some cases ascending hot waters wholly or in part of deep-seated magmatic origin are the probable agency of deposition, but in other occurrences of native copper in amygdaloid flows it would seem that the agencies were local and not necessarily derived from great depths.

The peculiar feature of this group is the constant association with hydrous calcium or calcium-aluminum silicates or calcium-iron silicates. The zeolites, that is, the hydrous alkaline silicates, appear in the later stages of the mineralization. Only in the few deposits which are found in sandstones or conglomerates are the calcium silicates lacking. Another curious feature is the practical absence of pyrite and the great scarcity of other sulphides. The ores are always low in sulphur. It would seem probable that the calcium is derived from the rocks themselves for all basalts are rich in that metal.

## THE ASSOCIATED MINERALS

Calcite and epidote are among the most common associated minerals. Prehnite ($H_2Ca_2Al_2Si_3O_{12}$) and datolite [($H_2Ca_2B_2Si_2O_{16}$) 21.8 per cent $B_2O_3$] likewise commonly occur with the copper; both minerals are generally foreign to ore deposits; they are considered to be allied to the zeolites. Their temperature of formation is probably low; datolite has been synthesized at 400° C.

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[1] W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 551–776.

[2] *Econ. Geol.*, 5, 1910, pp. 247–256.

The only magnesium minerals noted are chlorites of several varieties, in the altered rock or in blow-holes.

The zeolites formed towards the end of the mineralizing process and probably at low temperatures, perhaps as low as 100°.[1] They are mainly calcium-aluminum silicates with 8 to 15 per cent of water of hydration. Sodium or potassium may replace part of the calcium. Analcite is a sodium-aluminum silicate with 8 per cent of water. Other zeolites are natrolite, stilbite, chabazite, apophyllite (with fluorine), thomsonite, and laumontite. Besides occurring in amygdaloid lavas, they appear as very late products in some rare epithermal veins, contact-metamorphic deposits, and pegmatite dikes. A few have also been found in the orifices of hot springs. The zeolites seem to require quiet, stagnant conditions. They are generally absent in veins which seem to have been formed by rapidly ascending waters.

That zeolitization is far from being simply an effect of the leaching by surface waters is shown by the absence of the zeolites from large areas of basic flows, many of them full of vacuoles or blow-holes. There are, therefore, certain conditions[2] not yet fully elucidated—which are necessary for the deposition of zeolites. It is probable that their development would be greatly furthered if the eruption of the effusive rock took place under water; the sea water would cool the surface of the flow and a slow downward movement would be caused in the porous rock. Besides, these conditions would give rise to a system, cool at one end, hot at the other, in which circulation competent to effect concentration would be initiated.

C. N. Fenner finds that the zeolites of certain Triassic basalts of New Jersey, which cover land sediments and old playas or shallow desert basins of the same age, were formed mainly where the basalt flows covered the shallow lakes; he concludes that the circulation originated from the waters of these lakes. The general process, he says, was that of a slow cooling of the igneous rock, through which aqueous solutions were percolating. Material for solution was contributed by the basalt and by the previously evolved sublimates. The character of the minerals changed during the cooling. Pyrite and chalcopyrite are among the metallic minerals; native copper is not mentioned, but occurs at many places in

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[1] They may indeed form at considerable lower temperatures as shown by the occurrence of analcite and apophyllite in sedimentary rocks and of phillipsite in deep-sea deposits. See W. H. Bradley, The occurrence of analcite . . . in the Green River formation, *Prof. Paper* 158, U. S. Geol. Survey, 1930, pp. 1–8.

[2] J. Volney Lewis, *Ann. Rept.*, State Geologist, New Jersey Geol. Survey, 1907, p. 167.

Alfred Harker, The natural history of igneous rocks, 1909, p. 308.

C. N. Fenner, The Watchung basalt and the paragenesis of its zeolites, *Ann.*, New York Acad. Sci., 20, pt. 2, 1910, pp. 97–187.

these Triassic flows. Three periods of crystallization are distinguished. Beginning with the oldest they are as follows:

| | | |
|---|---|---|
| 1. Boric acid period | (a) | Albite, quartz, garnet, amphibole, specularite, sulphides. |
| | (b) | Datolite, prehnite, pectolite, amphibole, specularite, sulphides. |
| 2. Zeolite period | | Analcite, chabazite, heulandite, stilbite, natrolite, laumontite, apophyllite, amphibole, chlorite, specularite, sulphides. |
| 3. Calcite period | | Thaumasite, calcite, gypsum, amphibole, chlorite, specularite, sulphides. |

This association shows a peculiar combination of high-temperature minerals like garnet and amphibole with the zeolitic deposits. Extensive replacements were noted, similar to processes described long ago by Pumpelly from observations in the Lake Superior copper mines. Minerals stable under new conditions replace those formed in older crystallizations. Datolite, prehnite, pectolite, chabazite, stilbite, natrolite, apophyllite, and calcite—all replace the older albite. Quartz is replaced by calcite and various zeolites. Datolite is replaced by zeolites.

A similar series of events from the close of the magmatic period to the deposition of the zeolites was outlined by Earl Shannon,[1] who found the following sequence: albite, chlorite, amphibole, epidote, axinite, quartz, prehnite, datolite, chabazite, stilbite, laumonite, opal, apophyllite, calcite. Chalcopyrite, sphalerite, and galena were noted in diopside, prehnite, and apophyllite.

### SOURCE OF COPPER

Basic igneous rocks, such as gabbro, diabase, and basalt, probably always contain copper, in some cases as much as 0.2 per cent but more commonly about 0.02 to 0.04 per cent. According to the latest investigations based on polished sections, the copper is mainly present as chalcopyrite, bornite, or chalcocite. According to L. C. Graton the copper may also occur in native form. The suggestion that the whole or a part of the copper in the deposits was extracted by waters from the rock itself almost forces itself on the observer, and, although a number of investigators uphold the idea of ascending deep-seated waters of magmatic origin as a carrier of the copper, the suggestion noted above can not as yet be discarded.

### THE LAKE SUPERIOR COPPER DEPOSITS

In the following list of references on the copper deposits of the Lake Superior region only the more important works of the extensive literature are mentioned:

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[1] The mineralogy and petrology of intrusive Triassic diabase at Goose Creek, Virginia, *Proc.*, U. S. Nat. Mus., 66, Oct. 2, 1924, pp. 1–86.

R. Pumpelly, The metasomatic development of the copper-bearing rocks of Lake Superior, *Proc.*, Am. Acad. Arts Sci., 13, 1877–1878, p. 253.

C. Rominger, Geol. Survey Michigan, 5, 1895.

R. D. Irving, The copper-bearing rocks of Lake Superior, *Mon.* 5, U. S. Geol. Survey, 1883.

M. E. Wadsworth, Origin and mode of occurrence of the Lake Superior copper deposits, *Trans.*, Am. Inst. Min. Eng., 27, 1898, pp. 669–696.

L. Hubbard, Keweenaw Point, Geol. Survey Michigan, 6, pt. 2, 1898.

A. C. Lane, Geological report on Isle Royale, Geol. Survey Michigan, 6, pt. 1, 1898.

A. C. Lane, The theory of copper deposition, *Ann. Rept.*, Geol. Survey Michigan, 1903.

A. C. Lane, Native copper deposits, *Bull.* 13, Can. Min. Inst., February, 1911, pp. 81–87.

A. C. Lane, The Keweenaw series of Michigan, Geol. and Biol. Survey Michigan, Lansing, 1911.

T. A. Rickard, The copper mines of Lake Superior, 1905, p. 164. (Excellent brief description of geology and technology.)

L. C. Graton, *Mineral Resources*, U. S. Geol. Survey, Annual publication, 1906–1907.

C. R. Van Hise and C. K. Leith, *Mon.* 52, *idem*, 1911, pp. 578–592.

B. S. Butler, W. S. Burbank *et al.*, The copper deposits of Michigan, *Prof. Paper* 144, U. S. Geol. Survey, 1929.

T. M. Broderick, Zoning in Michigan copper deposits, *Econ. Geol.*, 24, 1929, pp. 149–162, 311–326. Fissure veins and lode relations, *idem*, 26, 1931, pp. 840–856.

**General Occurrence.**—The great deposits of native copper in the Keweenawan volcanic flows and conglomerates of the pre-Cambrian in Michigan form one of the principal items of the copper wealth of the United States. These deposits are mainly concentrated in Houghton County, on the Keweenaw peninsula of northwestern Michigan, on the southern shore of Lake Superior (Fig. 114). The copper belt continues, however, in a northeasterly direction to the point of the peninsula and southward into Ontonagon County and into northern Wisconsin and eastern Minnesota, having a total length of about 300 miles. The cupriferous formation is found also on Isle Royale and at Michipicoten in Ontario, Canada, opposite Keweenaw Point. The present productive belt extends in a northeasterly direction from the Lake mine, 43 miles southwest of Houghton, the center of the industry, to the Cliff mine, 22 miles northeast of that place.

The so-called Keweenawan series, the uppermost part of the pre-Cambrian, unconformably covers the Huronian, which in turn rests discordantly on the Archean greenstones and gneisses.

The Keweenawan series forms a huge synclinorium, bounding and underlying the western part of Lake Superior. The upper part of the series consists of 10,000 or 12,000 feet of red arkosic sandstones, shales, and conglomerates; the lower part consists of a vast accumulation of basaltic lavas, perhaps 25,000 feet in total thickness. On the Keweenaw peninsula the whole series strikes northeast parallel to the direction of the peninsula, and dips northwest from 30° to 75°. The sandstones follow the northwest coast. The rapidly alternating series of compact diabases or basalts (traps) and amygdaloid beds occupy the central belt. On the southeast the Keweenawan series is cut off by a long reverse fault parallel to the strike and dipping northwest. The southeast coast of the peninsula is underlain by horizontal non-productive Cambrian sandstones. Embedded in the volcanic flows are a few strata of sandstone, shale, or conglomerate. Quartzose porphyries (felsites) erupted in the same epoch enter conspicuously into the composition of these conglomerates (Figs. 206 and 207).

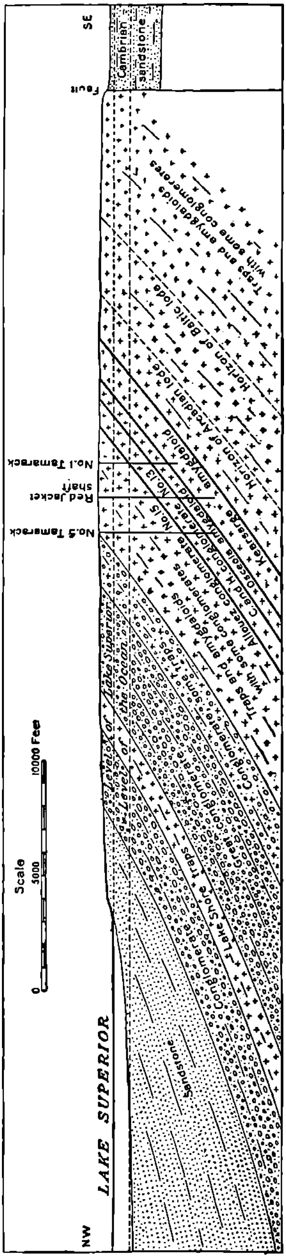

FIG. 206.—Section across the copper-bearing beds at Houghton, Michigan. (*After A. C. Lane.*)

Copper in visible grains is widely distributed through the amygdaloid rocks, but is of economic importance in only a few places. The deposits which now yield the larger part of the production are beds of amygdaloid rocks of great persistency. Almost as much copper is also mined from a felsitic or rhyolitic conglomerate (the Calumet conglomerate). Veins cutting across the strike of the beds were mined in the early years but are now of minor importance.

**The Calumet Conglomerate.**—This bed is worked mainly in the Calumet & Hecla and Tamarack mines; in the former mine it is opened for a distance of about 18,000 feet along the strike, and to a maximum depth of 9,300 feet down the dip. There are 10 inclined shafts and one vertical shaft, with an aggregate of about 200 miles of workings. The Red Jacket vertical shaft reaches to the fifty-seventh level of the incline. The Tamarack mine, in which the continuation of the shoot in depth has been found, reaches the conglomerate by four vertical shafts from 3,409 to 5,300 feet in depth, the latter being one of the deepest shafts in the world. The deepest workings, on the ninety-sixth level, are at 5,700 feet.

Only parts of the conglomerate are of profitable grade, the ore-shoot trending north on the bed. In depth it appears leaner and less regular in tenor. Along the surface only parts of the bed contain commercial ore; it is poor both northeast and southwest of the Calumet shoot. Also, in the upper levels the thickness of the conglomerate is decidedly less than in depth. The tenor of the ore at present has decreased from 4 per cent of copper in former years to 1.5 per cent. The apparent decrease is to some extent due to improvements in mining and milling that allow the handling of lower grades of ore.

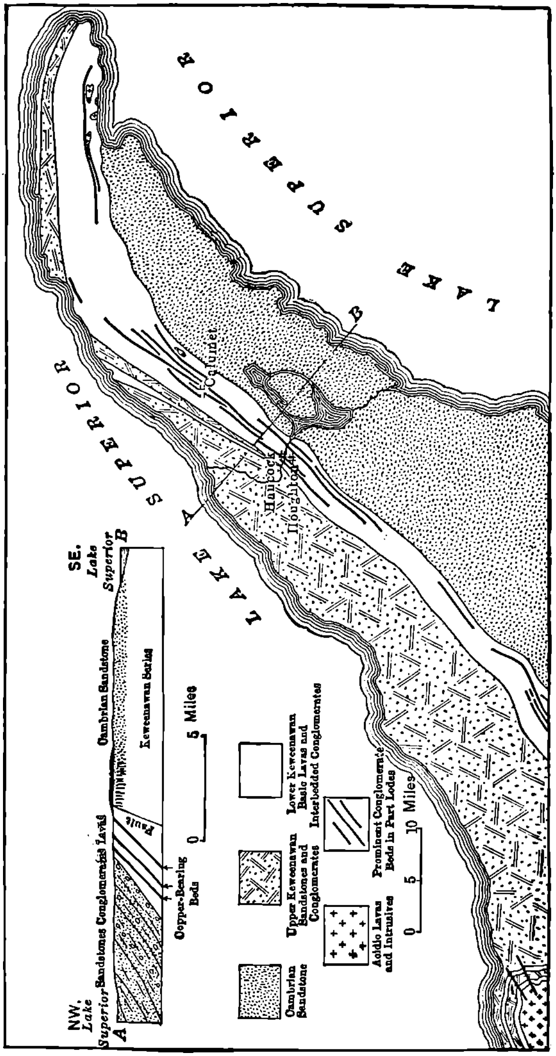

FIG. 207.—The Michigan copper region. (*After Seaman and Lane.*)

The conglomerate is 5 to 20 feet thick, dips 36° to 39° northwest, and forms a compact reddish-brown rock easily breaking across the pebbles. The copper occurs mainly as small particles in the cement between the cobbles, which are well rounded and consist of quartz porphyry with some basic igneous rocks. The hanging wall is a dark, fine-grained diabase, the footwall a thin layer of sandstone. About one-third of the copper production of Michigan is obtained from this conglomerate.

**The Amygdaloids.**—The amygdaloid copper-bearing beds, which occur at seven principal horizons, are named the Baltic, Kearsarge, Pewabic, Osceola, Isle Royale, Atlantic, and Winona amygdaloids and are worked by a dozen large mines. These beds are vesicular basalts, usually brownish in color, with earthy fracture and filled with amygdules of calcite, epidote, and zeolites (Fig. 208). The copper occurs in these, but also replaces the rock itself. Some native silver, bordering copper with sharp contacts,[1] occurs in the amygdaloids; scarcely any is found in the conglomerate. The flows are naturally more vesicular in the upper part than in the bottom part. Both width and distribution of copper are irregular. The Osceola bed is worked by the Calumet & Hecla, Tamarack, and Osceola mines; in the Osceola it is developed to a depth of 4,500 feet on the incline. In the Calumet & Hecla mine this amygdaloid is 30 to 35 feet thick, but the mineralization is mainly con-

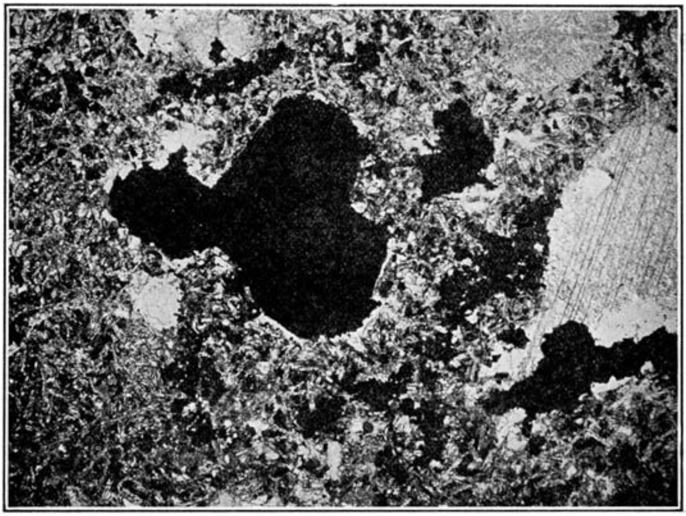

FIG. 208.—Amygdaloidal basalt, Houghton, Michigan. Black areas, native copper; larger areas represent fillings of blow holes, with calcite at right margin. Smaller black areas represent replacement of igneous rock by copper. Magnified 15 diameters.

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[1] H. C. N. Carpenter, and M. S. Fisher, *Bull.* 331, Inst. Min. and Met. (London), April, 1932. "Sharp boundaries between Cu and Ag would not be expected if the temperature had exceeded 200° C."

fined to a strip 8 or 10 feet thick along the hanging wall and also a streak along the footwall. In the Baltic mine the amygdaloid bed is from 15 to 80 feet in stoping width; this deposit has produced much coarse copper and contains veinlets of chalcocite, bornite, and copper arsenides. The Kearsarge amygdaloid is worked continuously for a distance of 12 miles. The Quincy mine has reached a vertical depth of 5,300 feet.

The amygdaloid ores yield an average recovery of 0.88 per cent copper, actually varying from 0.5 to 2.1 per cent (Champion mine, 1925). In 1916, the maximum output of silver, 716,640 ounces, was obtained, in small part as nuggets or "pickings," but mainly from the electrolytic refining process, to which a part of the copper produced was subjected. In the last 10 years the silver output has declined sharply. The largest lump of silver on record from the district weighed 12 pounds and was found in the Mass mine, in the southern part of the district.[1]

**The Veins.**—A third mode of occurrence of the copper ore is as veins following fracture zones, in the northern part of the peninsula and in Ontonagon County to the south. During the early years of the industry these veins yielded much copper, but are at present of little importance. Most of the veins cross the bedding and stand at a steep angle, though in Ontonagon County many strike veins are also found. In places they are also parallel to the dip of the strata. Some of them could be followed by the drifts for a distance of 2,000 or 3,000 feet. According to Credner, Pumpelly, and Irving, many of them were wide, though they averaged only 3 feet.

In part these veins were formed by filling, but they were chiefly the result of metasomatic replacement. Much of the native copper was coarse; some masses of unusual size were found, the most famous being that encountered in the Minnesota vein in 1880. The mass weighed 500 tons, was 46 feet long, 18.6 feet wide, and 8.5 feet thick.[2] At the Cliff mine many masses which weighed from 40 to 100 tons were discovered. From the vein the copper seems to have had a tendency to extend into the various amygdaloid flows. Most of the veins became impoverished at a depth of a few hundred feet.

The amygdaloid beds are cut by many minor cross fractures and slip faults, but according to the accounts these contain little or no copper.

**Mineral Association.**—The minerals occurring in the basaltic flows are plagioclase, pyroxene, olivine, and magnetite; also there is more or less glass. Immediately after consolidation a certain number of new (deuteric) minerals were formed by late magmatic processes: They were hematite (principally in the vesicular tops), serpentine, and chlorite. The glass was decomposed by magmatic gases, and some feldspar, pyroxene, and hematite were set free. Then began the ore-forming period,

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[1] T. A. Rickard, *op. cit.*, see p. 518.

[2] T. A. Rickard, *op. cit.*

the temperatures decreasing slowly. In the amygdaloid lodes the paragenesis of the ore-forming minerals was, according to C. Palache and A. Wandke, as follows: (Most abundant minerals italicized).

*Orthoclase, chlorite* (delessite), *epidote, pumpellyite;* later, prehnite, *copper,* datolite, silver, ankerite, *quartz,* sericite, *calcite* (over long period in middle of sequence), arsenides, adularia, chlorite, *laumontite,* analcite and other zeolites, barite, and anhydrite (the last two are scarce).

Pumpellyite ($6CaO.3Al_2O_3.7SiO_2.4H_2O$) is a bluish green, orthorhombic mineral allied to zoisite, often occurring in needles in the amygdules. The copper arsenides are whitneyite, algodonite, and domeykite in order of increasing arsenic. Under the microscope they prove to be complex intergrowths. Datolite, often enclosing copper, occurs as crystals and as dense botryoidal or colloform masses. According to Palache, the zeolites are not closely associated with the copper (although in part they certainly overlap). Hydrocarbons are found in the Nonsuch sandstone, Ontonagon County, associated with native copper, as shown by K. Nishio.[1]

The stages in alteration and filling were first studied by R. Pumpelly and later by A. C. Lane. The former distinguished as first stage the individualization of epidote, prehnite, and datolite and laumontite; then followed quartz, calcite, and copper, and, finally, came the alkaline silicates. Prehnite replaces other minerals extensively.

The paragenesis of the veins is very similar to that in the lodes. In the conglomerate the paragenesis is much simpler. Prehnite, datolite, and zeolites are absent. The succession is orthoclase (early), epidote, copper, silver, quartz, and calcite.

Discussing the evidence of zoning, Broderick finds that epidote, quartz, calcite, chlorite, and copper are persistent minerals. In the stratigraphically deeper parts of the lodes a higher temperature association of minerals tends to occur, as ankerite, sericite, occasional sulphides, and more arsenic; while the upper parts are characterized by prehnite, datolite, adularia, and zeolites. Only 25 per cent of the copper has come from the zeolite-prehnite-datolite zone. Broderick also finds that the ratio between arsenic and copper tends to increase in depth and thinks it probable that the mineralization will continue to far greater depths than yet attained. On the whole, the rock alteration is of the propylitic type. The veins are considered to have been fed by the solutions rising along the lodes.

**Origin.**—Various views have been expressed to explain the genesis of the Lake Superior copper deposits. A full account of the older theories may be found in *Professional Paper* 144, U. S. Geol. Survey. Whitney, Pumpelly, and Wadsworth believed that the copper was leached from the copper beds and carried down by surface waters. H. L. Smyth[2] thought

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[1] *Econ. Geol.,* 14, 1919, pp. 324–334.

[2] *Science,* 3, 1896, p. 251.

that the deposits were formed by ascending thermal waters, and Wright, Van Hise, Leith, and Steidtman have since implied or asserted a similar origin. A. C. Lane[1] assumed that after the submarine effusion of the lavas, sea water penetrated the beds decomposing the silicates and converting a part of the iron and all of the copper to chlorides. The reduction of the cuprous chloride was effected by calcium carbonate or silicate (calcite, datolite, or prehnite), with the formation of ferric oxide and calcium chloride. This reaction was confirmed by experiments of G. Fernekes.[2] That the deposits antedate the Cambrian was shown by U. S. Grant.[3] They were in existence when the Cambrian strata along the great fault sank to the level of the Keweenawan Series. The upper, doubtless oxidized, parts of the lodes have been swept away by the Quaternary ice sheet so that native copper now outcrops almost at the surface.

The authors of *Professional Paper* 144, U. S. Geol. Survey, have shown satisfactorily that the structure is earlier than the ores; the flows and the sandstones were practically in their present position before ore deposition began. Also that ascending thermal waters rose along the "tops" of the basaltic flows or along the likewise open felsitic conglomerates, and deposited copper underneath the impermeable barrier of the next overlying flow. They assume—which is not quite so evident—that the waters were of magmatic origin and reducing character, and that they contained copper and sulphur, likewise of magmatic origin. Further, that the hematite in the vesicular tops oxidized the sulphur to sulphates which were carried away as iron, calcium, and barium salts. A little of the latter two sulphates occur now in the mines. In Professional Paper 144 R. C. Wells explains this by the following equation, supported by experiments:

$$Cu_2S + Fe_2O_3 + 5H_2SO_4 = 2Cu + 6FeSO_4 + 5H_2O$$

However, it is to be noted that Wells's equations necessitate acidic solutions. The red tops are bleached, $Fe_2O_3$ being converted to FeO. The authors also assume that an underlying body of gabbro was the source of the emanations, this source being located at different depth for the various lodes.

The peculiar differences in mineralization between the amygdaloids and the conglomerate indicate definitely (1) that both kinds of lodes were open ducts, and (2) that the whole suite of calcium silicates (except epidote) were obtained from the basaltic rock during the ascent of the solutions.

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[1] Salt water in the Lake mines, *Proc.*, Lake Superior Min. Inst., 12, 1906; Native copper deposits, *Jour.* Can. Min. Inst., 14, 1911, pp. 316–325.

[2] *Econ. Geol.*, 2, 1907, pp. 580–584.

[3] *Bull.* 6, Wisconsin Geol. and Nat. Hist. Survey, 1901.

The theory outlined above still presents a number of difficulties: (1) The enormous amount of magmatic water required. It will no doubt be necessary to assume a very liberal mixture of meteoric water. Basaltic magmas are not rich in water. Their gases, as far as we know, carry some chlorine and much sulphur. (2) The chemistry of the process of copper deposition is not yet in satisfactory shape. (3) No fissure veins with similar mineralization and oxidation of sulphur are known from other parts of the world. (4) If strongly active hot waters ascended the lodes, it is not reasonable to assume that they could traverse distances of 20,000 to 40,000 feet in basaltic rocks containing from 0.01 to 0.03 per cent of copper without dissolving (and depositing) a large amount of this metal. The concentration required to produce ore would be only 100 times the original content of the rock. The presence of chlorides in the depositing waters can not be denied, and this would (in part) carry us back to the Lane and Fernekes equation.

**Mine Waters.**[1]—The present condition of the underground waters in the copper region is most interesting. Lane has shown that the water in the upper levels is soft and potable and has the normal composition of surface waters. It decreases in quantity as depth is gained and ceases at a depth of 1,000 to 1,500 feet below the surface. The salinity is about 90 parts per million, carbonates of calcium and magnesium predominating.

As depth is attained, chlorine and calcium increase very materially, but mine water is less abundant. At a depth of 3,000 to 5,000 feet, the mine waters are almost entirely absent; they constitute feeble drips and may collect in small quantities in the sumps. They are strong solutions of calcium chloride with bromine and other substances.

ANALYSIS OF MINE WATER, QUINCY MINE

From drippings on 55th level north of No. 6 shaft. G. Fernekes, analyst. Depth about 4,000 feet

Grams per liter

| | | | |
|---|---|---|---|
| Cl | 176.027 | $SiO_2$ | 0.020 |
| Br | 2.200 | $(Fe, Al)_2O_3$ | 0.010 |
| Ca | 86.478 | Mn | 0.004 |
| Na | 15.188 | Cu | 0.016 |
| K | 0.411 | Ni | 0.006 |
| $SO_4$ | 0.110 | Mg | 0.020 |
| | | | 280.490 |

Total solids determined 280.500.

Traces of boron and strontium. No barium, lithium, or carbon dioxide.

Some of these waters contain an appreciable amount of zinc and some strontium.

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[1] A. C. Lane, Salt water in the Lake Mines, *Proc.*, Lake Superior Min. Inst., 12, 1906, pp. 154–163.

In the most concentrated waters 99 per cent of the salts consist of the chlorides of calcium and sodium, and three-fourths of the remainder is sodium bromide.

These waters may possibly be of connate origin, that is, residual water from ancient seas. Van Hise and Leith believe that the deep mine water may represent the residuum of the ore-forming solutions but do not consider it as residual sea water.

**Mining and Smelting Operations.**—In the copper mines of Lake Superior mining operations are conducted on a large scale. The total amount of copper ore (including tailings) treated in Michigan in 1930 was about 6,600,000 tons. This is crushed with steam stamps, each one having a daily capacity of 500 to 700 tons; wet concentration is used with jigs, tables, etc., the resulting concentrates (locally called "mineral") amounting, in 1930, to about 129,000 tons, of an average copper content of 65.7 per cent, making an average yield of 1.27 per cent.

This concentrate of native copper is smelted and refined by a single operation in reverberatory furnaces, the smelting works being located in the district and at Buffalo. A part of the copper is electrolytically refined in order to eliminate the small amount of arsenic contained. The annual copper production of Michigan increased steadily to 1905 when it reached 230,000,000 pounds. In 1930, the mine output had decreased to 169,400,000 pounds. The reserves of amygdaloid copper-bearing rock are of great extent. The silver production in 1930 was 8,000 ounces, while in 1925 it had reached 140,000 ounces. The total production of the district from 1846 to the close of 1929 was 4,117,171 short tons of copper, of which about 1,800,000 tons have come from the Calumet conglomerate.

## THE COPPER DEPOSIT OF MONTE CATINI

The celebrated copper deposit of Monte Catini on the western coast of Italy, near Livorno (Leghorn) and the ancient Etruscan city of Volterra, has been described by many authors. A. Bergeat has given an excellent review of this literature in connection with his own observations.[1] Another detailed description is given by L. de Launay.[2] The mines have been worked to a depth of 850 feet.

Irregular laccolithic stocks of diabase, with some gabbro, break through Eocene marly limestones and siliceous shales; near the contacts, the igneous rock is in part glassy, so that the intrusions clearly took place near the surface. The ore occurs exclusively in the diabase, particularly in its lower part at or near the contact, but also reaches the surface. In the ore-body the diabase is crushed to reddish clayey masses seamed with zeolites and calcite. The ores contain native copper, in crevices and

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[1] A. W. Stelzner and A. Bergeat, Die Erzlagerstätten, 2, 1906, pp. 835–842.

[2] Métallogénie de l'Italie, *Compte Rendu*, 10th Internat. Geol. Congress, 1, 1907, pp. 603–621.

druses, with calcite, prehnite, datolite, analcite, and laumontite; also sulphides, especially chalcocite, bornite, and chalcopyrite, sometimes massive, but partly in large and small, rounded concretions surrounded by clayey, crushed rock and consisting of the several sulphides in concentric intergrowth. The tenor and distribution of the ores are very irregular.

The whole aspect of this unique deposit seems to indicate that the copper was concentrated from the diabase shortly after its consolidation and the crushing which followed. It is more difficult to point to the source of the concentrating waters, but it is probably safe to say that the present ground waters have had nothing to do with the formation of the ore.

There are remarkable similarities between the mineral association at Monte Catini and that of the amygdaloid flows of the Lake Superior region; and the processes of concentration may, in the main, have been identical.

### THE COPPER DEPOSITS OF CORO-CORO, BOLIVIA

The well-known and long-worked copper deposits of Coro-Coro,[1] in Bolivia, are contained in a series of sandstones of probably Pliocene age. Exact genesis is still in dispute. There are several cupriferous beds with disseminated native copper and much gypsum, also some native silver, domeykite, and chalcocite. The copper-bearing beds are much lighter in color than the prevailing deep-red sandstones.

The deposits are worked at the present time. One of the mines has a depth of 1,300 feet with native copper in beds 2 meters thick, the average being 2 per cent copper.

Ten miles away, diorite porphyry intrudes the beds and the sandstones contain much tuffaceous material.

### NATIVE COPPER WITH EPIDOTE IN BASIC LAVAS (CATOCTIN TYPE)

In some copper deposits contained in basic lavas the zeolites are absent and the mineral association is mainly native copper, epidote, quartz, and calcite. Such occurrences, which are of slight economic importance, have been found in the Appalachian region in Virginia and Pennsylvania.[2]

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[1] Older literature, see: Stelzner and Bergeat, Die Lagerstätten, 1, 1904, p. 419.

G. Steinmann, Rosenbusch Festschrift, 1906, pp. 335–368.

Miller and Singewald, Mineral deposits of South America, 1917, pp. 89–94.

J. T. Singewald, Jr. and E. W. Berry, Studies in Geology, 1, John Hopkins Univ., 1922, pp. 117.

F. A. Sundt, *Eng. and Min. Jour.-Press*, Mar. 29, 1924.

J. T. Singewald, Jr., A genetic comparison of the Bolivian and Michigan copper deposits, *Econ. Geol.*, 23, 1928, pp. 55–61. On Coro-Coro, see also F. Ahlfeld, *Centr. Mineral.*, *Abt. A.*, 1932, pp. 373–382. He considers these deposits of supergene origin.

[2] W. H. Weed, Types of copper deposits in the southern United States, *Trans.*, Am. Inst. Min. Eng., 30, 1900, pp. 449–504.

W. C. Phalen, Copper deposits near Luray, Virginia, *Bull.* 285, U. S. Geol.

The rocks are basaltic flows of pre-Cambrian age, in part amygdaloid, in part schistose. They contain, in irregular fractures and along shear zones, abundant epidote, native copper, calcite, and chlorite; in places chalcopyrite and bornite occur in the gangue or in the rock itself. Weed named this group of ores the "Catoctin type" and suggested that it owed its origin to infiltration from the present surface. This seems improbable; more likely the copper was deposited by hot waters shortly after the eruption and consolidation of the basalt.

Watson believes that these deposits were concentrated from copper contained in the rocks; they are not confined in the tops of the flows; the latter contain no hematite. His argument is directed against the views recently expressed by Graton, Butler and Broderick in relation to the Michigan ores.

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Survey, 1906, pp. 140–143.

G. W. Stose, Copper deposits of South Mountain, Pennsylvania, *Bull.* 430, U. S. Geol. Survey, 1909, pp. 122–131.

T. L. Watson, Native copper deposits of the south Atlantic states, compared with those of Michigan, *Econ. Geol.*, 18, 1923, pp. 732–752.

# CHAPTER XXVI

## MESOTHERMAL DEPOSITS

### METALLIFEROUS DEPOSITS FORMED AT INTERMEDIATE TEMPERATURES BY ASCENDING THERMAL WATERS AND IN GENETIC CONNECTION WITH INTRUSIVE ROCKS

### GENERAL FEATURES

It is exceedingly common to find metalliferous deposits in or near bodies of intrusive rocks. These deep-seated rocks have been exposed by long-continued erosion, and in places it is possible to arrive at a good estimate of the thickness of rocks removed, especially where the total thickness of the sedimentary series in which the intrusion occurred is known. That the deposits are not of recent development, but were formed a short time after the intrusion, can in most cases be proved conclusively, and from this it follows that they were developed under a great thickness of covering rocks. Their depth below the actual surface at the time of mineralization may, roughly speaking, be considered as ranging from 4,000 to 12,000 feet. In most of these deposits the absence of high-temperature minerals, such as magnetite, garnet, pyroxene, or tourmaline, shows that a high degree of heat did not prevail at the time of genesis. The depth below the surface indicates that the normal rock temperatures would be from 50° to 125° C., but in all cases the vicinity of recently intruded rocks had forced the temperature curves nearer to the surface. The heated waters which deposited the ores either emanated from the intrusive magma or, at least, derived their high temperatures from it. It is manifestly impossible to give accurate figures, but reasoning from what is known of the stability of minerals characteristic of this class of deposits, we may say with some confidence that the actual temperatures may have ranged from 175° to 300° C. When the high-temperature curves were near the surface, these deposits may have originated at a depth of only a few thousand feet; when the intrusions were deeper seated, the depth at which the deposits were formed may have exceeded 12,000 feet.

The pressure was necessarily strong; as calculated on a hydrostatic basis it ranged from 140 to 400 atmospheres. Communication with the surface was probably established in many places; where it was lacking the water and gases, propelled from the magma, may have been under still higher pressure.

When the temperature exceeded the upper limit stated above, silicate minerals undoubtedly developed and the resulting deposit is of a different type. When the temperature fell below the lower limit stated, the general type of mineralization must have approached that of the epithermal deposits.

The structure of the deposits is what might be expected from the opening of fissures under pressure at considerable depth. The fissures are fairly regular in strike and dip, having neither the extreme brecciated structure common to deposits formed close to the surface nor the lenticular form and irregular openings of the deep-seated deposits. Smooth walls and slickensides are abundant. As the fissures were opened in the zone of fracture, open spaces are present in many deposits, though the walls usually come together within short intervals. In calcareous rocks, more rarely in igneous rocks or quartzite, replacement deposits were often developed; they are more common here than in the deposits formed close to the surface, though the solutions, on the whole, spread much less widely through the igneous rocks in this group than in the shallow deposits.

The principal metals contained are gold, silver, copper, lead, and zinc. In the deep-seated deposits molybdenum, bismuth, tungsten, and arsenic are not uncommon associates; we find the same metals here, though they are much less prominent; in addition there is also much antimony, and in places tellurium. The ore minerals are sulphides, arsenides, sulphantimonides, and sulpharsenides. Pyrite, chalcopyrite, arsenopyrite, galena, sphalerite, tetrahedrite, tennantite, and native gold are the most common, and on the whole there is not much variety and complexity. Oxides such as magnetite and specularite if present occur in small amounts. The metallic minerals develop both in the filling and in the altered country rock, but in the fissure veins proper it is common to find the valuable ores mainly in the filled spaces. The predominating gangue mineral is quartz, but carbonates are also common, such as calcite, dolomite, and ankerite, more rarely siderite; fluorite and barite are occasionally of importance; adularia is rare; albite occurs in some deposits. Evidences of colloidal deposition are found in many places, though far less common than in the epithermal deposits.

Among the minerals of this type of metallization are found no biotite; no pyroxenes or amphiboles; no garnet, tourmaline, or topaz; no zeolites or kaolin.

Frequently these veins follow lamprophyric dikes, which are usually the last manifestation of igneous activity.

This class yields a large proportion of the gold production of the world, as well as much of its silver, copper, lead, and zinc. Its deposits are by far the most abundant in the Cordilleran region, as well as in other parts of the world where intrusive activity has been followed by deep erosion.

The gold-quartz veins of California and of Victoria (Australia) and many of those of the Cordilleran region, the zinc-lead-silver replacement deposits at Leadville, Park City, and Aspen, the Coeur d'Alene lead veins, and many other types belong to this class.

The intrusive bodies may be laccoliths, stocks, or batholiths; the last term being reserved for intrusive cross-cutting masses of large size like those of the Helena-Butte region in Montana, the great Idaho granite mass, or the enormous intrusive masses following the Pacific coast from Lower California to Alaska. It has long been known that the interior parts of great batholiths are relatively barren and that mineral deposits rarely occur in laccoliths. The volatile constituents of the magma, from which the ore deposits have been formed, have a tendency to rise to the roof of the intrusive or to accumulate in the cupolas of its upper parts.[1]

B. S. Butler[2] points out that in Utah those stocks which are truncated by erosion near their apices or points nearest to the surface contain ore deposits of great importance while those which have been truncated by erosion to deeper levels show relatively few deposits. When the mobile constituents reached a point where the magma was sufficiently consolidated to fracture they were guided by the fissures and on reaching favorable physical and chemical environments began to deposit the metals in solution.

Billingsley and Grimes[3] have applied the same theory to the Boulder Batholith, Montana, and appear to have shown that the maximum deposition took place near the roof of that batholith.

Many veins of this class show interesting changes in depth and laterally, generally referred to as "zoning." In general, the ore is likely to grow more siliceous and pyritic in depth. Gold, copper, and zinc appear to have been deposited by relatively hot waters while lead and rich silver deposits as a rule appear farther away from the intrusive suggesting deposition at lower temperatures (p. 121).

A great vertical range of up to 5,000 feet, or even more, is characteristic of many types described under this heading. Their upper limit is usually an erosion plane of some kind, and they can hardly ever be traced up to the actual surface at the time of vein formation. Physiographic investigations sometimes give a clue to their relations to that surface. Many of them continue to the greatest depth reached in mining.

Regarding the temperature at the time of ore formation there are few reliable data. Usually the temperature was higher in the initial stage than in the last phases. When, for instance, rhombic chalcocite of hypo-

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[1] R. A. Daly, Igneous rocks and the depths of the earth, New York, 1933, pp. 123 and 313.

[2] B. S. Butler, Relation of ore deposits to different types of intrusive bodies, *Econ. Geol.*, 10, 1915, pp. 101–122.

[3] P. Billingsley and J. A. Grimes, The ore deposits of the Boulder Batholith, *Trans.*, Am. Inst. Min. Eng., 58, 1918, pp. 284–368.

gene origin is found, it may be concluded that the temperature at this closing stage was not above 90° C.

On the other hand, the quartz is of the alpha variety which shows that 575° C. was not reached. In limestone no lime-magnesia-iron silicates are formed; the point at which these begin to be stable is generally held to be about 400° to 500° C. The minerals common in these deposits have mostly great persistency; their field of stability is extensive, but all of the typical high-temperature minerals are lacking.

The epithermal veins seem to have been formed by liquids ascending from the deep magma basins where active differentiation into flow rocks took place. The mesothermal deposits are probably largely derived from differentiation in congealing batholithic masses,[1] which were pressed up from the magma basins into higher levels.

## METASOMATIC PROCESSES

**General Character.**—Sericitization, with or without carbonates but always with pyrite, is the principal process of alteration in igneous rocks. In limestones the alteration is usually confined to silicification. The alteration of the country rock is usually very intense next to the ore, but seldom yields coarsely crystalline products as in some high-temperature deposits (p. 645). In feldspathic and ferromagnesian rocks the principal product is sericite, the fine-grained foliated form of muscovite; in many deposits carbonates, such as calcite, dolomite, and ankerite, develop in large amounts. The dark minerals are first altered, their iron being usually recombined as pyrite. The feldspars are also rather easily altered; even quartz grains are attacked and partly, at least, converted to an aggregate of sericite and carbonates.

While the orthoclase and the soda-lime feldspars are conspicuously absent as vein-forming minerals, albite is not uncommon, especially in some gold-quartz veins. This mode of alteration is frequently observed in amphibolitic rocks, which contain much sodium and presumably much albite developed during previous dynamo-metamorphism. In many cases the new albite and dolomitic or ankeritic carbonates form together. Pyrite is a common metasomatic mineral and is often associated with the ferromagnesian minerals, but may also form in quartz and feldspars. Other metallic minerals are not common; the apatite and zircon of the igneous rocks resist alteration; while titanite and ilmenite yield rutile. In many vein types of the interior Cordilleran province the metasomatic carbonates are scarce or absent, as in the copper veins of Butte, Montana,

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[1] W. H. Emmons, Primary downward changes in ore deposits, *Trans.*, Am. Inst. Min. Met. Eng., 70, 1924, pp. 964–994; Relations of metalliferous systems to igneous intrusives, *idem*, 74, 1926, pp. 29–70; Relations of the disseminated copper ores in porphyry to igneous intrusives, *idem*, 75, 1927, pp. 797–815.

and Clifton, Arizona. Serpentine is sometimes altered to magnesite and dolomite. Hypogene alunite is found in places.

Among sedimentary rocks, quartzite and sandstone are little affected, except in veins of the Coeur d'Alene type, where the quartz grains are replaced by siderite. Clay slates always contain metasomatically developed pyrite in cubes; whether they are otherwise altered or not depends upon their composition: if they contain feldspathic sediment, sericitic and carbonatic alteration will ensue; if only kaolin, sericite, and quartz are present there will be little noticeable alteration, except in some instances where almost complete silicification takes place.

Limestone and other calcareous rocks are almost always subject to silicification by the replacement of the carbonates with fine-grained quartz aggregates; the resulting rocks are usually called "jasperoids" and look more or less like chert (Figs. 62 and 63). For this alteration the presence of $CO_2$ appears to be necessary.[1] Ore minerals develop abundantly by metasomatic action in such rocks.

Along with or preceding this silicification, dolomitization[2] often takes place; the solutions apparently abstract a part of the calcite in the limestone and replace it by magnesium carbonate. Limestone is sometimes converted to magnesite, siderite, manganosiderite, or fluorite.

The alteration is accompanied by strong leaching of sodium and by concentration of potassium. Where there is little carbonatization much calcium and magnesium are also leached. Aluminum in most cases remains about constant.

**Alteration of Wall Rocks Adjoining Gold-Quartz Veins.**—In veins characterized by quartz filling with free gold and simple sulphides or arsenopyrite, the country rock next to the walls is usually rich in carbonates, sericite, and pyrite, but rarely contains much gold. Extensive alteration zones are not common, and sometimes fresh rock adjoins the vein. The relative quantity of alteration products may differ considerably, even in the same mine.

Clay slate, with more or less carbonaceous matter, is thought by some to have a precipitating and enriching influence on the vein; but to what extent this is true is doubtful. While this influence can apparently be recognized in some districts, like Gympie, in Queensland, it is not clearly shown along the Mother Lode of California. The black clay slates near the veins often contain much pyrite, ankerite, sericite, and albite.[3]

Metasomatic rocks containing albite result from the alteration of amphibolites at Angels Camp, Calaveras County, California, where they

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[1] G. H. Cox, R. S. Dean, and V. H. Gottschalk, *Bull.* 3 (2), Rolla School of Mines, Missouri, 1916.

[2] D. F. Hewett, Dolomitization and ore deposition, *Econ. Geol.*, 23, 1928, pp. 827–853.

[3] A. Knopf, *Prof. Paper* 157, U. S. Geol. Survey, 1929, p. 42.

constitute low-grade ores. The Utica, Lightner, and Melones mines have been working on these deposits.[1]

These altered rocks consist of sericite with embedded grains of calcium-magnesium-iron carbonates and pyrite. In a very common type, much of the sodium, abundant in the amphibolite, has been retained as albite in the altered rock. Large grains of carbonates are separated by a granular mass of quartz, albite, pyrite, and sericite. Such altered rock is too poor in gold to pay for working alone, but is often milled for the sake of the rich stringers which intersect it.

Fig. 209.—Altered granodiorite, Bellefountain mine, Nevada City, California. *m*, fine aggregate of sericite with some calcite and quartz, replacing orthoclase and andesine; *b*, original biotite altered to sericite; *q*, original quartz; black, pyrite with included sericite. Magnified 15 diameters.

In the ordinary course of alteration the ferromagnesian minerals are first converted into larger foils of sericite. A chlorite rich in iron is also formed, which during a later stage is converted into sericite. The feldspars are then attacked, along cracks and cleavage planes, and a finely felted aggregate of sericite and calcite invades the grains until the replacement is complete. The granular texture may be preserved (Fig. 209). A texture often observed consists of interlacing sericite foils, the interstices of which are filled with calcite (Fig. 66). Orthoclase is almost always more resistant than the soda-lime feldspars.

The quartz is also attacked, but with more difficulty, and is in no case completely replaced. Magnetite appears to be converted to siderite and titanite to rutile. A part of the ferromagnesian minerals is transformed into pyrite. Sharp cubes of pyrite develop, however, not only in the

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[1] F. L. Ransome, Mother Lode Folio, *Geol. Atlas* 63, U. S. Geol. Survey, 1900. See also W. Lindgren, *Econ. Geol.*, 1, 1905, p. 543.

TABLE I.—ANALYSES OF METASOMATIC ROCKS FROM CALIFORNIA GOLD-QUARTZ VEINS

| | A | $A_1$ | B | $B_1$ | C | $C_1$ | D | $D_1$ |
|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 65.54 | 46.13 | 45.56 | 37.01 | 66.65 | 56.25 | 51.01 | 45.74 |
| $TiO_2$ | 0.39 | 0.67 | 1.11 | 0.85 | 0.38 | 0.25 | 0.98 | 0.36 |
| $Al_2O_3$ | 16.52 | 15.82 | 14.15 | 12.99 | 16.15 | 17.65 | 11.89 | 5.29 |
| $Fe_2O_3$ | 1.40 | 0.89 | 1.20 | 0.43 | 1.52 | 0.76 | 1.57 | 0.13 |
| FeO | 2.49 | 2.27 | 9.83 | 3.57 | 2.36 | 2.64 | 6.08 | 2.06 |
| $FeS_2$ | ........ | 1.61 | 7.86 | 7.99 | 0.02 | 2.87 | [1]1.73 | 0.49 |
| $Cu_2S$ | ........ | ..... | 0.10 | trace | ...... | ...... | trace | ...... |
| MnO | 0.06 | 0.09 | 0.25 | 0.24 | 0.10 | none | trace | 0.26 |
| NiO, ZnO | ........ | trace | trace | trace | ...... | ...... | ...... | ...... |
| CaO | 4.88 | 10.68 | 2.30 | 9.78 | 4.53 | 4.46 | 10.36 | 23.85 |
| SrO | not det. | trace | trace | trace | trace | ...... | none | none |
| BaO | not det. | trace | trace | trace | 0.07 | 0.03 | none | trace |
| MgO | 2.52 | 2.13 | 6.76 | 5.49 | 1.74 | 1.69 | 8.87 | 0.94 |
| $K_2O$ | 1.95 | 5.30 | 1.18 | 4.02 | 2.65 | 6.01 | 0.15 | 1.29 |
| $Na_2O$ | 4.09 | 0.17 | 1.57 | 0.13 | 3.40 | 0.30 | 4.17 | 0.11 |
| $Li_2O$ | ........ | trace | trace | trace | trace | ...... | ...... | trace |
| $H_2O-$ | ........ | 0.12 | 0.23 | 0.13 | 0.18 | 0.30 | 0.24 | 0.22 |
| $H_2O+$ | 0.59 | 2.42 | 4.84 | 1.92 | 0.72 | 2.36 | 2.09 | 1.07 |
| $P_2O_5$ | 0.18 | 0.10 | 0.14 | 0.06 | 0.10 | 0.21 | 0.17 | 0.07 |
| $SO_3$ | ........ | ..... | 0.03 | 0.04 | ...... | ...... | ...... | ...... |
| $CO_2$ | ........ | 11.24 | 3.04 | 15.04 | ...... | 4.82 | ...... | 18.91 |
| Total | 100.61 | 99.64 | 100.15 | 99.69 | 100.57 | 100.60 | 99.31 | 100.79 |

[1] Probably present as $Fe_7S_8$.

A. Fresh granodiorite, Lincoln, Placer County, California. Though not adjoining the vein, it indicates closely the actual composition of the fresh wall rock. W. F. Hillebrand, analyst.

$A_1$. Altered granodiorite, Plantz vein, Ophir, Placer County. W. F. Hillebrand, analyst.

B. Amphibolite schist, Conrad vein, Ophir, Placer County. Fairly fresh, but contains pyrite and calcite. W. F. Hillebrand, analyst.

$B_1$. Completely altered amphibolite schist, Mina Rica vein, Ophir, Placer County. W. F. Hillebrand, analyst.

C. Fresh granodiorite, Nevada City, Nevada County. W. F. Hillebrand, analyst.

$C_1$. Altered granodiorite, Bellefountain mine, Nevada City. George Steiger, analyst.

D. Fresh diabase, Grass Valley. H. N. Stokes, analyst.

$D_1$. Altered diabase, North Star mine, Grass Valley. W. F. Hillebrand, analyst.

sericitic aggregate, but also in the fresh feldspars or even in the quartz. Arsenopyrite is almost the only other sulphide which is enabled to form in the altered rock, and it develops in sharp rhombic crystals.

From many analyses the eight given on page 535 are selected.[1]

From the chemical and microscopical data the following compositions may be calculated:

TABLE II.—CALCULATED MINERALOGICAL COMPOSITION OF THE ALTERED ROCKS OF TABLE I

| | $A_1$ | $B_1$ | $C_1$ | $D_1$ |
|---|---|---|---|---|
| Quartz | 16.00 | 24.00 | 25.00 | 35.00 |
| Sericite (with a little chlorite) | 41.76 | 46.97 | 61.46 | 21.20 |
| $CaCO_3$ | 17.53 | 18.87 | 7.23 | 42.15 |
| $MgCO_3$ | 9.67 | 2.93 | 2.70 | 0.71 |
| $FeCO_3$ | 5.76 | 3.67 | 0.58 | ........ |
| $MnCO_3$ | 0.42 | 0.14 | .......... | ........ |
| Rutile | 0.85 | 0.67 | 0.25 | 0.36 |
| Pyrite | 7.99 | 1.61 | 2.87 | 0.50 |
| Apatite | 0.13 | 0.22 | 0.46 | 0.15 |
| Total | 100.11 | 99.08 | 100.55 | 100.07 |

As it seems probable that the alumina has remained fairly constant in the first six analyses in Table I, they may be directly compared for an approximate review of the chemical changes.

Analysis $D_1$ differs from the rest in showing an exceptionally high percentage of introduced lime and carbon dioxide and a corresponding loss of magnesia. Moreover, the alumina is so low that it must be supposed to have been removed.

The characteristic features of the process seem to consist in the decrease of silica, magnesia, and soda and increase of lime, potash, and carbon dioxide—the calcitic altered rock strongly contrasting with the quartz-filled veins. There is some evidence of partial leaching of titanium and phosphorus. It seems probable that, in most cases, the added material has more than balanced the losses. There has been a strong addition of calcium and potassium; and the vein-filling process probably began with deposition by solutions extremely rich in these constituents, as well as in carbon dioxide. The quartz filling sometimes shows imprints of calcite crystals along its walls, from which it may be concluded that during the process of filling the nature of the solutions changed to the later phase, in which almost nothing but quartz was deposited.

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[1] W. Lindgren, The gold-silver veins of Ophir, Cal., *Fourteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1894, pp. 243–284.

W. Lindgren, Metasomatic processes in fissure veins, *Trans.*, Am. Inst. Min. Eng., 30, 1900, p. 666.

A very similar type of alteration has been described from certain gold-quartz veins from British Columbia,[1] though here more sodium remains in the metasomatic rocks.

TABLE III.—ANALYSES OF FRESH AND ALTERED ROCKS FROM VEINS IN IDAHO AND UTAH
(Analysts, W. F. Hillebrand $E$–$F_1$; E. T. Allen, $G$–$G_1$.)

| | E | $E_1$ | F | $F_1$ | G | $G_1$ |
|---|---|---|---|---|---|---|
| $SiO_2$ | 65.23 | 66.66 | 57.78 | 58.01 | 58.64 | 56.78 |
| $TiO_2$ | 0.66 | 0.49 | 1.01 | 1.08 | 0.83 | 0.81 |
| $Al_2O_3$ | 16.94 | 14.26 | 16.28 | 15.72 | 15.35 | 16.90 |
| $Fe_2O_3$ | 1.60 | 0.67 | 1.02 | 0.64 | 3.25 | 6.87 |
| FeO | 1.91 | 1.33 | 4.92 | 3.87 | 2.54 | 2.34 |
| CoO, NiO | ........ | ........ | 0.02 | 0.12 | ........ | ........ |
| MnO | trace | trace | 0.15 | 0.17 | ........ | ........ |
| CaO | 3.85 | 3.37 | 6.65 | 2.15 | 5.37 | 1.18 |
| SrO | ........ | ........ | 0.07 | none | ........ | ........ |
| BaO | 0.19 | none | 0.12 | trace? | 0.18 | 0.14 |
| MgO | 1.31 | 0.95 | 4.60 | 2.07 | 3.84 | 0.03 |
| $K_2O$ | 3.02 | 4.19 | 2.22 | 4.79 | 4.23 | 7.02 |
| $Na_2O$ | 3.57 | none | 3.25 | 0.10 | 3.60 | 0.37 |
| $H_2O-$ | 0.18 | 0.36 | 0.34 | 0.31 | 0.86 | 1.32 |
| $H_2O+$ | 0.88 | 2.16 | 0.92 | 2.71 | 1.50 | 2.23 |
| $P_2O_5$ | 0.19 | 0.17 | 0.30 | 0.31 | 0.02 | 0.04 |
| $CO_2$ | 0.25 | 3.67 | 0.15 | 2.86 | none | 0.26 |
| S | none | 0.95 | 0.02 | 1.25 | 0.05 | 5.93 |
| Fe | ........ | 0.84 | ........ | 1.52 | ........ | ........ |
| Pb | ........ | ........ | ........ | 0.86 | ........ | ........ |
| Cu | ........ | ........ | ........ | 0.05 | ........ | ........ |
| As | ........ | ........ | ........ | 1.65 | ........ | ........ |
| Total | 99.78 | 100.07 | 99.82 | 100.24 | 100.26 | 102.22 |
| Less O | ........ | ........ | ........ | ........ | 0.02 | 2.22 |
| | | | | | 100.24 | 100.00 |

E. Fresh granitic rock immediately adjoining the Silver Wreath quartz vein, Willow Creek, Idaho.

$E_1$. Altered rock adjoining the same vein.

F. Fresh quartz-pyroxene diorite adjoining the Croesus vein, Hailey, Idaho.

$F_1$. Altered rock adjoining the same vein.

G. Fresh monzonite, British tunnel, Last Chance mine, Bingham, Utah.

$G_1$. Altered monzonite, same locality, at wall of lode.

In the veins of Idaho, Utah, and Colorado genetically connected with Cretaceous or early Tertiary intrusions of quartz monzonite or similar

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[1] W. S. McCann, The gold quartz veins of the Bridge River district, B. C., *Econ. Geol.*, 17, 1922, pp. 350–369.

porphyries, the carbonatization is far less marked and both calcium and magnesium are leached. The accompanying analyses, Table III, illustrate the chemical changes in two prominent types.

E and $E_1$ represent the fresh and altered rock from the Willow Creek district, Boise County, Idaho, where the narrow quartz veins carry scarcely any free gold, but much auriferous galena, pyrite, arsenopyrite, and sphalerite. F and $F_1$ represent the fresh and altered rock from the Croesus mine, Wood River district, Blaine County, Idaho, where the narrow streaks of filling consist of quartz, siderite, pyrrhotite, and chalcopyrite, with a little galena, arsenopyrite, and sphalerite. Here again only a fraction of the gold is in the free state. The ore contains very little silver.[1]

The specific gravity of E is 2.714. From the mineralogical composition given in the report quoted the specific gravity is calculated to 2.720, which is a close agreement, the difference possibly indicating a very slight porosity.[2]

TABLE IV.—MINERALOGICAL COMPOSITION OF $E_1$ AND $F_1$. IN TABLE III

| | $E_1$ | $F_1$ |
|---|---|---|
| Quartz | 42.00 | 36.18 |
| Sericite | 46.84 | 38.18 |
| Chlorite | ........ | 11.76 |
| $CaCO_3$ | 4.80 | 3.11 |
| $MgCO_3$ | 1.96 | 1.26 |
| $FeCO_3$ | 1.45 | 2.19 |
| Rutile | 0.49 | 1.08 |
| Apatite | ........ | 0.72 |
| Pyrite | 1.78 | 0.58 |
| Pyrrhotite | ........ | 0.15 |
| Sphalerite | ........ | trace |
| Galena | ........ | 0.99 |
| Chalcopyrite | ........ | 0.15 |
| Arsenopyrite | ........ | 3.58 |
| Total | 99.32 | 99.93 |

The measured specific gravity of $E_1$ is 2.774, indicating that the rock alters to denser minerals. The calculation of the same specific gravity

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[1] For full calculations and description of E and $E_1$ see W. Lindgren, *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, p. 640; for F and $F_1$ see W. Lindgren, *Twentieth Ann. Rept.*, *idem*, pt. 3, 1900, pp. 211–232.

[2] In this calculation the following figures for specific gravity are used: quartz, 2.65; sericite, 2.83; biotite, 3.00; oligoclase, 2.65; orthoclase, 2.56.

from Table IV gives 2.796, which shows a decided porosity of the altered rock. Under these circumstances, no evidence of pressure being noted, it may be assumed that not much change in volume has taken place. By multiplying the percentages of E and $E_1$ by 2.714 and 2.774, respectively, and comparing the results, the absolute gains and losses per cubic meter may be obtained.

In the same manner the measured specific gravities of F and $F_1$ are compared with the calculated specific gravities.[1]

By multiplying the percentages of F and $F_1$ by the measured specific gravities, and comparing these data, the absolute gains and losses are again obtained.

The calculation shows that during the alteration of E to $E_1$ 291 kilograms were added and 229 lost per cubic meter, the net gain being 62 kilograms. During the alteration of F to $F_1$ 416 kilograms were added and 333 lost per cubic meter, the net gain being 83 kilograms.

Similar changes resulted in the two rocks: a moderate addition of silica and a strong gain of potash; nearly complete loss of sodium, barium, and strontium; partial loss of alumina, magnesia, and lime, F, however, losing much more lime than E. In $E_1$ the amounts lost of $Fe_2O_3$ and FeO are almost completely converted into Fe (in $FeS_2$). In $F_1$ these losses are less and not sufficient to account for the gain of Fe; consequently iron must have been added. Phosphoric acid is constant, consistent with the fresh state of the apatite.

These figures give some idea of the intensity of the transfer of material; the net change of material is comparatively small.

Similar alteration in the veins of Bingham, Utah, has been described by Boutwell.[2] The Last Chance lode is from 1 to 14 feet wide and contains galena, sphalerite, pyrite, and some calcite. The alteration and bleaching extend about 1 or 2 feet into the country rock, which is a monzonite, consisting of orthoclase, plagioclase, augite, biotite, and hornblende. The alteration begins by chloritization and dissemination of pyrite, but the end product consists largely of sericite and pyrite. The analysis indicates an unusual and almost complete removal of magnesia and extensive leaching of sodium and calcium. There has evidently been an addition of potassium, as there is considerably more than is called for in the ordinary composition of sericite. As usual, $TiO_2$ remains constant, and the altered rock contains practically no carbonates.

One of the great mineral belts of Colorado extends in a northeasterly direction from Leadville to Boulder by way of Park, Clear Creek, and Gilpin counties (p. 593). It is characterized, as a whole, by an abundance of heavy sulphide ores, principally pyrite, sphalerite, and galena, with

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[1] *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, pp. 221–232.

[2] J. M. Boutwell, Bingham mining district, *Prof. Paper* 38, U. S. Geol. Survey, 1905, p. 178.

subordinate chalcopyrite and a notable content of gold and silver. The gangue is subordinate and consists of a little quartz and more or less sideritic carbonate. The ores appear in replacement deposits and veins. At Leadville, where the ores replace limestone at the contacts with intrusive porphyry, the alteration of the carbonate rock is remarkably slight. There may be locally a little pyrite in the limestone near the ore; at other places the limestone is recrystallized and contains much manganosiderite, or is silicified. The contact between ore and limestone is usually very sharp—indeed, in spite of the completeness of the replacement, practically unaltered limestone may lie next to the ore.

At Georgetown and Central City, studied by Spurr, the deposits are fissure veins, generally filled with massive sulphides; and, in feldspathic rocks, they are adjoined by an altered zone from a few inches to 20 feet or more in width. The alteration products are quartz, sericite, and a sideritic carbonate, with more or less pyrite. The siderite is derived from biotite and magnetite, and the gangue minerals from the adjoining rocks. Kaolin is considered to have resulted from alteration by descending waters during the processes of weathering and sulphide enrichment.

At Breckenridge, Colorado, the alteration has been studied in detail by Ransome[1] at the Wellington lode, 350 feet below the surface and below the zone of oxidation. The vein is here 5 feet wide with good walls and contains sphalerite and galena in a little gangue of siderite and barite with more or less included country rock. The alteration spreads 20 feet from the vein. The fresh rock is a dark-gray monzonite porphyry, the ground-mass of which consists of labradorite, orthoclase, biotite, and diopside. The altered rock is light gray, with disseminated particles of sulphides. The rock, while retaining a faint trace of its structure, is changed to a mass of carbonate, sericite, and quartz. Probably no change of volume accompanied the alteration.

By multiplying the figures of the percentage compositions by the specific gravities of the rock mass, the constituents per 100 cubic centimeters of fresh and altered rock are obtained. These figures compared give the gains and losses for each constituent during the alteration of 100 cubic centimeters of fresh rock, and from these may be calculated the gains and losses in percentage of the original mass of fresh rock, 276.3 grams in column I. These gains or losses in percentages may be applied directly by addition or subtraction to the figures of the chemical analysis of fresh rock, and this, as shown in column III, will express the nature of the change more clearly. There has been a notable loss of silica, calcium, and sodium; to a less degree potassium. The additions comprise carbon dioxide, sulphides, ferrous iron, and magnesium, which would hardly bear out Spurr's assertion that the siderite in the vein is derived

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[1] F. L. Ransome, The Breckenridge district, *Prof. Paper* 75, U. S. Geol. Survey, 1911, pp. 95–101.

from the adjoining country rock. As usual apatite remains unaltered, and ilmenite is converted to rutile. Some paragonite has probably developed besides the sericite, if indeed the rock does not contain albite.

TABLE SHOWING ALTERATION OF DIORITE PORPHYRY AT BRECKENRIDGE, COLORADO

| | I | II | III |
|---|---|---|---|
| $SiO_2$ | 57.35 | 46.62 | 49.48 |
| $TiO_2$ | 1.07 | 1.01 | 1.07 |
| $Al_2O_3$ | 16.29 | 12.66 | 13.44 |
| $Fe_2O_3$ | 3.15 | trace | 0.02 |
| FeO | 4.36 | 11.15 | 11.78 |
| MnO | 0.12 | 0.92 | 0.97 |
| CaO | 5.66 | 1.55 | 1.66 |
| BaO | 0.10 | none | ............ |
| SrO | 0.05 | none | ............ |
| MgO | 2.41 | 4.02 | 4.25 |
| $K_2O$ | 3.39 | 1.68 | 1.79 |
| $Na_2O$ | 4.50 | 1.35 | 1.45 |
| $H_2O-$ | 0.15 | 0.31 | 0.33 |
| $H_2O+$ | 0.70 | 3.41 | 3.60 |
| $CO_2$ | 0.46 | 11.48 | 12.11 |
| $P_2O_5$ | 0.70 | 0.50 | 0.53 |
| $FeS_2$ | 0.09 | 1.99 | 2.10 |
| ZnS | none | 0.97 | 1.02 |
| PbS | ............ | 0.52 | 0.55 |
| Total | 100.55 | 100.14 | 106.15 |
| Specific gravity: | | | |
| Mass | 2.763 | 2.930 | |
| Powder | 2.799 | 2.940 | |

I. Diorite porphyry, 25 feet from vein, Wellington mine.

II. Altered porphyry, close to vein, Wellington mine.

III. Composition of same volume of altered rock in percentage of original rock mass.

An approximate calculation shows the altered rock to be composed as follows:

APPROXIMATE MINERALOGICAL COMPOSITION OF ALTERED DIORITE PORPHYRY AT BRECKENRIDGE, COLORADO

| | | | |
|---|---|---|---|
| Sericite | 30.5 | Apatite | 1.3 |
| Quartz | 31.6 | Pyrite | 2.0 |
| Kaolinite | 2.8 | Sphalerite | 1.0 |
| Carbonate | 29.3 | Galena | 0.5 |
| Rutile | 1.0 | Total | 100.0 |

The carbonate consists of 63.9 per cent $FeCO_3$, 29.6 per cent $MgCO_3$, 5.2 per cent $MnCO_3$, and 1.3 per cent $CaCO_3$, all in isomorphous mixture (Fig. 210).

The majority of the large copper desposits of the West are genetically connected with intrusives and have been formed at intermediate depths. Butte, Clifton, and Ely may serve as examples. The hydrothermal alteration of the feldspathic rocks results universally in sericite, quartz, and pyrite with practically no carbonates. Omitting minor constituents, the composition of the altered rocks would average, in per cent, about 65.0, $SiO_2$; 16.0, $Al_2O_3$; 2.0, FeO; 0.5, $Fe_2O_3$; 1.0, MgO; 0.25, CaO; 0.5, $Na_2O$; 5.0, $K_2O$; and 5.0, $FeS_2$.[1]

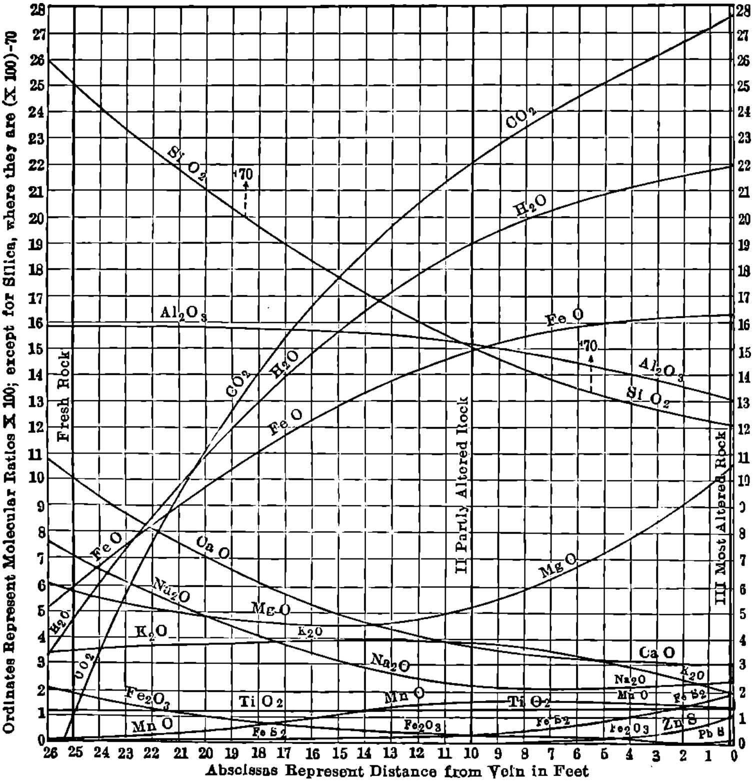

FIG. 210.—Diagram showing alteration of diorite porphyry by vein-forming solutions at Breckenridge, Colorado. (*After F. L. Ransome, U. S. Geol. Survey.*)

From all this it follows that the mineralization in mineral deposits of this class has been effected by solutions of relatively uniform character

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[1] Clifton-Morenci, W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905; Butte, W. H. Weed, *idem*, 74, 1912; Breckenridge, Colorado, F. L. Ransome, *idem*, 75, 1911; San Francisco district, Utah, B. S. Butler, *idem*, 80, 1913; Ely, Nevada, A. C. Spencer, *idem*, 96, 1917; Ray and Miami, Arizona, F. L. Ransome, *idem*, 115, 1919.

often capable of substituting $K_2O$ for $Na_2O$. In some deposits, particularly those carrying gold, silver, and lead, the alkaline earths have been fixed as carbonates, while in copper deposits there are usually no metasomatic carbonates.

The results indicate hot ascending solutions having a variable amount of alkaline carbonates and free carbon dioxide.

## PARAGENESIS

It has long been observed that the minerals are formed, in the main, in an orderly succession, which sometimes is repeated. This was first emphasized by Breithaupt, who recorded the series for different mineral deposits in his book on the paragenesis of minerals. The succession has an evident bearing on scientific and economic problems, and much work has been done lately on this subject in connection with the examination of ores in polished sections by metallographic methods. These have disclosed the wonderful extent to which metallic minerals are replaced by others during the process of metallization.

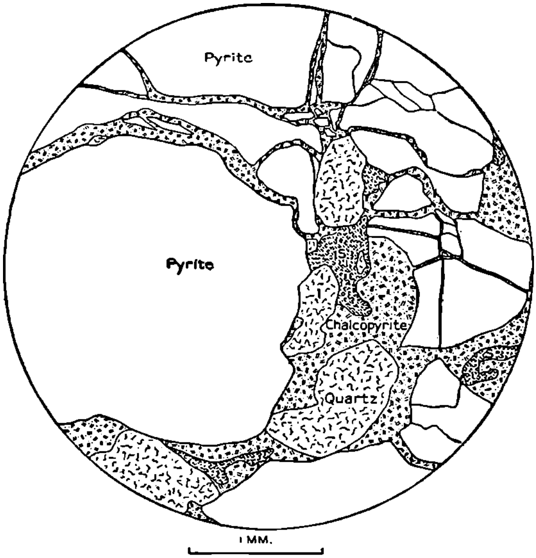

FIG. 211.—Drawing of polished surface of ore from Gilpin County, Colorado, showing earlier pyrite traversed by later veins of chalcopyrite, sphalerite (*s*), and quartz. (*After E. S. Bastin.*)

When both filling and replacement have been in action, it is natural that the country rock would first be attacked by the waters. There was first a process of dialysis by means of which the solutions were filtered through a porous rock which many elements found it difficult to penetrate. Thus we find that the composition of the metasomatic rocks near the fissure differs considerably from that of the filling. In gold-quartz veins, for instance, there is as a rule little gold in the metasomatic rocks while the filling may be rich. The carbonates, pyrite, and sericite in the rock seem to have developed about contemporaneously.

Quartz appears early in the vein filling and its deposition may continue long. Calcite, dolomite, and siderite are usually the latest gangue minerals, but may also be early. The sulphides usually form in the following order, exceptions from this sequence being rare.[1] Beginning with the oldest we have pyrite, arsenopyrite, cobalt and nickel arsenides, pyrrhotite, sphalerite, enargite, tennantite, tetrahedrite, bornite, chalcopyrite, galena; the latest minerals are argentite, gold, sulphantimonides, and sulpharsenides of silver and lead. Each mineral may replace any of the preceding ones (Fig. 211). Sulphides freely replace gangue minerals while the latter rarely replace the sulphides. Some disseminated minerals, *e.g.*, argentite in galena or chalcopyrite in bornite and sphalerite, are believed to originate by the "unmixing" at lower temperature of a solid solution deposited at a higher temperature.[2]

The general features and paragenesis of the gold-quartz veins have been summarized by F. Buschendorf.[3]

## GOLD-QUARTZ VEINS OF THE CALIFORNIA AND VICTORIA TYPE

**Principal Characteristics.**—As quartz and gold may be deposited together within a considerable range of temperature, there are several types of gold-quartz deposits. The deposits formed at higher temperatures, distinguished by such gangue minerals as tourmaline, apatite, garnet, biotite, and amphibole, will be described in subsequent pages. Those formed near the surface at temperatures not much above 150° C. have been discussed in the preceding chapter. Between the two kinds stands the large group of important deposits whose geological relations point to development at considerable depth and whose mineral association points to moderate temperatures—perhaps 200° to 300° C.

The first type is represented by certain Appalachian, Canadian, and Brazilian gold-quartz veins; the second by veins like those of the Comstock, Bodie, and Tonopah, usually appearing in Tertiary lavas. Between the two stand the gold-quartz veins of California, eastern Australia, and many localities in the interior Cordilleran region of North America.

The general characteristic of these intermediate deposits is the association of a preponderant gangue of milky, coarsely crystalline quartz, sometimes drusy, though rarely showing comb structure, with free gold and auriferous simple sulphide minerals. Where the country rock is suitable

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[1] W. Lindgren, Magmas, dikes and veins, *Trans.*, Am. Inst. Min. Met. Eng., 74, 1926, p. 88.

[2] H. Schneiderhöhn, Entmischungserscheinungen, etc., *Metall und Erz*, 19, 1922, pp. 501–526.

[3] *Zeitschr. prakt. Geol.*, 1926, pp. 1–11.

for replacement, carbonates and sericite appear with pyrite in the altered rocks.

The veins occur in deeply eroded regions and in or surrounding intrusives of quartz monzonitic or dioritic or gabbroitic kind. The only silicates present are albite and chlorite, and these only locally. The destruction of the outcrops by erosion usually results in rich placers, in which large nuggets of gold are often found.

The free gold always contains a little silver, the average fineness being 800; the sulphides are likely to carry more silver in proportion than the native gold. Some types of these veins carry a notable amount of silver, but scarcely ever such amounts as are common in the Tertiary veins deposited near the surface in Tertiary lavas.

**Gold-quartz Veins of the Sierra Nevada.**[1]—The greatest development of the gold-quartz veins is found in California; they begin in the southern end of the state in San Diego County and continue, with interruptions, to the northern end, where, in Trinity and Siskiyou counties, there is a productive area of no small value. The gold belt also continues into southwestern Oregon, but farther north disappears under the Tertiary lavas and Cretaceous and Tertiary sediments.

Most typically, the veins are developed in the Sierra Nevada, which, with its gentle western slope and abrupt eastern escarpments, separates the deserts of the Great Basin from the central valleys of California (Fig. 212).

The crest and main mass of this range form parts of an enormous batholith of massive granodiorite and allied rocks, intruded into Mesozoic and Paleozoic metamorphosed sediments. These sedimentary rocks are

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[1] H. W. Fairbanks, Geology of the Mother Lode region, *Tenth Rept.*, California State Min. Bur., 1890, pp. 23–90.

W. Lindgren, Characteristic features of the California gold-quartz veins, *Bull.* Geol. Soc. Am., 6, 1896, pp. 221–240.

W. Lindgren, Gold-silver veins of Ophir, *Fourteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1893, pp. 243–284.

W. Lindgren, The gold-quartz veins of Nevada City and Grass Valley, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, pp. 1–262.

Ernest Howe, The gold ores of Grass Valley, California, *Econ. Geol.*, 19, 1924, pp. 595–622.

F. L. Ransome, The Mother Lode district, Folio 63, U. S. Geol. Survey, 1900.

W. H. Storms, The Mother Lode region of California, *Bull.* 18, California State Min. Bur., 1900.

W. H. Storms, Possibilities of the Mother Lode in depth, *Min. and Sci. Press*, Nov. 18, 1911.

W. H. Storms, The occurrence of gold at intersections, *Min. and Eng. World*, Nov. 25, and Dec. 2, 1911.

H. W. Turner and F. L. Ransome, Sonora folio, *Geol. Atlas* 41, U. S. Geol. Survey, 1897, and other folios of the same region by Lindgren and Turner.

A. Knopf, The Mother Lode system of California, *Prof. Paper* 157, U. S. Geol. Survey, 1929.

closely folded and compressed and occupy a belt on the western slope, which gradually widens and, in Plumas County, spreads over a width of 60 miles. The great batholith itself contains extremely few quartz veins; mineralization is confined to the belt of metamorphic rocks on the western slope and often begins abruptly at the contact; this is shown, for instance, by the river gravel, which becomes auriferous where the streams enter the metamorphic areas. The highly productive part of the belt does not, usually, adjoin the granitic rocks, but appears lower down in the foothill region near smaller intrusive areas.

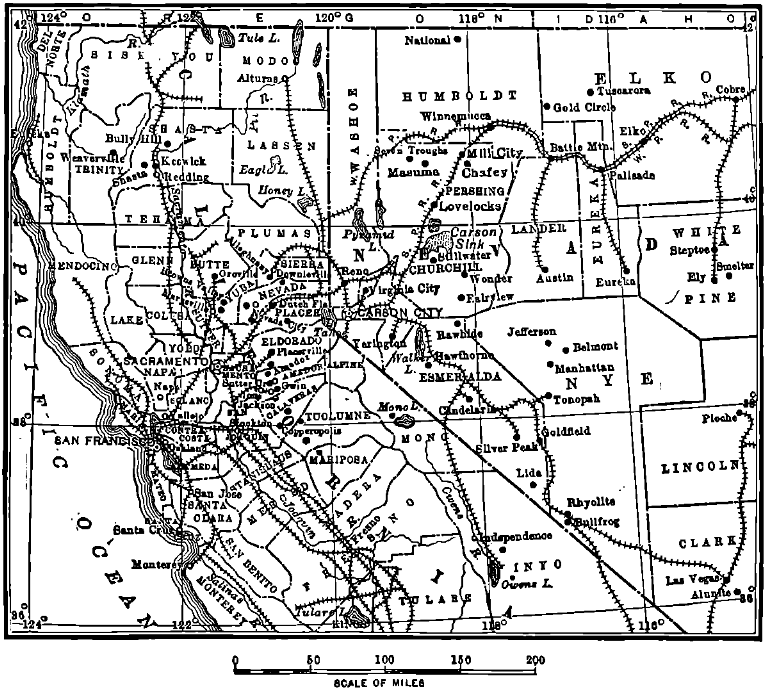

FIG. 212.—Map of Nevada and northern part of California, showing prominent mining districts.

The metamorphic rocks are a complex body, for besides the prevailing Paleozoic slates with occasional lenses of limestone and, in the lower foothills, a narrow belt of late Jurassic Mariposa slate, they contain Paleozoic lava flows and a vast quantity of tuffs, diabases, and old andesites erupted by volcanoes of Jurassic age.

Later than these rocks, and probably dating from earliest Cretaceous time, are numerous smaller intrusions of gabbro, diorite, and granodiorite, which are massive and, in a general way, contemporaneous with the main batholith of the range. The basic intrusions appear to be somewhat older than those containing more silica.

In and around these smaller intrusions, as, for instance, at Grass Valley, Nevada City (Fig. 215), Ophir (Fig. 213), and West Point, the gold-bearing veins often cluster and may occur in any kind of rocks. There are also several long lines along which fracturing and subsequent mineralization have taken place. One of these follows the so-called "serpentine belt," a dike-like intrusive mass 70 miles long; another extends from the Forest Hill divide, in Placer County, up into Sierra County, passing the town of Washington. The most important line is that followed by the Mother Lode, in the foothills of Mariposa, Tuolumne, Calaveras, Amador, and Eldorado counties, for a distance of 130 miles. The Mother Lode is by no means a single vein but, rather, a system of linked veins, placed, however, within a narrow belt about a mile wide and maintaining a remarkably straight course; it cuts Paleozoic slates and greenstones, but on the whole follows fairly closely a narrow belt of the Jurassic Mariposa slate and in places lies between this slate and the greenstone. The Mother Lode fissures usually show a reverse fault in which the maximum displacement in the line of dip is 375 feet (Fig. 214).

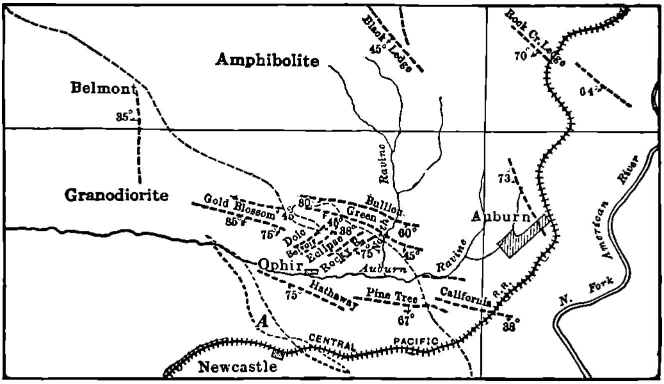

FIG. 213.—Map of principal vein systems near Ophir and Auburn, California. *A*, small area of amphibole. Scale 1 inch = 2.7 miles.

The strike of the veins is predominantly north-northwest, parallel to the range and to the strike of the steeply inclined strata; but the dip usually intersects that of the beds and, in the Mother Lode, is about 60° east. In many districts other directions of strike and dip prevail. The veins are easily traceable by prominent quartz outcrops, and many of them are remarkably straight and continuous in strike and dip. It is not uncommon to find veins continuous along the strike for 1 or 2 miles.

Many of the veins have been successfully worked to a vertical depth of 2,000 feet. In the Kennedy mine, on the Mother Lode belt, a vertical depth of 4,950 feet has been attained, good ore appearing in the lower levels. The Argonaut now has reached 6,200 feet in vertical depth with good ore in the lowest levels. At the Central Eureka mine, near the Kennedy, new rich ore-bodies were found below the 1,000-foot level, though there was very little ore above that horizon. Again, new ore-shoots were opened from the 3,500- to the 4,400-foot levels. The North Star vein, at Grass Valley, has been followed along its flat dip for about 9,000 feet to about 4,000 feet vertically. Mainly from the deepest levels, 94,600 tons of ore was produced in 1925, averaging $8.19 per ton.

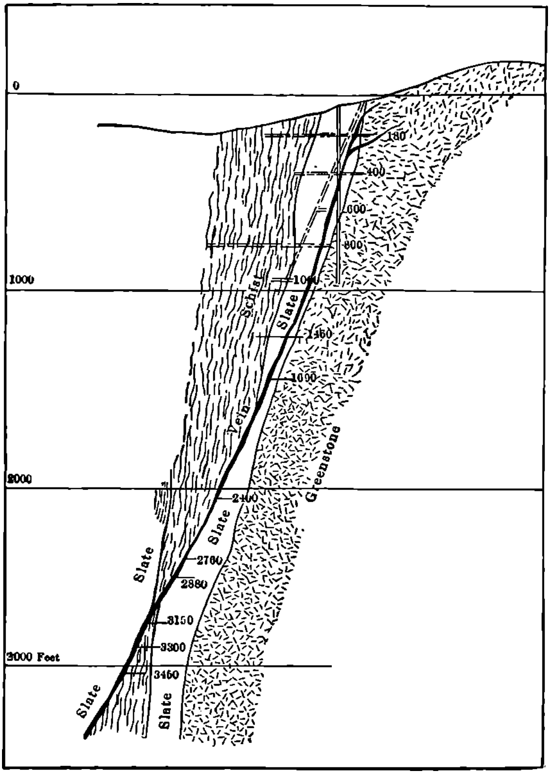

FIG. 214.—Vertical cross-section of the Mother Lode near the Argonaut shaft, showing reverse fault along vein fissure. "Schist" is amphibolite, white area, Mariposa clay slate (Jurassic).

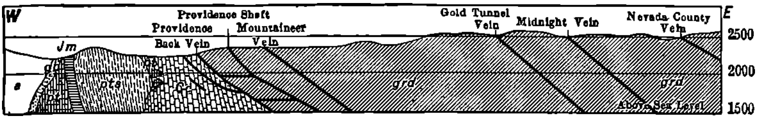

FIG. 215.—Geological section at Nevada City, California. *Cc*, Carboniferous slate: *Jm*, Jurassic slate; *pt*, porphyrite; *gb*, gabbro; *pts*, amphibolite; *s*, serpentine; *grd*, granodiorite. Scale 1 inch = 2,400 feet.

There are many structural types; the most common is the simple filled vein (Fig. 216), which may range from a mere film of quartz to a thickness

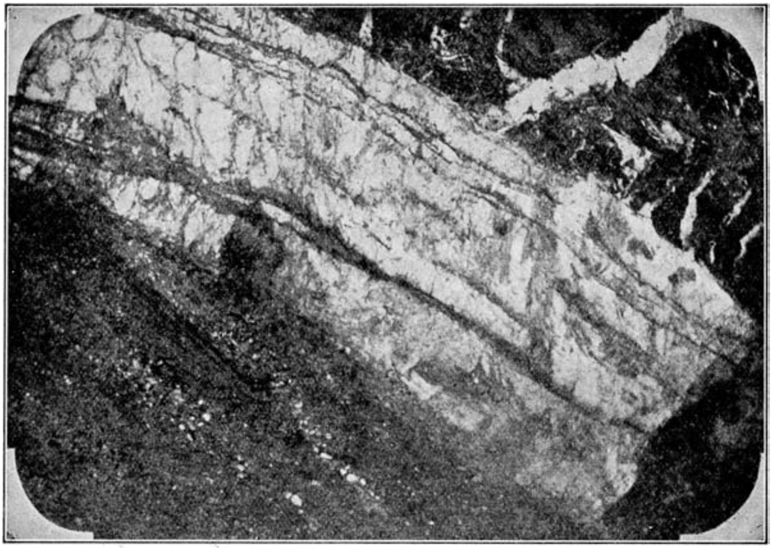

FIG. 216.—Argonaut vein in slate country rock, Amador County, California, at 650-foot level. (*Photograph by O. H. Packer.*)

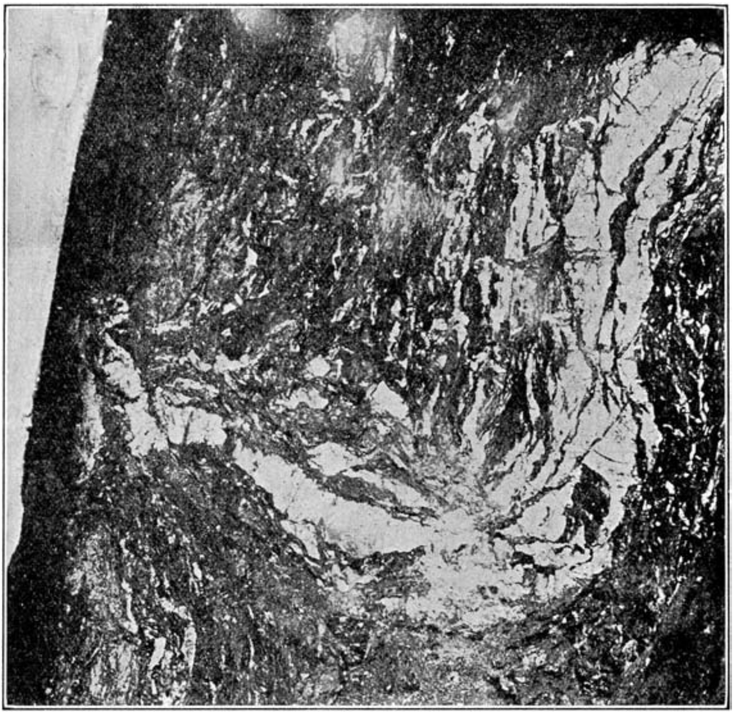

FIG. 217.—Bunker Hill vein, Amador County, California, showing folded vein in crushed clay slate.

of 10 or 20 feet. Many of the outcrops appear to be much thicker than the figures just given, but these large masses are poor in gold. Again, there are composite veins or lodes in which certain belts of country rock are filled by branching veinlets of quartz or which may contain altered slabs of country rock (Fig. 218). In crushed clay slates the veins are sometimes broken and folded (Fig. 217). Large bodies of rock changed by replacement to gold ores are comparatively rare; such ores are mined in several places at Angels Camp, Calaveras County, but even here the gold is mainly contained in thin quartz seams in the altered rock. Again, gold-bearing quartz seams may follow joints of certain direction in large masses of rock; many such masses have been worked by the simple process of hydraulic washing of the upper, weathered part. Such deposits are called seam diggings.

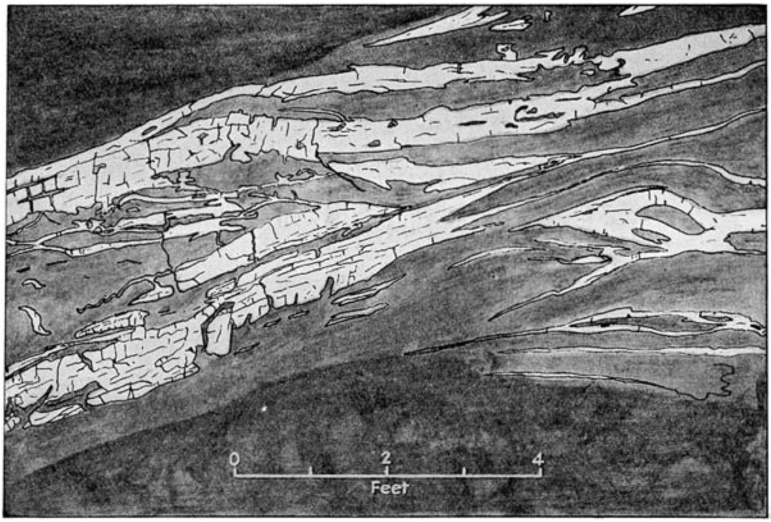

FIG. 218.—North Star vein, Grass Valley, California, near 1,800-foot level, showing quartz vein in brecciated and altered diabase.

More rarely the veins follow narrow dikes of albite aplite; or they are developed on joint planes across the strike of thicker dikes in the manner of ladder veins.

The association of gold with dikes consisting mainly of albite rock has been described by Turner[1] and Reid.[2] Turner describes such dikes on

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[1] H. W. Turner, Notes on the gold ores of California, *Am. Jour. Sci.*, 3d ser., 47, 1894; *idem*, 3d ser., 49, 1905; Replacement deposits in the Sierra Nevada, *Jour. Geol.*, 7, 1899, pp. 389–400.

[2] John A. Reid, The east country of the Mother Lode, *Min. and Sci. Press*, March 2, 1907.

Moccasin Creek, in Tuolumne County,[1] at the Shaw mine, in Eldorado County, and at other places; but the associated ores are generally of low grade and the mineralization is everywhere later than the dike.

Reid observed numerous thin dikes in Calaveras slate near Blue Canyon, Placer County, consisting largely of albite, which are cut or followed by seams or veins containing quartz, albite, pyrite, arsenopyrite, and native gold.

In feldspathic and femic rocks there is more or less replacement extending a few inches or a few feet from the vein, with dissemination of pyrite, calcium-magnesium-iron carbonates, and much sericite (p. 532). Occasionally, in sodic amphibolites, much albite develops; and, in the vicinity of Angels Camp, on the Mother Lode, such replacements may contain enough gold to be called ore. In serpentine the alteration to a coarse aggregate of ankerite and bright-green chromium mica (mariposite) is characteristic; this product of replacement constitutes ore in only a few places, such as the Rawhide mine, southeast of Angels Camp, where it was penetrated by gold-bearing quartz stringers.

The ore-shoots are irregularly distributed; many veins are of pockety character, containing rich bonanzas at certain points, which may be determined by intersections or by the crossing of certain beds of the schist series. Other veins have large and regular shoots generally with a steep pitch, and sometimes with a pitch length of many thousand feet. In isolated cases, such as the Idaho-Maryland vein at Grass Valley, the pitch of the rich pay-shoot was flat on the plane of the vein (p. 194). In many districts, especially at Grass Valley, the rule is that the shoot pitches to the left of an observer looking down the dip.

A. Knopf,[2] writing of the Mother Lode, believes that the larger shoots represent the easy passage ways of the ascending solution, and that the poor or barren bodies of quartz indicate more stagnant solutions which were constantly enriched in silica derived from the alteration of the country rock. The evidence as to the supposedly favorable influence of the carbonaceous Mariposa slates is contradictory.

Including the placer gold yielded by the outcrops disintegrated during Tertiary and Quaternary time, the production of the California gold-quartz deposits is exceedingly large, being $1,852,000,000 to 1931 inclusive. The actual mining of the quartz veins has yielded about 32 per cent of this amount.[3] In 1930 the lode mines produced 60 per cent and the placer mines 40 per cent of the gold. A long list of celebrated mines could be cited, each one having yielded from $5,000,000 to $30,000,000. Among them are the North Star, Empire, and Idaho-Maryland, of Grass Valley, and the Plymouth Consolidated, Kennedy, Keystone,

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[1] H. W. Turner, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 1, 1896, p. 664.

[2] *Op. cit.*, p. 32.

[3] J. M. Hill, *Econ. Geol.*, 21, 1926, pp. 172–179.

Eureka Consolidated, Gover, and Zeile on the Mother Lode.[1] The annual production from lode mines in California has decreased since the World War to about $7,200,000 in 1925 and $5,700,000 in 1930.

The principal and almost exclusive gangue mineral is milk-white quartz with coarse massive texture, occasionally drusy. In thin section the quartz shows partly idiomorphic forms (Fig. 54), and some individuals include earlier slender prisms. Comb structure is sometimes seen, but never the delicate banding of the veins formed near the surface. In places sulphides encrust rock fragments enclosed in quartz. A rough banding may result from irregular distribution of the sulphides, from the inclusion of narrow strips of black slate, or from subsequent shearing of the vein (Fig. 55); the last is not uncommon and is indicated in thin section by the crushing of the primary individual crystals (Fig. 56). Fluid inclusions are plentiful, and seem to consist of an aqueous solution. Calcite, dolomite, and ankerite are formed in subordinate quantities, though they may be present abundantly in the replaced country rock adjoining the vein. Barite, fluorite, and tourmaline are absent, as are biotite, garnet, amphibole, epidote, zeolites, rhodonite, and rhodochrosite No bituminous material has been reported. Mariposite, a chromium mica, is common near serpentine in the altered rock; roscoelite, a vanadium mica, is sometimes associated with native gold. Rutile is generally confined to the altered rock. Specularite, cinnabar, and magnetite are absent, except in isolated cases. Scheelite is known to occur at several places.[2]

The native gold is the principal ore mineral and occurs in all ores and at all depths. Sometimes large masses are found. A mass of solid gold valued at $40,000 was taken out from the Bonanza mine, near Sonora, in a pocket which yielded $360,000. This mine produced more than $2,000,000 in gold, the greater part of which was pounded out of the quartz in hand mortars. Still heavier masses of gold were found in the Monumental mine, Sierra County, and below the outcrops of the Carson Hill veins on the Mother Lode. In some veins the gold is distributed in microscopic particles; in others it is visible (Fig. 219) and occurs in threads and plates. Coarse gold, replacing quartz and arsenopyrite, is described by Ferguson and Gannett[3] from the pocket mines of Alleghany, in Sierra County (Fig. 219). Very rarely, in some pocket mines, gold of a fineness exceeding 900[4] is encountered, but the average fineness is 800

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[1] The Kennedy mine to the close of 1915 yielded $6,378,000 from 729,000 tons of ore; later yields are not disclosed. The North Star mine from 1884 to 1926 inclusive yielded $28,924,488 from 2,704,810 tons of ore.

[2] C. D. Hulin describes fluor-apatite with gold and quartz from the lower levels of the Kennedy mine, *Econ. Geol.*, 25, 1930, p. 348.

[3] H. G. Ferguson and R. W. Gannett, Gold-quartz veins of the Alleghany district, *Prof. Paper* 172, U. S. Geol. Survey, 1932, 139 pp.

[4] Gold from the San Giuseppe mine, near Sonora, was 990 fine.

to 830, and it is rarely as low as 700, the remainder being principally silver.

Variable, but always comparatively small, quantities of metallic minerals accompany the gold, ordinarily making up 2 to 3 per cent of the mass. Pyrite is universally present; pyrrhotite rarely, and then only in veins in granite rocks. Chalcopyrite, sphalerite, and galena are most

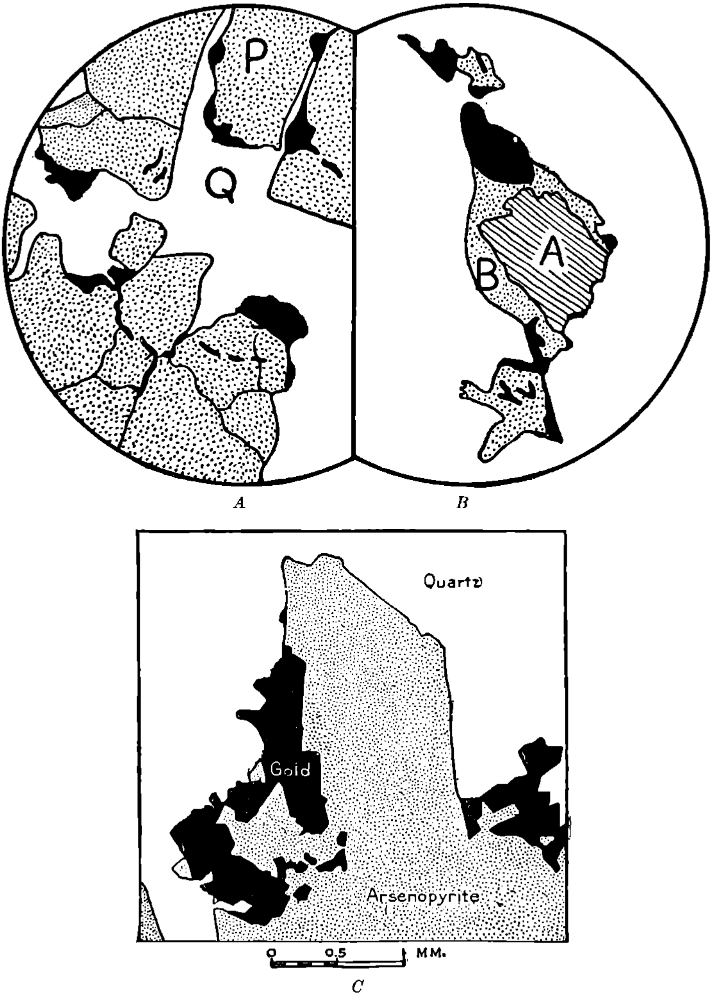

Fig. 219.—*A*. Thin section of gold-bearing quartz, Keltz mine, Tuolumne County, California. *Q*, quartz; *P*, pyrite; black, gold, deposited later than pyrite. Magnified 70 diameters. (*After W. J. Sharwood.*)

*B*. Drawing showing succession arsenopyrite (*A*), telluride (*B*), and gold (black) in quartz from Bridge River district, British Columbia. Magnification 12 diameters. (*After W. S. McCann.*)

*C*. Gold replacing arsenopyrite and quartz, Alleghany district, California. (*After H. G. Ferguson.*)

abundant next to pyrite; arsenopyrite is also common. Tetrahedrite is frequently found, while stibnite and molybdenite are rare. The gold is, as a rule, later than the sulphides. Compounds of tin, uranium, boron, phosphorus, and fluorine are lacking. Tellurides like altaite, hessite, calaverite, petzite, and melonite are sometimes associated with native gold, as at Nevada City and Carson Hill.

The sulphides obtained by concentration from the ore are usually rich, often having a value of $100 to $300 per ton, but their value is only a small part of the value of the ore. J. M. Hill[1] gives the following figures for the Mother Lode mines. In the five Mother Lode counties 437,409 tons of ore were mined in 1925, with a total recovery of $3,137,150. The gold recovery on the amalgamating plates averaged $5.63 per ton, while 9,465 tons of concentrates (mainly pyrite) obtained from the ore averaged in value $57 per ton; the total value recovered in gold (with a very small quantity of silver) averaged $7.17 per ton.

The gold-quartz veins of the Sierra Nevada were formed shortly after the intrusion of the granodiorite batholith in latest Jurassic or earliest Cretaceous time. They have, with the surrounding rocks, been subjected to an intense erosion, the vertical measure of which amounts to several thousand feet. The exposures by unequal erosion or by mining operations show, in many districts, that the vertical range of gold deposition without notable change in richness of shoots was over 4,000 feet; the relations in some districts lead to the conclusion that the deepest parts now mined were formed 7,000 feet or more below the surface.

By metasomatic processes, the wall rocks absorbed carbon dioxide, potassium, sulphur, and lime from the solutions. On the other hand, sodium has been leached, and undoubtedly much silica has been extracted from the walls. The result of this would probably be a concentration of the ascending solution and a progressive accumulation of silica,[2] available for deposition in the open spaces. A. Knopf[3] believes, indeed, that practically all of the quartz filling was derived from this source. This is probably going too far. The solutions ascending from the magmatic source were most likely siliceous in the first place. The mode of deposition was predominantly by filling during successive re-openings of the fissures.

The gold and other metals were undoubtedly derived from magmatic emanations from the congealing granodioritic masses, which probably underlie the region at no excessive depths. In places the evidence points to more basic rocks or to differentiated albite dikes as the source of at least a part of the gold.

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[1] *Mineral Resources*, U. S. Bur. Mines, pt. 1, 1925, p. 301.

[2] W. Lindgren, The gold-quartz veins of Nevada City and Grass Valley, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, p. 184.

[3] *Prof. Paper* 157, *idem*, 1929, p. 45.

As to the state in which gold was carried by the solutions, no definite answer can be given. Hot waters containing carbon dioxide, alkaline carbonates, and hydrogen sulphide would be competent to effect the changes noted in the rocks. Gold is soluble to some extent in a solution containing sodium carbonate, sodium silicate, and carbon dioxide. It is relatively easily soluble in sodium sulphide, which may well have been present. Further, it is easily dispersed in silica gels and sols and can be easily precipitated from these dispersions. On the whole, it is most probable that the gold was carried in such colloidal form. The presence in the solution of any gold salt, such as the chloride, is extremely improbable.

**The Gold-quartz Veins of the Interior Cordilleran Region.**[1]—A great number of intrusive masses of quartz monzonitic or granodioritic type are found in the interior Cordilleran region of the United States. They are, as a rule, of more recent age than the great coast batholith, their epoch of intrusion falling at the end of the Cretaceous or the beginning of the Tertiary. In or around these intrusives gold-quartz veins are often found, clearly related to the California type, but differing from it in some respects. Many of them occur with east-west strike in the central parts of the Idaho batholith—a rather unusual position; the majority are, however, located near contacts or roof pendants. Frequently they follow lamprophyric dikes (Fig. 220). They contain more sulphides, though of the same kinds, and they carry, as a rule, more silver in the sulphides than the veins of the California type; there is less free gold, and, in some instances, only a small proportion of the total gold is amenable to direct amalgamation. The gold contains silver and rarely has a fineness above 700. Rich silver minerals often form in the oxidized zones. The

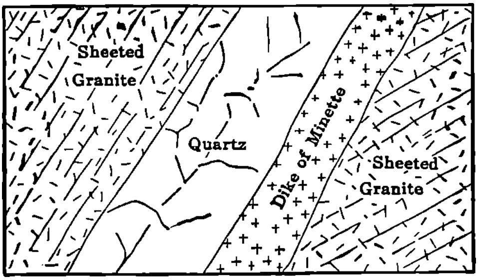

FIG. 220.—Section of Hidden Treasure vein, Neal district, Idaho.

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[1] W. Lindgren, The mining districts of Idaho Basin, etc., *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1897, pp. 617–744.

W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, *idem*, pt. 2, 1901, pp. 551–776.

C. E. Weaver, The Blewett mining district, *Bull.* 6, Washington Geol. Survey, 1911.

precious metals are contained chiefly in the quartz filling; but the altered rock adjoining the veins sometimes carries gold and silver, which means that it is in part replaced by gold- and silver-bearing sulphides. In feldspathic and ferromagnesian rocks sericitic and pyritic alteration (Fig. 221) affects the wall rocks; carbonatization is rarely as intense as in the veins of the Sierra Nevada. Pyrite, arsenopyrite, chalcopyrite, galena, and sphalerite are the common ore minerals, but tetrahedrite is also plentiful and cinnabar is known to occur. Tellurides are sometimes present and are almost always intergrown with native gold. Quartz is the prevailing and usually the only gangue mineral. Quartz with coarse native gold has been observed to replace an earlier calcite gangue (Fig. 222). Tourmaline, magnetite, and pyrrhotite are not known. The grade of the ore is from $5 to $15 per ton.

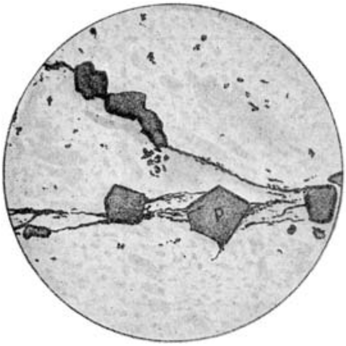

FIG. 221.

FIG. 222.

FIG. 221.—Pyrite (*p*), forming by replacement along calcite veinlets (black); calcite forms lining around pyrite crystals. In chloritic diabase Great Northern mine, Canyon, Oregon. Magnified 10 diameters.

FIG. 222.—Quartz (*q*) with native gold (black), replacing vein of calcite (*c*); Great Northern mine, Canyon, Oregon. Magnified 10 diameters.

**Victoria, Australia.**[1]—The principal gold-bearing region of Victoria, though of much smaller extent than the California gold belt, is believed

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[1] E. J. Dunn, Report on the Bendigo gold field, Dept. of Mines, Melbourne, 1896.

T. A. Rickard, The Bendigo gold field, *Trans.*, Am. Inst. Min. Eng., 20, 1891, pp. 463–545.

J. W. Gregory, The Ballarat East gold field, *Mem.* 4, Victoria Geol. Survey, 1907.

W. Baragwanath, The Castlemaine gold field, *Mem.* 2, *idem*, 1903.

O. A. L. Whitelaw, The Wedderburn gold field, *Mem.* 10, *idem*, 1911.

W. Lindgren, Characteristics of gold-quartz veins in Victoria, *Eng. and Min. Jour.*, March 9, 1905.

F. L. Stillwell, Replacement in the Bendigo quartz veins, etc., *Econ. Geol.*, 13, 1918, pp. 100–111.

N. R. Junner, The geology of the gold occurrences of Victoria, Australia, *idem*, 16, 1921, pp. 79–123. The analyses given indicate that the altered wall rocks (unusually) contain much sodium, suggesting the presence of albite.

to have produced about the same amount, namely, $1,500,000,000 in gold. Here, too, the placers have yielded by far the greater production. Both gravel deposits and quartz veins still yield a sharply diminishing output. In 1915 the production of gold from quartz mines was only about 218,660 ounces, and in 1930 had declined to 25,119 ounces. The ores averaged $7.50 per ton.

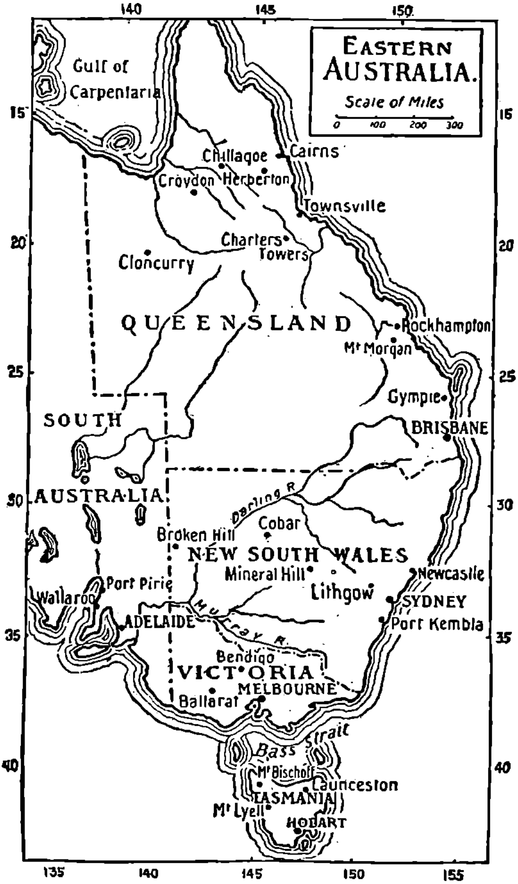

FIG. 223.—Sketch map of eastern Australia, showing location of important mining districts.

This most productive region includes the celebrated districts of Ballarat and Bendigo and is situated in the low ranges of the mountains rising between the basaltic and Tertiary terranes on the south and the Murray Plains on the north (Fig. 223).

Little altered Ordovician slates and sandstones prevail and form sharply compressed folds. Intruded in them are two batholiths of granitic rock, probably quartz monzonite, the largest being that between Bendigo and Castlemaine; there are also many smaller bodies of the same

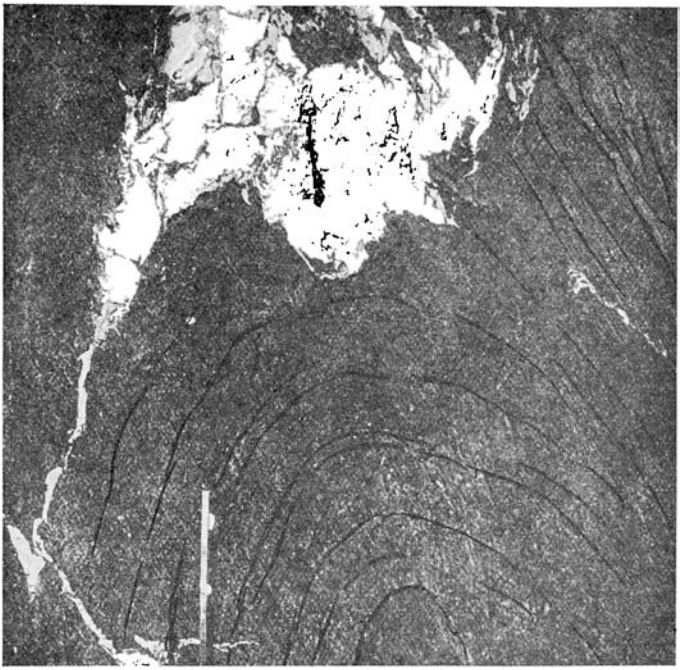

FIG. 224.—Saddle reef in slate and sandstone, Bendigo, Victoria. (*After E J. Dunn.*)

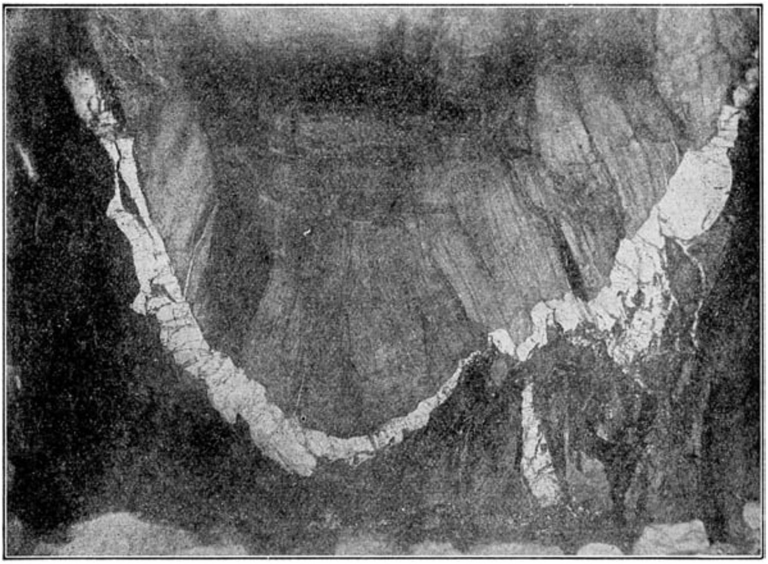

FIG. 225.—Trough reef, in slate and sandstone, Bendigo, Victoria. (*After E. J. Dunn.*)

kind. The intrusions are probably of Devonian age, and erosion of at least 3,000 feet has planed the region to an undulating surface.

Within the folded Ordovician rocks quartz veins are abundant and generally follow the strike of the strata, being massed along certain productive "reef lines." Frequently they are conformable between shale and sandstone, but some of them cut across the strike. A common type has one well-defined wall from which flat and irregular bodies of quartz project into the hanging or footwall. These flat "makes" are particularly characteristic and usually contain the best ore at Ballarat East and other places. The saddle reefs constitute an interesting division, in which masses of quartz fill cavities produced at anticlines (Fig. 224) or less commonly at synclines (Fig. 225) by stresses subsequent to the principal folding; they are often connected with irregular "makes" and spurs (Fig. 226) of quartz. These open cavities, subsequently filled by quartz, are the necessary result of stresses applied to folded masses of little altered sediments, the strata of which vary considerably in hardness.

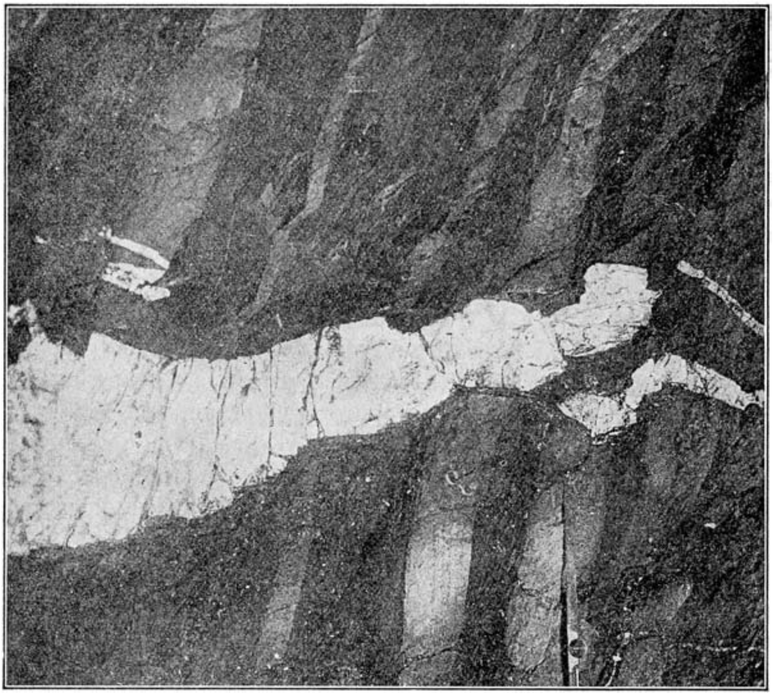

FIG. 226.—Spur reef in slate and sandstone, Bendigo, Victoria. (*After E. J. Dunn.*)

The best instances of saddle reefs, many of them superimposed upon and following three or four distinct lines of anticlines, are found at Bendigo (Fig. 38) and Castlemaine. The Bendigo veins have been worked to a depth of 4,600 feet in the Victoria reef, situated on the New Chum reef line, but sinking has been suspended. A body of quartz, containing at best $17 per ton, was mined at a depth of about 4,200 feet, but it is said that on the whole little profitable mining has been done at Bendigo below a depth of 2,500 feet. The granitic rocks rarely contain quartz veins. The vein-filling is a massive milk-white, sometimes glassy, quartz of coarse, crystalline texture. It contains native gold, often coarse, and also a little pyrite and arsenopyrite; sometimes also a little galena, sphalerite, molybdenite, stibnite, and bournonite. No tellurides are reported. The gold is later than the sulphides and about 900 fine. There is neither barite nor fluorite. Ankerite or calcite with some magnesium and iron is common but subordinate, usually appearing near the walls. Albite and a vermicular chlorite are present in places, the former in vugs, the latter enclosed in massive quartz. There is little evidence of banded structure, except that near the walls of the veins thin lamellæ of slate may be interlaminated with quartz.

At Ballarat, rich ore-bodies occur at the intersection of flat bodies of quartz with certain thin pyritic and carbonaceous seams of slate, the so-called "indicators." It has been held that the gold has been precipitated by the carbon in the indicator. A more plausible view is that the indicators are narrow fissures, later than the flat "makes" and enriching them at the intersection. Similar features have been noted at other points in Victoria and seem to point to a process of enrichment, although probably not caused by surface waters. At Ballarat, the developments at depths below 1,500 feet have not been encouraging, and the mines are now closed.

The granitic intrusions and the formation of the quartz veins were closely associated events. The fact that so few lodes occur in the granitic rocks is probably explained by the great resistance of the hard intrusive bosses to compressive stresses, compared with the yielding nature of the soft sedimentary rocks.

J. R. Don (p. 7) has shown that the sediments away from the veins contain no gold, and that the increasing traces of gold found as the veins are approached are dependent upon the amount of pyrite introduced from the veins.

The metasomatic processes have been studied by Junner in the Walhalla-Woods Point belt where diorite porphyries are found close to the veins. Numerous analyses indicate that the changes consist, in addition of potash and carbon dioxide, in the reduction of soda, and in small but constant losses of silica, alumina, and ferrous oxide. The quartz is mainly formed by filling, but Stillwell has advanced the view that the laminated "ribbon quartz" is a product of replacement.

**New South Wales and Queensland.**—A large number of well-known and productive districts are found in New South Wales and Queensland, in which the gold occurs in quartz veins associated with intrusive rocks. Some of these veins carry quantities of sulphides besides free gold; occasionally fluorite and barite are reported. The almost universal conditions are a deeply eroded region with diorite or granodiorite or their basic dikes intruded into Paleozoic sediments, which are usually more highly altered than in Victoria; the veins occur either in intrusive or sedimentary rocks or in both. Placers are usually present.

At Hill End,[1] north of Bathurst in New South Wales, folded Silurian slates and tuffs are intruded by dikes and sills of quartz porphyry. The lenticular quartz veins lie in slate or at the contact with the intrusive rocks. Coarse gold prevails; one mass, extracted in 1872, consisting of solid gold mixed with some quartz weighed 630 pounds and was valued at $60,000. Five and one-half tons of solid gold were recovered at this place from 10 tons of quartz, the value of the gold being $3,300,000. Similar geological conditions exist at Hargraves, but the quartz here occurs as saddle reefs. Here, as at Ballarat, flat "makes" are present and are enriched where they are crossed by "indicators" or narrow bands of dark-greenish slate.

At Hill Grove,[2] in the New England district, in the northeastern part of New South Wales, near the wolframite deposits mentioned elsewhere (p. 760), slates and quartzites are intruded by quartz-mica diorite; and the veins occur in the sedimentary rocks, often near lamprophyric dikes. The veins, which average 6 feet in width, contain, besides quartz and free gold, scheelite, arsenopyrite, and much stibnite. Andrews regards them as due to the last emanations from the same granitic magma, the earlier high-temperature emanations having produced the cassiterite-molybdenite-wolframite deposits.

At Charters Towers,[3] in Queensland, the veins intersect granitic rocks ranging from granites to quartz-mica diorites. The veins have been highly productive and have been worked to a depth of 3,000 feet along the dip. They contain about 7 per cent of sulphides (pyrite, galena, sphalerite, pyrrhotite, and arsenopyrite). As usual in granitic rocks, a considerable part of the gold of the ore is contained in the sulphides. The veins are regular but narrow, averaging about 3 feet in thickness. The average value of the ore is probably less than $15 per ton.

**Nova Scotia.**—The gold-quartz veins of Nova Scotia,[4] from which during the last 50 years a moderate production has been derived, are, in many respects, of special interest. The veins are contained in folded

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[1] E. F. Pittman, Mineral resources of New South Wales, Geol. Survey N. S. W., 1901, p. 31. Comprehensive summary in Maclaren, Gold, 1908, pp. 341–358.

[2] E. C. Andrews, *Records*, Geol. Survey N. S. W., 8, 1909, p. 143.

[3] Jack, Rands, and Maitland, *Ann. Rept.*, Geol. Survey Queensland, 1892.
W. E. Cameron, *Publ.* 224, Geol. Survey Queensland, 1909.

[4] E. R. Faribault, The gold measures of Nova Scotia and deep mining, *Jour.*

sedimentary rocks—slate and quartzite—probably of Cambrian age, which are intruded by granitic rocks of Silurian age. The gold belt extends for a distance of 280 miles along the south coast, and its average width is about 30 miles. The numerous quartz veins, many of which can be traced for long distances, often occur in the manner of saddle reefs along the anticlines. The anticlinal axes are in places marked by structural elliptical domes in which the strata pitch both ways on the strike, and gold-bearing quartz veins are usually found in such domes. The veins are ordinarily parallel to the stratification, but some of them, while parallel in strike, cut across the dip. Corrugated and crenulated veins are common (Fig. 227) and the term "barrel quartz" is used to describe the material in them; the corrugation is believed to have been caused by deformation subsequent to the deposition. The gangue is always quartz; arsenopyrite, pyrite, chalcopyrite, galena, and sphalerite are fairly common, but the principal valuable mineral is native gold.

Fig. 227.—Folded quartz vein in slates, Mexican mine, Goldenville, Nova Scotia. (*After T. A. Rickard.*)

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Can. Min. Inst., 2, 1899, pp. 119–161.

J. E. Woodman, Geology of Moose River gold district, N. S., Nova Scotia Inst. Nat. Sci., 11, 1903, pp. 18–88.

W. Malcolm, Goldfields of Nova Scotia, *Mem.* 20-E, Canada Dept. Mines, Geol. Survey Branch, 1912.

T. A. Rickard, The domes of Nova Scotia, *Trans.*, Inst. Min. and Met. (London), 21, 1912, pp. 506–560.

Veins with stibnite occur in the auriferous belt, and scheelite is rather frequently present.[1] The gold-bearing veins were probably formed soon after the granitic intrusion.

Under the microscope the glassy quartz shows intense deformation, and the corrugated veins are probably simply the result of the crumpling of harder beds in a plastic medium (*cf.* Fig. 19). Two epochs of folding are indicated, one preceding and the other following the deposition of quartz.

## GOLD-BEARING REPLACEMENT DEPOSITS IN LIMESTONE

Deposits in which limestone is replaced by jasperoid or fine-grained silica and which carry gold or silver or both are sometimes found in the Cordilleran States where intrusive porphyries invade calcareous sediments. Few examples are known elsewhere. These ores, which are usually very poor in sulphides, are at several places of economic importance.

In the Mercur district,[2] situated in the Oquirrh Range in Utah, siliceous silver ores are found at the lower contact of a thin sheet of granite porphyry with Carboniferous limestone. The jasperoid rock, in places 55 feet thick, contains more or less silver throughout but has not been extensively worked. It carries barite and calcite in places.

In the same district, below an upper sheet of porphyry which like the lower is greatly decomposed by processes of weathering, is found a sheet of jasperoid rock, locally 25 feet thick, which contains minutely divided, generally invisible gold with some fine-grained pyrite, a little barite, and some realgar and cinnabar. In part the porphyry itself constitutes ore, and the ore may extend into the limestone above the porphyry. Spurr suggests that the ores gained access to the sheet through vertical fissures, now filled with calcite.

From 1890 to the end of 1913 about 4,900,000 tons of this gold ore averaging about $3.58 per ton in gold have been mined in the Mercur district. The total yield had a value of $19,000,000. The mines are now closed and dismantled.

W. H. Weed[3] described similar deposits in the Moccasin district, in Montana, where rhyolite porphyry and phonolitic dikes intrude Carboniferous limestone. Near these intrusives the limestone is replaced by fluorite (Fig. 69) and by jasperoid; the replaced rock contains gold and has been mined successfully for a number of years. Some of the ore deposits here, too, lie at the lower contacts of intrusive sheets.

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[1] V. G. Hills, Tungsten mining in Nova Scotia, *Proc.*, Colorado Sci. Soc., 10, 1912, pp. 203–210.

[2] J. E. Spurr, *Sixteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1895, pp. 349–455.

[3] W. H. Weed and L. V. Pirsson, Geology and mineral resources of the Judith Mountains of Montana, *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, pp. 437–616.

The so-called refractory siliceous ores of the Black Hills of South Dakota, described by J. D. Irving,[1] constitute the best examples of this type of replacement ores. For many years these ores yielded annually about $2,000,000 in gold and 100,000 ounces of silver, from about 600,000 short tons of ore. At present the production is very small. The ores are treated by the cyanide process. The deposits form replacements of dolomite at two horizons in the Cambrian section of the Black Hills, in a region which is intruded on a large scale by dikes, sheets, and laccoliths of rhyolite porphyry, syenite porphyry, and phonolite of probable Eocene age.[2] The more important lower horizon is 15 to 25 feet above the basement of pre-Cambrian schists.

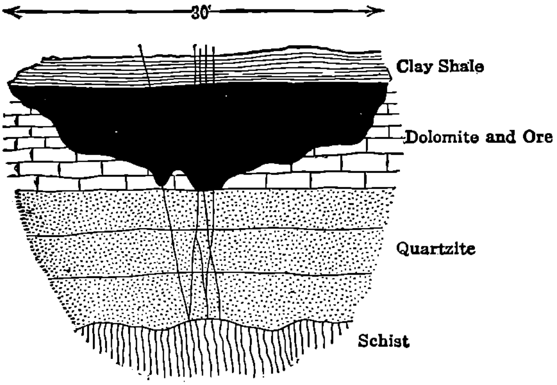

FIG. 228.—Cross-section of shoot of siliceous ore (black) replacing Cambrian dolomite, Black Hills, South Dakota. Spread of ascending solutions on under side of impervious shale makes shoot wider at top. (*After J. D. Irving.*)

The ores occur immediately below more or less impervious beds of shale or below sills of intrusive rocks. While the richest ore replaces the dolomite, ores of lower grade may also replace the underlying basal Cambrian quartzite and the overlying shale; the replaced bodies are at most 18 feet thick, averaging 6 feet. These channel-like ore-bodies have a width attaining 300 feet but averaging much less. Their length is considerable, one shoot having been followed for three-fourths of a mile. Many parallel shoots may be found in one locality, each shoot corresponding to a fissure or series of fissures ("verticals") which intersect the basal beds but which do not carry the ore below the quartzite and rarely above the shale (Fig. 228).

The ore is a hard, brittle fine-grained siliceous rock, often reproducing the dolomite texture with great fidelity (Fig. 61). The fresh ore is locally bluish and contains finely divided pyrite; much of it contains solution cavities lined with quartz crystals. Fluorite is always present, frequently

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[1] *Prof. Paper* 26, U. S. Geol. Survey, 1904.

[2] T. A. Jaggar, *Prof. Paper* 26, U. S. Geol. Survey, 1904, pp. 24–26.

also barite. Other associated minerals are stibnite, occasionally wolframite, and probably arsenopyrite and tellurides in fine distribution. Much of the ore is mined at shallow depths and is largely oxidized. Interesting data as to the form and distribution of the ore-shoots are also given by J. D. Irving in a later paper.[1]

### GOLD-BEARING REPLACEMENT DEPOSITS IN PORPHYRY

Larger bodies of aluminous rock are more rarely replaced. W. H. Emmons[2] has described an example of this in the Little Rocky Mountains of northeast Montana, a small outlier on the Great Plains. Stocks and sheets of syenite porphyry are intruded in a Paleozoic sedimentary complex. Broad zones in this porphyry are replaced and cemented by quartz, pyrite, secondary orthoclase, and fluorite. The deposits are really wide replacement lodes, some of them traceable for 1,200 feet and varying from a few feet to 100 feet in width. The gold is finely distributed and probably occurs as a telluride, the ores averaging about $3 per ton in gold and one ounce of silver. The operations have thus far been confined to the oxidized zone, which has a depth of 200 feet. The relationship to the epithermal Cripple Creek ores is evident.

### THE SILVER-LEAD VEINS

**General Features.**—The mesothermal silver deposits include many types between which so many transitions exist that a classification is difficult. Certain forms occurring as fissure veins parallel closely the gold-bearing quartz veins; but many contrast with them in being usually associated with carbonate gangue, more frequently ankerite or other magnesium-calcium-iron carbonates than calcite or siderite.

The replacement deposits in limestone very often contain rich silver ores, the three most common minerals being galena, tetrahedrite, and sphalerite. Galena and sphalerite may so predominate that the base metals yield the principal value of the deposits. Chalcopyrite, pyrite, and arsenopyrite play subordinate parts. Native silver is here probably never a primary mineral, although abundantly formed by secondary reactions effected by descending waters; and rich sulphantimonides like proustite, pyrargyrite, and polybasite are also partly though not wholly, of similar secondary origin.

The following types merely serve as centers around which the descriptions may be grouped.

**Quartz-Tetrahedrite-Galena Veins.**—Prominent veins carrying milky quartz and sparsely disseminated tetrahedrite, galena, and sphalerite, with subordinate pyrite, are common in the Cordilleran region in or near

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[1] Replacement ore-bodies, *Econ. Geol.*, 6, 1911, pp. 527–561.

[2] *Bull.* 340, U. S. Geol. Survey, 1908, pp. 96–116. See also W. H. Weed and L. V. Pirsson, *Jour. Geol.*, 4, 1896, pp. 399–428.

intrusive bodies of granitic texture. Many such veins are found in the great batholiths of Idaho and Montana. The deposits are on the whole poor and rarely worked, although from 1870 to 1890 the enriched surface zones in many places yielded much silver chloride, native silver, and ruby silver. The Granite-Bimetallic vein in Montana is a famous representative (p. 864).

The Premier mine,[1] at the head of Portland Canal, British Columbia, is located in a country of high relief and strong mineralization near the eastern contact of the great batholith which follows the coast for several hundred miles.

Since 1918, it has produced 1,870,411 tons of ore, yielding 1,100,000 ounces of gold and 27,000,000 ounces of silver. In 1930 the mine still produced 2,500,000 ounces of silver.

The veins are mainly formed by replacement along a strong fracture zone and occur in tuffaceous Jurassic rocks intruded by dikes of granodiorite porphyry. The depth reached is only about 800 feet but the upper part of the deposit is evidently the richer. Sericitic alteration of the country rock is marked. The gangue minerals comprise adularia, quartz, chlorite, and sericite. The ore minerals in their order of succession are as follows: Pyrite, sphalerite, galena, tetrahedrite, chalcopyrite, electrum, argentite, polybasite, and pyragyrite. The supergene minerals are polybasite, chalcopyrite, covellite, and native silver; but undoubtedly hypogene silver minerals have yielded the large part of the production. Burton estimates that the deposit was formed at a minimum depth of 4,500 feet. In many ways this deposit recalls features of the epigenetic veins.

**Tetrahedrite-Galena-Siderite Veins (Wood River Type).**—The association of siderite gangue with galena and sphalerite and with a smaller quantity of tetrahedrite rich in silver (freibergite) is not uncommon in veins associated with intrusions of quartz monzonite, granodiorite, diorite, or lamprophyric dikes. The last dikes are the latest igneous rocks, and the ores appear to have been introduced shortly after their intrusion.

The deposits are usually veins in which the ores appear in part as filling, but largely as replacements of the country rock. Siderite is the characteristic gangue mineral, but calcite and intermediate carbonates of calcium, magnesium, and iron are often present; quartz enters into the gangue when the veins intersect the granitic rocks. Among the ore minerals tetrahedrite is the principal carrier of silver and is often intimately intergrown with galena. The galena is mostly coarse grained and also carries silver, while the sphalerite, with 4 or 5 per cent of iron, is

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[1] W. D. Burton, Ore deposition at the Premier Mine, B. C., *Econ. Geol.*, 21, 1926, pp. 586–604.

relatively poor in silver, but is sometimes recovered as a by-product in concentration.

Chalcopyrite is less abundant than tetrahedrite; pyrite is not conspicuous; while arsenopyrite and pyrrhotite occasionally appear, particularly in granitic country rock.

The ore-bodies often replace calcareous shales along the vein; but these shales appear to have been little altered, except for the introduction of metallic minerals and some siderite. In granular, feldspathic rocks close to the vein, sericite, carbonates, and a chlorite rich in iron develop in large amounts and sulphides are introduced. A complete replacement by sulphides is unusual. Sodium is almost wholly removed, but potassium fixed as sericite and calcium fixed as carbonate remain.

The structure of the ore is generally massive, and large bodies of galena are common. Smaller veins may show banded structure, but rarely comb structure. Sometimes a thin layer of quartz may be found along the wall, then a narrow comb of calcite, while the mass of the vein consists of massive galena, alternating with bands of sphalerite and containing intergrown tetrahedrite. The galena frequently shows phenomena of pressing, gliding, and recrystallization.

The width of the veins rarely exceeds a few feet, and part of this is usually crushed country rock. Their outcrops are inconspicuous.

The ore-shoots are markedly irregular and the cost of mining is therefore high. A marked deterioration may often be observed in depth; the large bodies of rich silver ore—aside from those affected by the surface enrichment—are found comparatively near the surface and the lower levels commonly show pinched veins or a predominance of sphalerite, pyrite, and quartz.

The upper, oxidized parts of the veins are usually enriched by secondary silver chloride, native silver, and pyrargyrite.

*Wood River, Idaho.*[1]—The silver-lead veins near Hailey, Idaho, on the Wood River, north of the Snake River lava plains, discovered in 1864, yielded a production of more than $25,000,000, most of it from 1880 to 1896. A small output is still maintained.

The district lies a few miles east of the eastern contact of the great granitic batholith of central Idaho, and the prevailing rocks are calcareous shale, quartzite, and limestone of Carboniferous age, compressed in northwestward-striking folds (Fig. 230). These sedimentary rocks are intruded by a minor batholith of diorite and quartz monzonite, following the general direction of the strata and from 2 to 3 miles wide. The deposits are fissure veins arranged in two parallel linked systems (Fig. 229) along

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[1] W. Lindgren, Wood River mining district, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, pp. 218–231.

J. B. Umpleby, L. G. Westgate, C. P. Ross, and D. F. Hewett, *Bull.* 814, *idem*, 1930.

the contacts of the batholith and in places cutting across the contacts into the granitic rock. Some of the veins follow lamprophyric dikes. Their strike generally cuts the stratification at an acute angle and their dip is prevailing 50° southwest. The outcrops are inconspicuous. Most of the veins are in calcareous shale.

The ore consists of galena, sphalerite, and tetrahedrite, with but little pyrite and chalcopyrite; the gangue is siderite, or intermediate calcium, iron, and magnesium carbonates, with a little quartz. The ore minerals have massive structure, sometimes roughly banded. Second-class ore consists of seams of carbonate gangue with small grains of galena. As there are no smelting works in the district, the ores and concentrates must be shipped; the shipments consist of high-grade ore containing 40 to 50 per cent of lead and 50 ounces of silver per ton; a little gold is usually present.

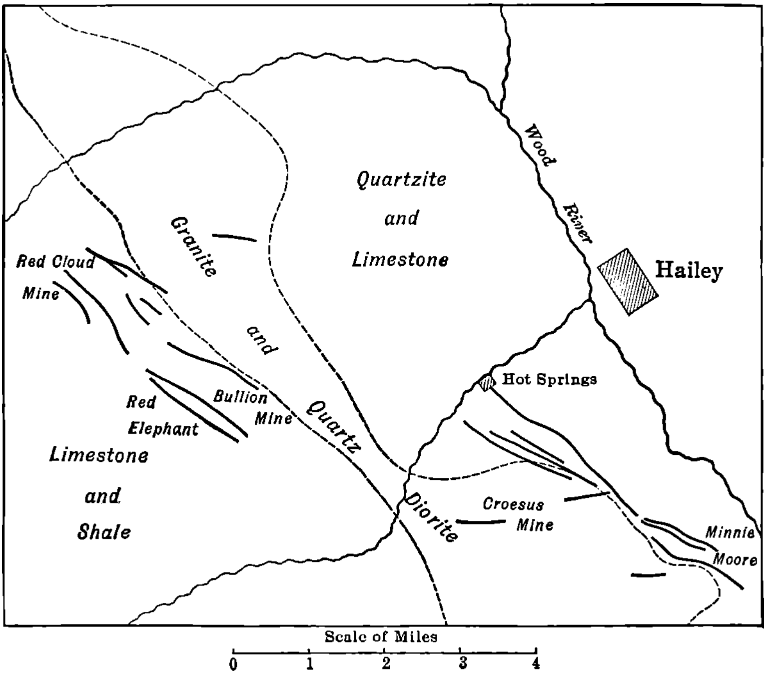

FIG. 229.—Vein system near contact of intrusive mass in Carboniferous sediments, Wood River district, Idaho.

Most of the veins are narrow, although they may in places widen out into bodies of galena many feet wide.

The ore-bodies are irregularly scattered along the veins and are for the most part replacements of calcareous shale by galena. Some of these replacement bodies lie obliquely across the strike of the vein and may be several hundred feet long and 10 to 30 feet wide. In a few places the developments have been carried far below the adit levels, but on the whole, the levels below a depth of 600 to 800 feet have shown fewer and poorer ore-bodies than the upper parts of the veins. There is little indication of sulphide enrichment, and the oxidized zone is shallow.

The post-mineral erosion has been very deep; the ore-bodies have no relation to the present surface. Deeper exploration might well encounter new ore. The Minnie Moore, one of the most productive veins, was cut off by intense post-mineral faulting (Fig. 230).

While some of the veins in the granitic rocks have the same character as those in the shales, others carry gold as the principal metal; they contain quartz, calcite, siderite, pyrrhotite, arsenopyrite, and chalcopyrite. The metasomatic alteration results in sericitization and carbonatization.

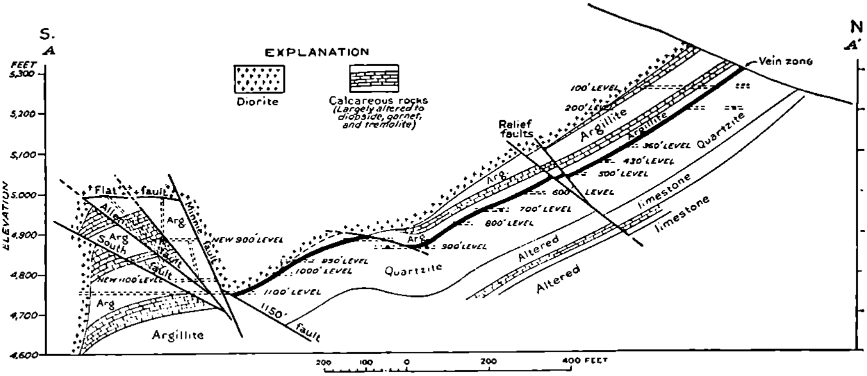

FIG. 230.—Geologic cross-section of the Minnie Moore mine, Wood River district, Idaho. Black represents vein zone. (*After D. F. Hewett, U. S. Geol. Survey.*)

*Slocan, British Columbia.*[1]—The veins of the Slocan district have yielded lead and silver to a value of about $45,000,000. They are mainly contained in the clay slates of the Slocan series, the age of which is possibly Carboniferous. The sedimentary rocks are intruded by granite (the Nelson batholith), quartz porphyry, and lamprophyric dikes. The fissure veins have a general northeast direction and high southeast or northwest dip. Where the veins intersect the igneous rocks quartz is the prevailing gangue mineral. In the sedimentary rocks the gangue is mainly siderite or manganosiderite. A specimen gave, for instance, 59 per cent $FeCO_3$, 27 per cent $MnCO_3$, 12 per cent $MgCO_3$, and 2 per cent $CaCO_3$. The ore minerals are sphalerite, galena, and tetrahedrite, rich in silver. Pyrite and chalcopyrite are fairly common; pyrrhotite is less

[1] W. L. Uglow, *Econ. Geol.*, 12, 1917, pp. 643–662.

A. M. Bateman, (Slocan), *idem*, 20, 1925, pp. 554–572.

M. E. Bancroft, Geol. Survey Canada, *Summ. Rept.*, 1919, pt. B, pp. 41–46; *idem*, 1917, pp. 28–41.

C. W. Drysdale, *idem*, 1916, pp. 56–57, with map.

abundant and is confined mainly to the vicinity of intrusive rocks. Native silver, of secondary origin, is present in the zone of oxidation.

Some gold-bearing veins occur together with the silver-lead veins and are apparently of the same age.

Many of the Slocan veins have proved less rich in depth than near the surface and in depth contain more siderite, quartz, and pyrite. The succession of minerals is siderite, sphalerite, galena, tetrahedrite.

Bateman finds that ore-bodies occur at very different elevations and have no definite relations to the original surface. Therefore, the local impoverishment in an ore-shoot is no proof of a zoning with reference to that surface.

**Galena-Siderite Veins.**—The galena-siderite veins form a small but important type, represented in the United States in the Coeur d'Alene district,[1] Idaho. In contrast to the Wood River type, these veins contain little tetrahedrite and are poor in silver; they yield about one-third to one-fifth of the lead production of the United States and in the aggregate also much silver. In 1930, the ore production was about 1,800,000 tons yielding 142,000 tons of lead, 33,000 tons of zinc, and 8,800,000 ounces of silver. Among the principal mines are the Bunker Hill and Sullivan, the Hecla, and the Morning. Two of the mines have been worked to a depth of 4,500 feet below the outcrops. The total production of the district since 1884 reaches the enormous value of over $700,000,000. As in other districts the production decreased materially after 1930.

The prevailing country rock is a fine-grained sericitic quartzite, referred to the Burke and Revett formations of the thick, folded and faulted pre-Cambrian Belt series of northern Idaho. The large faults[2] are not mineralized; the veins follow subordinate fissures of small throw. Two masses of monzonite of probable Cretaceous age, the larger not more than 3 miles in length, intrude the Belt series and cause some contact metamorphism by the development of biotite, garnet, and pyroxene in the quartzites. The last phase of the intrusion is represented by a few lamprophyric dikes. These are later than the mineralization and intersect the ore.

It is held probable that the intrusions of monzonite connect and widen below the surface. The ore deposits are composite veins or lodes, often of considerable thickness, formed partly by filling, but largely by replacement of the country rock along vertical shear zones with northwesterly

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[1] F. L. Ransome and F. C. Calkins, The geology and ore deposits of the Coeur d'Alene district, Idaho, *Prof. Paper* 62, U. S. Geol. Survey, 1908.

J. B. Umpleby and E. L. Jones, Jr., Geology and ore deposits of Shoshone County, Idaho, *Bull.* 732, U. S. Geol. Survey, 1923.

[2] One of these, the Osborne fault, has, according to Hershey, a horizontal throw of 12 miles.

trend. The longest of the veins is the Bunker Hill, which is traceable for 7,000 feet.

The ore-shoots are large, and many of them are roughly vertical; some have been followed to a depth of over 4,500 feet. At the Bunker Hill and Sullivan mine[1] (Fig. 231), the ore-bodies do not always follow the main wall, which dips 38° SSW., but may lie in the shattered country rock within 250 feet above it. The width of the ore is in places as much as 40 feet, 9 feet being the average in some of the larger mines.

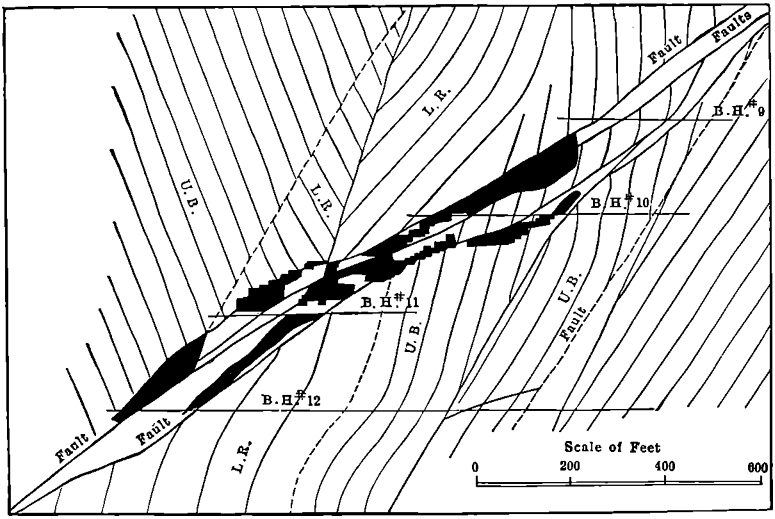

Fig. 231.—Vertical cross-section of part of Bunker Hill and Sullivan vein, showing relation of ore-bodies (black) to structure. *L.R.*, Lower Revett formation; *U.B.*, Upper Burke formation. Lines between faults indicate intersection of stratification planes with vertical plane. (*After Oscar H. Hershey.*)

Galena, with some pyrite and sphalerite, and in places a little tetrahedrite rich in silver are the principal ore minerals. Chalcopyrite and probably secondary boulangerite are present in small amounts; in some mines pyrrhotite takes the place of pyrite. Siderite and quartz are the predominant gangue minerals; barite, calcite, and dolomite are rare. Some of the siderite contains several per cent of manganese.

The ores are in large part formed by replacement of sericitic quartzite along the tight shear planes of the lodes. The siderite develops first, replacing both sericitic cement and quartz grains in the quartzite. Rhombohedrons of siderite may often be seen cutting across the clastic grains. The galena is later than the siderite and replaces that mineral. Replacement of quartzite by galena is shown in Fig. 65. Some ore-bodies consist of almost massive galena, but the ordinary ore is an

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[1] Mining methods at the Bunker Hill and Sullivan mine, *Information Circ.* 6407, U. S. Bur. Mines, 1931.

aggregate of siderite and galena, which must be concentrated. The bulk of the ores range from 3 to 14 per cent lead and from 2.5 to 6 ounces of silver to the ton. These are concentrated to a product containing about 50 per cent lead. The lowest grade which can be worked at present carries 5 per cent lead and 3 ounces of silver to the ton. A few of the

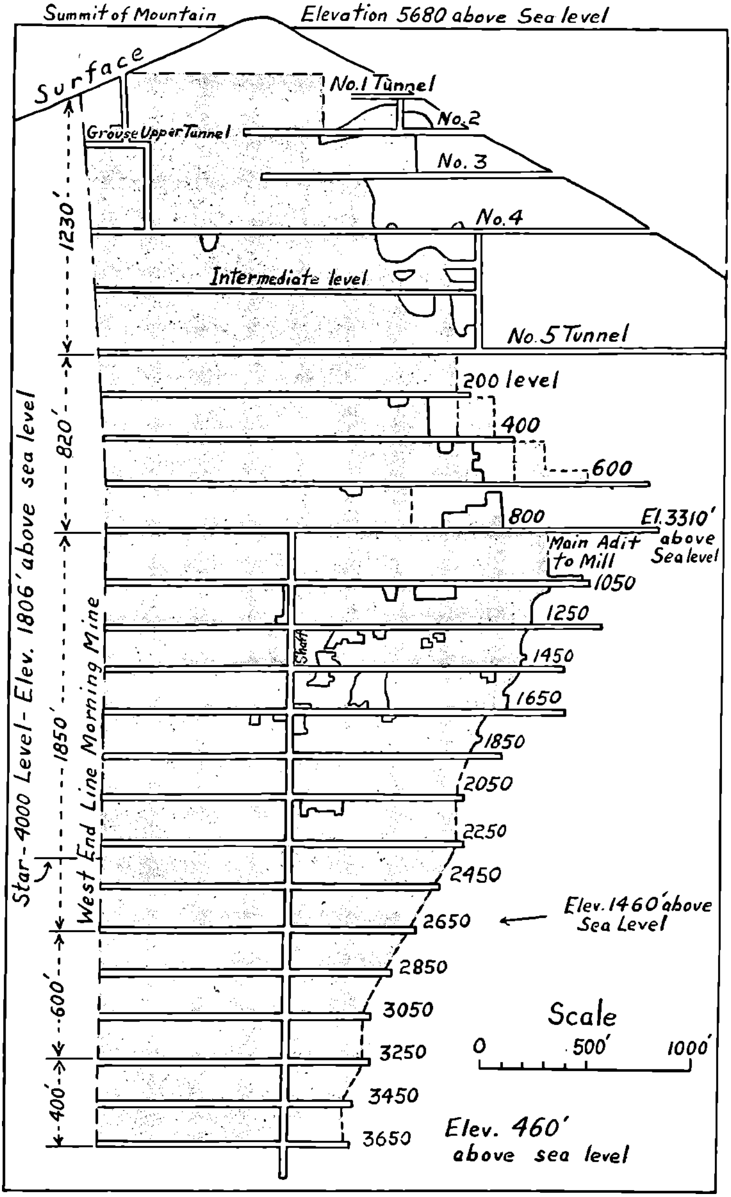

FIG. 232.—Longitudinal section showing ore-shoot in the Morning mine, Idaho.

mines, like the Bunker Hill, the Interstate-Callahan, and the Morning, yield much sphalerite.

The metasomatic action, indicated by the presence of siderite in the quartzite, often spreads for 100 feet or more beyond the ore. There are some indications of change of ores in depth; Ransome finds that in the lower levels of many mines pyrite, pyrrhotite, and sphalerite become more abundant.

The great shoot of the Morning mine is about 1,800 feet long and has been followed to a depth of 5,100 feet below the outcrops with undiminished width and tenor (Fig. 232). In 1925, the ore averaged 9.5 per cent lead and 5.3 per cent zinc with 4.3 ounces of silver per ton. Where exposed on the 3,050 level the ore shows little or no difference from that of the upper levels. The ore was evidently deposited in the cool country rock several miles from the intrusive rock. As the outcrop only represents the Eocene erosion surface, it is evident that the total depth of the ore-shoot must considerably exceed the figure of 5,100 feet, just mentioned. Actually the shoot is considerably longer than shown in Fig. 232, for it extends westward into the adjoining ground owned by the Hecla Company. Another Hecla shoot and the Bunker Hill and Sullivan ore-body are of similar dimensions and show no deterioration of the ore. The latter mine has now developed its shoot on the seventeenth level below the tunnel, a total vertical depth of 4,500 feet below the outcrop.

On the other hand, several mines—The Hercules, Mace, Tiger-Poorman, and Helena-Frisco—have been impoverished at considerably higher elevations, probably in part because the deposits entered the unfavorable slates of the Pritchard formation.

The zone of oxidation is irregular and shallow.

Ransome, Umpleby, and Jones trace a genetic connection between the ore deposits and the intrusive monzonite. Contact-metamorphic ores in irregular bodies are found in two mines close to the monzonite, in the contact zone. These ores contain galena, sphalerite, pyrite, pyrrhotite, chalcopyrite, and magnetite, with a gangue of garnet, biotite, and diopside. In veins near the intrusive mass, pyrrhotite and magnetite, as well as garnet, biotite, and tourmaline are found. Siderite occurs only outside of the contact zone. In the Wardner mines, which are several miles from the contact, siderite is most plentiful. The deposits were formed within the epoch of granitic intrusions, as shown by the occasional intersection of ore-bodies by lamprophyric dikes; and since the time of ore formation, erosion has probably removed several thousand feet of rock.

The facts briefly set forth are of highest importance and serve to connect the high-temperature deposits with those of intermediate conditions.

O. H. Hershey believes that the metals were originally disseminated in the Belt sediments and later concentrated into ore-bodies by hot waters ascending on thrust faults. He also holds that the contact-metamorphic deposit of the Success mine has been formed prior to the monzonite intrusion and was invaded and metamorphosed by it.[1]

**Lead-Silver Veins with Calcite, Siderite, and Barite.**—Veins containing galena and sphalerite with a gangue of calcite, siderite, or barite are abundant in many mining regions and are frequently connected with replacement deposits in limestone. In many places they have a distinct connection with intrusive rocks and were formed shortly after the irruption.

*Lead-Silver Veins of Clausthal.*[2]—The mines of Clausthal, in the Harz Mountains of Germany, which have been in operation since the thirteenth century and still maintain a moderate production, are working on a vein system which intersects a folded complex of Devonian and Carboniferous sedimentary beds, the prevailing rocks being clay slate and graywacke. The general strike of the veins is east-west, and the dip is steep. The numerous veins extend over an area 15 miles in length and 5 miles in width; they are in general composite veins, or lodes, and the important fissures are also faults of considerable throw. Mining operations in this district have attained a depth of 3,000 feet.

The ores contain chiefly galena and sphalerite, with some marcasite, pyrite, chalcopyrite, and tetrahedrite. Arsenical minerals are generally absent. In one group of veins calcite and quartz predominate; in another barite and siderite. Most of the galena contains about 0.05 per cent silver. Symmetrical banding is exceptional, the normal ore having an irregularly massive structure. The tendency appears to be toward an increase in the percentage of zinc in depth. Within the lodes the clay slates are altered by mechanical and chemical processes. The change in composition is slight[3] and appears to consist largely of an increase in sericite at the expense of an original chloritic mineral in the clay slate.

There are no intrusive rocks in the immediate vicinity aside from a dike of kersantite, which is faulted by the vein fissures, and the mass of "Brocken" granite in the eastern part of the district. A genetic connection of these intrusives with the veins seem probable, but cannot be regarded as proved. The age appears to be late Paleozoic. The mineral

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[1] J. B. Umpleby, Genesis of the Success zinc-lead deposits, *Econ. Geol.*, 12, 1917, pp. 138–153; O. H. Hershey, *idem*, pp. 348–558.

[2] A. von Groddeck, Ueber die Erzgänge des Oberharzes, *Zeitschr.* deutsch. geol. Gesell., 1866, pp. 693–776.

B. Baumgürtel, Oberharzer Gangbilder, Leipzig, 1907, pp. 23.

F. Dahlgrün, Zur Klassification der Jung-Palaeozoischen Erzgänge des Harzes, *Neues Jahrbuch*, Ref. III (1), 1930, pp. 1–2.

[3] A. von Groddeck, *Jahrb.* Preuss. geol. Landesanstalt, 1885, pp. 1–52.

association would indicate deposition at less depth or at lower temperature than the veins of the Coeur d'Alene district, for instance.

*Lead-Silver Veins of Przibram, Bohemia.*[1]—The mines of Przibram, which have been worked for several hundred years and still maintain a small output, are situated 40 miles south-southwest of Prague, in the "Silurian syncline," well known in the early history of geology. The predominating rocks are Cambrian (?) graywacke, and clay slate, a folded and faulted complex intruded by a stock of diorite. Dikes of diabase are numerous and are followed by the veins; dikes of diorite and kersantite are also present. The intrusive diorite produced a decided contact metamorphism in the Paleozoic sediments.

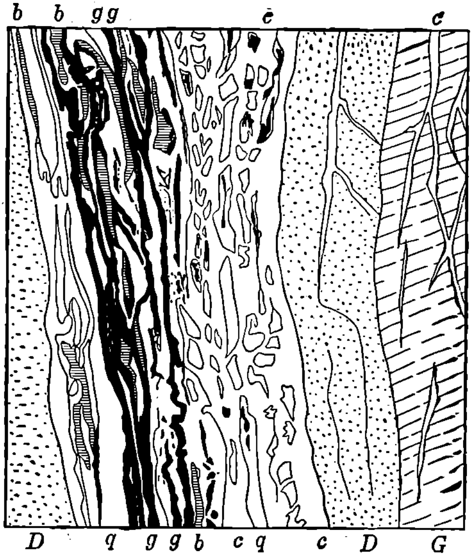

FIG. 233.—Section of the Adalbert vein at Przibram, Bohemia. *G*, graywacke; *D*, diorite; *q*, quartz; *c*, calcite; *g*, galena; *b*, sphalerite. (*After J. Zadrazil and J. Schmidt.*)

The veins have a steep dip and have been followed down to a depth of 3,773 feet; about forty of these veins have been worked, and are contained within a narrow area 4 or 5 miles in length. The width of the veins attains 25 feet, but averages much less. Figure 233 gives an idea of their structure. The ore minerals consist of galena and sphalerite with some pyrite and chalcopyrite and occasionally many other minerals, like arsenopyrite, stibnite, uraninite, cobalt and nickel minerals, wurtzite, and millerite. Rich silver minerals, like argentite and pyrargyrite, as

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[1] J. Schmidt, Bilder von den Erzlagerstätten von Przibram. Published by Austrian Agricult. Dept., Vienna, 1887.

F. Posepny, *Archiv für prakt. Geol.*, 2, Freiberg, 1895, pp. 609–745.

A. Hofman and F. Slavik, Ueber Dürrerze von Przibram, *Bull. internat.* 15, Acad. Sci. de Bohême, 1910.

well as native silver were plentiful in the oxidized zone. The galena is the carrier of silver and contains about 0.5 per cent of this metal. Among gangue minerals calcite, siderite, and quartz predominate, but barite and ankerite are also known. The structure is in part banded and drusy.

The quartz and sphalerite appear to increase in depth and the ores become "dry." These dry ores contain about 50 per cent quartz, 17 per cent siderite, 17 per cent galena, 0.26 per cent silver, also primary boulangerite, tetrahedrite, pyrargyrite, diaphorite, specularite, chlorite, and cassiterite.

The deep workings are practically dry, but there existed formerly a rich zone of oxidation descending, in spite of a high present water level, to depths of 200 to 900 feet.

The genetic connection of the veins with the intrusive diorite and its satellites of diabasic and lamprophyric dikes appears to be clearly indicated.

*Other Regions.*—The Bawdwin mine,[1] long worked by the Chinese, is situated in northern Burma near the Chinese frontier. During the last 20 years it has yielded a great production in silver, lead, and zinc. It is a replacement deposit along a strong shear zone in rhyolite tuffs, which are covered and underlain by sediments. The ore consists of argentiferous galena, and sphalerite in extremely fine intergrowth, with silicification and gangue of calcite and siderite.

The Buchans deposit in Newfoundland[2] was discovered by electrical methods of prospecting and has developed into a very important mine. The ores occur in a thick early Paleozoic series of andesite, rhyolite, and basalt with large amounts of rhyolite tuff. It is intruded by rhyolite porphyry and probably by granite. The ore replaces the tuff and consists of a fine-grained mass of early barite followed by sphalerite, chalcopyrite, and galena. The largest body, irregular in shape but following stratification to some extent, is estimated to contain 5,750,000 tons of ore with the following tenor: Gold 0.033 ounce per ton, silver 3.75 ounces per ton, copper 1.4 per cent, lead 7.65 per cent, zinc 17.85 per cent. The ore-bodies lie on the flank of an anticline. In part the ore shows colloform texture, and the deposit suggests epithermal affinities.

In the eastern Alps of Austria,[3] there is a widespread but rarely important lead-zinc type with galena, sphalerite, carbonates, fluorite, and barite, best known from the Bleiberg and Raibl occurrences. Their age ranges from Cretaceous to Pliocene, and they are connected with deep-seated intrusions. They are generally metasomatic, because the

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[1] M. H. Loveman, *Trans.*, Am. Inst. Min. Eng., 55, 1917, pp. 170–194.

[2] W. H. Newhouse, *Econ. Geol.*, 26, 1931, pp. 399–414.

[3] A. Tornquist, *Jahrbuch*, Austria geol. Bundesanstalt, 81, 1931, pp. 143–175.

constant northward pressure prevented open cavities. The localization depends on impermeable covers.

In the Pennine region (north of England) lead mining was formerly important. The deposits, containing galena, barite, fluorite, and carbonates, are of immediate post-Carboniferous age and are connected with diabase intrusions (Whin Sill) in Carboniferous sediments.

**Pyritic Galena-Quartz Veins.**—In the vicinity of granitic and dioritic intrusions a certain type of lead-bearing veins is especially common, distinguished by pyrite, galena, dark sphalerite, and some chalcopyrite, with subordinate arsenopyrite, in a gangue of quartz, with a small amount of calcite or dolomite.

*Freiberg, Saxony.*—The type just mentioned corresponds closely to the "Kiesige Bleiformation" of Freiberg,[1] there represented by numerous veins of considerable persistency contained in a flat dome of biotite gneiss. K. Dalmer and other geologists have pointed out their probable genetic connection with the intrusive (Carboniferous) granites of the Erzgebirge and their well-established relationship to the tin veins which are situated closer to or within the intrusives. The Freiberg veins of this type are narrow, being seldom 3 feet wide, and have been mined to a depth of 2,100 feet. The pyrite, arsenopyrite, and sphalerite are poor in silver, but the galena contains 0.1 to 0.2 per cent of this metal.

The vein structure is irregularly massive, without marked banding or crustification.

The ores are of low grade, and after a period of activity extending over nearly 750 years the mines are now practically closed.

The silver-lead deposits of Freiberg comprise a complicated system of fissure veins of different types and ages, which have been carefully studied by such men as A. G. Werner (1791), A. von Weissenbach (1836), J. C. Freiesleben (1843), F. C. von Beust (1840), B. von Cotta (1861), and H. Müller (1849–1901).

The veins are classified as follows:

(1) Older Veins.—Pyritic lead formation: Comprises an early high-temperature pyrite, arsenopyrite, cassiterite stage, followed by a quartz, pyrite, sphalerite, chalcopyrite (magnetite, hematite), galena, chalcocite stage. After fracturing, a third stage followed with marcasite, pyrite, tetrahedrite, ruby silver, argentite, and ankerite. After renewed fracturing, came the last stage with galena and rhodochrosite. Late iron oxides indicate temporary increase in temperature. Noble[2] quartz formation: Fine-grained quartz with argentite, pyrargyrite, native silver, pyrite, and arsenopyrite.

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[1] Herman Müller, Die Erzgänge des Freiberger Bergrevieres, *Erläuterungen zur geol. Special-Karte Sachsens*, Leipzig, 1901, p. 350.

O. Oelsner, Beiträge zur Kenntniss der kiesigen Bleierzformation Freibergs, *Jahrbuch.* Berg-u. Hüttenw. Sachsen, 104, 1930, pp. *A*3–50.

[2] The word "edel," or noble, refers to the high-grade silver ores.

Noble lead formation: Quartz, ankerite, rhodochrosite, sphalerite, galena, pyrite, tetrahedrite, ruby silver, and polybasite.

(2) Younger Veins.—Barytic lead formation: Barite, fluorite, quartz, calcite, galena (poor in silver), chalcopyrite, tetrahedrite, sphalerite. These veins are often of considerable width.

The barite veins are distinctly later and show beautifully banded and drusy structure. Their age is probably Tertiary, and they are possibly connected with basaltic eruptions. They sometimes carry nickel and cobalt and may perhaps be correlated with the veins of Annaberg.

The older group appears to be genetically connected with the granitic intrusions of Carboniferous age, or perhaps also with the Permian and Carboniferous porphyries (intrusive and effusive) of the same region. The "noble quartz formation" alone is intersected by dikes of quartz porphyry, while the other veins appear to be later than the porphyry. The granite stocks of the region are intersected by veins similar to those of Freiberg; but no granite occurs in the Freiberg district. In parts of the district dikes of kersantite and minette are plentiful, and the veins are later than these dikes.

Between the various members of the older group many transitions exist, and it seems justifiable to regard them as genetically connected with the granitic eruptions of Carboniferous age and as formed shortly after the last lamprophyric dikes of that parent magma had been intruded. The mineral association of the "noble quartz formation" and the "noble lead formation," with apparently primary argentite and pyrargyrite continuing to the greatest depth reached, far beyond the zone of oxidation, seems to suggest that these veins have been formed at relatively low temperature. They do not correspond to the types usually associated with intrusive masses.

The ore-shoots of the Freiberg veins are irregular; the richest parts were often at intersections of fissures (Fig. 86). The oxidized ores worked in the early history of the mines were rich in argentite and native silver.

*Pyritic Galena-Quartz Veins in the United States.*—Near intrusive areas in the central and eastern Cordilleran states are many veins of the Freiberg type, just described, although they ordinarily also carry gold together with silver. Few of them are, however, of the first importance. More common, perhaps, are veins which carry mainly massive galena and sphalerite associated with but little pyrite, or veins in which the pyrite entirely predominates. Examples of this kind are given in the description of the Leadville region (p. 590).

Excellent examples of the type, described by F. C. Schrader, occur near Kingman,[1] in northwestern Arizona, in the Wallapai mining district. The upper levels yielded rich silver ores, but of late years the silver

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[1] F. C. Schrader, Mineral deposits of the Cerbat Range, Black Mountains, etc., *Bull.* 397, U. S. Geol. Survey, 1909.

E. S. Bastin, Studies of polished sections, *Bull.* 750, *idem*, 1924, pp. 17–39.

production has declined as the leaner primary sulphides were encountered, and in the ores now extracted sphalerite is the most valuable constituent. The greatest depth attained is about 1,400 feet. The rocks are pre-Cambrian granite, gneiss, and schist intruded by granite porphyry, probably of Mesozoic age, and by a great number of lamprophyric dikes of minette and vogesite, which in part are followed by the veins.

The deposits are well-defined fissure veins with steep dip, forming conjugated systems with northwesterly strike; they are straight and have well-defined walls, and some of them are traceable for considerable distances. The gangue is quartz, in places shattered and cemented by a later generation of calcite, occasionally also siderite. Among the primary sulphides are pyrite, arsenopyrite, galena, sphalerite, chalcopyrite, tennantite, proustite, and pearceite. The ore may contain $10 in gold and silver, 8 per cent lead, and 5 to 16 per cent zinc. It is in part shipped crude, in part concentrated.

The structure is irregularly massive, in places with rough banding by arrangement of the sulphides. The veins are narrow, though in some places ore-bodies 20 feet wide have been worked. The pay-shoots are irregular, but often coincide with intersections of veins. The water level is from 100 to 400 feet below the surface, and above it were rich oxidized lead ores, horn silver, native silver, argentite, and ruby silver. The decrease of galena and increase of chalcopyrite noted in the lower levels suggest a gradual change in the primary filling.

The ore is mainly deposited by filling of cavities; the wall rocks contain little ore but are sericitized and filled with pyrite close to the veins.

### THE SILVER TIN VEINS OF BOLIVIA[1]

This important and peculiar type is confined to the southern part of the Bolivian tin belt, which traverses the state along the eastern range of

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[1] A. W. Stelzner, Die Silber-Zinnlagerstätten Bolivias, *Zeitschr.* deutsch. geol. Gesell., 49, 1897, pp. 51–142.

W. Myron Davy, Ore deposition in the Bolivian tin-silver veins, *Econ. Geol.*, 15, 1920, pp. 463–496.

B. L. Miller and J. T. Singewald, Jr., Mineral deposits of South America, New York, 1919.

M. J. Buerger and J. L. Maury, Tin ores of Chocaya, Bolivia, *Econ. Geol.*, 22, 1927, pp. 1–13.

W. Lindgren and J. G. Creveling, The ores of Potosi, *idem*, 23, 1928, pp. 233–262

W. Lindgren and A. C. Abbott, The silver-tin deposits of Oruro, *idem*, 26, 1931, pp. 453–479.

F. Ahlfeld, *Zeitschr. prakt. Geol.*, 37, 1929, pp. 216–219 (Carguaicollo); *idem*, 39,1931, pp. 33–38 (Vila Apacheta).

For abstracts of later papers on Potosi, Oruro, and Chocaya by S. Jaskolski and R. Kazlowski, see *Annot. Bibliogr.*, *Econ. Geol.*, 5 (2), no. 412, 1932; 6 (2), no. 388, 1933; 8 (1), no. 423, 1935.

the Andes, known as the Cordillera Real. Genetically it is intimately connected with the hypothermal tin veins of the same country (p. 655). Broadly speaking the veins are found on a late Tertiary peneplain of Paleozoic slates, in or near small intrusions of rocks allied to quartz-monzonite porphyry. More rarely (Chocaya) they occur in Tertiary acidic flow rocks. They are well-defined replacement veins, with more or less filling and are usually frozen to the walls. The country rock is pyritized and sericitized near the veins. Pyrite greatly predominates in the ores. At Potosi, many veins have been worked over a vertical interval of nearly 2,000 feet. The ores appear to be poorer in depth. The present hypogene ores contain about 3 per cent tin and 12 ounces of silver per ton. The gangue is quartz with some alunite and the succession is pyrite (oldest), arsenopyrite, cassiterite, sphalerite, chalcopyrite and stannite, tetrahedrite, andorite, ruby silver, and jamesonite. The oxidized zone was extremely rich; and the district is reported to have produced 30,000 tons of silver since 1545, more than any other district in the world.

At Oruro similar veins have been worked. They have a steep dip and average 70 centimeters in width. The veins in the Socavon de la Virgen have been followed to a depth of 300 meters below the tunnel level. They still contain rich silver minerals as later streaks in prevailing pyrite. The average content is about 1 per cent tin and 12 ounces of silver to the ton. The Itos vein, belonging to the same vein system, has been worked over a vertical range of 580 meters. Below the oxidized zone a strong silver enrichment is reported. The veins carry, in approximate order of deposition, quartz, pyrite and cassiterite, stannite, tetrahedrite, andorite, zinkenite, and jamesonite.

Another type, represented by mines at Huanuni, contains primary arsenopyrite, sphalerite, and galena, with some tin mineral, probably cylindrite. This poor protore is strongly enriched by franckeite and a silver-bearing jamesonite (owyheeite), whether by supergene or hypogene concentration is not quite certain.

At Carguaicollo are many crustified veins, which in their upper levels were rich in franckeite, cylindrite, teallite, and zinc teallite.

Still another type, represented by the veins at Chocaya, is of an epithermal type. Contained in acidic flow rocks, they are largely formed by filling and are rich in silver. The succession is pyrite (earliest); arsenopyrite; cassiterite; chalcopyrite; stannite; sphalerite; galena; tetrahedrite; matildite;[1] jamesonite.

At Vila Apacheta, in Cretaceous Puca sandstone, the veins contain cassiterite and tetrahedrite and have pronounced colloform structure. They must likewise be classed as epithermal.

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[1] Perhaps rather aramayoite ($Ag_2S.(Sb,Bi)_2S_3$). L. J. Spencer, *Mineralog. Mag.*, 21, 1926, pp. 156–162.

## THE SILVER-LEAD REPLACEMENT DEPOSITS IN LIMESTONE

**General Features.**—Limestones, dolomites, and calcareous shales are easily soluble. Waters circulating above the water level along stratification planes, joints, fissures, or zones of brecciation will produce caves and open passages. Below the water level more slowly moving solutions often replace limestone by dolomite or cherty or jasperoid silica. If the solutions carry metallic sulphides these may be precipitated, and by a simultaneous operation the carbonate goes into solution while a corresponding volume of sulphides takes its place (Fig. 236). Some of these replacement deposits that have no certain genetic connection with igneous rocks have been described above (p. 423).

In districts where metallization is caused by igneous activity, the limestone is often replaced close to the contact by sulphides, particularly copper sulphides, associated with high-temperature minerals; these deposits are described in Chapter XXVII. Frequently, however, replacement by sulphides, is also found at greater distances from the igneous rock; but the solutions which caused the replacement, while probably derived in part from the magma, had a lower temperature and therefore no high-temperature minerals could form. Such deposits, which contain mainly lead, zinc, and silver, may form relatively close to the surface, but are more common in the vicinity of intrusive rocks now exposed by erosion. The process is, therefore, favored by higher temperature and pressure.

For the development of replacement deposits, pathways that can be followed by the solutions are necessary. Joints and seams may provide them, but more commonly the fissures which were formed during or after the intrusion guide the solutions to the limestone. When the waters have entered a fissure the processes of replacement begin immediately, but the products of interchange are not confined to this fracture. On the contrary, they spread in all directions, guided by minor structural planes, and replacement deposits in limestone are therefore characteristically irregular. It often happens that the original fissure may be difficult to discover, though genetically it is the key to the extent and the continuation of the deposit. The mining of such deposits demands thorough knowledge of the geological structure.

There are a great number of such deposits in the Cordilleran region of the Americas. Many of them are small and are soon exhausted, while others are among the great ore deposits of the world. The districts of Aspen and Leadville, Colorado; Eureka, Nevada; Lake Valley, New Mexico; Elkhorn, Montana; Park City and Tintic, Utah; and Santa Eulalia, Mexico, may serve as examples.

At some places these silver-lead deposits follow dikes or intrusive sheets. At other places they are dependent upon impervious overlying beds, like shale. The latter condition is exceedingly common (Fig. 84) and indicates that the solutions were ascending and that deposition followed the ponding or stagnation of the solution or at least was favored by less rapid movement. Long pipes and chimneys are common forms and may result from replacement along intersections of fissures or along intersections of fissures with favorable beds. Sometimes their location is difficult to account for, and the solutions appear to have bored their way through the limestone for long distances. The replacements are generally wider than the stope maps indicate, by reason of more or less extensive silicification. Sections of two smaller replacement deposits are shown in Figs. 234 and 235.

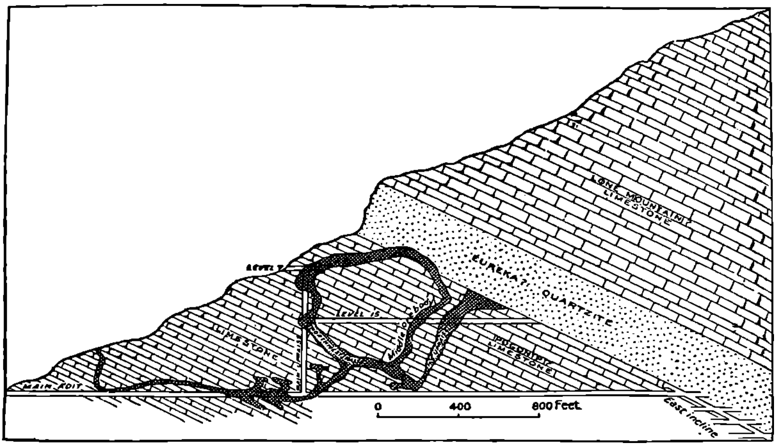

FIG. 234.—Irregular replacement deposit in the Garrison mine, Cortez, Nevada. Ore consists of galena, sphalerite, pyrite, stromeyerite, etc., and their oxidation products. (*After W. H. Emmons, U. S. Geol. Survey.*)

The primary minerals of these replacement deposits are comparatively few and simple. Deep oxidation is, however, common in limestone, and descending waters may effect many changes and develop a great number of rare oxidized minerals in the oxidized zone, while complex secondary sulphides may form in the lower parts of the deposit. The gangue minerals are few: Dolomite is often present as a coarse aggregate, and at many places the process of replacement was begun by a dolomitization of the limestone. Dense, cherty quartz, in part formed by colloidal replacement, is exceedingly common, much more so than coarser crystalline quartz. In accordance with the suggestion of Spurr, this siliceous gangue is called jasperoid, though this term is really a misnomer, for the rocks are gray rather than red or brown. Other gangue minerals are calcite, barite, sometimes fluorite, various carbonates allied to ankerite, and more rarely rhodochrosite. The common primary ore minerals are pyrite, galena, sphalerite, chalcopyrite, and more rarely arsenopyrite.

Tetrahedrite, tennantite, enargite, bornite, bismuthinite, and stibnite are of local importance. Argentite, ruby silver, stephanite, polybasite, and native silver as well as various sulphantimonides of lead may be hypogene or supergene. Gold is sometimes present as a primary mineral, but the ores carry ordinarily much more silver than gold. Galena is most common and is often rich in silver. The silver content of galena is usually caused by small disseminated grains of argentite which may result from unmixing of a solid solution of $Ag_2S$ in PbS. In many so-called lead deposits the lead really predominates only in the oxidized zone, while the primary (hypogene) ore carries far more pyrite and sphalerite than galena. Such are the relations at Leadville, for instance.

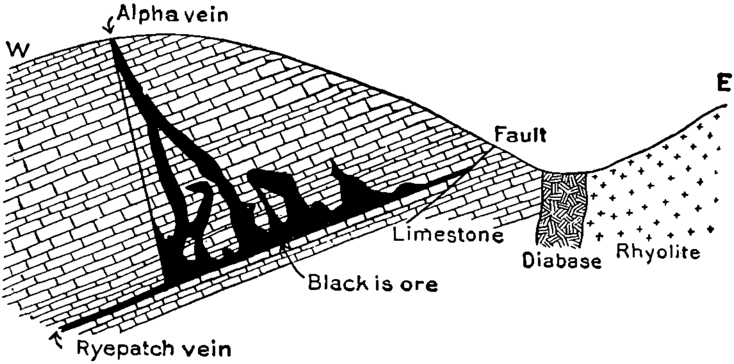

FIG. 235.—Section of Ryepatch mine, Unionville, Nevada. Ore-body between faults 250 feet wide; consists of calcite, quartz, pyrite, galena, sphalerite, tetrahedrite, etc. (*After F. L. Ransome, U. S. Geol. Survey.*)

The ore minerals replacing limestone are usually coarse grained, while the replacement of limestone by silica yields rocks with fine grain. Crustified or drusy structures are common only in the oxidized parts of these deposits. Before the importance of replacement as a geological process was recognized, many of these ores were considered as fillings of limestone caves. Some of the deposits consist of massive sulphides, while in others, presumably formed at lower temperature, the gangue may prevail.

Replacement deposits are not confined to calcareous rocks. They occur also in quartzite, shale, and igneous rocks, but they are certainly more common in carbonate rocks. Very hot solutions may replace any rock, but most of the deposits described in this chapter were probably laid down by solutions having a temperature of less than 250° C.; and under such circumstances limestone would be replaced while other rocks would be little affected. Siliceous rocks are more easily replaced than aluminous material; it is evidently difficult to carry away large amounts of alumina even at high temperatures and ores in aluminous rocks contain much residual material.

**Park City, Utah.**[1]—The Park City district lies near the summit of the Wasatch Range. Since 1870, it has yielded 195,189,175 ounces of silver and 948,410 tons of lead; lately, also much copper and zinc, and it still remains one of the most important districts of the United States. Its ores are in part shipped as mined, but much is also concentrated, the total output of ore for 1929 being 526,681 tons. The concentrating ore contains from 6 to 8 per cent lead, 6 to 8 per cent zinc, 6 to 10 per cent iron, and 9 ounces of silver per ton; also some gold and copper. The deepest shafts attain 1,500 and 2,000 feet.

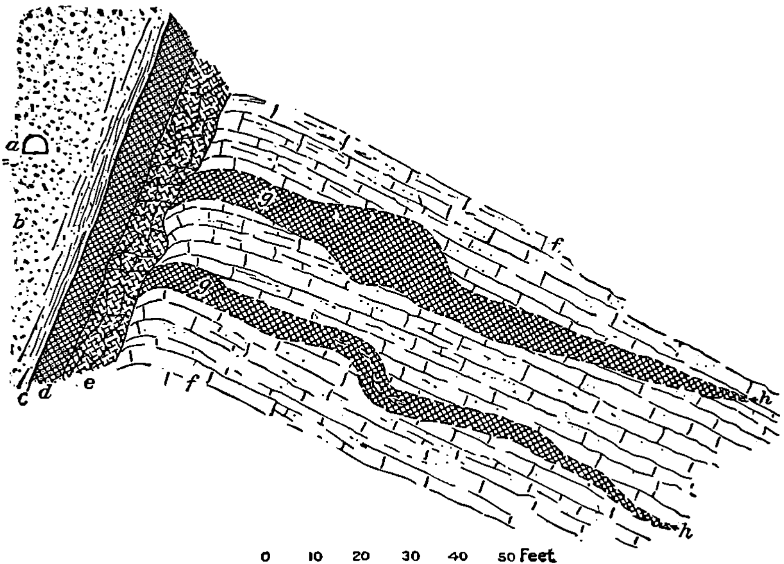

FIG. 236.—Vertical section of rich lead ore occurring in veins and in replacement deposits, Kearns-Keith mine, Park City, Utah. *a*, tunnel; *b*, diorite porphyry, sheeted and pyritic; *c*, hanging-wall fissure; *d*, lead ore in siliceous gangue; *e*, breccia zone with ore fragments; *f*, marmorized limestone. Thaynes formation; *g*, *h*, banded replacement ore, in part oxidized. (*After J. M. Boutwell, U. S. Geol. Survey.*)

A huge anticline of late Carboniferous, Permian, and Triassic sediments, mainly limestone, quartzite, and shale, the total thickness of beds exceeding 8,000 feet, is intruded by laccolithic stocks of diorite porphyry, probably of late Cretaceous age, which have caused contact metamorphism in the adjoining limestone and shales.

The ores occur as lode deposits in wide and strong master zones, and closely associated bedded deposits; they lie in two parallel zones extending northeastward. The bedded deposits, mainly in limestone, have been mined to a depth of 900 feet; the lode deposits continue to the greatest depths attained and are here poorer but wider. The lode deposits inter-

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[1] J. M. Boutwell, *Prof. Paper* 77, U. S. Geol. Survey, 1912.
B. S. Butler, *Prof. Paper* 111, *idem*, 1920, pp. 294–318.

sect the sediments and the porphyry as well, have a steep dip, and often lie in quartzite or between limestone and quartzite. The ores are in part deposited by filling of seams in shattered ground, in part by replacement. The stopes are as much as 30 feet in width.

The bedded deposits are massive sulphides replacing limestone strata in two of the calcareous formations and are from a few inches to 10 feet thick, 500 to 800 feet in the direction of the strike and, at most, 200 feet along the dip. The relation between the two types is shown in Fig. 236. The layers of the bedded ore are made up of ore and gangue minerals in granular texture exactly like that of the original limestone. There is evidence of two epochs of deposition, for some of the bedded areas near the porphyry contacts contain garnet with calcite as gangue, while the lode deposits and the bedded ores associated with them are free from garnet and were formed after the cooling of the porphyry.

The bedded ores often show a granular intergrowth of sulphides with the carbonates of the limestone, while the walls are strongly silicified. The ore minerals are galena, sphalerite, tetrahedrite, and a little chalcopyrite. Jamesonite and bournonite have been noted. The gangue is mainly quartz and jasperoid; fluorite, calcite, and rhodonite occur locally. Sericitization is noted where the lodes intersect porphyry. The richest ore was formed in the bedded deposits; the ore in depth is of leaner grade, but carries more copper and zinc.

The Park City mines are very wet and the water level is high. The oxidation is deep and partial oxidation has been noted to a depth of 1,200 feet suggesting changes in the water level. The oxidized zone contained apparently but little native silver and cerargyrite.

**Tintic, Utah.**[1]—The replacement deposits of the Tintic district, situated in a desert range 70 miles south of Salt Lake City, exemplify another type, which has been so modified by oxidation that the original character of the ore is sometimes difficult to interpret. Folded Paleozoic limestones are intruded by a small monzonite stock that formed the core of a volcano of early Tertiary age, the surface flows of which are largely eroded. Possibly these deposits should rather be referred to the epithermal group. A number of narrow fissures traverse both monzonite and limestone; in the former the deposits are pyritic veins with sericitized walls, while in the limestone the inconspicuous fractures widen locally into large or small ore-

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[1] G. W. Tower, Jr., and G. O. Smith, Geology and mining industry of the Tintic district, Utah, *Nineteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, pp. 603–785.

W. Lindgren, G. F. Loughlin, and V. C. Heikes, *Prof. Paper* 107, *idem*, 1919.

W. Lindgren, Processes of mineralization and enrichment in the Tintic district, *Econ. Geol.*, 10, 1915, pp. 225–240.

G. W. Crane, Geology of the ore deposits of the Tintic district, *Trans.*, Am. Inst. Min. Eng., 54, 1917, pp. 342–355.

C. A. Dobbel, Deep-hole prospecting at the Chief Consolidated mines, *Trans.* Am. Inst. Min. Met. Eng., 72, 1925, pp. 677–689.

bodies characterized as chambers, chimneys, pipes, pockets, or pods, which in large part replace the adjacent rock or follow, for a distance, stratification planes, fissures, or joints (Fig. 237). In the Iron Blossom mines the galena ore forms a horizontal pipe-like ore-shoot with a greatest width of 150 feet (Fig. 239). This has been mined for nearly 8,000 feet, and there seems to be no well-defined fissure underneath it. Towards the southern

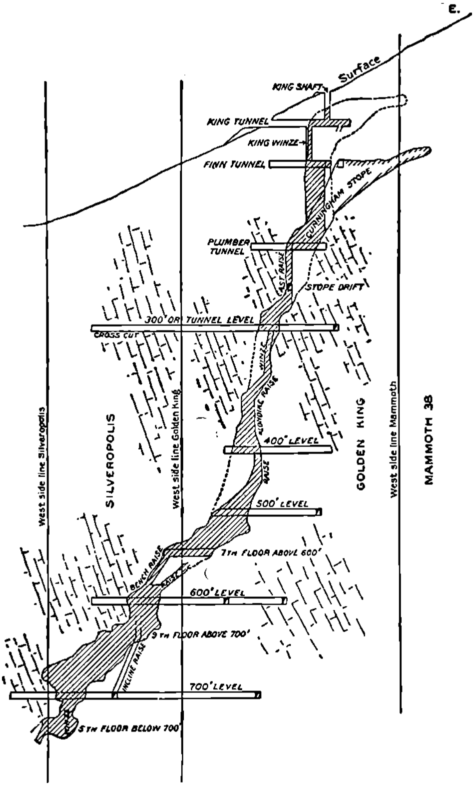

FIG. 237.—Vertical section showing Silveropolis column, Mammoth mine, Tintic, Utah.

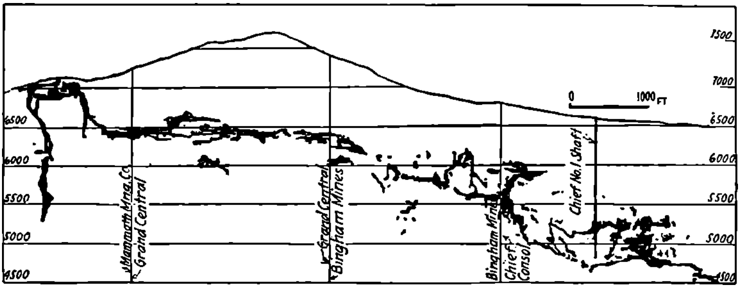

FIG. 238.—Section of western ore channel through the Tintic district, looking west. (*After Charles A. Dobbel.*)

end, after crossing the Sioux fault, it assumes the form of a well-defined fissure which has been followed to a depth of 1,000 feet; the ore in depth is poorer and contains copper and lead, while lead predominates in the upper part. North of the fault the horizontal pipe follows a bed of pure limestone in the Pine Canyon formation, until, near the Beck tunnel shaft No. 2 its northward course ceases and the ore turns laterally towards the Godiva channel (Fig. 239). A more westerly ore zone of less regularity is shown in Fig. 238. It is contained in steeply dipping limestones and gradually descends towards the north. Its beginning is in the steep chimney of the Mammoth mine.

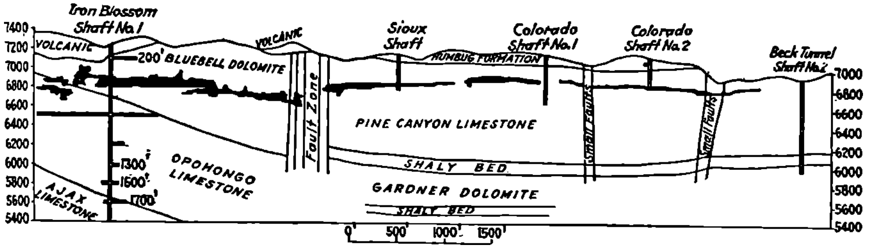

FIG. 239.—Longitudinal section along the fourth ore zone from Beck tunnel to Iron Blossom No. 1. Ore-bodies shown in black. (*After Lindgren and Loughlin, U. S. Geol. Survey.*)

The primary ore minerals are galena and sphalerite with very little pyrite. The gangue minerals are fine-grained quartz or jasperoid and barite. The galena is rich in silver (20 to 50 ounces per ton). Other large replacement bodies consist mainly of quartz and jasperoid and carry gold and silver with a little lead. Near the intrusive mass the ore contains mainly enargite ($Cu_3AsS_4$) with a little gold.

The jasperoids along the deposits result from the replacement of limestone and dolomite by colloidal silica, which later crystallized to chalcedony and quartz. The crushing of this silica resulted in brecciation, and silica of a second phase was deposited in open spaces as colloidal material which later crystallized to quartz. The development of crystalline barite, galena and other sulphides accompanied both phases.

Complete or partial oxidation continues to a depth of about 2,000 feet, which coincides with the water level. In the mines in monzonite the water level stands much higher.

During the deep oxidation cerussite and anglesite formed in the lead deposits. Secondary zinc minerals, mainly smithsonite, usually develop in the limestone outside of the primary lead shoots.[1] Complex copper and copper-lime arsenates are found in the copper-bearing deposits and are usually accompanied by more or less chalcocite and covellite.

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[1] G. F. Loughlin, The oxidized zinc ores of the Tintic district, *Econ. Geol.*, 9, 1914, pp. 1–19.

The mines of Tintic yield annually $7,000,000 to $14,000,000; their complex smelting ores contain gold, silver, copper, lead, and zinc. The annual ore production is about 400,000 tons. The total value of the gold, silver, lead, copper, and zinc produced from 1869 to 1929, inclusive, was approximately $325,000,000, the silver making a total of about 217,000,000 ounces. Since 1929 the output has decreased, as in most other districts. In recent years the Tintic Standard[1] was the most productive silver mine in the country if not in the world. In 1925, it yielded ore valued at $7,117,000, including 4,000,000 ounces of silver and 24,000 tons of lead. (Fig. 240.) In 1930, silver production had decreased to 2,500,000 ounces.

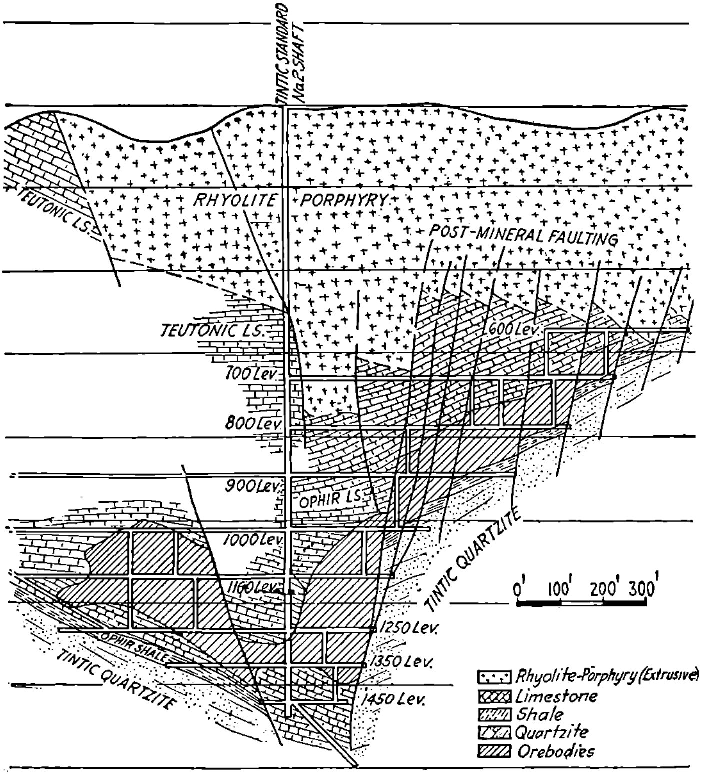

FIG. 240.—Vertical section of Tintic Standard mine. (*After G. W. Crane.*)

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[1] A. B. Parsons, The Tintic Standard Mine, *Eng. Min. Jour.-Press*, Aug. 8, 1925.

G. W. Crane, Notes on the Geology of East Tintic, *Trans.*, Am. Inst. Min. Met. Eng., 74, 1926, pp. 147–162.

J. W. Wade, Mining Methods and Costs at the Tintic Standard, *Information Circ.* 6300, U. S. Bur. Mines, 1930.

**Tsumeb, Southwest Africa.**[1]—Schneiderhöhn's efficient description of the Tsumeb mine discloses an interesting similarity to the Tintic deposits. It is contained in folded limestones and dolomites of Paleozoic or pre-Cambrian age, and the ores form replacements. The dolomite is silicified but not contact metamorphosed. The ore minerals are, in order of deposition, pyrite, bornite, sphalerite, enargite, tetrahedrite, galena, and hypogene chalcocite. Some barite is present, also much supergene chalcocite. The mine is well known for the beauty of its oxidized minerals.

**Aspen, Colorado.**[2]—The ore deposits at Aspen, in the central part of Colorado (Fig. 194), for many years yielded a large amount of lead and silver, but the annual output is now small. The ores average 5 ounces of silver per ton and 6¼ per cent lead. The geological column at Aspen includes 200 to 400 feet of Cambrian quartzite; 250 to 400 feet of Silurian dolomite; 60 feet of Devonian quartzite and shale ("parting quartzite"); 250 feet of Lower Carboniferous dolomite and 150 feet of limestone of the same age; 1,000 feet or more of thin-bedded Carboniferous limestones and shales called the Weber formation; and a great thickness of Carboniferous, Triassic, and Cretaceous sandy and calcareous sediments. The entire series is sharply upturned. A sheet of diorite porphyry intrudes the lower Paleozoic formations and a sheet of rhyolite porphyry, with aplite, lies at the base of the Weber formation. Granodiorite porphyry appears to the south. The intrusives are of early Tertiary age. Complicated faulting and local doming accompanied the intrusion. During the short epoch of ore deposition sulphides were deposited along the Silver fault and the Contact fault, the most important horizon of mineralization being at the base of the Weber shales, where the depositing waters were dammed by the relatively impervious shale, which may also have acted as a precipitant.

The ore minerals occur in disseminated form replacing limestone, either without gangue or with a little jasperoid and dolomite. The only other gangue mineral is barite, which in places is quite abundant and generally older than the ore minerals. There is little pyrite. In order of deposition we have silver-bearing tennantite cut by later bornite, chalcopyrite, galena, and lastly, pearceite (Fig. 241). There is also much native silver, which is held to be of supergene origin though it is found at depths of 1,000 to 1,500 feet, as nests in shale or as "teeth" in pearceite; sometimes minute crystals of barite are suspended by silver wires.

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[1] H. Schneiderhöhn, Das Otavibergland und seine Erzlagerstätten, *Zeitschr. prakt. Geol.*, 37, 1929, pp. 85–116.

Mineralische Bodenschätze im südlichen Afrika, Berlin, 1931.

[2] J. E. Spurr, *Mon.* 31, U. S. Geol. Survey, 1898.

J. E. Spurr, Ore deposition at Aspen, Colo, *Econ. Geol.*, 4, 1909, pp. 301–320.

E. S. Bastin, *Bull.* 750, U. S. Geol. Survey, 1924, pp. 41–62.

Adolph Knopf, *Bull.* 785, *idem*, 1926, pp. 1–28.

A remarkable shoot of "polybasite" ore yielding many million dollars was mined in the Mollie Gibson mine at a depth of a few hundred feet below the surface. It was ascertained later that this ore was really a complex mixture of tennantite and pearceite, with more or less native silver.

The mine water is abundant and the water level stood originally about 300 feet below the surface. The total production of the Aspen district, since 1880, has a value of about $100,000,000, almost wholly in silver and lead.

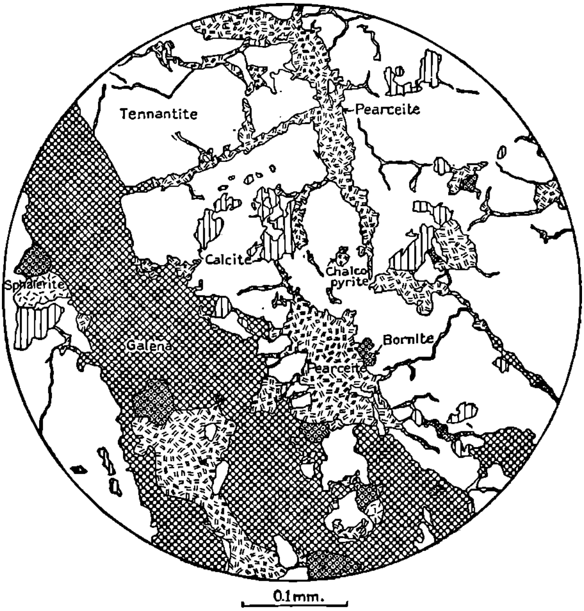

Fig. 241.—Tennantite traversed by late primary replacement veinlets of pearceite, base-metal sulphides, and calcite. Smuggler Hill workings, thirteenth level, Aspen, Colorado. Camera lucida drawing from polished specimen. (*After E. S. Bastin, U. S. Geol. Survey.*)

**Leadville, Colorado.**[1]—Since the discovery of the ore deposits of Leadville, Colorado, in 1875, this district has yielded an enormous production of lead and silver, also much gold, copper, and zinc. Previous to that date placers were worked in the district and gold to the value of several million dollars was washed from the gravel of the gulches. For a long time after 1875 the oxidized lead ores, containing much iron and

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[1] S. F. Emmons and J. D. Irving, The Downtown district, *Bull.* 320, U. S. Geol. Survey, 1907.

G. F. Loughlin, Guides to ore in the Leadville district, *Bull.* 779, *idem*, 1926.

C. W. Henderson, Mining in Colorado, *Prof. Paper* 138, *idem*, 1926.

S. F. Emmons, J. D. Irving, and G. F. Loughlin, Geology and ore deposits of the Leadville district, *Prof. Paper* 148, *idem*, 1927.

manganese, were worked. At the present time the main product is heavy sulphide ore containing pyrite, sphalerite, galena, and chalcopyrite. Bodies of oxidized zinc minerals such as calamine and smithsonite were discovered about 1910. The total value of gold, silver, copper, lead, and zinc recovered from 1859 to 1926 in the Leadville district was $431,000,000.

Out of a total of 477,000 tons mined in 1916, 310,000 tons were sulphide ores, and 18,000 tons iron-manganese ores of the oxidized type containing a little silver and used for flux. The heavy sulphide ores are in part smelted directly and in part concentrated; they vary considerably but consist mainly of pyrite and sphalerite, with less than 1 per cent of copper, 1 to 4 per cent of lead, and 2 to 9 ounces of silver per ton. The silver is mostly contained in the galena. The crude zinc ore shipped averaged 30 per cent of that metal. In 1930, the production had declined considerably. The yield was 5,035 ounces of gold, 616,800 ounces of silver, 252,000 pounds of copper, 6,800 tons of lead, and 11,500 tons of zinc—a total value of $2,502,400.

The geological section consists, according to S. F. Emmons, of Paleozoic rocks resting on granite and gneiss. The following formations are important in the study of the ore deposits:

| | |
|---|---|
| Weber shales and grits, Upper Carboniferous | 2,500 feet |
| Blue limestone, Lower Carboniferous | 200 feet |
| Parting quartzite, Ordovician | 40 feet |
| White limestone, Ordovician | 160 feet |
| Lower quartzite, Cambrian | 150 to 200 feet |

These formations are intruded by numerous sheets of porphyry, which in the main lie parallel to the bedding, but in places cut diagonally across it. Some sheets are thin, and others are nearly 1,000 feet in thickness. The earlier "white porphyry" is a siliceous granite porphyry, forming a thick sheet. The "gray porphyry" is similar but a little lower in silica. The white porphyry is normally intruded in the blue, or Leadville limestone; the gray porphyry forms thinner sheets at various horizons.

The intrusions and the ore deposition were followed by a gentle folding and reverse faulting, and this again by fissuring, in part radially arranged about the Breece Hill intrusive stock. Then came the main epoch of ore deposition. After this followed much more extensive normal faulting, so that the district now consists of numerous blocks successively dropping off toward the Arkansas Valley (Fig. 242). The ore deposits are found mainly in the blue limestone at or near the contact with the overlying white porphyry (Fig. 243). The upper surface of the ore is often remarkably regular and sharp, being formed by the porphyry contact, while the lower surface is irregular. Although this is the normal development, replacement ores are also found in other positions, along fissures in limestone, along fault planes or below the gray porphyry, or in fissure veins extending below the sedimentary beds. The fissure veins and stockworks are confined to the part of the district near the Ibex mine.

Some of the ore-bodies are of large size, especially in a horizontal direction. Owing to their mode of occurrence and to the great quantity of mine water, a depth of only 1,500 feet has been attained; it was considered useless to go below the basement of the Paleozoic formations.

The usual ore is a massive, granular mixture of sulphides, among which pyrite and sphalerite prevail. Banded textures of pyrite and sphalerite, suggesting diffusion, occur in places. There is a scant gangue of quartz, jasperoid, and little barite. The limestone near the deposits often contains much manganosiderite spreading from the ore-bodies. The contact between ore and limestone is usually surprisingly sharp, though irregular. Among the rarer minerals are native gold, molybdenite, wolframite, and scheelite; the ores contain a little antimony, arsenic, and bismuth. Along the Tucson overthrust fault ore rich in gold was found containing bismuthinite and argentite. Ores containing magnetite, specularite, and epidote, besides sulphides, occur in the Breece Hill area. The oxidized ores of the upper levels contain limonite, manganese oxide, cerussite, and zinc carbonates. They were frequently rich in silver chloride.

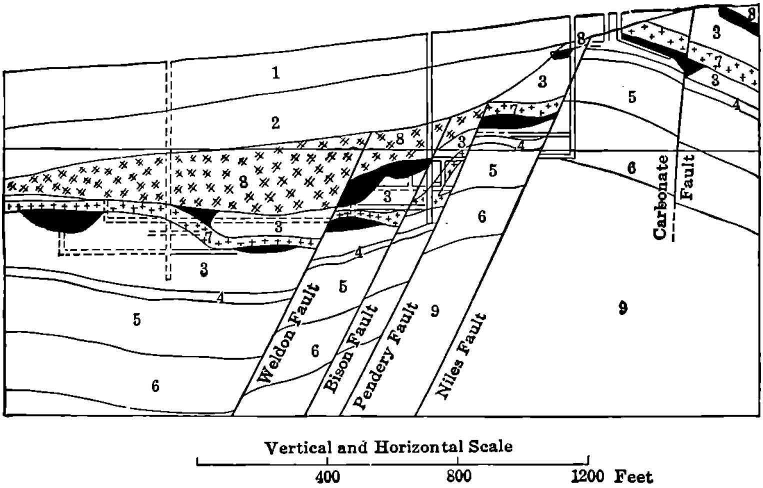

FIG. 242.—Vertical section showing geological structure and occurrence of ore-bodies at Leadville, Colorado. 1, wash; 2, lake beds; 3, Leadville blue limestone (Carboniferous); 4, parting quartzite (Ordovician); 5, white limestone (Ordovician); 6, lower quartzite (Cambrian); 7, Gray porphyry (early Tertiary); 8, white porphyry (early Tertiary); 9, granite (pre-Cambrian); ore-bodies in black. (*After Emmons and Irving, U. S. Geol. Survey.*)

The genesis of the Leadville ores has been discussed extensively. Emmons held in 1886 that they were formed by aqueous solutions coming from above and that they derived their mineral content mainly from the igneous rocks. He also fully recognized that at the time of ore formation the present deposits were covered by about 10,000 feet of overlying rocks.

In 1907, Emmons recognized the possible deep-seated origin of the ore. A. A. Blow, Philip Argall, and other writers have insisted that the deposits were formed by ascending solutions, and in *Professional Paper* 148 (U. S. Geol. Survey) the same view is expressed. The solutions came up along fissures, such as the reverse faults of Tucson and Ibex (Fig. 243), and then spread into the flat limestone to produce blanket deposits. The earlier deposits were of the high-temperature type with magnetite and specularite, but as the temperature dropped the main ore-bodies assumed the normal mesothermal type. The intrusive stock of Breece Hill is believed to represent the center of the metallization.

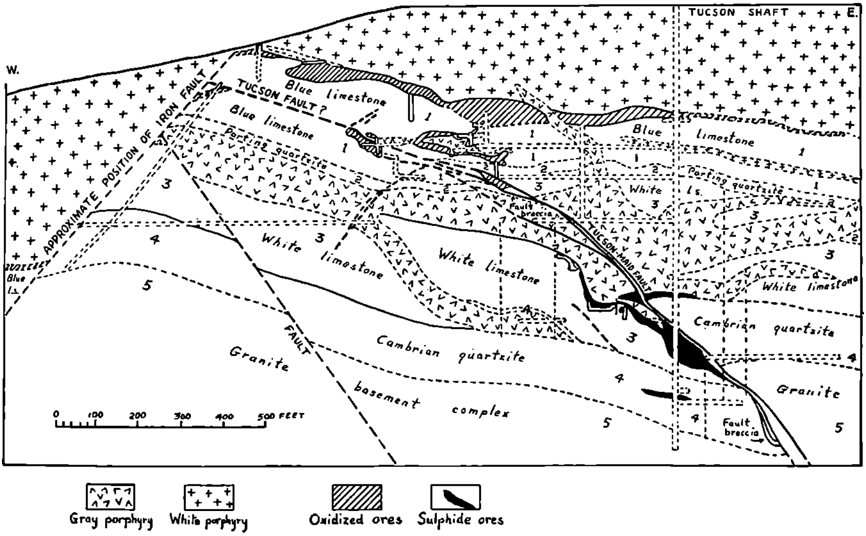

FIG. 243.—Section through Tucson fault at Leadville, Colorado, looking northwest. (*After Loughlin and Aicher, U. S. Geol. Survey.*)

**The Leadville-Boulder County Belt.**—The Leadville deposits form only a single unit in a belt of deposits which extends for 80 miles in a northeasterly direction and comprises a great many districts, including the Kokomo, Alma, Fairplay, Breckenridge, Montezuma, and Argentine, and continuing through Clear Creek, Gilpin, and Boulder counties (Fig. 194). The deposits include replacement bodies and veins and are found in rocks of the most diverse kinds. A common feature of the whole belt is a series of intrusives, appearing as sheets in the sedimentary formations and dikes or smaller stocks in the pre-Cambrian granite and schists. The inference that these intrusives are genetically connected with the deposits seems well founded. S. F. Emmons and Whitman Cross first called attention to this belt of intrusives; S. H. Ball,[1] F. L. Ransome,[2] and A. D. Crawford[3] have discussed the petrography of the porphyries and Ransome has presented a diagram showing the composition of all analyzed varieties. Ball has indicated on a map all the occurrences of porphyry within this belt. The most abundant intrusives are alaskite porphyry, granite porphyry, bostonite porphyry, monzonite porphyry, and quartz monzonite porphyry. The dikes commonly extend in a northerly direction, but show no great individual continuity. Emmons held that the intrusions were probably of Jurassic age, but later evidence discovered by Cross and others has shown that they are late Cretaceous or early Tertiary. The ore deposits are later than the porphyries but were formed shortly after their consolidation.

The prevailing type is a sulphide ore with abundant pyrite and sphalerite and lesser amounts of galena and tetrahedrite. Chalcopyrite is subordinate, arsenopyrite rare. Telluride ores occur occasionally in the eastern end of the belt. Silver prevails in the southeastern part and gold is the important metal in Gilpin and Boulder counties. The gangue is made up of quartz, siderite, manganosiderite, and other carbonates, but not much rhodochrosite is present. In places there is considerable barite.

The replacement deposits of Leadville, in Carboniferous limestone below porphyry sheets, have already been mentioned. In the Tenmile district[4] replacement deposits and fissure veins appear in the Upper Carboniferous formations, with pyrite, sphalerite, and galena in a gangue of quartz, calcite, rhodochrosite, and barite. At Alma, there are base-metal blankets in the Leadville limestone and also gold-quartz veins in the "white porphyry" close to the London fault. The Red Cliff district[5] has ores similar to those of Leadville, also replacement deposits carrying gold in Cambrian quartzite. The Breckenridge district[6] contains fissure veins intersecting Cretaceous shale and monzonite sheets, with pyrite, sphalerite, and galena as the principal ore minerals. Remarkable pockets of crystallized gold are thought to be deposited by descending waters. Northeast of Breckenridge is the Montezuma district,[7] in pre-Cambrian rocks with northeasterly trending veins carrying pyrite, chalcopyrite, galena, and sphalerite, with later ruby silver or similar rich silver minerals. The gangue consists of quartz, siderite, and barite.

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[1] *Prof. Paper* 63, U. S. Geol. Survey, 1908, pp. 67–70; see Pl. XI.

[2] *Prof. Paper* 75, *idem,* 1911, pp. 60–62.

[3] *Am. Jour. Sci.,* 5th ser., 7, 1924, pp. 365–388.

[4] S. F. Emmons, Tenmile district folio, *Geol. Atlas* 48, U. S. Geol. Survey, 1898.

[5] A. H. Means, *Econ. Geol.,* 10, 1915, pp. 1–27.
R. D. Crawford and R. Gibson, *Bull.* 30, Colorado Geol. Survey, 1925.

[6] F. L. Ransome, *Prof. Paper* 75, U. S. Geol. Survey, 1911.

[7] H. B. Patton, *First Rept.,* Colorado Geol. Survey, 1909, pp. 112–144.

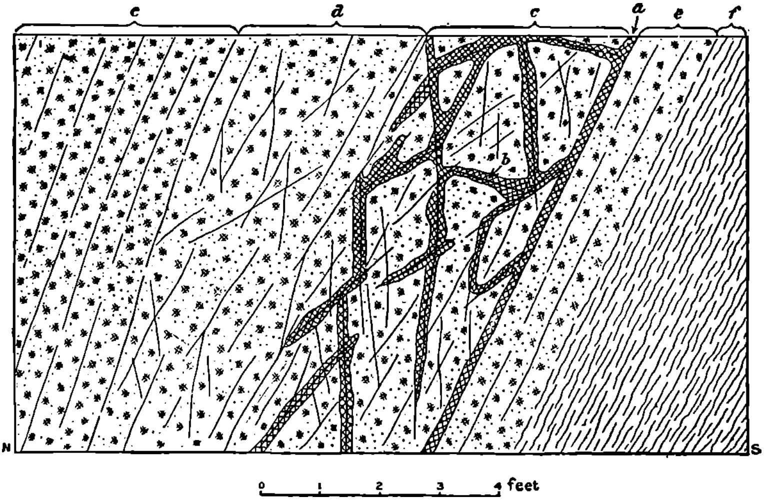

FIG. 244.—Section of Pelican vein, Georgetown, Colorado. *a*, main vein; *b*, ore, mainly sphalerite; *c*, ore zone of brecciated and silicified alaskite porphyry; *d*, fractured and silicified porphyry; *e*, sheeted porphyry; *f*, altered gneiss. (*After J. E. Spurr, U. S. Geol. Survey.*)

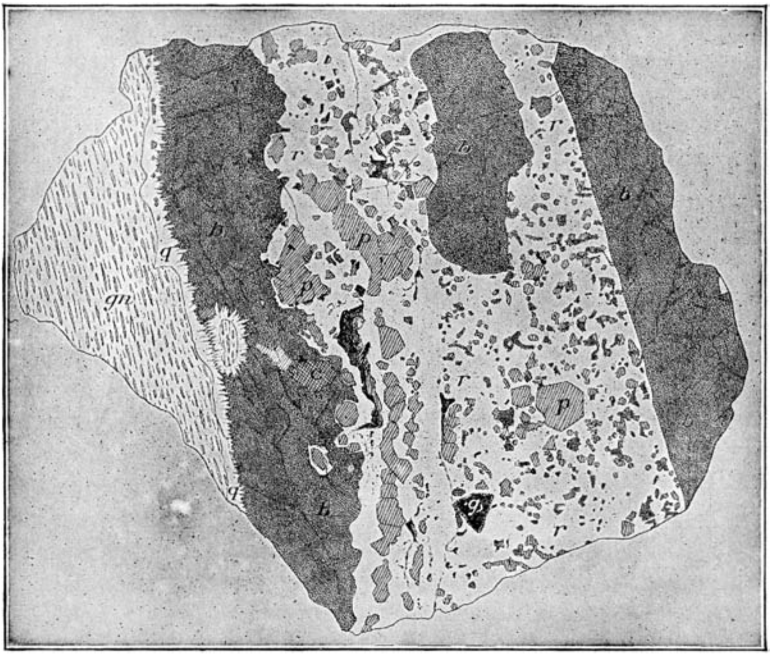

FIG. 245.—Drawing of specimen, Griffith mine, Georgetown, Colorado, showing sulphide vein split open, and the new fissure filled with manganese-iron carbonate and pyrite. *gn*, gneiss; *q*, quartz (comb structure); *b*, sphalerite; *p*, pyrite; *g*, galena; *r*, manganese-iron carbonate with pyrite. Natural size. (*After Spurr and Garrey.*)

The rocks in the Clear Creek district[1] comprise the pre-Cambrian Idaho Springs formation, intruded by granite, diorite, and pegmatite. A complex system of dikes of the kinds mentioned above is followed by the veins, which are principally silver deposits containing galena, sphalerite, pyrite, tetrahedrite, and chalcopyrite, in a more or less scant gangue of earlier quartz, and later siderite, ankerite, and calcite (Figs. 244 and 245). Silver, gold, copper, lead, and zinc are produced.

In Gilpin County,[2] gold-bearing veins prevail; they carry abundant pyrite, with some chalcopyrite, tennantite, and enargite, rarely pitchblende, in a scant quartz and siderite gangue. The ores average about $8 in gold to the ton and contain a small amount of silver. A later vein formation carries galena and sphalerite, with silver, in a quartz, siderite, and calcite gangue. Quartz-telluride ores are also known (Fig. 47).

In all the districts the alteration of the feldspathic country rock adjacent to the vein is of the sericitic type. The lowest workings in Clear Creek and Gilpin counties are 2,000 feet below the outcrops. In depth the veins are generally poor.

In Boulder County,[3] the present production of which is small, there are some gold-bearing veins similar to those of Gilpin County, and also some veins which have produced rich silver ores; but the most interesting types are the telluride veins, which are rare in the other districts mentioned, and the tungsten veins, which are absent elsewhere.

The Enterprise vein,[4] which is typical of the telluride deposits, consists of several narrow seams, forming a sheeted zone along which filling and replacement have occurred. The width of this zone is from 1 to 3 feet and the filling is often beautifully banded with abundant druses. The country rock is pre-Cambrian granite. The minerals are crystallized quartz and dense jasperoids with barite, adularia, and roscoelite (vanadium mica); there are a little pyrite and much molybdenite; the telluride is probably sylvanite.

**The Tungsten Deposits of Boulder County.**—Some wolframite occurs in the gold-bearing veins of Boulder County, but there is a fairly well-defined area near Nederland[5] which is characterized by tungsten ores. In 1918, 5,020 tons of tungsten concentrate were produced in the United

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[1] J. E. Spurr, G. H. Garrey, and S. H. Ball, Economic geology of the Georgetown quadrangle, *Prof. Paper* 63, U. S. Geol. Survey, 1908.

[2] E. S. Bastin and J. M. Hill, Economic geology of Gilpin County, etc., *Prof. Paper* 94, U. S. Geol. Survey, 1917.

[3] T. S. Lovering, *Proc.*, Colorado Sci. Soc., 13 (3), 1932, pp. 77–88.

[4] W. Lindgren, *Econ. Geol.*, 2, 1907, pp. 453–463.

T. A. Rickard, *Trans.*, Am. Inst. Min. Eng., 33, 1903, pp. 567–577.

[5] W. Lindgren, *op. cit.*

R. D. George, The main tungsten area of Boulder County, *First Ann. Rept.*, Colorado Geol. Survey, 1908, pp. 9–103.

States of which the larger part came from the deposits of Boulder County. The present production is small. The domestic production comes from Atolia, Kern County, California, where scheelite ($CaWO_4$) occurs in gold-bearing quartz veins of the mesothermal type; in part, also, from contact-metamorphic deposits in Nevada and California (p. 728).

The veins of Boulder County are narrow, often brecciated fissures in granite, some of them following porphyry dikes. The gangue is made up of quartz and fine-grained silica. Kaolinite is abundant as a secondary mineral. There is a little pyrite, but the principal mineral is the iron tungstate, ferberite, which occurs in a fine-grained mixture with quartz or as beautiful crystals coating druses, associated with quartz crystals. The concentrates contain a trace of gold and a little silver. Apparently the present ore-shoots have been bottomed at a depth less than 1,000 feet.

**Summary.**—The mineral belt described in the preceding paragraphs illustrates well the intimate connection between veins and replacement deposits. With all their differences the deposits are evidently of approximately contemporaneous origin. The southwest end, at Leadville, represents the deposits formed at considerable depth, probably not less than 10,000 feet. Toward the northeast end the sedimentary series is absent and may not have covered the Front Range of Colorado. There is evidence in this range of a Tertiary surface higher than the present but of less relief; and Spurr believes that effusive rocks once rested on this surface. S. H. Ball[1] estimates that in Clear Creek County the total erosion since the intrusion of the porphyries has been about 5,000 feet, and it is probable that it was less in Boulder County. Corresponding to this difference in geologic conditions is a difference in structure and composition of the ore deposits. At the southwestern end of the belt heavy sulphide ores prevail, mostly as replacements. In Clear Creek and Gilpin counties the banded and drusy structure of the veins begins to be apparent, and in Boulder County we find the telluride and the tungsten veins, which distinctly resemble the deposits formed at slight depth below the surface (epithermal veins), having drusy, banded structure and containing fine-grained quartz and tellurides. These relations are, to say the least, very suggestive of progressive change in original depth of deposition from the western to the eastern end of the belt.

The rich silver minerals, which are found in some of the districts, are regarded by Spurr and also by E. S. Bastin[2] as products of deposition by descending surface waters. Spurr believed that the metals in the Clear Creek deposits were derived from deep-seated magmatic sources and, dissolved in magmatic waters, ascended the fissures after intrusion of porphyry. He also thought that most of the gangue minerals—quartz,

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[1] *Prof. Paper* 63, U. S. Geol. Survey, 1908, p. 145.

[2] *Econ. Geol.*, 8, 1913, p. 51.

carbonates, and barite—were derived from the leaching of the country rock by the ascending waters.

Ransome points out that many of the fissures at Breckenridge are short and die out before reaching great depth, so that there was probably considerable lateral movement during which the solutions had opportunity to search the surrounding rocks more or less thoroughly. The gangue minerals are regarded by him as derived from the country rock, while he considers a magmatic origin possible for the metals and elements like sulphur and fluorine. In view of the faint contact-metamorphic effects shown at the western end of the belt in the limestones, it is indeed improbable that the mineral-bearing solution was derived from the small bodies of intruding porphyries; the metals must have been derived from deep-seated magma basins.

**The Lead-Silver Replacement Deposits of Mexico.**[1]—Lead-silver replacement deposits are very numerous in Mexico and are generally found in the thick Mesozoic limestone which covers a large part of the Mexican plateau. The geologic column at Santa Eulalia is as follows, according to Prescott:

| | | |
|---|---|---|
| Shales | 3,000 feet | Upper Cretaceous (?) |
| Upper Bufa limestone | 3,000 feet | Cretaceous |
| Intermediate beds | 500 feet | Cretaceous |
| Lower fossil limestone | 100 to 400 feet | Lower Cretaceous (Comanchean) |
| Blue limestone | 1,500 feet | |
| Sandstone and shale | 5,000 feet | Jurassic ? |

The deposits stand in genetic connection with intrusive masses of smaller extent, usually of diorite, granodiorite, and various porphyries, and frequently are developed in the form of long, continuous pipes. They are mined in many states. Excellent examples were studied, for instance, near Zimapan, Hidalgo. There are here two old and productive mines, the Lomo del Toro and the Balcones. The latter deposit trends northwest and follows a dike system of quartz monzonite porphyry, which extends out from the main mass of intrusive rock in the Toliman canyon. These ores carry lead, zinc, and silver and contain, in part, lime silicates, and are thus allied to the hypothermal deposits. Most of the ores are oxidized. The Lomo del Toro deposit follows a wide zone of fracturing in the horizontal limestones and contains few dikes. In the lower part

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[1] Basil Prescott, *Trans.*, Am. Inst. Min. Met. Eng., 51, 1916, pp. 57–99. The underlying principles of the limestone replacement deposits of the Mexican province, *Eng. Min. Jour.-Press*, Aug. 14 and 21, 1926.

A. R. Fletcher, Mexico's lead-silver manto deposits and their origin, *idem*, Mar. 30, 1929, pp. 509–513.

M. W. Hayward and W. H. Triplett, Occurrence of lead-zinc ores in dolomitic limestones in northern Mexico, *Tech. Publ.* 442, Am. Inst. Min. Eng., 1931, 31 pp.

of the mine, but still above the Toliman canyon, are flat bodies of sulphide ore containing $1 in gold, 24 ounces silver, 8 per cent lead, and 18 per cent zinc. No lime silicates are present. Across the Toliman canyon, farther away from the intrusive, are many small, more or less horizontal pipes of oxidized lead ores accompanied by silicification. In cross-section they range up to 30 feet in diameter, and some of them are 1,300 feet in length, usually with a winding course.

Preeminent localities are Santa Eulalia, Chihuahua, Mapimi, Durango, and Sierra Mojada, Coahuila. The Santa Eulalia has yielded lead and silver, valued at from $300,000,000 to $500,000,000. The deposits, described by Prescott, carry mainly oxidized ores and form flat, tabular (mantos), or irregular bodies and pipes in Cretaceous limestone and have been worked to a depth of 2,000 feet. The favorable limestones are the Lower Fossil Horizon and the Upper Bufa. The course of the pipes is only in minor degree determined by fissures. According to Prescott, there are no vertical fissures feeding the pipes though the solutions seem to have ascended through the chimneys and worked their way outward (Fig. 246). It is interesting to compare these pipes with the somewhat longer pipes of the Iron Blossom mines at Tintic. Naturally, there are many difficult problems in ascertaining how the solutions could have managed these long penetrations. The ores are apparently similar throughout the length of a pipe. The average composition of the Santo Domingo pipes

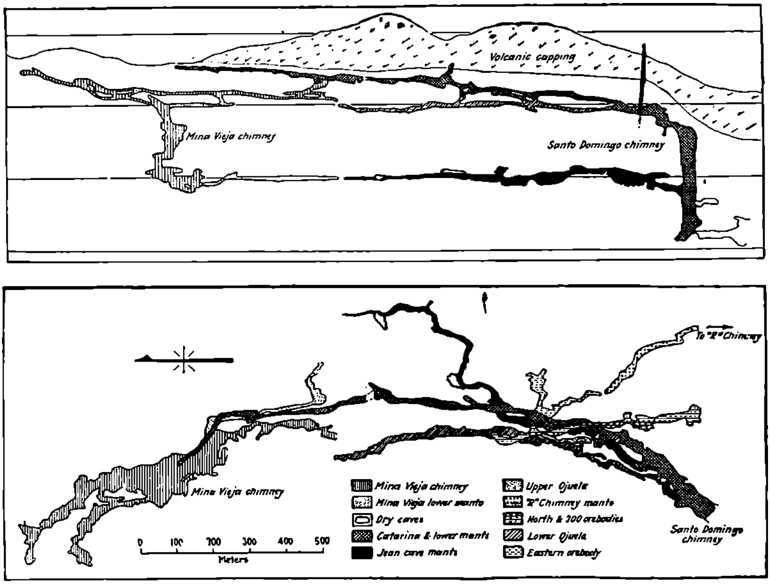

FIG. 246.—Plan and longitudinal section showing chimneys and pipes in flat limestone in the Santa Eulalia district, Mexico. (*After Basil Prescott.*)

is, in per cent, 13 lead, 2 to 6 zinc, 25 iron, and 8 lime. The silver averages about 300 grams per ton. In Tintic there is a great deal of silicification surrounding these pipes. At Santa Eulalia the silicification appears less well marked. Two main types of ore are found: (1) an older, complex silicate gangue (ilvaite, hedenbergite, fayalite, and chlorite), with argentiferous pyrite and little lead and zinc; (2) the predominating Santa Eulalia ore containing pyrite, galena, and sphalerite, with lime silicates in minor amount or absent. At Sierra Mojada the "mantos" occur in sedimentary dolomite.

## DEPOSITS WITH NATIVE SILVER

Native silver is not, as a rule, a primary mineral in the deposits which contain it, nor is it restricted to any particular class of deposits. As a secondary (supergene) product due to reactions within the oxidized zone, it is common in many kinds of deposits—for instance, in argentiferous galena ores, in tetrahedrite ores, and in the argentite veins. It is ordinarily found some distance below the surface; cerargyrite (AgCl) is more abundant in the outcrops. The native silver often occurs at depths far below the zone of oxidation, properly speaking. At Aspen, Colorado, it is abundant in fissures and vugs of limestone and shale 900 feet below the surface and is distinctly later than the primary lean galena-sphalerite ores.

In certain deposits the native silver is, however, the predominating ore mineral and is evidently of primary (hypogene) origin. These occurrences may be divided into two groups:

1. The silver-bearing deposits with zeolites.
2. The silver-bearing cobalt-nickel deposits.

**The Zeolitic Deposits.**—Zeolites are ordinarily foreign to ore deposits connected with igneous rocks. Their occurrence in some contact-metamorphic deposits and in certain fissure veins is mentioned on page 516. Zeolites are found in the veins of Kongsberg, Norway; Andreasberg, Germany; Arqueros and Rodaito, Chile; and Guanajuato, Mexico, all of which are worked for silver. They are rarely noted in the western part of the United States. In general, it seems certain that the zeolites were deposited later than the other minerals. A concentration of silver ores often accompanied their development. Calcite, quartz, barite, and fluorite are the principal gangue minerals in the typical localities; the presence of antimony and quicksilver is often mentioned.

The renowned silver mines of Kongsberg, in southern Norway, which have been worked for several hundred years and still remain productive, have been described by Vogt.[1] The deposits are narrow east-west veins

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[1] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1899, pp. 113–123; 177–181.

A. Bugge, *Norsk. Geol. Tidsskr.*, 12, 1931, pp. 123–147; *Norges Geol. Undersök.*, 133, 1929, pp. 100–108.

in gneiss and mica schist, often breaking up when entering amphibolite. Along certain lines following the schistosity the rocks contain disseminated sulphides, mainly pyrite and pyrrhotite, and the veins become enriched where crossing these "fahlbands," probably on account of their precipitating influence. Sections indicate a steeply dipping, conjugated vein system, the pay-shoots pitching northward and the veins becoming closely spaced in depth. The greatest depth attained is 3,000 feet. The veins are held to be of Late Paleozoic age and genetically connected with the Kristiania igneous rocks. The erosion surface below the Cambrian was probably not far above the present land surface. Possibly the Paleozoic slates formed an impervious cover below which the ore was deposited. Quartz, chlorite, and axinite crystallize next to the walls, but the prevailing gangue is calcite with some barite and fluorite, rarely adularia and albite. Zeolites also occur and are among the latest gangue minerals, prehnite, stilbite, harmotome, and laumontite being among those identified. The principal ore mineral is native silver, mostly in wire form; this is believed to be derived from the more scarce argentite. Less prominent are ruby silver, stephanite, pyrite, pyrrhotite, arsenopyrite, chalcopyrite, sphalerite, and galena, the last poor in silver. A certain part of the silver is believed by Vogt to result from primary deposition. The native silver contains quicksilver. Anthracite with 95.5% carbon is also one of the gangue minerals deposited during the early stages.

Vogt supposes the native silver to be derived from argentite and proustite by action of water according to Bischof's reactions which as shown by Moesta can take place at 100° C.

$$Ag_2S + H_2O = 2Ag + H_2S + O.$$
$$Ag_3AsS_3 + 3H_2O = 3Ag + As + 3H_2S + 3O.$$

The list of minerals given shows clearly that the veins have had a complicated history, beginning with the deposition of high-temperature minerals like axinite and ending with that of minerals like zeolites, probably formed at about 100° C. This history has evidently not yet been followed in detail; it is stated by Vogt that axinite crystallized together with the zeolites, but this seems a curious association.

The other notable occurrence is at Andreasberg, in the Harz Mountains, best described by A. Bergeat.[1] The veins at Andreasberg are simple filled fissures, at most 0.5 meter thick, chiefly in Silurian clay slates and quartzites. They appear not far from the intrusive mass of the "Brocken" granite. The veins are included between two great divergent dislocations, forming impermeable walls, against which the silver veins split and cease. The mines, which attained a depth of 2,700 feet, are now closed.

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[1] Stelzner and Bergeat, Die Erzlagerstätten, 2, 1906, pp. 718–720.

In general, the veins carry argentiferous galena and tetrahedrite; sometimes they yield large druses full of rich silver ores, calcite, and zeolites. Bergeat distinguishes five phases:

1. In crevices near the veins, in part also in the fissures themselves, are garnet, epidote, axinite, and albite.
2. Earliest bituminous calcite with simple antimonides, arsenides, and sulphides: Niccolite, smaltite, löllingite, breithauptite (NiSb), sphalerite, galena, pyrite, pyrrhotite, chalcopyrite. There are also narrow veins of quartz, cobaltite, and selenides in sericitized greywacke.
3. Tetrahedrite with quartz (replacing calcite) and fluorite, chalcopyrite, galena, sphalerite; native silver and millerite, encrusting tetrahedrite.
4. Sulphantimonides and sulpharsenides: Pyrargyrite, proustite, miargyrite, polybasite, stibnite, argentite (in part from pyrargyrite), fluorite.
5. Native silver, realgar, calcite, apophyllite, analcite, chabazite, heulandite, brewsterite, harmotome, stilbite, natrolite; also fluorite and chalcopyrite.

This clear exposition of paragenesis indicates a long epoch of deposition with gradually diminishing temperature and emphasizes the connection of the native silver with zeolitization.

Zeolites occur in a number of Chilean silver veins, particularly at Arqueros and Rodaito, in association with native silver, calcite, barite, and silver amalgam. At the South Republic mine, Republic, Washington, where the gold selenide veins in andesite are filled by closely banded, chalcedonic quartz (p. 497), the vein has in part been replaced by a loose aggregate of calcite and laumontite much richer in silver than the original quartz.

The occurrence of native silver in the zeolitic copper and silver deposits of the Lake Superior region is described on page 521. In short, zeolitization, probably effected by warm or tepid waters, seems particularly adapted to the concentration of silver and the deposition of both native silver and rich silver minerals.

**The Silver-bearing Cobalt-nickel Veins of Saxony.**—In different parts of the world occur narrow veins with calcite, or barite gangue and arsenides or sulphides of cobalt and nickel; the cobalt and nickel minerals contain silver, and this metal is often separated as seams of native silver enclosed in the older metallic minerals. Here, the silver is evidently primary and deposited by similar solutions from which the earlier arsenides were formed.

The cobalt veins of Annaberg,[1] in Saxony, appear in gneiss intruded by dikes of granitic and lamprophyric character; they are younger than

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[1] H. Müller, Die Erzgänge des Annaberger Bergrevieres, Geol. Landesanstalt, Leipzig, 1894.

the veins in the same region carrying cassiterite and those yielding pyritic ores with galena. The gangue minerals are barite, calcite, fluorite, quartz, and dolomitic carbonates. The principal ore minerals are chloanthite, smaltite, bismuthinite, also rich silver minerals such as argentite, pyrargyrite, and native silver; the latter are distinctly later than the primary nickel-cobalt-bismuth ores.

Most of the rich silver ores appear where the veins intersect certain flat crushed zones in the gneiss, which contain carbonaceous material and finely divided sulphides, like pyrite and chalcopyrite. The greatest depth attained was 1,400 feet.

The veins of Schneeberg, in Saxony,[1] are contained in contact-metamorphic clay slates and tend to impoverishment in the underlying granite. The primary gangue consisted of calcite, ankerite, barite, and fluorite; but these minerals are now largely replaced by hackly and platy fine-grained quartz by a process similar to that to which many later Tertiary gold-silver veins have been subjected. The ore minerals are smaltite, chloanthite, niccolite, bismuthite, and native bismuth. Native silver and rich silver minerals are subordinate in the silicified veins, but appear in the primary barytic veins. From this it is perhaps permissible to draw the conclusion that the silicification has been accompanied by solution and removal of silver by ascending siliceous solutions.

Uranium ores, mainly uraninite or pitch blende, are found at Schneeberg and, more abundantly, in the somewhat similar veins at Joachimsthal,[2] Bohemia. The geological relations at the two places are similar. At both places the cobalt and nickel minerals are the older and the rich silver minerals the younger. Between them in point of age lie the uranium ores.

In 1930, veins were discovered on the east shore of Great Bear Lake,[3] in pre-Cambrian sediments. They contain uraninite, cobalt and nickel arsenides, chalcopyrite, native silver, bismuth, etc., in a gangue of quartz and rhodochrosite. Their importance remains to be proved.

**The Silver-bearing Cobalt-nickel Veins of Ontario, Canada.**[4]—Many of these veins appear to be connected genetically with igneous rocks of

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[1] K. Dalmer, E. Köhler, and H. Müller, Section Schneeberg, *Geol. Spez. Karte Sachsen*, 1883.

K. Keil, Beiträge zur Kenntniss der Co, Ni, Bi, Ag gänge, *Jahrbuch*, Berg-u. Hüttenw. Sachsen, 105, 1931, pp. *A* 95–132. (Author erroneously considers silver earlier than the arsenides.)

[2] J. Step and F. Becke, *Wien. Sitzungsber.*, *k.* Akad. Wiss. 113, 1904, pp. 585–618; also E. S. Bastin and J. M. Hill, *Prof. Paper* 94, U. S. Geol. Survey, 1917, p. 122.

[3] D. F. Kidd, *Econ. Geol.*, 27, 1932, pp. 145–159.

Ellis Thomson, Univ. Toronto Studies, Geol. Ser., No. 32, 1932.

[4] E. D. Ingall, Report on mine and mining on Lake Superior (Silver Islet), *Ann. Rept.*, Canada Geol. Survey, pt. H, 1887; also C. H. Chadbourne, *Econ. Geol.*, 18, 1923, pp. 77–82.

Willet G. Miller, Cobalt-nickel arsenides and silver deposits of Temiskaming, Keweenawan age; and the occurrence of native silver in the copper mines working the amygdaloids of Michigan may be recalled.

The veins of Silver Islet and those at points southwest of Port Arthur, on the north shore of Lake Superior, have been known for many years. The Silver Islet vein, at one time a heavy producer, intersects greenstone and contains, in order of deposition, calcite, ankerite, and quartz, niccolite, sphalerite, chalcopyrite, galena, calcite, native silver, and argentite. Graphite also occurs in the vein.

In 1903, the silver veins of Cobalt, Ontario, were discovered and they soon became extraordinarily productive. At the end of 1930, the district had produced 365,000,000 ounces silver, or 15,208 short tons, besides much cobalt, arsenic, and nickel; the maximum output, in 1911, was 31,500,000 ounces. During the last years, the output has declined owing to approaching exhaustion and the low price of silver. Cobalt is the fourth district of the world in regard to silver production, being only preceded by Potosi, Guanajuato, and Zacatecas. But while Potosi has been worked for 379 years, the output of Cobalt has been achieved in 27 years. The ore is extremely rich, its tenor often reaching several thousand ounces per ton. Some of the veins contained slabs of native silver of great size. One specimen, now in the Parliament building at Toronto, is 5 feet long and weighs 1,640 pounds; it contains 9,715 troy ounces of silver. The La Rose vein, in a horizontal distance of 100 feet and above the 60-foot level, yielded 532 tons, which contained 658,000 ounces of

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Ontario Bur. Mines, *Nineteenth Ann. Rept.*, pt. 2, 1913.

H. V. Ellsworth, A study of certain minerals from Cobalt, *Twenty-fifth Ann. Rept.*, *idem*, 1916, pp. 200–243.

E. S. Bastin, Significant mineralogical relation in silver ores of Cobalt, *Econ. Geol.*, 12, 1917, pp. 219–236; silver ores of South Lorraine, *idem*, 20, 1925, pp. 1–24.

C. Palmer and E. S. Bastin, Metallic minerals as precipitants of silver, *idem*, 8, 1913, pp. 140–170.

W. H. Collins, Geology of Gowganda mining division, *Mem.* 33, Canada Geol. Survey, 1913.

C. W. Knight, Cobalt and South Lorraine silver areas, *Thirty-first Ann. Rept.*, Ontario Bur. Mines, 1924, pp. 1–238. Literature on geology, *idem*, pp. 321–358. Read, Denny, and Hutchinson, Mining and metallurgical practice at Cobalt, *idem*, pp. 241–320.

E. W. Todd, Gowganda vein minerals, *Thirty-fifth Ann. Rept.*, *idem*, 1926, pp. 62–78.

W. L. Whitehead, *Econ. Geol.*, 15, 1920, pp. 103–136; A. R. Whitman, *idem*, pp. 136–149; J. M. Bell, *idem*, 18, 1923, pp. 684–694 (South Lorraine).

T. L. Walker, Arsenides from South Lorraine, Univ. Toronto Studies, 20, 1925; Ellis Thomson, Mineralographic notes, etc., *idem*, pp. 54–58; T. L. Walker and A. L. Parsons, A comparison of the Port Arthur, Cobalt, South Lorraine, and Gowganda silver minerals, *idem*, pp. 59–62.

Ellis Thomson, A qualitative and quantitative determination of the ores of Cobalt, *Econ. Geol.*, 25, 1930, pp. 470–505, 627–652.

H. C. Boydell (Keeley mine) *Bull.* 230, Can. Min. Met., 1931, pp. 726–750.

silver. So-called low-grade ores averaged about 200 ounces per ton, while that which was concentrated before shipping contained about 30 ounces per ton. Some milling ores worked contained only 10 ounces per ton. Owing to the complex character of the ore the smelting charges were very high. Later, the cyanide, amalgamation, and flotation processes have been adopted with much success.

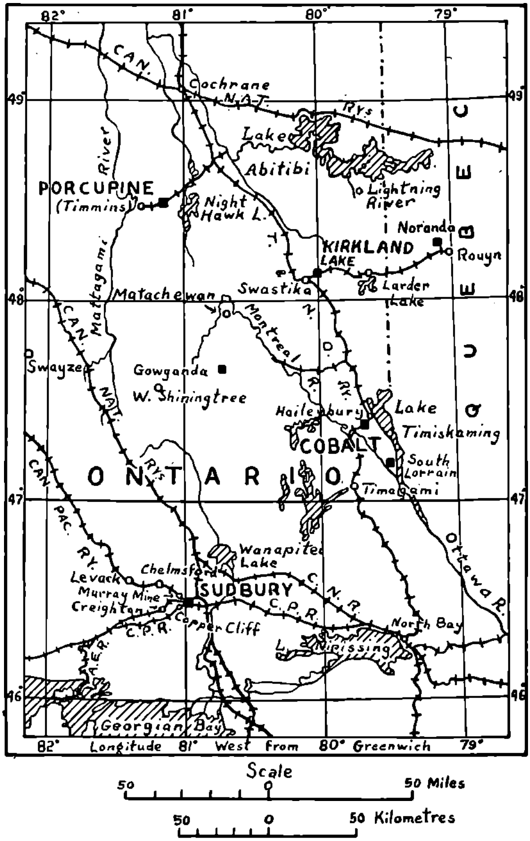

FIG. 247.—Map showing the location of Cobalt in relation to the other important mining areas of northeastern Ontario. (*After Ontario Bur. Mines.*)

Similar cobalt-silver veins have been discovered in other portions of Ontario, particularly at Gowganda and South Lorraine.

The geology of the Cobalt region is summarized by W. G. Miller as follows: The oldest rocks, known as the Keewatin series, are basic volcanic rocks, greenstones, and schists, with more or less cherty or jaspery sediments. On the eroded Keewatin were deposited the conglomerate and graywacke (arkose sandstone) of the Cobalt series (Animikie; Upper Huronian). A thickness of 300 feet of these gently dipping strata is exposed at Cobalt.

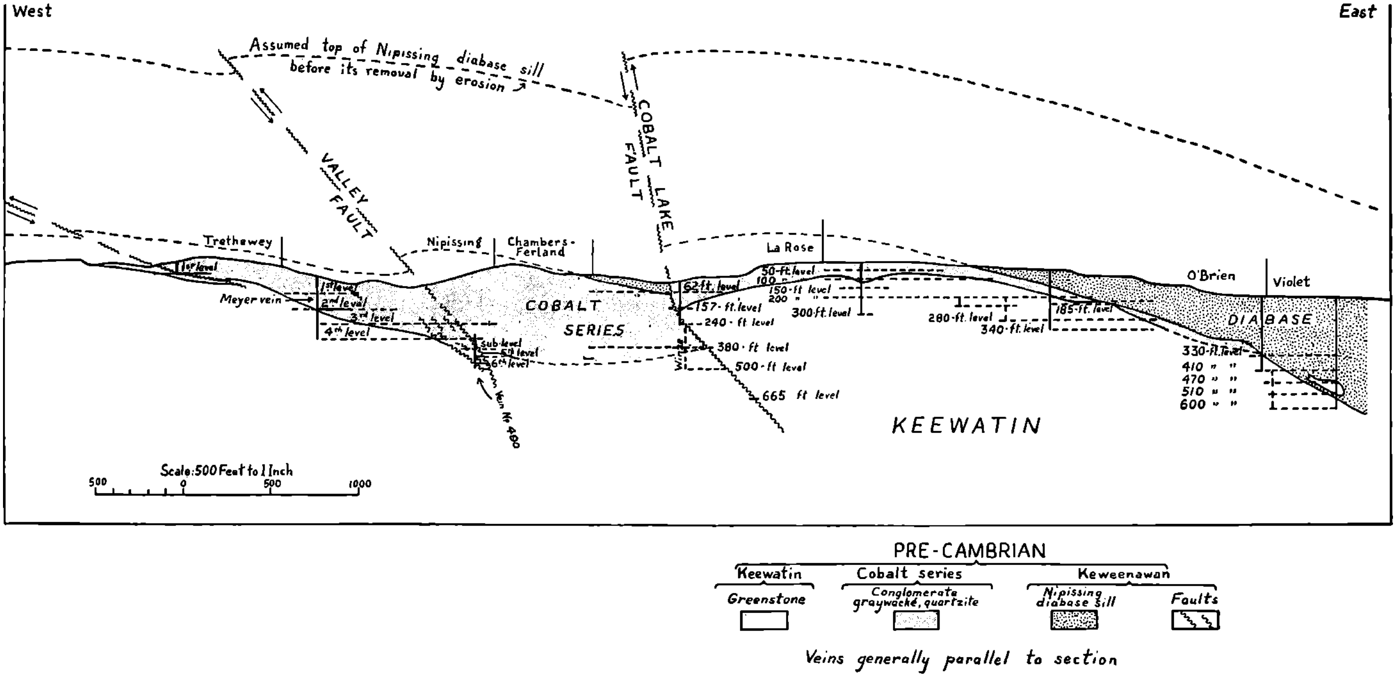

FIG. 248.—Vertical section through north part of Cobalt, District of Timiskaming, Ontario. (*After C. W. Knight.*)

After the deposition of the Huronian beds an irruption of diabase took place, assuming the form of sills up to 1,000 feet thick. The veins were formed after the intrusion of this diabase, which is regarded as of Keweenawan age; the fractures occupied by the veins were probably caused by folding and fracturing of the diabase. Figure 248 illustrates the geological relations and the subsequent erosion of the region to the present surface. Most of the productive veins occur in a shallow zone, 100 to 200 feet thick, near the base of the Cobalt series, and they rarely contain ore in the Keewatin, though one fissure (Nipissing 64) filled with calcite has been followed to a depth of 800 feet in the Keewatin. In few places does the ore descend to more than 300 feet. In exceptional cases, some ore was found at a depth of 1,000 to 1,600 feet (Beaver mine). In general, the smaller fissures could not maintain themselves in the tough Keewatin formation and in the diabase. Some ore has been mined at Cobalt from the upper contact of the diabase sill, and all the ore at Gowganda and South Lorraine has come from this contact.

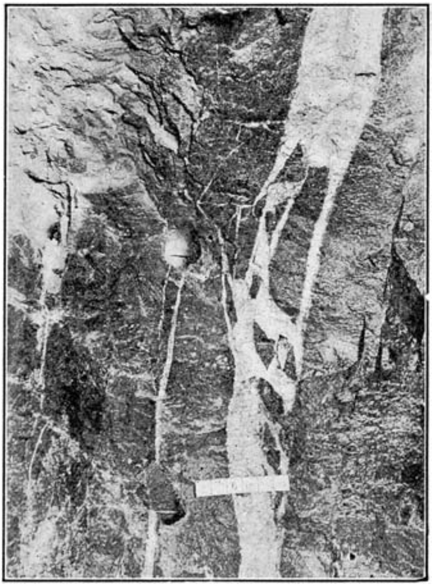

FIG. 249.—Photograph of vein of arsenides with inclusions of graywacke, Cobalt district. (*After W. L. Whitehead.*)

Several large faults (Fig. 248) intersect the district, and on these some pyrite-sphalerite-chalcopyrite ore may occur. These faults may have had some influence in damming of the ore solutions. Silver ore occurs at a few places at Cobalt on the faults, and most of the ore from South Lorraine has come from Wood's fault (a reverse dislocation with a throw of 30 feet). In general, the veins do not continue for long distance in the diabase. At Cobalt, the small non-continuous, ore-bearing veins number about 100. Few of them have been followed for more than 500 feet in length.

In the veins, which are seldom more than 6 inches in width, the filling is often "frozen to the walls" (Fig. 249). In some places the veins form a network of stringers. The arsenides of cobalt and nickel may fill the whole vein or may be distributed in calcite or dolomite gangue. The silver occurs irregularly distributed but frequently appears in sheets at the walls; the adjoining country rock is not greatly altered but may contain carbonates of dolomitic character and small specks of the various minerals, including native silver. The cobalt and nickel minerals occurring in the veins are:

| | | |
|---|---|---|
| Smaltite | $CoAs_2$ | isometric |
| Chloanthite | $NiAs_2$ | isometric |
| Löllingite | $FeAs_2$ | orthorhombic |
| Safflorite | $CoAs_2$ | orthorhombic |
| Rammelsbergite | $NiAs_2$ | orthorhombic |
| Skutterudite | $CoAs_3$ | isometric |
| Arsenopyrite | $FeAsS$ | orthorhombic |
| Gersdorffite | $NiAsS$ | isometric |
| Cobaltite | $CoAsS$ | isometric |
| Glaucodot | $(Co, Fe) AsS$ | orthorhombic |
| Niccolite | $NiAs$ | hexagonal |
| Breithauptite | $NiSb$ | hexagonal |

In addition there are native silver, dyscrasite,[1] and some chalcopyrite, galena, pyrite, pyrrhotite, sphalerite, and tetrahedrite. Among the rare minerals are bismuth, polybasite, stephanite, ruby silver, stromeyerite, and argentite. Cobaltite, safflorite, arsenopyrite, and native silver appear to be most abundant. Smaltite is not common. Erythrite, (cobalt bloom ($Co_3As_2O_8.8H_2O$), annabergite (nickel bloom ($Ni_3As_2O_8.$-$8H_2O$)), and scorodite ($FeAsO_4.2H_2O$), and some supergene silver occur close to the surface. At the Keeley mine, South Lorraine, probably pre-glacial oxidation has proceeded to a depth of 500 feet.

An analysis by A. R. Ledoux of two carloads shipped to the smelter shows, besides about 20 per cent Ca-Mg carbonate and 3.34 per cent silica, the following constituents, in per cent: Co 8.63, Ni 13.87, Ag 5.31, Sb 1.46, As 42.46, Fe 1.78.

The arsenides are intimately intergrown, and opinions as to paragenesis are conflicting. E. Thomson (1930) suggests that the di-arsenides, tri-arsenides, and sulpharsenides of cobalt and nickel were the earliest, the isometric forms apparently crystallizing first. These were followed by the arsenides and sulpharsenides of iron; and lastly came the monoarsenides, particularly niccolite. Later, the calcite gangue was intro-

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[1] The silver and the dyscrasite contain in places a little mercury, rising in exceptional cases to 5 per cent. It is believed to be present as silver amalgam. G. H. Clevenger, *Econ. Geol.*, 10, 1915, p 770.

duced, followed by silver, argentite, and bismuth. Still later came the scant sulphides of zinc, copper, and lead, the tetrahedrite, and the silver sulpharsenides.

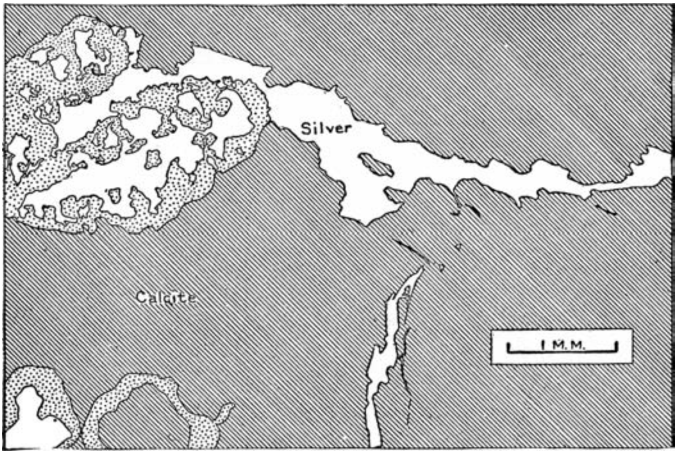

FIG. 250.—Replacement veinlet of silver traversing ferruginous calcite and smaltite (with niccolite). Arsenides are dotted. (*After E. S. Bastin.*)

W. Lindgren holds, with emphasis, that the calcite gangue was the earliest mineral, though additional deposition took place after the arsenides. The calcite is replaced by all arsenides, sulphosalts, and native metals. Among the arsenides, cobaltite is probably the earliest product and niccolite the last, often found replacing the interior part of the earlier minerals or the calcite included by them.

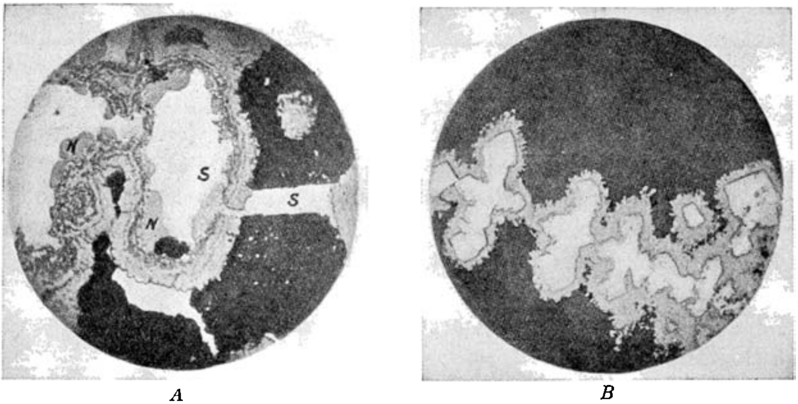

FIG. 251.—*A*. Native silver (*S*) within and outside of concretionary smaltite. *N*, Niccolite; black, calcite. Townsite Mine, Cobalt. Magnification, 22.5 diameters. (*After E. S. Bastin.*) *B*. Silver enclosed in smaltite. Black is calcite. Casey Mine, Cobalt. Magnification, 20 diameters. (*After E. S. Bastin.*)

A moderate amount of crystallized arsenides, also some silver, replace the country rock. In the filled veins fine-grained granular textures are common in which a succession is difficult to establish. A dendritic development of di-arsenides with the interior occupied by calcite or silver is also common. The dendrites are often covered by successive colloform deposits of later arsenides; some textures are plainly colloform, the arsenides showing rounded outlines and, after removal of the calcite, excellent contraction cracks (Gowganda).

The acid emanations of chlorides and sulphides were rapidly neutralized by the calcite and colloidal solutions developed, from which the arsenides and sulphides were precipitated as gels which soon crystallized to fine granular or radial aggregates. The silver is distinctly later than the arsenides and replaces them or the included calcite (Figs. 250 and 251). It was deposited in crystalline form and by hypogene solutions. The so-called "concretionary" or "tubercular" texture, almost universally present (Fig. 252), results, according to Lindgren[1] and Boydell,[2] from gel replacement of calcite and crystallization of the gel.

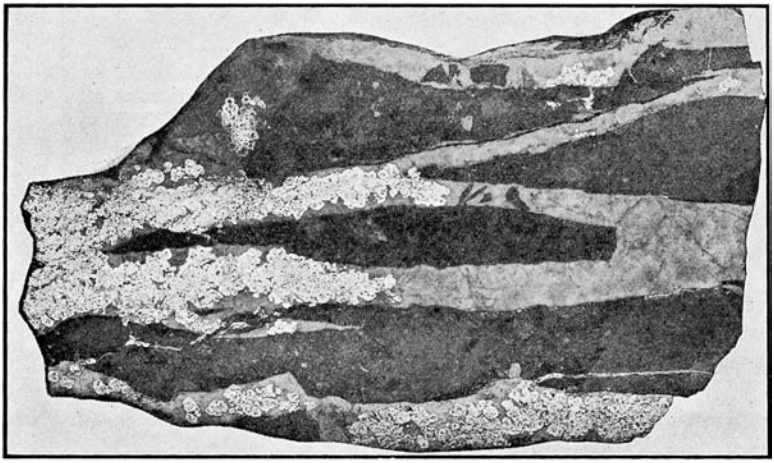

FIG. 252.—Photograph of specimen from Cobalt showing concretionary arsenides (white) in calcite veins. One-half natural size.

Many authors believe that the ore-forming solutions were derived from the diabase sills.

## QUARTZ-ADULARIA-ZEOLITE VEINS (ALPINE TYPE)

**Occurrence and Mineral Association.**—The so-called Alpine veins have little or no importance as a source of metalliferous ores, but many of them contain beautifully developed crystals, represented in all large mineral collections, and are prospected and worked to some extent for

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[1] *Bull.* Geol. Soc. Am., 36, 1925, p. 259.

[2] *Trans.*, Inst. Min. and Met. (London), Dec. 18, 1924, p. 219.

such specimens. They occur in the crystalline rocks in Tyrol, Switzerland, and the French Alps. There are several types. One class, found in the Zillerthal, in Tyrol, contains pyrite and galena with quartz, adularia, albite, epidote, calcite, prehnite, desmine, and laumontite. Another, in the French Alps, yields axinite, titanite, ilmenite, and many other minerals which indicate deposition at high temperature.

The veins in Switzerland have been studied in detail by J. Königsberger.[1]

The "veins" are approximately horizontal filled crevices in a biotite gneiss; they are really local openings in joints, which can be followed for considerable distances and which lie perpendicular to the schistosity of the rock. These crevices are surrounded by zones of altered rock not more than double the width of the opening; they usually contain open cavities into which the crystals project.

The biotite of the gneiss is altered to chlorite, epidote, and quartz; the plagioclase alters to sillimanite, kaolin, and epidote; albite and adularia crystallize in vugs; quartz and orthoclase are not attacked but the latter is often covered by secondary adularia. The kaolin is probably secondary, for the veins have for a long time been within reach of oxidizing waters.

It will be observed that the type of alteration is not that of ordinary veins; it seems most closely related to some veins formed near the surface—for instance, the Cripple Creek deposits. Sillimanite has not been shown to form in the wall rocks of metalliferous veins.

The general succession is: (1) smoky quartz and adularia (oldest); (2) calcite; (3) zeolites. The long list of other minerals includes fluorite, chlorite (often dusting the faces of quartz and adularia), apatite, albite, stilbite, heulandite, apophyllite, laumontite, chabazite, specularite, and rarely galena, chalcopyrite, and molybdenite.

**Origin.**—Königsberger concludes that the minerals were derived chiefly from the rock itself, and were deposited by the cooling of hot solutions containing carbon dioxide. The crystallization in the crevices took place at temperatures of 290° to 130° C. The Alpine veins may have been produced by intrusive after-effects at various temperatures.

Some of the quartz crystals contain large fluid inclusions; the liquid contains, in per cent, 85 $H_2O$, 5 $CO_2$, and 10 of solids, consisting of alkaline carbonates, chlorides, and sulphates in the order named.

As the smoky quartz is bleached at 300° C., it is assumed that it could not have been formed above this temperature. The bubbles of the

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[1] J. Königsberger, Die Minerallagerstätten im Biotitprotogin des Aarmassivs, *Neues Jahrbuch*, Beil. Bd. 14, 1901, pp. 43–49.

Ueber Alpine Minerallagerstätten, *Schweiz. Min. u. Pet. Mitt.*, Bd. 5, 1, 1925; *Abhandl.* d. Bayr. Akad. Wiss., 28, Bd. 12, 1919. See also review of R. L. Parker's work by Ian Campbell in *Am. Mineral*, 12, 1927, pp. 157–167.

inclusions disappear at 225° C. Because the liquid must have filled the cavity at the time of formation and because the pressure did not differ much from that of saturated water gas, the temperature of formation is about the same as that at which the bubbles disappear.

The red color of the fluorite disappears at 175° C., independent of the surrounding medium and pressure. The zeolites were probably formed at about 130° C.

## THE COPPER VEINS

The copper-bearing veins in which filling has been the principal process do not have the importance of the corresponding class of gold-quartz veins. Only when replacement has converted wider bodies of rock into ore and when enrichment by descending surface waters has acted upon the primary ore, do we find large and productive members in this class.

Many of the great copper deposits are mainly pyritic replacements of igneous or sedimentary rocks and are described on the following pages or with the high-temperature deposits in Chapter XXVII. Still others derive their importance from the accumulation of secondary chalcocite, either in wide replacement veins or in broad mineralized zones. Many of these, in fact, also belong to the high-temperature deposits as far as the poor, primary mineralization is concerned.

**Chalcopyrite-Quartz Veins.**—Simple veins containing mainly chalcopyrite, pyrite, and quartz, with some bornite, more rarely tetrahedrite and siderite, are common enough in many districts, but are rarely of great importance unless also containing gold and silver.

J. B. Umpleby[1] describes the Lost Packer vein, in central Idaho, which cuts through mica schist and which, in a gangue of quartz and siderite, contains chalcopyrite and some pyrrhotite and pyrite. The ore contains about 3 ounces of gold per ton. The vein is intersected by dikes and is probably a high-temperature deposit.

**Bornite-Quartz Veins.**—Quartz veins containing bornite are not uncommon but are rarely of great economic importance. Veins of this kind occur in the Virgilina district,[2] Virginia and North Carolina, where F. B. Laney noted interesting intergrowths of bornite and chalcocite (p. 840). The deposits are, however, lenticular veins and probably were formed at high temperatures.

**Pyrite-Enargite Veins.**—The enargite-bearing veins constitute a less common yet in places a most important class.

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[1] *Bull.* 530, U. S. Geol. Survey, 1913, pp. 70–73.

[2] W. H. Weed, *Bull.* 455, U. S. Geol. Survey, 1911, pp. 67–93.

F. B. Laney, *Bull.* 21, North Carolina Geol. and Econ. Survey, 1910; and *Bull* 14, Virginia Geol. Survey, 1917.

Enargite is, on the whole, a rare mineral and rather favors the mesothermal deposits. At Mancayan, in Luzon, in the Philippine Islands, enargite is found in a large replacement deposit in "quartz porphyry" and "andesite," formerly worked on a large scale and containing, in a quartz gangue, tetrahedrite, bornite, and enargite (luzonite). Another deposit is that of Bor,[1] in Serbia, which is also an irregular quartz-ore replacement in a rock that is called andesite but is really an intrusive porphyry with holocrystalline ground mass. The principal minerals are pyrite, enargite, quartz, barite, and secondary covellite.

The real home of the enargite group of minerals is in the Cordilleran region of North and South America. In Argentina, in the Sierra Famatina,[2] veins carrying enargite and famatinite ($Cu_3SbS_4$) break through clay slate, intruded by Tertiary granitic rocks and "dacite." Many similar veins are found near Chuquicamata, Chile; and this great deposit itself, though now mainly worked for its oxidized ores, carries enargite in depth. Enargite veins also probably occur at Collahuasi, Chile, and they are worked extensively at Cerro de Pasco,[3] at Morococha, and at Quiruvilca in Peru. No good descriptions of these Peruvian mines have been published. At Morococha, quartz monzonite, partly sericitized, is intrusive in Mesozoic limestones and sandstones; the ores occur as veins and blankets (mantos); the minerals are pyrite, tetrahedrite, enargite, in places with sphalerite, galena, rhodonite, and rhodochrosite. There is distinct evidence of zoning with copper ores in the central and deeper parts and lead-zinc-silver ores farther out.

The ores of the Cerro de Pasco district[4] occur principally along the eastern and southern margins of a body of tuff and breccia filling a volcanic vent about 1½ mile in diameter, and along a major zone of faulting which separates an area of Mesozoic limestone on the east from altered pre-Cambrian or early Paleozoic sediments and Tertiary tuffs and intrusives on the west. Tabular replacement bodies extend into the limestone roughly parallel to the eastward dipping bedding.

A small stock of quartz monzonite porphyry intrudes the tuffs on the western side of the district. The porphyry is locally sericitized and pyritized.

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[1] M. Lazarevic, *Zeitschr. prakt. Geol.*, 1912, pp. 337–370.

[2] G. Bodenbender, El Nevado de Famatina, *Anales* Minist. Agricult., Seccion Geol., 16, No. 1, 1922.

[3] G. Steinmann, Gebirgsbildung und Massengesteine in der Kordillere Südamerikas, *Geol. Rundschau,* 1, 1910, pp. 13–35. Über gebundene Erzgänge in der Kordillere Südamerikas, Int. Congress, Düsseldorf, 1910.

D. H. McLaughlin, Geology and physiography of the Peruvian Cordillera, *Bull.* 35, Geol. Soc. Am., 1923–1924, pp. 591–632.

[4] Personal communication from the Cerro de Pasco Copper Corporation through D. H. McLaughlin and L. C. Graton.

The ores are mined from numerous separate bodies of common origin. The most important types are: (1) enargite-famatinite, tetrahedrite, and pyrite veins with quartz gangue in bleached and silicified phyllite; (2) portions of a very large pyritic body along the major fracture zone in which enargite-famatinite and luzonite ores (with bismuth), silver-bearing galena-sphalerite ores, and pyritic silver ores (containing tetrahedrite, famatinite, pyrargyrite, stephanite, etc.) occur with a gangue of pyrite and cherty quartz with subordinate epidote and barite; (3) rich chalcocite ores formed by supergene enrichment in portions of the pyritic body; and (4) oxidized siliceous silver ores (pacos) in an extensive blanket over the pyritic ores.

The enargite occurs both as large conspicuous crystals and as fine-grained, slightly pinkish "luzonite." Chalcopyrite in small amounts is widely distributed.

Tetrahedrite is probably the chief silver-bearing mineral. In the greater part of the pyritic silver ore, however, silver minerals are inconspicuous.

Galena and sphalerite occur in abundance in parts of the massive pyritic body and with pyrite in replacements in limestone on the eastern side of the district. These minerals are uncommon in the richer enargite veins.

Oxidation extends to depths between 50 and 150 feet, and locally to 500 feet. Most of the oxidized material over the pyritic body is a low-grade silver ore (pacos), high in silica, with copper and zinc thoroughly leached.

Supergene chalcocite with a little covellite is abundantly developed in parts of the pyritic ore-body. The richest ore is black, pulverulent material. It contains much pyrite, but the chalcocite is chiefly a replacement of sphalerite. Pyrite grains are thinly coated or tarnished but not extensively replaced. The secondary copper ores lie immediately below the oxidized ground but become reduced to small size before the 500-foot level is reached and pass with depth into pyritic material, usually lean in copper but in places high in zinc and lead.

Silver enrichment in the sulphide ores is believed to be relatively slight.

The mineralization probably occurred shortly after the intrusion of the porphyry in the early Tertiary. Folded overlying rocks may have formed a cover of over 10,000 feet when the ores were introduced, but no certain estimate of post-ore erosion can be made.

In the United States, enargite occurs in mesothermal and epithermal veins, in the San Juan district and in Gilpin County, Colorado; at Bingham and, especially, in the Tintic district, Utah, and many other places.

**Butte, Montana.**[1]—The most prominent example of the enargite type of deposit is found in the Butte district, which for some time has been the most productive copper camp in the world.

In recent years, the Butte district has produced annually about 124,000 tons of copper, 10,000,000 ounces of silver, and $550,000 in gold, besides some lead and 64,000 tons of zinc, a total value per annum of about $50,000,000. The largest production was achieved in 1918, when the output exceeded 150,000 tons of copper and 15,000,000 ounces of silver with a total value of $103,331,000. The copper production began in 1879. The ores are concentrated and smelted at Anaconda and at Great Falls; they average 3 to 4 per cent copper and 4 ounces of silver per ton. The total value of the output of the Butte district (comprising only a few square miles) from 1882 to the end of 1929 is calculated to have a value of $2,119,410,164. In point of metallic wealth extracted, it is probably only exceeded by the Witwatersrand gold district of South Africa.

The copper deposits form a system of steeply dipping veins cutting quartz monzonite as well as a few dikes of aplite and granite porphyry (Modoc porphyry) intrusive into this granular rock. They are mainly pyritic replacements along fissures and contain pyrite with enargite, tennantite, bornite, chalcocite, sphalerite, and a little chalcopyrite and covellite. There is a scant gangue of quartz.

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[1] S. F. Emmons, W. H. Weed, and G. W. Tower, Jr., Butte Folio, *Geol. Atlas* 38, U. S. Geol. Survey, 1897.

W. H. Weed, Geology and ore deposits of the Butte district, Montana, *Prof. Paper* 74, *idem*, 1912.

J. F. Simpson, *Econ. Geol.*, 3, 1908, pp. 628–635.

Reno H. Sales, Superficial alteration of Butte veins, *idem*, 5, 1910, pp. 15–21; 3, 1908, pp. 326–331. Ore deposits of Butte, Montana, *Trans.*, Am. Inst. Min. Eng., 46, 1914, pp. 3–106.

C. T. Kirk, Conditions of mineralization in the copper veins at Butte, *Econ. Geol.*, 7, 1912, pp. 35–820.

J. C. Ray, Paragenesis of the ore minerals in the Butte district, *idem*, 9, 1914, pp. 463–481.

D. C. Bard and H. M. Gidel, Mineral associations in the Butte district, *Trans.*, Am. Inst. Min. Eng., 46, 1914, pp. 123–127.

A. P. Thompson, The occurrence of covellite at Butte, *idem*, 52, 1916, pp. 563–596.

W. W. Atwood, The physiographic conditions at Butte, etc., *Econ. Geol.*, 11, 1916, pp. 697–740.

Many authors, Mining methods in the Butte district, *Trans.*, Am. Inst. Min. Met. Eng., 72, 1925, pp. 234–287.

A. Locke, D. A. Hall, and M. N. Short, *idem*, 70, 1924, pp. 933–963 (Chalcocite).

W. M. Agar, Minerals of the intermediate zone, *Econ. Geol.*, 21, 1926, pp. 695–707.

W. Lindgren, Paragenesis of minerals in the Butte veins, *idem*, 22, 1927, pp. 304–307.

Extensive sericitization along the veins is characteristic. The age of the intrusive and of the veins is Eocene or late Cretaceous. The Butte district covers an area 5 miles square. The copper-producing area, 2 by 3 miles, occupies the southern and central part of this area. The principal zinc mines border the copper area on the north, west, and south. There is a central copper area of zinc-free ores surrounded by a zone of copper-zinc ores. There is very little copper in the zinc area proper.

The ores occur in a fissure system of great complexity (Figs. 51 and 52). The principal vein system strikes east-west. A second system known as the Blue, or northwesterly, veins intersect and fault, with moderate throw, the east-west veins which are also known as the Anaconda system. The dips are generally to the south at high angles. These two vein systems are cut by northeasterly trending fault fissures which dip northwest or southeast and which have caused dislocation of the older veins amounting in places to 200 or 300 feet. The fault veins sometimes carry ore as "drag" but do not contain large bodies of ore.

Experiments by W. J. Mead[1] suggest that the complicated vein system may have originated by a progressive shearing movement along parallel vertical planes.

The ore shoots are irregular; while some continue to the greatest depth reached in mining (3,000 to 4,000 feet), others have been bottomed at less depth.

The stope width is from 6 to 30 feet. Closely spaced mineralized fissures coming from the east-west veins may produce what is known as a "horse-tail structure" (Fig. 51) and form together a stockworks; this is particularly noticeable in the West Colusa area.

While the east-west veins are wider and more productive, there is not a great difference in mineralization between the two systems. Even when the veins are sharply defined, as is the case in the Badger mine and many other places, much of the ore has resulted from the replacement of country rock. A zone of sericitic and pyritic granite borders the veins, however small they may be. In the central part of the copper area there is much spreading replacement by ore and the outlines of the ore-bodies are indefinite.

The ore is a heavy, coarse-grained sulphide aggregate in which pyrite predominates; the quartz gangue is not prominent. There are few druses and little banding or crustification. The succession of the minerals is as follows, beginning with the oldest: quartz (long continued), pyrite, sphalerite, enargite, tennantite, bornite, chalcopyrite, chalcocite; scant magnesian and manganese carbonates; finally calcite closes the succession. Each sulphide mineral may replace any of the preceding sulphides. The

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[1] *Jour. Geol.*, 28, 1920, p. 512.

chalcocite is massive, rarely crystallized, but then always in orthorhombic forms. At the deposition of these crystals the temperature cannot have exceeded 90° C.

The silver-lead-zinc metallization is apparently later and the sulphides consist of pyrite, sphalerite (chalcopyrite), and galena. These minerals are usually shattered and brecciated by a quartz-rhodochrosite-rhodonite invasion, with or without fluorite, which in turn is followed by a scant second generation of pyrite and chalcopyrite.

In contrast to the hypogene chalcocite, there is an upper, deep zone in which descending cool solutions have deposited supergene chalcocite. This has been effected by replacement of the other sulphides by solutions of cupric sulphate. This zone of "sooty chalcocite" yielded the rich ores mined in the earlier years of the camp. However, much of this ore was not "sooty" but compact and mixed with bornite.

The outcrops are not prominent and the copper is leached from them; in some places they contain chrysocolla. The depth of this oxidized barren zone of honeycombed quartz veins ranges from 10 to 400 feet. In the central copper area, where the granite is greatly altered, the upper limit of the sulphides is practically a plane in spite of surface inequalities of nearly 300 feet, evidently depending more upon the alteration of the country rock than on the topography. In the leached zone there is a slight enrichment of silver, the material containing as much as 30 ounces per ton, in contrast to 2.5 ounces per ton in the sulphide ore. Supergene argentite and native silver are rare. The gold tenor is uniform throughout the copper area, indicating that practically no secondary concentration has taken place. The ores average 30 to 50 cents per ton in gold, with little difference between oxidized ore and that of the sulphide zone. A sharp line of demarcation separates the two zones, the change in many places occurring within 2 or 3 feet vertically. At this level the ores contained much chalcocite and averaged 8 per cent or more of copper. Solid masses of chalcocite ore, 15 feet or more in thickness, were found. Covellite is less abundant and secondary chalcopyrite is rare. In depth the enriched ore gradually decreases in value, but supergene chalcocite persists in places to depths of 3,000 feet or more, particularly along places where the circulation of water was energetic. In general, chalcocitization in the upper levels was accompanied by more or less kaolinization.

Below this, enargite began to come in but chalcocite still persisted and is even now found in the deepest levels (3,500 feet below the surface) though there is much pyrite, enargite and other older minerals. Covellite occurred in considerable amounts in the upper levels and as deep as 2,000 feet or more. Reno Sales, J. C. Ray and others long ago expressed the belief that this deeper and more massive chalcocite is of hypogene origin though later than the first succession, which is represented by quartz, pyrite, and enargite. Some geologists also believe that the covellite may be of hypogene origin.

It is conceded that the present barren zone was created by leaching of the former upper part of the chalcocite zone and that during this process the copper was sharply concentrated in the upper part of the present enriched zone. Therefore the latter, like many other bodies of secondary chalcocite, must be of considerable geological age. When it was accumulated the water level was assuredly much higher than at present. This view is supported in a notable paper by W. W. Atwood in which physiographic principles are applied to the study of supergene enrichment.

## THE PYRITIC REPLACEMENT DEPOSITS

While pyrite is a persistent mineral, crystallizing within a wide range of temperatures, it is easily apparent that the deposits containing large masses of pyrite have not been formed close to the surface, but rather at considerable depths and at temperatures well above 100° C. The deposits of this type are mainly confined to regions that have been deeply eroded since the deposits were formed, and many of them show a mineral association indicating high temperature.

In many textbooks the pyrite deposits are treated as a distinct class and are assumed to have a similar origin. We know now that they comprise deposits of widely differing origin and history.

In a broad way we may distinguish (1) those associated with silicates such as amphibole, pyroxene, epidote, tourmaline, and garnet, the iron sulphide being in part present as pyrrhotite, and (2) those associated with calcite, barite, and quartz as gangue minerals. The deposits of the first class were undoubtedly formed at considerably higher temperatures than those of the second and, in general, probably also at greater depth.

Class 1 comprises (A) some deposits of magmatic origin like those of Sudbury, Ontario; (B) a large division, of contact-metamorphic type, like the deposits of Ducktown, Tennessee; Granby, British Columbia; and the Highland Boy mine at Bingham, Utah; (C) a third division, difficult to interpret but thought by many to represent a phase of igneous injection. In each of these three divisions the deposit may have been subjected to dynamic metamorphism, with the attendant development of amphibole and garnet and of schistose structure. Many of these metamorphosed deposits have had a complicated history and are among the most difficult to interpret and classify.

Class 2 is also connected with the eruption of igneous rocks, but the high-temperature minerals are absent; the deposits of Rammelsberg in the Harz, Germany, of Mount Lyell, Tasmania, of Rio Tinto, Spain, and of Shasta County, California, may serve as examples. The deposits at

Kyschtim,[1] in the Ural Mountains, at Tyee,[2] on Vancouver Island, at the Britannia mine, B. C.,[3] at Besshi,[4] and Hitachi[5] in Japan, as well as certain occurrences in Arizona, described below, also appear to belong to this class. The ores, while consisting mainly of pyrite, derive their value from a small percentage of chalcopyrite; there are usually minute quantities of gold and silver, and frequently also sphalerite and a little galena; other sulphides are rare. The ores are fine-grained and massive. In many of the deposits mentioned a genetic connection with rhyolite porphyries, alaskite porphyry, or keratophyre can be established, and deposition by hydrothermal replacement at moderate depth and temperature seems the most reasonable explanation.

Some pyritic deposits of Class 2 will be briefly characterized in the following paragraphs.

### Copper Deposits of Jerome, Arizona[6]

The pyritic replacement deposits of the Bradshaw Mountains and Jerome are contained in pre-Cambrian schists of sedimentary and igneous origin. Their development followed the intrusion of the, also pre-Cambrian, Bradshaw granite. The most important mine is the

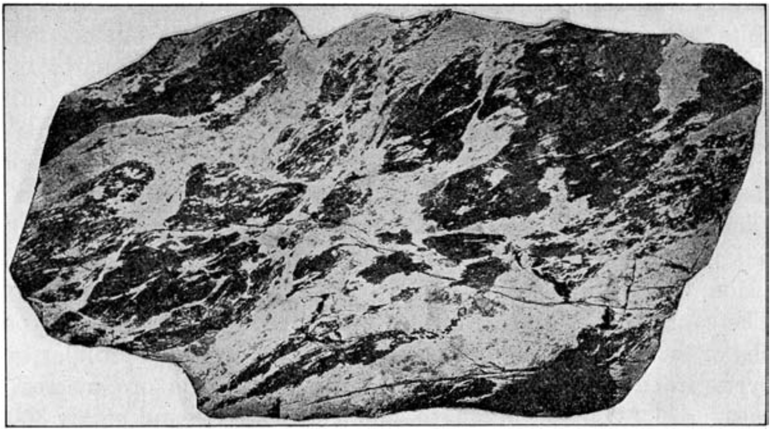

FIG. 253.—Chalcopyrite (white) replacing chloritic schist (dark gray). United Verde mine, Arizona. One-half natural size.

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[1] A. W. Stickney, *Econ. Geol.*, 10, 1915, pp. 593–633.

[2] C. H. Clapp, *Mem.* 13, Canada Geol. Survey, 1912, pp. 180–187.
V. Dolmage, *Econ. Geol.*, 11, 1916, pp. 390–394.

[3] S. J. Schofield, *idem*, 21, 1926, pp. 271–284.

[4] M. Watanabe, *idem*, 18, 1923, pp. 174–178. This mine is developed to a depth of 4,000 feet.

[5] Watanabe and Landwehr, *idem*, 19, 1924, pp. 434–454.

[6] L. E. Reber, Jr., *Trans.*, Am. Inst. Min. Met. Eng., 66, 1922, pp. 3–26.
W. Lindgren, *Bull.* 782, U. S. Geol. Survey, 1926.
J. L. Fearing, *Econ. Geol.*, 21, 1926, pp. 757–773.

United Verde, which from 1888 to the end of 1929 had produced over 700,000 tons of copper with some gold and silver. The deposit forms a pipe, with a diameter of several hundred feet, and has been followed to a depth of 3,500 feet. The fine-grained ore replaces rhyolite schist (Fig. 253) and contains from 4 to 6 per cent of copper. Pyrite is followed by sphalerite, chalcopyrite, and tennantite. The gangue minerals are quartz, chlorite, dolomite, calcite, and sericite. There is no barite, nor are there any distinctive high-temperature minerals present.

### Copper Deposits of Shasta County, California[1]

Copper deposits which have been actively mined and smelted since 1895 are found in a number of districts in Shasta County, California; among the more prominent mines are Iron Mountain, Bully Hill, Mammoth, and Balaklala. The production of copper in 1929 was 6,034,000 pounds.

The sedimentary rocks consist of Devonian and Carboniferous closely folded slates and contain intrusions of a highly siliceous and sodic alaskite porphyry, which is the country rock of almost all the important copper deposits. Somewhat later than the alaskite porphyry, but belonging to the same (early Cretaceous) period of intrusion, is a quartz diorite, probably equivalent to the granodiorite of the Sierra Nevada. No copper deposits occur in the quartz diorite, but it contains workable gold-bearing quartz veins. Deep erosion has taken place since the period of intrusion; Graton estimates the depth of rocks removed as not less than 5,000 or 6,000 feet. The rocks have been subjected to some shearing and brecciation, but to little extensive dynamo-metamorphism, since the intrusion.

The copper deposits were formed during the interval between the two epochs of intrusion. The ore-bodies are large, irregular tabular masses of pyrite with some chalcopyrite (Fig. 254); single ore masses have dimensions of 1,200 feet in length, 300 feet in width, and nearly 300 feet in thickness, and some of them contain many million tons of ore. The Iron Mountain mass before a great part of it was converted to gossan probably contained 20,000,000 tons of ore, exclusive of the large amount which has been removed by erosion. Many of the bodies lie flat and are easily accessible by tunnels.

The ores contain chiefly pyrite with about 3 per cent of copper as chalcopyrite, and as much as $2 per ton in gold and silver, about equally divided between the two metals. Sphalerite is present in varying

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[1] J. S. Diller, Redding Folio, *Geol. Atlas* 138, U. S. Geol. Survey, 1906.

L. C. Graton, The occurrence of copper in Shasta County, Cal., *Bull.* 430, U. S. Geol. Survey, 1910, pp. 71–111.

A. C. Boyle, The geology and ore deposits of the Bully Hill mining district, *Trans.*, Am. Inst. Min. Eng., 48, 1915, pp. 67–117.

amounts, sometimes in later veins and the ore contains a little bismuth, arsenic, and selenium. The gangue minerals include quartz, calcite, barite, gypsum, and anhydrite; the succession is in general pyrite (oldest), sphalerite, chalcopyrite, quartz, and barite. In places there is a deep oxidized zone with sulphide enrichment.

The alaskite porphyry near the ore-bodies is more or less altered and contains sericite (probably also paragonite), secondary quartz, chlorite, pyrite, carbonates, and epidote. The pyritic ores occur in sheared and brecciated zones and are replacements of the surrounding porphyry caused by hot solutions emanating from the cooling alaskite porphyry. The action of surface waters in the ore deposition is probably negligible, for at that time the alaskite porphyry was everywhere covered with a blanket of impermeable shales.

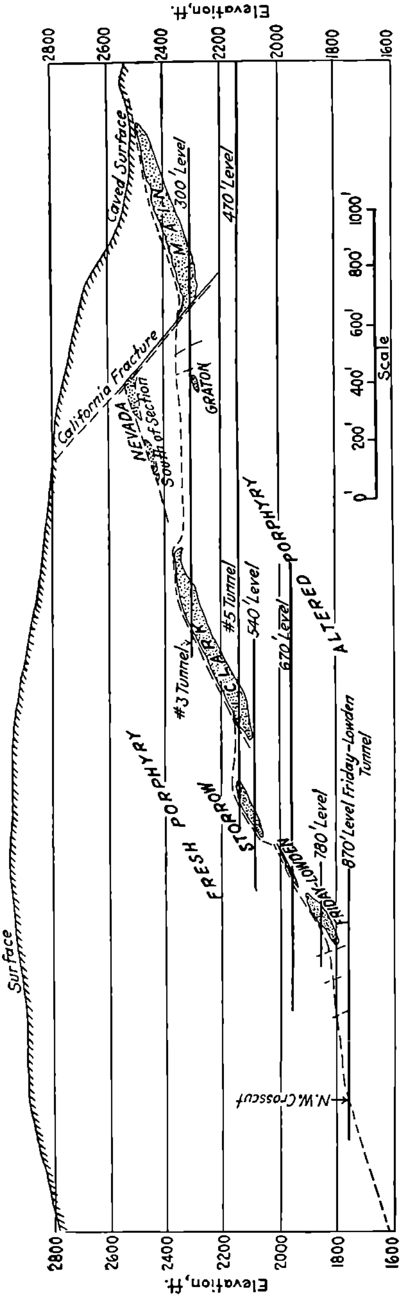

Fig. 254.—Vertical section of ore-bodies at Mammoth, Shasta County, California.

### THE COPPER DEPOSITS OF BRITANNIA MINES, BRITISH COLUMBIA

For many years the Britannia mines[1] have produced large amounts of copper. In 1930, 2,200,000 tons of ore were mined yielding 22,150 tons of copper with minor amounts of gold and silver. The locality is a short distance north of Vancouver and the main tunnel is 1,700 feet above the sea. The deposits occur in a belt of highly sheared rocks, 5 miles long

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[1] S. J. Schofield, The Britannia Mines, British Columbia, *Econ. Geol.*, 21, 1926, pp. 271–284.

H. T. James, Britannia Beach Map-Area, *Mem.* 158, Canada Geol. Survey, 1929, 159 pp.

by 2,000 feet wide and have been mined to a depth of 2,500 feet. The slates, tuffs, and intruded dacites and latites are regarded as early Mesozoic and now constitute a roof pendant in the Coast Range batholith.

The various deposits are vein-like replacements in this shear zone. They contain pyrite, chalcopyrite, quartz, sericite, and anhydrite. This mineralization is regarded as a later phase, preceded by a less important sphalerite-barite mineralization not recognized below the 1,050-foot level.

## The Copper Deposit of Mount Lyell, Tasmania[1]

At Mount Lyell, on the west coast of Tasmania, is one of the large copper deposits of the world. According to Gregory, the ore-bodies replace sericite schists, which are dynamo-metamorphic forms of perhaps Paleozoic acidic porphyries. The Devonian conglomerates are probably of later age.

The main ore-bodies are lenticular and lie in part in the sericite schist and in part at its contact with the conglomerates, which have been brought against the schist by faulting. The largest ore-body is that of the Mount Lyell mine; this is several hundred feet long and 200 feet wide, but appears to be limited in depth by a fault.

The ore consists mainly of fine-grained pyrite with some gangue of quartz and barite. It also contains 5 to 6 per cent of copper in the form of chalcopyrite, more rarely bornite and enargite.

The ore yields very little gold and 1.33 ounces of silver to the ton. The output of copper for many years has been about 6,000 tons per annum. Pyrite is the oldest mineral; it was followed by chalcopyrite, bornite, enargite, and tetrahedrite.

Enrichment near the surface and in the footwall of the deposit has added much to the wealth of the property. The outcrops at one place contained quartz, barite, hematite, and about 15 ounces of silver and $15 in gold per ton. The presence of hematite in the outcrops would suggest that at some time during the weathering of the deposits the climate was warmer than now.

## The Pyritic Deposits of Rio Tinto, Spain[2]

**General Features.**—The pyritic ore-bodies of the southern Spanish province of Huelva, more generally known as the deposits of Rio Tinto,

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[1] J. W. Gregory, The Mount Lyell mining field, Tasmania, *Trans.*, Austral. Inst. Min. Eng., Melbourne, 10, 1905, pp. 26–197.

C. G. Gilbert and J. E. Pogue, The Mount Lyell copper district of Tasmania, *Proc.*, U. S. Nat. Mus., 45, 1913, pp. 609–625.

[2] J. Gonzalo y Tarin, Descripción, etc., de la provincia de Huelva, *Mem.*, Com. Mapa Geol. España, Madrid, 1–3, 1886–88.

J. H. L. Vogt, Das Huelva Kiesfeld, etc., *Zeitschr. prakt. Geol.*, 7, 1899, pp. 241–254.

are probably the greatest in the world and have been mined since Phoenician and Roman times. The deposits are in the main lenticular; there are at least 50 of these pyritic lenses, whose length varies from 1,200 to 6,500 feet, while the width, in general proportional to the length, reaches a maximum of 250 feet and the depth ranges from 500 to 1,800 feet. The vertical range of deposition, according to Finlayson, probably in no case exceeded 3,300 feet, and few of the deposits attain a depth of 1,000 feet.

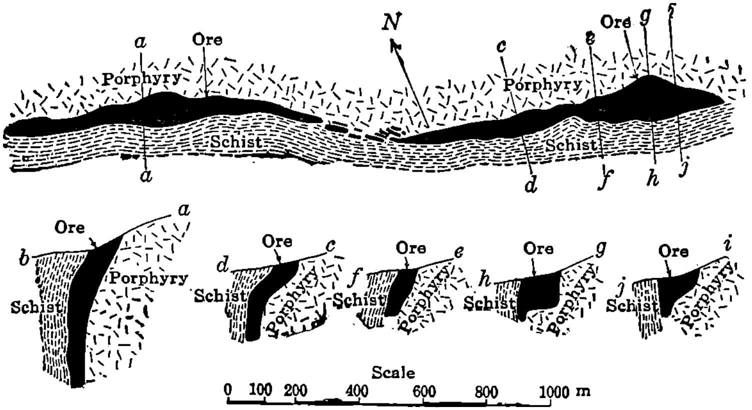

FIG. 255.—Plan and cross-section of ore-bodies, Rio Tinto, Spain. (*After Gonzalo y Tarin.*)

In the slates the deposits often taper downward to a point, while in the porphyry a flat or rounded lower surface is not uncommonly observed. On the whole they appear to lie conformably between slates and porphyry or in either porphyry or slate.

The production of these deposits has always been large and is well maintained; in 1926 it was 40,000 metric tons of copper, all sources considered. Besides the regular copper ore with more than 2 per cent Cu, up to 4,000,000 metric tons of pyrite, poor in copper, are shipped for sulphuric acid manufacture. A part of the copper is recovered as a precipitated cement or a sulphate.

**Geological Formations.**—The rocks consist of (1) a uniform series of folded and compressed clay slates and graywacke, striking east and west and believed to be of Devonian and Carboniferous age; (2) granites and granodiorites intrusive into the Carboniferous rocks, north of the district; (3) several varieties of porphyry, including rhyolite porphyry and kerato-

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Bruno Wetzig, Beiträge zur Kenntniss der HuelvanerKieslagerstätten, *Zeitschr. prakt. Geol.*, 14, 1906, p. 173.

A. Moncrieff Finlayson, The pyritic deposits of Huelva, Spain, *Econ. Geol.*, 5, 1910, pp. 357–372, 403–437.

H. F. Collins, *Trans.*, Inst. Min. and Met. (London), 31, 1922, pp. 61–84.

A. M. Bateman, Ore deposits of the Rio Tinto District, *Econ. Geol.*, 22, 1927, pp. 569–614.

phyre, some varieties with granophyric structure; in places the porphyry is affected by shearing and schistosity; (4) basic dikes and sills, mainly of diabase, but including also camptonites and diorites. Finlayson shows quite conclusively that the porphyries are intrusive masses, occurring in belts and lenses throughout the field (Fig. 255). The basic rocks cut both granites and porphyries.

Displacements of considerable throw occur along many ore-bodies. According to Finlayson, the last event in this series of igneous and dynamic disturbances consisted in the development of the mineral deposits.

The lodes that occur in the slate are in the main conformable with the bedding, but the ore sometimes, according to Finlayson, cuts across it; the lode walls are usually well defined and smooth; the deposits occur, as a rule, along contacts or other lines of weakness and crushing. According to the same author, the adjoining rock shows effects of hydrothermal action in marked degree, the porphyry being transformed into an aggregate of chlorite, sericite, quartz, carbonates, and pyrite. Analyses show extremely well-marked carbonatization and sericitization, entirely similar to the alteration occurring in the California type of gold-quartz veins, and undoubtedly of hydrothermal origin. The alteration is illustrated by the following analyses, quoted from Finlayson:

ANALYSES OF FRESH AND ALTERED PORPHYRY FROM THE SAN DIONISIO MINE, RIO TINTO, SPAIN

| | Fresh. | Altered. |
|---|---|---|
| $SiO_2$ | 76.21 | 70.68 |
| $Al_2O_3$ | 12.66 | 11.45 |
| $Fe_2O_3$ | 2.98 | 1.31 |
| $FeO$ | 1.46 | 0.72 |
| $MnO$ | 0.08 | 0.05 |
| $CaO$ | 1.15 | 2.28 |
| $MgO$ | 0.10 | 0.17 |
| $K_2O$ | 3.27 | 4.85 |
| $Na_2O$ | 1.64 | 0.65 |
| $H_2O+$ | 0.18 | 0.23 |
| $H_2O-$ | 0.35 | 1.41 |
| $CO_2$ | 0.09 | 5.08 |
| $FeS_2$ | | 1.27 |
| | 100.17 | 100.15 |

**The Ores.**—The ores consist chiefly of almost massive pyrite, with but a small amount of quartz and few other gangue minerals, although barite occurs in some localities. There are also disseminated ores which clearly represent replacements of slate or porphyry. Banded or pressed ore is rarely seen. The primary ore carries from 48 to 50 per cent of sulphur. Chalcopyrite is present in minute scattered grains, or as threads and strings traversing the granular pyrite and filling interstices in the ore. Sphalerite and galena occur in small amounts, and there are traces of bismuth, selenium, and tellurium. Arsenic varies from 0.25 to 1 per cent. The succession is pyrite (oldest), chalcopyrite, sphalerite, galena. Tetrahedrite and enargite are present.

Especially interesting are the changes in the ore produced by weathering. The outcrops are gossans of hematite carrying 10 to 15 per cent of silica and alumina and little or no copper. The average depth of this gossan is 100 feet. The lower limit of the gossan is well-defined; and the line of contact between it and the sulphide ore is sometimes marked by a thin earthy material, which, as described by Vogt, is rich in gold and silver. The top portion of the sulphide zone, for a thickness of 3 feet or more, is composed of leached pyrite with traces of copper (Finlayson). Below this commences the zone of enriched sulphides, in which the ore assays from 3 to 12 per cent of copper. This enrichment is effected by deposition of chalcocite, and its influence may be traced to depths of 300 feet below the surface. In the San Dionisio lode, enrichment was noted even at a depth of 1,000 feet, indicating, according to Finlayson, that the secondary changes extend far down into what is usually regarded as primary ore. Wetzig states that in the Cabeza de Pasto mine the ore at the 40-meter level contained 3.5 per cent of copper, at the 60-meter level 3 per cent, and at the 80-meter level 2 per cent. The tenor of the primary ore ranges from 0.5 to 2 per cent of copper.

**Genesis.**—The origin of these deposits has been the subject of long discussion among geologists. The papers presented at the Fourteenth International Congress in Madrid (1926) well illustrate the diversity of opinions. The earlier geologists believed in a sedimentary origin, and this view is still held by some. Gonzalo y Tarin, as well as de Launay, held them to be veins or lodes deposited in open cavities by ascending solutions. Later, Vogt considered the deposits to be of pneumatolytic nature, formed as an after effect of the extrusions of porphyry. In the latest contributions by Finlayson and Bateman it appears to be firmly established that the deposits were formed by the hydrothermal replacement of crushed and sheared zones. Finlayson believes that the deposits were formed after the intrusion of the basic dikes and sills, which are considerably later than the porphyry, and thinks that the concentration of the ores was in the first place due to a process of magmatic concentration of sulphides, accompanying the differentiation of the series of intrusive rocks and dependent, with the latter, on the Hercynian tectonic movements.

Evidence of pneumatolytic deposition seems to be lacking, nor do the deposits seem to indicate injection of magmatic molten sulphides. On the contrary, the processes of replacement probably proceeded at moderate temperature and at moderate depth. Magnetite and pyrrhotite are present in only a few deposits, such as those at Cala, and opinions differ as to whether these minerals are due to later dynamo-metamorphic or contact-metamorphic processes, or to original deposition.

### The Pyritic Deposit of Rammelsberg, Germany[1]

The Rammelsberg deposit (Fig. 256), lies on the northern slope of the Harz Mountains, in Germany, near the town of Goslar. It has been worked for copper and lead since ancient times, the first records dating back to the tenth century. Its geological structure has been investigated by a number of authors, but its complete and detailed description is as yet a problem of the future. The most diverse explanations have been offered as to its mode of origin. Many geologists, like Bergeat and Klockmann, lately also G. Frebold and P. Ramdohr, consider it a sedimentary deposit. Vogt, following Freiesleben and Lossen, explains it as a deposit from magmatic solutions. Dahlgrün says it is not syngenetic but formed by plutonic hydrothermal agencies and altered by regional metamorphism.

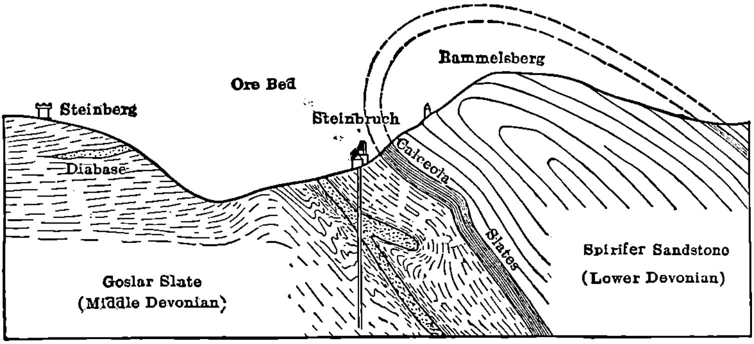

Fig. 256.—Cross-section of Rammelsberg, showing overturned anticline and supposed conformable position of the ore deposit. (*After F. Klockmann.*)

The deposit is enclosed, apparently conformably, in Middle Devonian slates, which form an overturned anticline and dip toward the north. These slates are overlain by a thick series of Lower Devonian Spirifer sandstone, which makes up the summit of Rammelsberg, at the foot of which the mine is located.

The slates have suffered considerable deformation and the ore-body follows their contortions more or less closely. The underground developments extend for 2,000 meters and have attained a vertical depth of

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[1] F. Klockmann, Berg- und Hüttenwesen des Oberharzes, 1895, p. 57.

J. H. L. Vogt, Über die Genesis der Kieslagerstätten vom Typus Röros-Rammelsberg, *Zeitschr. prakt. Geol.*, 1894, p. 173.

W. Lindgren and J. D. Irving, The origin of the Rammelsberg ore deposit, *Econ. Geol.*, 6, 1911, pp. 303–313.

F. Dahlgrün, Abstr., *Neues Jahrbuch*, Bd. 3 (1), 1930, pp. 1–2.

W. E. Schmidt, Das Rammelsberger Lager, *Zeitschrift für Berg-Hütten-und Salinenwesen* 81, 1933, pp. 247–270.

380 meters. Mining has not yet penetrated to great depths, in spite of the fact that the deposit has been worked for nearly 1,000 years.

The thickness of the "ore bed" swells to dimensions of as much as 30 meters, due to folding and local enlargement. In most places the thickness is not over 1 to 3 meters. The dip is uniformly 45° to the southeast.

The banded structure of the ore is in part unconformable with the stratification of the slate; but everywhere follows with great faithfulness the outlines of the sulphide mass, whether these be smooth or irregular.

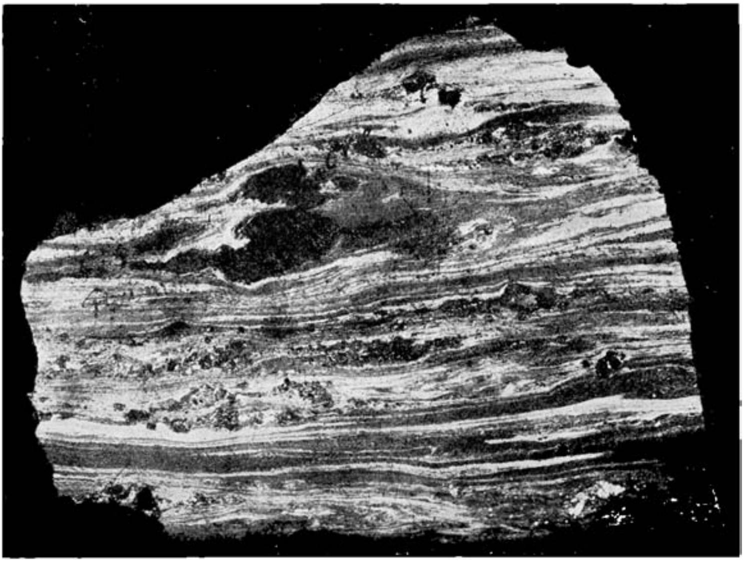

FIG. 257.—Pressed ore, Rammelsberg, Germany, with alternating streaks of chalcopyrite (light) and sphalerite (dark gray). Nodules of pyrite, black. Natural size.

The principal minerals are sphalerite, chalcopyrite, galena, pyrite, and arsenopyrite, which are abundant approximately in the order enumerated. The gangue is barite, but it is often inconspicuous. The ores themselves are entirely or predominantly composed of sulphides. Alteration of the enclosing slates is rarely observed. At most there is a slight impregnation of pyrite.

The texture and composition of the ore vary with the locality, so that in one place copper will predominate, while in other places the ore carries mostly sphalerite with a little galena.

The intergrowths of these minerals are fine-grained and intimate, which adds to the metallurgical difficulties of treating the ore. The most common texture is that of the banded ores, consisting of dominant sphalerite with narrow and gently curved streaks of chalcopyrite and galena. In places the ore contains rounded nodules, generally of pyrite, around which the fine-grained streaks of sphalerite and chalcopyrite bend in regular curves. Pyrite shows a strong resistance to such deformation (Fig. 257).

Similar phenomena of flowage have been described from the Slocan district.[1]

The structural relations of the ore-body indicate that the deposit is a bedded vein—that is, a fissure vein lying in part, at least, conformable to the surrounding slates. The distribution and structure of the ore itself are inconsistent with the theory of sedimentary deposition. The sulphides do not occur with their primary texture; their structure is that of a dynamo-metamorphic rock, in which all the constituents, except pyrite, have been drawn out into streaks which are intricately mingled. The appearance shown in the figure could be easily duplicated from any area of fine-grained schist resulting by pressure from an original granular rock. The different constituents have recrystallized and flowed under pressure (p. 749).

While the surrounding slates are soft they evidently behaved quite differently from the sulphide mass.

The association of minerals with barite tends to show that the deposit is not of the deep-seated type but was formed at a moderate depth below the original surface, probably within a few thousand feet of it.

The deposit may have been formed by ascending solutions derived from the neighboring batholith of granite, which is only 3 kilometers distant from the mine.

## DISSEMINATED COPPER DEPOSITS

The group just described carries a massive fine-grained ore, consisting of pyrite and chalcopyrite. The so-called disseminated copper ores are very different. They also are products of replacements but instead of forming veins or bed veins they have a large horizontal extension and their principal ore mineral is chalcocite. They are often referred to as "chalcocite blankets" or "porphyry ores" though they by no means always are contained in such rocks.

Large masses of schist, quartzite, monzonite, quartz monzonite, quartz diorite, or their porphyries have been weakly mineralized by quartz, sericite, pyrite, and chalcopyrite; such masses contain but a fraction to one half per cent of copper and the rock texture is very apparent. These weak mineralizations are called "protore," following Ransome. In their original form they have no value, but when acted upon by descending waters, chalcocite is deposited at and below the water level, thus giving rise to large, flat bodies of low-grade ore ranging from 1 to 4 per cent of copper. As they can be cheaply mined and concentrated, they form most valuable deposits which have been actively worked since 1900, and now furnish a large proportion of the copper production in the United

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[1] W. L. Uglow, *Econ. Geol.*, 12, 1917, pp. 643–662.

States. The Utah Copper, the Nevada Consolidated, the Morenci, the Ray, and the Miami are mines working such deposits. At a varying depth the supergene chalcocite ceases. On account of the importance of supergene processes in these deposits, they will be described in Chapter XXXII.

W. H. Emmons[1] has pointed out that these deposits are controlled by fracturing and that they are situated high on the cupolas or upward extensions of batholiths. To use Emmons' classification of deposits in their relation to the main masses of the batholiths, they are "acrobatholithic." They are derived from the ascending emanations from such cupolas.

### THE COPPER DEPOSITS OF THE BELGIAN CONGO AND NORTHERN RHODESIA

In this region, mining history has been made rapidly during the last 10 years. It is now clear that the copper deposits are of great extent and value and will prove a most important factor in the world production.

From the Cape to the Congo, Africa is chiefly a high plateau about 4,000 feet above the sea. On a basement of crystalline schists, largely of sedimentary origin and intruded granite, both probably old pre-Cambrian, rest unconformably various series of sediments.[2] In this region, that is, on both sides of the Belgian Congo-North Rhodesia boundary, which follows the Zambesi-Congo watershed, the most important series is the Katanga System (early Paleozoic or late pre-Cambrian), which again is divided into an upper part, the Kundelungu with about 12,000 feet of limestone and quartzite, and a lower, apparently conformable part, called the "Série des Mines," *or* the "Série schisto-dolomitique-cherteux" (Katanga); *or* the Bwana M'Kubwa Series, *or* the Roan Series (Rhodesia). Almost all of the copper deposits lie in the "Série des Mines" and this is divided into: (1) An upper part (M'washia, Christmas), 1,500 feet thick, of arkose, shales, and oölite; (2) the Upper Roan (1,500 feet), of dolomite and interbedded sandstones—these are the ore-bearing dolomites of Katanga; (3) the Lower Roan (1,000 feet), of shales, arkose, dolomites, and conglomerates. In Rhodesia, the last section contains the Roan Antelope, Mufulira, Chambishi, N'Kana, and N'Changa deposits, all recently discovered. The replaced beds generally lie below impervious shale beds. The persistence of the ore-bearing beds is amazing.

After the close of the Kundelungu deposition followed a period of folding, close in the Katanga, more open in Rhodesia, and an epoch of

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[1] Relation of the disseminated copper ores to igneous intrusions, *Trans.* Am. Inst. Min. Met. Eng., 75, 1927, pp. 797–815.

[2] Anton Gray, The correlation of the ore-bearing sediments of the Katanga and Rhodesian copper belt, *Econ. Geol.*, 25, 1930, pp. 783–804.

H. Schneiderhöhn, Mineralische Bodenschätze im südlichen Afrika, Nem-Verlag, Berlin, 1931, 111 pp. Excellent summary.

intrusion of granite, which on the Rhodesia sections is shown as just touching the "Série des Mines" but not intruding it extensively; pegmatite dikes are also mentioned. Some geologists, among them H. Schneiderhöhn, doubt the actuality of these intrusions. Basic rocks certainly intrude the Roan Series in many places. Then followed, according to the Rhodesian geologists, the metallization of the folded beds.

Above the folded Katanga Series rest unconformably the flat Lubilashi beds (Permian to Jurassic, Karroo, coal-bearing series). Erosion and oxidation then took a hand in the game, the oxidation penetrating deep below the present water level. In Katanga, the oxidized ores predominate. In Rhodesia, supergene chalcocite occupies the upper part of the deposits.

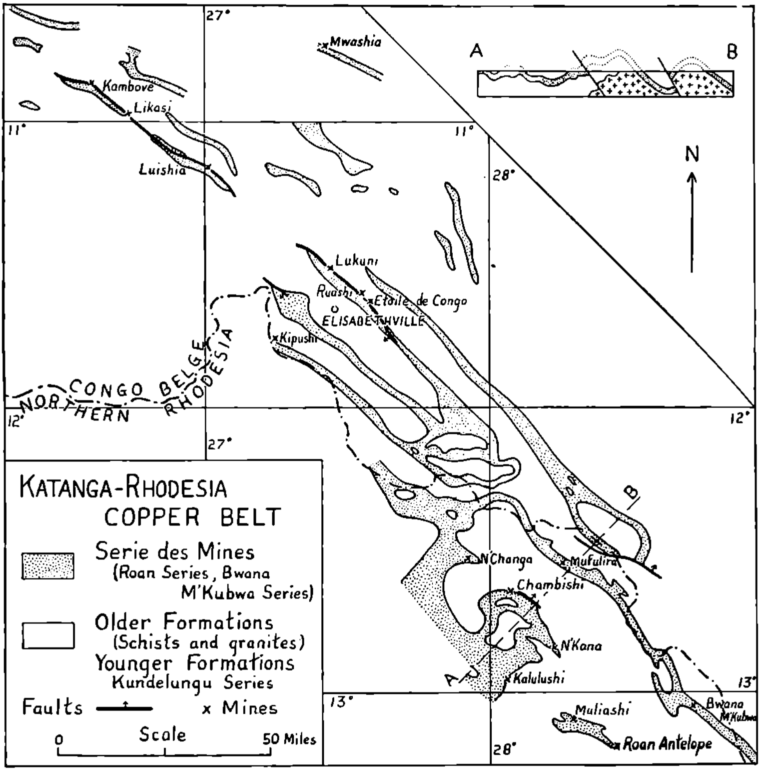

FIG. 258.—Geologic map of the Katanga-Rhodesia Copper belt. (*After Anton Gray.*)

**The Katanga Districts.**[1]—In Katanga the deposits extend for about 150 kilometers northwest from Elizabethville and are worked by the Union Minière du Haut Katanga. The production has risen rapidly

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[1] P. Fourmarier, Carte géologique du Congo Belge, *Revue Universelle*, 1930.

M. Robert, Carte du Katanga, 1:200,000; Comité Spéciale du Katanga; also, since 1920, now (1931) reaching about 140,000 metric tons of copper a year. As elsewhere, adverse conditions began in 1930. The reserves are estimated variously from 40 to 80 million tons of 7 to 8 per cent ore, practically all oxidized. Deposits now worked are the Kipushi, Ruashi, Luishia, and Kambove. There is difficultly decipherable close folding with overthrusts and "nappes." The Kipushi mine is worked by underground methods, the others in huge open cuts; and, owing to the structure, the thickness of the ore is often great.

The ore carries malachite, chrysocolla and other silicates, copper phosphates, and copper-pitch ore; in part it contains cobalt. At Kipushi, chalcocite occurs and bore-holes have shown chalcopyrite (pyrite), bornite, sphalerite, and tetrahedrite. The ores are contained in dolomite, dolomitic sandstone, and cherty rocks not highly metamorphosed.

**The North Rhodesia Districts.**[1]—In the open folds of the Lower Roan Series, about 150 kilometers southeast of Elizabethville, five or six very important districts have been opened since 1924, mostly by drilling. The beds are thus lower than those of Katanga, and the ore-bearing beds lie within the lower 1,500 feet of the Lower Roan but not all at the same horizon. They are generally little metamorphosed arkosic and dolomitic sandstones.

When the ore beds are in narrow synclines, as at Roan Antelope, N'Kana, and N'Changa, the deposits lie on both limbs; where they are on one limb only, as at Mufulira and Chambishi (Fig. 259), they take the form of tabular lodes. At the Roan Antelope mine, which now is beginning to produce, the syncline is 2,000 feet deep and the ore bed has been traced for thousands of feet. The thickness of the ore beds ranges from 20 to 80 feet, and the grade of the ore from 3¼ to 4½ per cent. The possible production of copper per annum, with all of the mines in operation, would be several hundred thousand tons.

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*Econ. Geol.*, 26, 1931, pp. 531–539.

Guidebooks for the Fifteenth Internat. Geol. Congress, Pretoria, 1929.

A. Timmerhans and H. J. Schuiling in papers published in *Comptes Rendus* of the Fifteenth Internat. Geol. Congress, *Comptes Rendus*, 2, 1930, pp. 446–453 and 688.

J. Thoreau, *idem*, pp. 482–484 (Examination of drill cores).

[1] A. M. Bateman, The Rhodesian Copper Deposits, *Trans.* Inst. Min. and Met. Canada, 33, 1931, pp. 173–213.

A. M. Bateman, The ores of the North Rhodesian Copper belt, *Econ. Geol.*, 25, 1930, pp. 365–418.

H. Schneiderhöhn, *op. cit.*

J. A. Bancroft and R. A. Pelletier, *Mining Magazine* (London), Dec. 1929; Jan., Feb., Mar. 1930.

J. C. A. Jackson (N'Changa) *Econ. Geol.*, 27, 1932, pp. 247–280.

Anton Gray (Mufulira), *idem*, pp. 315–343.

D. M. Davidson (Chambishi), *idem*, 26, 1931, pp. 131–152.

The Roan Antelope is believed to have at least 75 million tons of reserves. The total amount of ore reserves in the various Rhodesian mines is held to be something like 300 million tons.

The ore is an inconspicuous material—sandy shale, dolomitic shale, or arkose—mineralized by uniformly distributed sulphide particles, few over 2 millimeter in diameter. They unquestionably replace the grains of sand or dolomite, but there are few gangue minerals, little quartz, little sericite; many of the ores look, however, like biotitic hornfels with moderate metamorphism, but certainly this biotite is not confined to the sharply defined ore bed. The origin of the tourmaline sometimes observed seems doubtful. Bateman mentions tremolite and garnet with later sulphides from the Muliashi beds, but this seems an exceptional occurrence.

The ore minerals are linnaeite ($Co_3S_4$) (earliest), followed by chalcopyrite, bornite, and chalcocite, the last two often in pseudo-eutectic intergrowth. Down to a depth of several hundred feet there is much, probably supergene, chalcocite. Cobalt in minor quantities is characteristic of all of the Katanga and Rhodesian deposits.

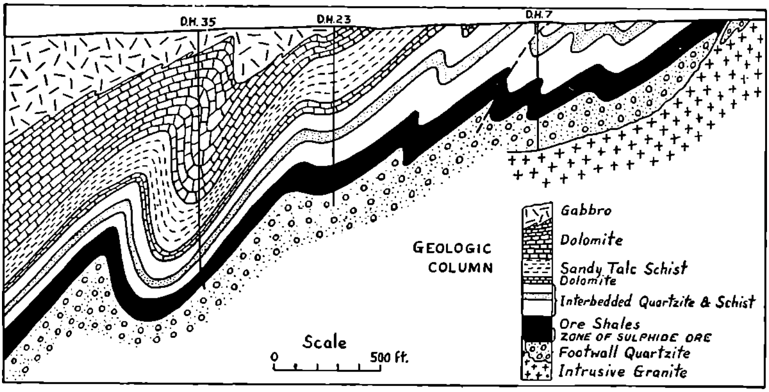

FIG. 259.—Vertical section showing geology at Chambishi Mine. (*After D. M. Davidson.*)

**Origin.**—Most of the South African geologists call the deposits mesothermal, formed by granitic emanations. This theory has apparently not yet been proved. A theory of sedimentary origin with subsequent slight metamorphism can not yet be wholly discarded. There are no mesothermal deposits in the world which are definitely comparable to these enigmatic South African ores.

## CADMIUM ORES

Almost the only cadmium mineral is the yellow greenockite (CdS), which is fairly common in the Joplin region, Missouri, as a yellow coating

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[1] C. E. Siebenthal, The cadmium supply of the United States, *Trans.*, Am. Inst. Min. Met. Eng., 60, 1919, pp. 185–197.

N. F. Budgen, Cadmium, its metallurgy, properties, and uses (London), 1924.

on sphalerite and disseminated in oxidized zinc ores coloring them yellow. Cadmium, probably as sulphide, is almost universally present in sphalerite. In the Joplin region, the metal is contained to the extent of a trace up to 1 per cent, averaging 0.3 per cent, but many western ores also contain this metal in noteworthy amounts. It is found in high-temperature deposits, for instance, in those of Bodenmais, Bavaria, and also, as just stated, in zinc ores in limestone without igneous affiliations. The zinc deposits of Santander, Spain (p. 428), contain 0.4 per cent or more Cd. Traces of cadmium have been found in some coals and in mine waters. The metals, indium, thallium, germanium, and gallium generally accompany cadmium in small amounts.

Cadmium has been produced in the United States since 1907 in increasing amounts, about 60 tons in 1924 and 1,500 tons in 1931. In addition, a similar amount of cadmium sulphide was manufactured. The metal being more volatile than zinc is obtained from the dust of the bag houses of lead smelters that treat zinc-bearing lead ores. Much cadmium is also produced from the Sullivan ores, British Columbia, from the Broken Hill ores, New South Wales, Australia, and from the Silesian deposits in Poland and Germany. The price of cadmium was 56 cents per pound in 1931.

Cadmium is used in the manufacture of silverware and alloys of many kinds, in electroplating, for easily fusible alloys, as a substitute for tin, etc. Cadmium sulphide is a brilliant yellow pigment.

## ARSENIC DEPOSITS

Arsenic is of widespread occurrence, in volcanic exhalations, in the sea water, in many spring waters and spring deposits, and in some products of organic life. Nevertheless, we rarely find pure arsenical ores in such abundance as to be of economic importance. About the only plentiful ore mineral of arsenic is arsenopyrite ($FeAsS$), which is found in many veins, usually associated with quartz and gold. More rare are löllingite ($FeAs_2$), smaltite ($CoAs_2$), niccolite ($NiAs$), and other cobalt and nickel arsenides; they occur, for instance, in the silver veins of Cobalt, Ontario (p. 608).

The minerals of arsenic are found in practically all classes of sulphide deposits of igneous affiliations, but are most plentiful in veins of the intermediate and high-temperature type.

In a few deposits like that of Hastings, Ontario[1] and Brinton, Virginia,[2] arsenious oxide ($As_2O_3$) has been recovered by roasting of arsenopyrite; but the great bulk is obtained as $As_2O_3$ in the flue dust of smelters using mixed ores in which occur the above-mentioned arsenical minerals or some of the numerous sulpharsenides, such as tennantite or

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[1] J. W. Wells, *Eleventh Rept.*, Ontario Bur. Mines, 1902, pp. 101–122.

[2] F. L. Hess, *Bull.* 470, U. S. Geol. Survey, 1911, pp. 205–122.

enargite. There are several sulpharsenides of lead but they are rare in the United States. In 1929, 16,600 tons of arsenious oxide was produced from flue dust. The price was about 3 cents per pound, and the use is largely for killing insect pests, fungi, the preservation of wood, etc. The largest amount of arsenic is produced at the Boliden gold mine, Sweden (p. 689).

## FLUORITE DEPOSITS

Fluorite ($CaF_2$) is the only simple fluoride occurring in nature.[1] It is a persistent mineral (p. 90) occurring in almost all deposits and formed at all temperatures. It rarely occurs in great abundance. Most of the

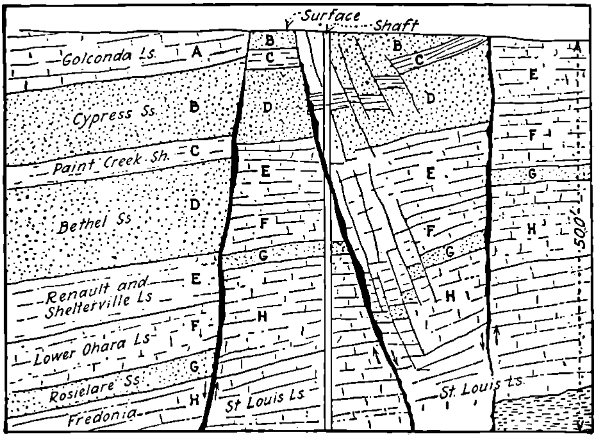

FIG. 260.—Idealized section to show vein relationship and stratigraphy at Blue Diggings shaft, Rosiclare, Illinois. Heavy black represents vein material. (*After S. Weller.*)

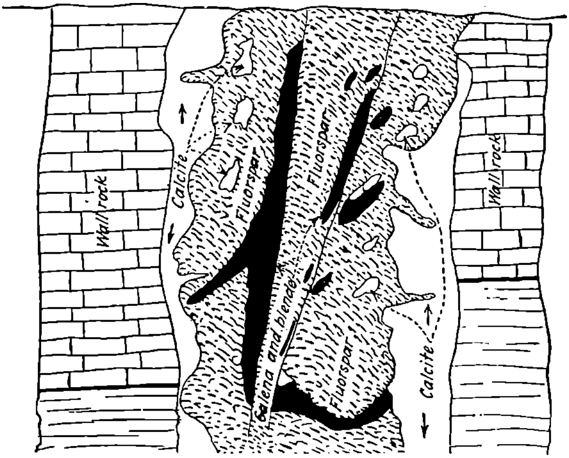

FIG. 261.—Idealized section of fluorspar veins, showing general relationship between wall rock, minerals, and center slips. (*After S. Weller.*)

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[1] Excepting the very rare sellaite ($MgF_2$).

fluorite deposits worked are epithermal or mesothermal veins. Among the associated minerals galena, pyrite, sphalerite, quartz, and barite are most common.

In the United States, fluorite is chiefly mined in Illinois and Kentucky[1] from wide fault veins and from flat replacement deposits in Carboniferous limestone, shale, and sandstone (Fig. 260). The paragenesis is (1) coarse-grained calcite; (2) fluorite; (3) sphalerite and galena (Fig. 261). The texture is massive, sometimes banded. Fine diffusion banding has been described from the replacement deposits. The deposits are believed to stand in genetic connection with dikes of peridotite.

Fluorite is also mined from veins in Colorado, in Boulder and Custer counties, and from many smaller veins in New Mexico.[2] The interesting hot-spring deposit of Wagon Wheel Gap, in Mineral County, Colorado (p. 71), is of some economic importance.

The English deposits in Derbyshire and Durham are likewise lead-bearing veins.

Fluorite is used as fluxing material, mainly in the basic open-hearth steel furnaces, for opalescent glass, enamels, and for the manufacture of hydrofluoric acid. Lenses are made from optically perfect material which is rare. In 1930, 96,000 short tons of fluorite were mined in the United States, mainly in Illinois and Kentucky, and brought an average price, f.o.b., of $17 per ton. Regarding cryolite see page 765.

## SIDERITE DEPOSITS

Siderite occurs as a gangue mineral in many veins (p. 570). When it predominates and the veins are wide, as in southern Westphalia,[3] the vein filling is used as an iron ore. No such deposits are mined in the United States.

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[1] E. O. Ulrich and W. S. T. Smith, *Prof. Paper* 36, U. S. Geol. Survey, 1905.

H. Foster Bain, The fluorspar deposits of southern Illinois, *Bull.* 255, U. S. Geol. Survey, 1905.

J. F. Fohs, *Bull.* 9, Kentucky Geol. Survey, 1907; *Econ. Geol.*, 5, 1910, p. 377.

S. Weller, *Bull.* 41, Illinois State Geol. Survey, 1920.

L. W. Currier, Fluorspar deposits of Kentucky, *Bull.* 13, Kentucky Geol. Survey, 1923.

J. E. Spurr, The Kentucky-Illinois district, *Eng. Min. Jour.-Press*, Oct. 30, Nov. 6, 1926.

R. B. Ladoo, Fluorspar, *Bull.* 244, U. S. Bur. Mines, 1927, 184 pp.

E. S. Bastin, *Bull.* 58, Illinois State Geol. Survey, 1931, 116 pp.

[2] E. F. Burchard, Fluorite deposits in western United States, *Tech. Pub.* 500, Am. Inst. Min. Eng., 1933, 26 pp.

[3] W. Bornhardt (with P. Krusch), Ueber die Gangverhältnisse des Sieger Landes, *Archiv. Lagerstättenforsch.*, Berlin, 1912. The history of these veins is complicated; they contain besides siderite, hematite (early), pyrite, marcasite, bornite, and gold, also rare bismuth minerals.

Near deposits in limestone this rock is often replaced by siderite (p. 592). Extensive replacements may result in deposits of siderite of economic value. At Eisenerz,[1] in Styria, a celebrated deposit of this kind is worked. Crushed and folded Paleozoic and Triassic limestones form the material for the replacement. The ore, which averages 39 per cent Fe, is worked in vast open cuts. Opinions as to genesis are far apart; the relations are very complicated, but the deposits are certainly epigenetic and probably related to intrusions. The annual output is nearly 2,000,000 tons.

## COBALT DEPOSITS

Cobalt has less pronounced siderophilic tendencies than nickel, and little of it is contained in the magmatic and pyrometamorphic deposits. It is rather abundant in some mesothermal ores, here usually associated with arsenic, nickel and silver (Cobalt, Ontario, Schneeberg and other places in Saxony, pp. 602–610). In combination with sulphur it occurs as linnaeite ($(Co, Ni)_3S_4$, some varieties without nickel) in mesothermal, polymetallic veins of the Siegen district, Germany, and other places; also in the Mississippi Valley lead-zinc deposits (p. 439), and in the copper deposits of Katanga and Northern Rhodesia (p. 629). Cobalt seems also to have a tendency to concentrate in deposits ascribed to the action of meteoric waters.

The present supply of cobalt, about 1,200 metric tons of metal per annum, is derived from Katanga (700 tons), Cobalt, and various other places.

For a long time cobalt has been employed in the ceramic and paint industries for the beautiful color of some of its oxide compounds, *e.g.*, cobalt blue, $CoO.Al_2O_3$, but its principal ever-widening use is for rustless alloys of the stellite type (Co, Cr, W), and for cobalt steel, used in cutting tools and steel magnets. The New York price in 1931 for cobalt metal (97 to 99 per cent Co) was about $2 per pound.

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[1] T. Redlich, Iron ore resources of the world, Stockholm, 1, 1910, p. 153. Beyschlag, Vogt, Krusch, Ore deposits, translated by S. J. Truscott, London, 2, 1926, p. 817.

K. A. Redlich, Der Steyerische Erzberg, in "Der Bergbau Steiermarks," Fasc. 9, Leoben, 1916.

K. A. Redlich, Die Geologie der innerösterreichischen Eisenerzlagerstätten, Wien-Berlin, 1931, 165 pp.

K. A. Redlich, Zur Tektonik und Genesis des steirischen Erzberges, *Jahrbuch*, Geol. Bundesanstalt, 80, 1930, pp. 231–260.

# CHAPTER XXVII

# HYPOTHERMAL DEPOSITS

## VEINS AND REPLACEMENT DEPOSITS (EXCEPTING CONTACT-METAMORPHIC DEPOSITS) FORMED AT HIGH TEMPERATURE AND PRESSURE AND IN GENETIC CONNECTION WITH INTRUSIVE ROCKS

### GENERAL FEATURES

**High-temperature Minerals.**—In the ore deposits described in previous chapters such minerals as the pyroxenes and amphiboles, the garnets, ilmenite, magnetite, specularite, pyrrhotite, tourmaline, topaz, the brown and green micas, and the spinels are generally absent. In the veins and replacement deposits formed at high temperature one or more of these minerals are commonly present, besides many other persistent ore and gangue minerals which are formed under widely varying conditions. In general, simple sulphides and arsenides prevail and are in many deposits associated with oxides, such as magnetite, specularite,[1] ilmenite, and cassiterite.

In the presence of water and other volatile substances, crystallization may take place at a much lower temperature than in dry fusion. In a magma, high pressure is necessary to hold these substances in the fluid melt which then is really a magmatic solution. The conceptions of solvent and solute are inapplicable, the various constituents of the magma being dissolved in one another. Under diminishing pressure, as during the ascent of magmas to higher levels, water and other mineralizers separate from the magmatic solution and carry with them certain constituents of the magma such as silica, some heavy metals, and alkaline metals. This "magmatic extract" may be in a state of aqueo-igneous fusion as a pegmatite magma; or when the temperature is lowered the crystallization of some constituents may convert it into an aqueous solution.

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[1] Specularite and magnetite develop in small quantities in some epithermal and mesothermal deposits (Mogollon, New Mexico; San Juan region, Colorado; the Mother Lode mines, California). It will also be remembered that specularite may form in the outcrops of deposits, in tropical regions. Magnetite occurs in small amounts in certain sedimentary iron ores. It is believed, however, that these two minerals only appear in quantity in deposits formed at high temperature. Just what are the lower limits of their temperature fields is not known. Pyrrhotite has been synthesized by Allen, Crenshaw, and Johnston at temperatures of from 80° to 225° C., but as a matter of fact the mineral does not appear in ore deposits until the lower limit of the mesothermal zone is reached.

The hypothermal deposits are formed, we believe, from aqueous solutions at a comparatively high temperature, say from 300° to 500° C. These solutions were formed by differentiation mainly from batholithic magmas at no excessive depth.

What part the critical phenomena play in these solutions is not well known (p. 117). On one hand there are many fluorides and chlorides in which the critical temperature is relatively low—not much higher than that of water. But it must be remembered that these haloid salts are only parts of the solution which must contain an abundance of salts with exceedingly high critical points. Among these constituents the most important is silica which, it would seem, must be carried in colloidal condition. The evidence tends to prove that the solutions are liquids rather than gases and that "pneumatolytic" conditions (if by this much abused term is meant solutions above their critical temperatures) play no important part. Therefore, the term *pneumatolytic* will be avoided in the following discussions.

These inductions are supported by field evidence of the strongest kind. Practically all these deposits occur in or near bodies of intrusive rocks and have been exposed by deep erosion. They were, therefore, certainly formed at considerable depths below the surface. The contact-metamorphic (pyrometasomatic) deposits, for which cogent proof of magmatic origin can be given, grade into these hypothermal veins. On the other hand, the high-temperature veins at many places imperceptibly grade into those in which the magmatic origin is less clear, thus giving in such places an almost complete line of transition from the rocks congealed from the magma, such as the pegmatite dikes, to the metal-bearing veins of the mesothermal type.

In a given district these phenomena—the pegmatitic dikes, contact-metamorphic deposits, deep-seated veins, replacement deposits, and veins of the common type—all developed very soon after the intrusive activity and during a rather short and sharply defined epoch of metallization, during which the temperature gradually declined.

Some of the minerals enumerated on the previous page are dependent upon temperature only; such are the pyroxenes, spinel and magnetite; others, like tourmaline, topaz, chondrodite, the micas, etc., contain a volatile constituent and require pressure and the presence of mineralizers for their formation. Some high-temperature deposits may be formed comparatively near the surface and even in lava flows, like certain rare tin deposits in rhyolite. Similar results may follow in the case of intrusions near the surface, where the temperature of the solutions is raised to such a degree that the vapor tension overcomes the pressure and fumaroles and "soffioni" result. In general, however, it will be found that most deposits with a mineral association indicating high-temperature have been formed at considerable or great depths.

Fumarolic action might result if, for instance, at a depth of 1,000 meters and consequent hydrostatic pressure of 100 atmospheres, there existed a solution temperature of 330° C. The vapor pressure at this temperature would exceed the hydrostatic pressure. It is probable, however, that in such a case there could be no delicate banding of alternating minerals as so frequently characterizes the veins of shallow depth deposited by ascending hot waters.

If no connection with the surface existed, the pressure would be measured by the height of the rock column and a much higher solution temperature would be necessary to produce fumarolic conditions. Generally, we may say that the pressure will suffice to keep the solutions in liquid form.

When a mineral deposit carries only persistent minerals, like quartz and pyrite, there is no mineralogical criterion present to decide whether or not it belongs to the high-temperature class. In many cases, though not always, mode of crystallization and geological criteria may solve the problem.

**Metasomatic Processes.**—The minerals enumerated in the beginning of this chapter generally appear in the metasomatically altered country rock and, to a less extent, in the fillings of the open cavities. The metasomatic action is often intense and leads to the development of coarse-grained aggregates. Sericitization still persists in some of these deposits though the foils of white mica may be larger and usually are associated with biotite, tourmaline, and similar minerals. The total changes in feldspathic rocks are, however, often less pronounced than in the veins produced at lower temperature, and there is less leaching of alkalies. Sericitization and propylitization are less prominent. On the other hand, the iron-magnesia metasomatism begins to acquire importance: iron-rich chlorites, tourmaline, biotite, and amphibole begin to appear. The processes will be described in detail under the various classes of deposits.

The carbonate rocks are always peculiarly susceptible to metasomatic processes and usually absorb large quantities of material from the solutions. Silicates rich in iron, like epidote, andradite (garnet), hedenbergite (pyroxene), cummingtonite (amphibole), and certain varieties of biotite, are frequently found in these deposits.

Andalusite rarely develops in hypothermal ore deposits. B. S. Butler[1] describes such a case in the Beaver lake mining district, Utah, where a latite has been converted to quartz, pyrite, muscovite, and andalusite.

From the very beginning of the replacement there develops an active interchange of constituents between solutions and country rocks. The composition of the solutions is continually changing.

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[1] *Prof. Paper* 80, U. S. Geol. Survey, 1913, pp. 78–81.

**Temperature and Pressure.**—As yet we have but few determinations which will serve as safe guides to the estimation of the temperature prevailing when the hypothermal deposits were formed. Generally speaking, it began by a maximum and the temperature gradually sank during the later stages.

The best "geological thermometer" (p. 207) for the present purpose is the enantiotropic inversion from *alpha* (trapezohedral-tetartohedral) quartz to *beta* (trapezohedral-hemihedral) quartz. This is known to take place at 575° C., which temperature will be but slightly raised by heavy pressure, probably by not more than 20° to 30° C. Wright and Larsen[1] have shown that the quartz of igneous rocks and of most pegmatite dikes is *beta* or high-temperature quartz, but that during the later stages of pegmatite formation the *alpha* quartz appears; and also that all quartz of veins and replacement deposits thus far examined is of the *alpha* or low-temperature kind. It seems certain, therefore, that most of the hypothermal quartz was formed below 575° C. Quartz is usually one of the earliest minerals developed.

Pyroxene is stable at a higher temperature than amphibole. Becke places the transition point between the two stability fields at about 550° C. It is true, indeed, that pyroxene is scarce in the hypothermal deposits while the amphiboles are common. The latter may probably be formed at a temperature of 300° C. or less.

The synthesis of amphibole,[2] tourmaline, and topaz has proved very difficult, while pyroxene may be easily obtained by fusion. Neither tourmaline nor topaz has been artificially produced. They are generally held to be characteristic high-temperature minerals, topaz probably forming at higher temperature than tourmaline. The absence of tourmaline in the epithermal veins is significant, especially when we consider that the hot waters which formed these veins contain boron, sometimes in considerable quantities. On the other hand, it is true that tourmaline in authigenetic form is not uncommon in contact aureoles far from the intrusive and in some regionally metamorphosed rocks where the temperature hardly could have reached 300° C. In both cases the crystals are very small. V. M. Goldschmidt found boron present in sediments and coal in larger amounts than in igneous rocks.

A. Brammal and H. F. Harwood[3] concluded that the temperature of crystallization of tourmaline in the Dartmoor granite was about 575° C. and that the range might be extended upward towards 800° C. The lower limit was not determined.

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[1] F. E. Wright and E. S. Larsen, Quartz as a geologic thermometer, *Am. Jour. Sci.*, 4th ser., 27, 1909, pp. 421–427.

[2] Obtained by Chrustchoff and by Allen and Wright. See Geo. W. Morey and P. Niggli, The hydrothermal formation of silicates, *Jour.* Am. Chem. Soc.. 35, 1913, pp. 1086–1130.

[3] *Mineralog. Mag.*, 21, 1927, pp. 205–220.

One of the best criteria would be the beginning of the development of silicates of calcium, magnesium, and iron from limestone or dolomite. It is generally held that this action begins to take place about 500° C., provided an avenue of escape is open for the $CO_2$ set free. Our knowledge is deficient in regard to this vital point. We do know that in a long line of deposits of epithermal to mesothermal character the limestone is not converted to silicates but replaced by silica. Until further data are secured, it may be assumed that below 300° C. no silicates are formed in the carbonate rocks. Probably the temperature is not the same for iron, calcium, and magnesium.

The heat necessary for the development of these deposits is usually derived from bodies of igneous rocks. The possibility cannot be denied that the same effect may be produced simply by the natural increment in temperature due to increase in depth. If a surface temperature of +25° C. and an increment of 1° C. for every 30 meters[1] are assumed, a depth of 10,200 meters, or about 33,600 feet, will be required for a temperature of 365° C. Van Hise has shown that down to this depth even the hydrostatic pressure is sufficient to hold the water in the form of a liquid. Such observations as have been made in the Cordilleran region show that contact-metamorphic and other deposits of the type here called deep-seated have been formed much nearer to the surface, some of them at depths of 3,000 or 4,000 feet, the criterion being a rough measurement of the amount of erosion on the basis of known thickness of strata. It may be true for some problematical deposits of the Archean (for instance, the zinc deposit in limestone of Franklin Furnace, New Jersey), that the rocks have been buried to a depth approximating 10,000 meters and that, at that depth, they have been exposed to the metasomatic influence of magmatic gases, while they were at a considerable distance from igneous intrusions. Such deposits would be connecting links between igneous and regional metamorphism, and such a condition would explain the occasional occurrence of deposits of the contact-metamorphic type at a distance from known igneous bodies. The copper deposits of Ducktown, Tennessee, are clearly of the contact-metamorphic type, but there are no adequate igneous masses in the immediate vicinity which could have produced the metamorphism. The district is one of intense regional metamorphism, and it is possible that magmatic gases of distant origin might have searched out the limestone beds and transformed the calcareous rock into ore.

As to the pressures actually existing our knowledge is slight. The hydrostatic pressure calculated by Van Hise would have little applicability, for at a relatively short distance below the surface the paths of

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[1] C. R. Van Hise, *Mon.* 47, U. S. Geol. Survey, 1904, p. 567.

underground water are probably effectively closed, and even where they are open the friction would be a factor of no mean importance. The pressure, therefore, at any considerable depth is probably far higher than that calculated from the weight of the water column. At a depth of 3,000 meters the hydrostatic pressure would be 300 atmospheres. Under purely static conditions the greatest pressure at any given point would be that indicated by the weight of the overlying rock column, or, for the depth just mentioned, equal to 810 atmospheres. Arching of resistant rocks might make this figure smaller; on the other hand, if the conditions are those of lateral stress it is possible that the actual pressure might be considerably higher and would then be measured by the strength of the buttress against which the pressure was applied.

If magmas and their differentiated gases invade the crust their pressure would be hydrostatic and could not exceed that of the static pressure of the overlying rock column without rupture of the rock. A contact-metamorphic deposit developing at a depth of 1,000 meters under a covering of limestones could, therefore, have been formed at a pressure of not more than 271 atmospheres.

**Mode of Fissuring and Filling.**—The question whether open spaces exist in the high-temperature veins has been discussed extensively. Under certain conditions, at least, it would seem improbable that open spaces could have existed where we now find deposits. Many investigators, as W. O. Crosby, E. J. Dunn, F. B. Laney, and Stephen Taber (p. 170), hold that the action of crystallization of minerals has forced the walls apart and thus provided space for the reception of ores. But aside from the problematic intensity of this force, and the certainty that pressure would cause increased solubility, such crystallization could hardly have produced perfect crystals or drusy structure. L. C. Graton,[1] in his description of the gold-quartz veins of the southern Appalachians, has suggested that the vein-forming solutions, representing the final products of emanation of a granitic magma, were injected under heavy pressure into the surrounding rocks along lines of weakness and so, like pegmatite dikes, made a space for themselves by opening their own fissures. This reasoning, which has much to commend it, would not be applicable where earlier fissures had established connection with the surface. Finally, it must be remembered that there is always a strong hydrostatic pressure helping to hold fissures open and also that any fault fissure would be sure to be partly or wholly filled by attrition products which would support the walls while offering little difficulty for the movement of ascending solutions.

The texture of the veins is generally coarse grained and irregular. There may be a rude banding by deposition but nothing to equal the delicate concentric banding of the veins formed near the surface. Com-

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[1] *Bull.* 293, U. S. Geol. Survey, 1906, p. 59.

pare, for instance, Figs. 180 and 193 to Fig. 268. The structure of the composite veins, or lodes, is much like that of the deposits formed at intermediate depth. Compare Figs. 46 and 266.

**Classes of Deposits.**—The veins and replacement deposits formed at high temperature may be divided as follows:

1. Cassiterite, wolframite, and molybdenite veins.
2. Gold-bearing veins and replacements.
3. Copper-tourmaline deposits.
4. Lead-tourmaline deposits.

In these deposits we note again the remarkable connection of certain metals with certain igneous rocks. The tin, tungsten, and molybdenum veins, for instance, almost always appear in or near intrusions of acidic granites and porphyries.

The veins and replacement deposits carrying gold, copper, and iron are mainly connected with intrusive rocks of monzonitic or granodioritic character. In general, deposits of gold, copper, iron, tin, tungsten, and arsenic are much more common among the hypothermal class, than those of silver, lead, zinc, and antimony.

## THE CASSITERITE VEINS[1]

### Mineral Association

The cassiterite veins form a rather sharply defined group, connected by transitions on the one hand with the copper-tourmaline veins and on the other hand with the wolframite and molybdenite veins. Apparently they also grade over into pegmatites, though there are few deposits of importance in this transition zone. They present the constant association of such ore minerals as cassiterite, molybdenite, arsenopyrite, wolframite (also scheelite), bismuth, and bismuthinite, with much less abundant pyrite, pyrrhotite, chalcopyrite, galena, and sphalerite. Among the gangue minerals quartz always predominates and is accompanied by lithium mica, fluorite, topaz, tourmaline, axinite, and apatite; more rarely beryl. Specularite, magnetite, and ilmenite are sometimes present. Siderite is reported from the cassiterite veins, and calcite sometimes appears as one of the very last minerals. On the other hand, the pyroxenes and amphiboles, as well as magnesium micas and garnets, are absent. Orthoclase is reported from several localities but does not assume the form of adularia. Chlorite is occasionally present in considerable amounts. Kaolin and allied hydrous aluminum silicates are

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[1] H. G. Ferguson and A. M. Bateman, Geologic features of tin deposits, *Econ. Geol.*, 7, 1912, pp. 209–262.

J. T. Singewald, Jr., Some genetic relations of tin deposits, *Econ. Geol.*, 7, 1912, pp. 263–279.

W. R. Jones, Tin fields of the world, London, 1925.

often recorded, but are probably products of secondary alteration near the surface, as are various hydrous arsenates and phosphates.

Cassiterite ($SnO_2$) is the principal ore mineral; it is often well crystallized, sometimes almost transparent, usually zoned in various shades of brown. Stannite ($Cu_2FeSnS_4$) is seldom found in pegmatites and in the cassiterite veins proper, but is an important ore in certain Bolivian silver-tin veins. There are also some rarer sulphostannates, like cylindrite and teallite.

Small quantities of tin, probably as cassiterite, are sometimes contained in pyrite or sphalerite of other classes of veins—for instance, at Freiberg, Saxony, and at the Sullivan mine, British Columbia.

Cassiterite is extremely resistant to weathering, as shown by its occurrence in placers. It is held by some authors that the so-called fibrous tin ore or "wood tin" which is often found in placers is a product of alteration of cassiterite, but the question does not seem to be definitely settled. If secondary, it is probably derived from stannite.

R. Beck gives the following succession of minerals in the cassiterite veins of Saxony: Molybdenite, lithium mica, quartz and topaz, wolframite and cassiterite, arsenopyrite, fluorite, apatite, and siderite.

In Cornwall, the general paragenesis is: quartz, mica, tourmaline, cassiterite, wolframite, stannite, and molybdenite. Or: quartz, chlorite, fluorite, chalcopyrite, arsenopyrite, pyrite, sphalerite. In Bolivia: quartz, tourmaline, pyrite, arsenopyrite, cassiterite, wolframite, bismuthinite, sphalerite.

The tin-bearing veins appear in or near granites (though not all granites contain them), or in their acidic porphyries, or, occasionally, in rhyolite. In Bolivia they are found with quartz monzonite and its porphyries; in Japan with diorite. The tenor of the ores is low, ranging from 1 to 7 per cent.

Some cassiterite veins contain bismuth, tungsten and copper in commercial quantities. A little silver and a trace of gold are found occasionally. In some Bolivian veins silver occurs in important amounts (p. 579).

## PRODUCTION AND USE OF TIN

The world's production of tin[1] was 174,000 tons in 1930. This comes from widely separated countries and is greatly concentrated in certain localities. Seventy per cent of the tin production is derived from placer deposits and of this, 60 per cent comes from one remarkable field, the Malayan province which extends from the Dutch islands off Sumatra

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[1] J. B. Umhau, Summarized data on tin production, *Econ. Paper* 13, U. S. Bur. Mines, 1932, 34 pp.

C. W. Merrill, World reserves and resources of tin, *Information Circ.* 6249, U. S. Bur. Mines, 1930.

through Malaya and up into Siam and Burma. In 1929, the Dutch East Indies, from the islands of Banka, Billiton, and Singkep, contributed 30,586 tons, all deposits being owned or controlled by the government. The Federated Malay States yielded in the same year 67,042 tons, or 37 per cent of the world's production. Of this, 95 per cent came from placers. This tin field occupies an area of 200 by 20 miles. Siam yields 7,000 to 9,000 tons, China from 7,000 to 10,000 tons, all from Yunnan province. Nigeria contributes 10,000 tons. Bolivia, finally, almost wholly from lode mines, adds annually 30,000 to 45,000 tons, of which 80 per cent comes from six properties.

Other regions are extremely poor in tin. In the whole of North America the annual production is but a few tons, in spite of many efforts to encourage the industry. South Africa and Australia yield a small production.

At first used as an alloy with copper in bronze, tin still finds an extensive use for alloys of many different kinds, but the principal purpose for which tin is produced is plating of sheet iron and this again is used principally in the canning industry. This industry, one may say, centers in the United States. Comments on the situation are unnecessary.

The tin fields of the world are being exhausted rapidly. Little, if any, of the tin used for plating is recovered.

## Metasomatic Processes[1]

**General Features.**—The tin ores generally appear in distinct fissure veins or composite lodes; in part they fill open cavities and in such ores a rough banded structure may appear. Often, however, the fissures are merely narrow breaks and the ore replaces the adjoining country rock. In ores of this kind also, a rude banding may result from the accumulation of tourmaline or cassiterite along lines parallel to the fissure.

The metasomatic alteration is strong and characteristic, resulting in coarse-grained rocks which contain muscovite, quartz, and topaz or tourmaline and to which the name greisen is usually applied (Fig. 262). Where tourmaline takes the place of topaz we may speak of tourmaline greisen, or luxullianite (Cornwall). Granite, granite porphyry, clay slate, calcareous shale, limestone, and diabase are affected by this mode of alteration where they form the country rock of the veins, but the development differs in the different rocks. Complete silicification of the wall rocks is a phase of subordinate importance.

While the total changes in composition of the original rock may be much less pronounced than in other veins, the metasomatic process is evidently more intense, pointing to a greater degree of heat and especially energetic action of mineralizers.

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[1] Besides the special papers, see F. Zirkel, Lehrbuch der Petrographie, 2, 1894, pp. 118–127.

In granites and porphyries adjoining the more common, or Saxon, type of veins the feldspars and the brown mica are replaced by quartz, topaz, and muscovite, in large crystals or foils. Chlorite is sometimes present. Topaz may also replace quartz grains. Sometimes crystals or radial aggregates form. Cassiterite appears as an accessory in the greisen, which spreads out from the fissure plane for a few inches or a few feet.

The quartz porphyry dikes of Mount Bischoff, in Tasmania, are, in part, replaced by cassiterite, with much topaz and subordinate tourmaline. The groundmass is changed to a topaz-quartz aggregate, while the feldspars are transformed to cassiterite, pyrite, pyrrhotite, arsenopyrite, and fluorite. The quartz phenocrysts remain intact. In the final product quartz and minute crystals of zircon are the only constituents which have withstood the altering processes. Siderite appears in places as a metasomatic product.

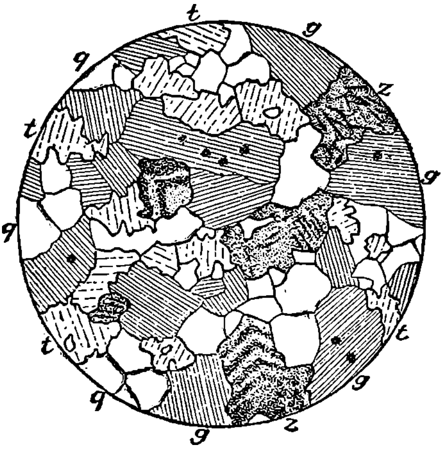

Fig. 262.—Thin section of greisen from Banka, derived from granite. *g*, lithium mica; *q*, quartz; *z*, cassiterite; *t*, topaz; stippled spots in mica consist of zircon and rutile, surrounded by pleochroic rings. Magnified 45 diameters. (*After R. Beck.*)

**Metasomatic Processes in the Deposits of Cornwall.**—The Cornwall granites,[1] which consist mainly of quartz, acidic feldspars, muscovite, and biotite, also carry some tourmaline, topaz, and fluorite. The greisen along the Cornwall veins consists chiefly of granular quartz and muscovite, often with aggregates of topaz spreading through the partly altered feldspars. Fluorite is occasionally present. The albite is more resistant than orthoclase or perthite. Some of the secondary quartz is filled with liquid inclusions containing small cubical crystals and mobile bubbles.

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[1] J. S. Flett, Explan. Sheet 347, Geol. Survey England, 1909.

The elvans, or quartz porphyries, are also altered to quartz, tourmaline, topaz, and fluorite. Kaolin, where present, appears to be due to a later process.

The greisen is essentially a vein formation in Cornwall and is not known to occur in broad masses or in patches either within the granite or along the contact. The occurrence along the contact is typical of the tourmaline rock, which, however, in places also appears along the veins. Tourmalinization is frequently superimposed upon the normal contact-metamorphic rocks, of which hornfels is the most common. Both greisen and tourmaline rock are held to have been formed before the interior of the granite was fully crystallized.

Considerable portions of the granites of Cornwall have been altered by kaolinization, but Flett believes, with good reason, that this process took place at a lower temperature than the development of topaz, tourmaline, and white mica. Kaolinized rocks appear mainly in the central parts of the granite masses, while the tourmaline rock, usually called "schorl," lies mostly along the peripheral parts. The kaolinized portions often form pipes having rudely circular outlines; the granite is altered to kaolin, muscovite, and quartz, and the product rarely contains cassiterite. The composition differs little from that of the granites. It seems probable that these kaolin deposits are due to the oxidation of distinctly later pyritic impregnations, by the action of the liberated sulphuric acid on the feldspar and sericite.

**Examples of Alteration.**—In the following table four sets of analyses are given showing the composition of fresh and altered wall rocks in Saxony, in New South Wales, in Cornwall, and in Bolivia. If, as seems probable, there has been but little change in the total mass of the alumina, the analyses are roughly comparable in their present form. Silica has also remained fairly constant. The additions consist of iron, lithium, tin, fluorine, and boron, the iron evidently entering in a silicate. Calcium, of which but little is present, is strongly leached in the rock represented by analyses IV and VI; the evidence as to magnesia is less conclusive. Sodium and potassium have both been abstracted, the former more than the latter.

The composition of a normal greisen from the Erzgebirge, in Saxony, is, according to Dalmer, as follows: quartz, 50.28; topaz, 12.14; lithium mica, 36.80; and cassiterite, 0.43 per cent.

**Alteration of Sedimentary Rocks.**—The alteration of sedimentary rocks proceeds somewhat differently. In Cornwall, the argillaceous slates are tourmalinized, forming "cornubianite" (Fig. 263), the biotite and muscovite being replaced by that mineral. Andalusite and cordierite also disappear, but rutile, ilmenite, and magnetite remain. The result is an aggregate of quartz and tourmaline, which well preserves the original structure of the rocks. In places tourmaline-albite rocks are formed.

ANALYSES OF GRANITES AND GREISENS DEVELOPED FROM THEM

| | I | II | III | IV | V | VI | VII | VIII |
|---|---|---|---|---|---|---|---|---|
| $SiO_2$....... | 74.68 | 70.41 | 70.17 | 69.42 | 76.69 | 78.47 | 65.39 | 63.82 |
| $TiO_2$....... | 0.71 | 0.49 | 0.41 | Trace | ...... | 0.22 | 0.48 | Not det. |
| $SnO_2$...... | 0.09 | a0.49 | ...... | ...... | ...... | 0.08 | ...... | Trace |
| $Al_2O_3$...... | 12.73 | 14.86 | 15.07 | 15.65 | 10.89 | 11.50 | 16.15 | 16.03 |
| $Fe_2O_3$...... | ...... | 1.42 | 0.88 | 1.25 | 0.76 | 2.64 | 1.08 | None |
| FeO....... | 3.00 | 5.09 | 1.79 | 3.30 | 0.39 | 1.05 | 3.29 | 8.79 |
| CuO....... | 0.50 | ...... | ...... | ...... | ...... | ........ | ...... | ........ |
| MnO...... | ...... | 0.29 | 0.12 | 0.39 | ...... | ........ | ...... | Trace |
| CaO....... | 0.09 | 0.21 | 1.13 | 0.63 | 1.73 | Trace | 3.07 | 0.37 |
| MgO...... | 0.35 | 0.09 | 1.11 | 1.02 | 0.18 | 0.49 | 1.87 | 2.49 |
| $K_2O$....... | 4.64 | 3.01 | 5.73 | 4.06 | 2.97 | 1.17 | 5.35 | 3.74 |
| $Na_2O$...... | b1.54 | 0.98 | 2.69 | 0.27 | 5.35 | 1.99 | 2.28 | None |
| $Li_2O$....... | ...... | ...... | 0.11 | 0.81 | ...... | ........ | ...... | ........ |
| $H_2O+$ ... | 1.17 | ...... | 0.70 | 0.54 | 0.13 | 0.23 | c1.13 | c4.77 |
| $H_2O-$ ... | | | 0.18 | 0.06 | 0.37 | 1.17 | 0.11 | ........ |
| $P_2O_5$....... | ...... | ...... | 0.34 | 0.40 | ...... | ........ | 0.40 | 0.36 |
| $MoS_2$...... | ...... | ...... | ...... | ...... | ...... | 0.80 | ...... | ........ |
| Cl......... | ...... | ...... | 0.06 | Trace | ...... | ........ | ...... | ........ |
| F......... | ...... | 3.10 | 0.15 | 3.36 | ...... | Present | ...... | ........ |
| S.......... | ...... | ...... | 0.04 | ...... | ...... | ........ | ...... | ........ |
| $B_2O_3$...... | ...... | ...... | Strong trace | 0.59 | ...... | ........ | ..... | Trace |
| Less O for F | | | 100.68 | 101.75 | | | 100.60 | 100.37 |
| and Cl... | ...... | ...... | 0.07 | 1.41 | ...... | ........ | ...... | ........ |
| Total.... | 99.50 | 100.44 | 100.61 | 100.34 | 99.46 | 99.81 | ...... | ........ |

*a* As cassiterite 0.43; in mica, chemically combined, 0.06.

*b* Including lithia.

*c* Ignition.

I. Fresh granite, Altenberg, Saxony. K. Dalmer, Explanations to the section Altenberg-Zinnwald. Geol. map Saxony.

II. Greisen, Altenberg, Saxony. K. Dalmer, *idem.*

III. Fresh Lamorna granite, Lands End, Cornwall. W. Pollard, analyst.

IV. Greisen, with tourmaline and topaz. St. Michaels Mount, Lands End, Cornwall. W. Pollard, analyst. Explan. Sheets 351 and 358, Geol. Survey England.

V. Fresh "acidic granite," New England, New South Wales. L. A. Cotton, analyst. *Proc.*, Linnean Soc. N. S. W., 34, pt. 2, 1909, pp. 220–238.

VI. Greisen, cassiterite vein near Inverell, same locality. L. A. Cotton, analyst. *Idem.*

VII. Fresh quartz monzonite, Caracoles, Bolivia. Helen Vassar, analyst.

VIII. Altered quartz monzonite close to Llamero vein, same location. Helen Vassar, analyst.

The calcareous rocks, as well as the greenstones, yield mainly axinite in large brown crystals, also pyroxene, actinolite, epidote, garnet, sphalerite, pyrite, apatite, specularite, titanite, and tourmaline, but no topaz.

H. Tronquoy[1] describes tin-bearing veins from Villeder, Morbihan, France, which are accompanied by albitization of granite, without tourmaline or topaz, while tourmaline develops in adjacent clays late.

The metasomatic development in the sedimentary rocks is of great interest, for it connects in the closest manner the effects of the ore-bearing solutions with those in contact-metamorphic deposits. Regarding the cassiterite deposits of Alaska and Pitkäranta and their close association with contact-metamorphism, see page 727.

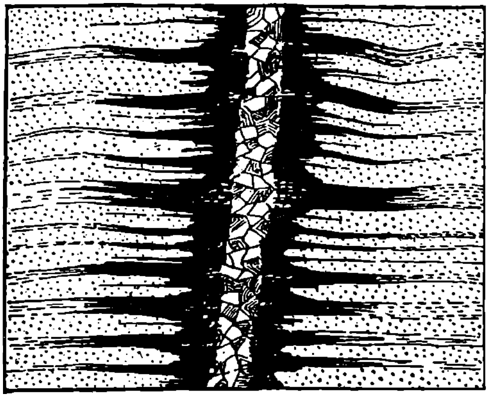

FIG. 263.—Vein of quartz, cassiterite, and tourmaline traversing Paleozoic slates, in which the argillaceous bands are replaced by tourmaline; the siliceous bands in the slate are not altered. Belowda Beacon, Cornwall. Natural size. (*After D. A. MacAlister.*)

## Origin of Tin-bearing Veins

The occurrence of cassiterite as a primary constituent of granite is recalled, as well as its occasional appearance in the pegmatitic druses of granite—for instance, those in Elba described by G. vom Rath.[2] We may further recall the appearance of cassiterite in economic quantities in pegmatite dikes (p. 760) and its occasional occurrence in base-metal veins—for instance, in those of Freiberg and in the lower levels of the Przibram veins. The occurrence of tourmaline in cassiterite veins is also important in view of the transitions to the chalcopyrite-tourmaline type.

These facts were realized at an early date by A. Daubrée and E. de Beaumont.[3] To J. H. L. Vogt[4] belongs much credit for his resuscitation

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[1] *Comptes Rendus*, 154, 1912, p. 899.

[2] G. vom Rath, *Zeitschr.* deutsch. geol. Gesell., 1870, p. 646.

[3] A. Daubrée, Sur le gisement, etc., des amas de minerai d'étain, *Ann. des Mines*, 3d ser., 20, 1841, and other later papers.

E. de Beaumont, Notes sur les émanations volcaniques et métallifères, *Bull.* Soc. Min. France, 2d ser., 4, 1847, p. 65.

[4] J. H. L. Vogt, Zinnstein-Ganggruppe, *Zeitschr. prakt. Geol.*, 1895, pp. 145–156, and other places.

of these meritorious ideas and the addition of important investigations. The extensive literature regarding the occurrence of tin deposits has been summarized by F. L. Hess and L. C. Graton.[1]

Daubrée and de Beaumont argued, from the close association of cassiterite veins and acidic granites, that there must be some genetic connection between them and concluded that the veins were deposited by emanations from these magmas. This conclusion has been adopted and confirmed by almost all geologists who have studied these deposits.

The general occurrence of cassiterite in well-defined veins shows clearly that the granite was consolidated when these fissures were broken, even if it is probable that the whole granitic intrusion had not congealed and that liquid magma still existed below the veins. The veins occur usually either in the center or along the contacts of the granitic masses, but some of them extend into the adjoining sedimentary rocks. Finally, it is clear that the development of these veins cannot be identified with the contact-metamorphism, for they are distinctly later and their metasomatic effects are superimposed upon the products of contact-metamorphism.

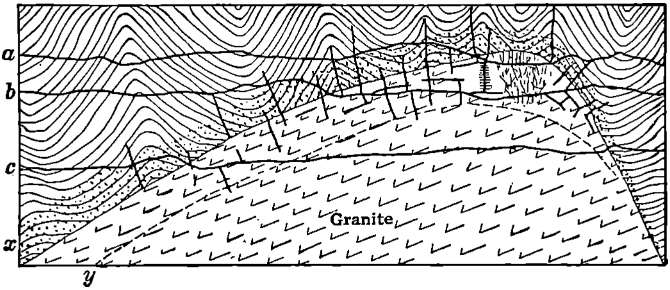

Fig. 264.—Ideal section of granite intrusion showing probable development of tin-bearing veins and the effect of successive erosion levels, *a*, *b*, and *c*; *x*, metamorphic aureole; *y*, inner limit of mineralization. (*After Ferguson and Bateman.*)

The general distribution of tin-bearing veins in relation to an intrusive mass of granite gradually eroded is shown in the diagram, Fig. 264.

The constant presence of boron and fluorine compounds, as well as those of phosphorus, in the tin veins is, of course, of the highest importance. During the short but intense epoch of metallization the temperature must have been high, probably above the critical temperature of water, and it is usually stated that the deposition took place under "pneumatolytic" conditions. In the absence of exact data as to the behavior of the solutions at this critical temperature the insistence upon

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[1] F. L. Hess and L. C. Graton, The occurrence and distribution of tin, *Bull.* 260, U. S. Geol. Survey, 1905.

F. L. Hess and Eva Hess, Bibliography of the geology and mineralogy of tin, *Smithsonian Misc. Coll.*, 58, No. 2, 1912.

W. R. Jones, Tinfields of the world, London, 1925. Later literature.

pneumatolysis has, however, little value. Just how much water was present is problematical; certainly there was some, as shown by the hydroxyl radicle in muscovite and topaz, by the presence of chlorite and by the occurrence of aqueous inclusions in quartz crystals. We may say that the agencies were aqueous solutions.

Daubrée's reaction, inferred as probable from experiments with the chloride, is as follows:

$$SnF_4 + 2H_2O = SnO_2 + 4HF.$$

The metasomatic development of topaz has been taken to indicate that free hydrofluoric acid was active in the solution. We may better frankly state that at present we are in the dark as to the exact equations for the reaction. The solutions probably did not contain much chlorine.

Both in Saxony[1] and in Cornwall lead-silver veins occur at some distance from the center of intrusions and the cassiterite veins, and transitions are also found between them. This would tend to show that these metals were less easily precipitated and were carried farther away from their sources than were the tin, tungsten, etc.

### The Cassiterite Veins of Cornwall, England

**Literature.**—An extensive literature exists on the subject of the tin lodes of Cornwall, for they have been repeatedly studied by geologists since they were first described in 1839 by H. T. de la Bèche, and in 1843 by W. J. Henwood. Many articles by J. H. Collins and C. Le Neve Foster were published about 40 years ago. The deposits have also been studied by the Geological Survey of England, and the results are published in a series of memoirs.[2]

The mining region of Cornwall, known to the ancients for the treasures of tin which it contained, furnishes one of the most remarkable instances of the dependence of ore deposits on the distribution of igneous rocks. The folded Paleozoic slates and sandstones are intruded by four main granitic batholiths or stocks (Fig. 265), and the tin deposits cluster

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[1] K. Dalmer, *Zeitschr. prakt. Geol.*, 1894, p. 321.

[2] D. A. MacAlister, Geological aspects of the lodes of Cornwall, *Econ. Geol.*, 3, 1908, pp. 363–380.

J. B. Hill, and D. A. MacAlister, Geology of Falmouth and Truro, and the mining district of Camborne and Redruth, *Mem.*, Geol. Survey England, Explan. Sheet 352, 1906.

Ussher, Flett *et al.*, The geology of Bodmin and St. Austell, *idem*, Explan. Sheet 347, 1909.

C. Reid and J. S. Flett, The geology of the Lands End district, *idem*, Explan. Sheets 351 and 358, 1907.

E. H. Davison, Mineral association in Cornish tin veins, *Mining Mag.*, 43, 1930, pp. 143–149; The variation in composition of Cornish granites and its relation to the occurrence of tin lodes, Royal Geol. Soc. Cornwall, 16, 1928, pp. 15–27.

characteristically in the marginal zones of these granitic intrusions, in both slates and granites. The slates, or killas, as they are locally called, in part overlie the granite, forming in places the roof of the batholiths. The intrusion is post-Carboniferous and pre-Triassic in age, and the veins were formed shortly after the intrusive activity, probably while the rocks still remained hot. Before the vein-forming epoch a series of dikes of granite porphyry (elvans) intersected granite and slate. The granite is a typical rock of its kind, containing, in order of crystallization, zircon, apatite, magnetite, biotite, muscovite, oligoclase, albite, perthite, and quartz. Tourmaline, topaz, and fluorite are occasional accessories in the crystallization of the magma.

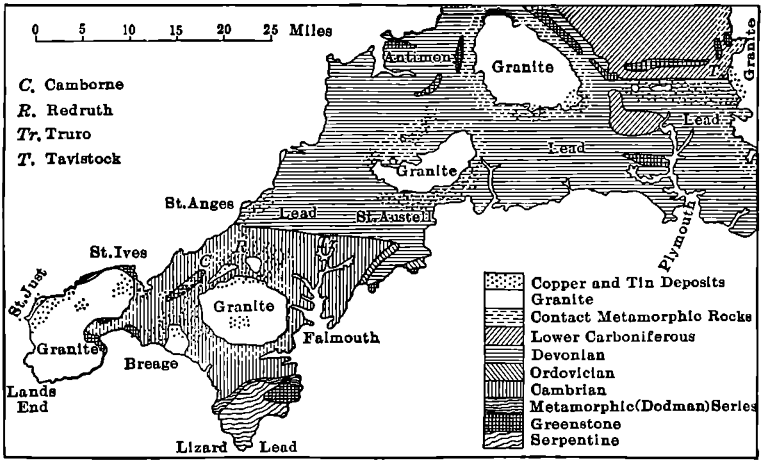

FIG. 265.—Geological map of the peninsula of Cornwall, England. (*After D. A. MacAlister.*)

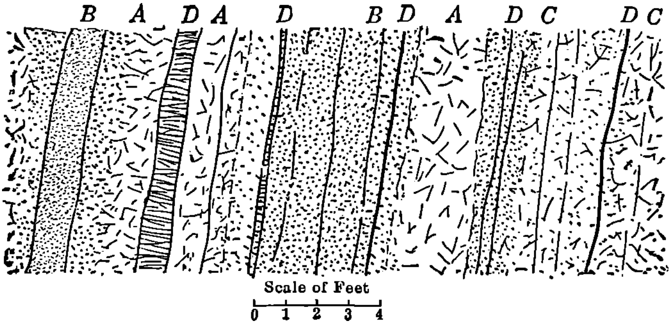

FIG. 266.—Diagrammatic section of the main lode at the 300-foot level, Bunny mine, St. Austell, Cornwall. *A*, kaolinized granite; *B*, stanniferous greisen; *C*, silicified granite; *D*, veins of quartz and cassiterite. (*After Ussher, Flett, et al.*)

The tin and copper veins are the older and were followed by a later though much less important series of veins, containing lead, silver, and sometimes also cobalt, nickel, and uranium. A little gold is present, both in the lead-silver veins and in the cassiterite veins. Many of the veins are of complex structure and may properly be called lodes. Some of them are traceable for 2 or 3 miles, or even more. The width of infilled fissures and altered rocks averages 3.5 feet, but in places, especially in the slates, reaches 50 feet. The general direction of the veins is northeast. Stockworks of irregular veinlets also occur. Faulting has taken place along many of the fissures. The veins are simple or composite (Fig. 266); reopening of fissures was evidently a common phenomenon. The filling is mainly of quartz, frequently with comb structure and, in the upper parts of the lodes, also with drusy cavities. Chlorite, fluorite, scheelite, sphalerite, molybdenite, and bismuthinite are also found in these veins. Magnetite and specularite are reported, but are rare.

The alteration spreading into the country rock from the fissures is characteristic and, when effected from a series of closely spaced fissures, may produce a banded appearance (capel); tourmalinization is the usual mode of alteration in the granite, resulting in an aggregate of tourmaline and quartz. In calcareous rocks or greenstones the altered zone along the veins contains axinite, pyroxene, garnet, and some tourmaline.

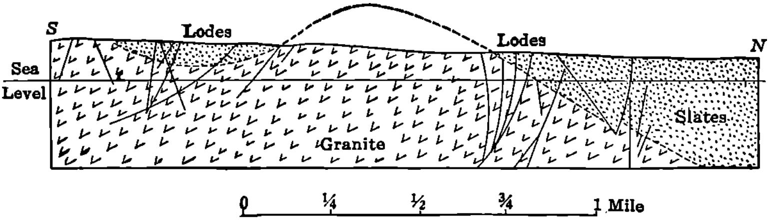

FIG. 267.—Section from Feltrick to New North Pool, Cornwall, showing relation of granite and slate and the lodes intersecting them. (*After Hill and MacAlister.*)

In the slates the lodes often contain much crushed and brecciated material; sometimes cassiterite and tourmaline following bedding planes impregnate the slates for some distance on both sides of the lode. On the whole the copper ores are confined to the lodes in the slates. Where the lodes break into the granite, the tin ore makes its appearance (Fig. 267).

The main lode of the Dolcoath and Carn Brea contained copper ores down to a depth of 1,000 feet, mainly in the slates. Below this depth the tin ore predominated and was worked until recently at a depth of 3,000 feet. According to Hill and MacAlister, the lode at the bottom of the mine is 42 feet wide and contains 1 to 3 per cent of cassiterite. The lodes of Wheal Vor were of enormous value in the killas, but proved worthless in the granite. At the Great Work mine, not far distant, these relations were reversed. The great ore-shoots of both mines plunge eastward approximately parallel to the subterranean contact surface of the granite.

## Cassiterite Veins of Saxony[1]

Saxony and the adjacent part of Bohemia contain several tin-producing districts, the ores occurring in or near granites of post-Carboniferous age. The deposits are now of little importance, but have an interesting place in the history of ore deposits.

At Altenberg the ores occur in a stockwork, about 3,000 feet in diameter, of small veins cutting across the granite and the adjacent, older granite porphyry; the whole mass of rock is altered to a greisen, containing a little cassiterite and arsenopyrite. The characteristic minerals occurring in the veinlets are quartz, bismuth, bismuthinite, pyrite, chalcopyrite, molybdenite, sphalerite, wolframite, fluorite, tetrahedrite, magnetite, and specularite. The ore, according to Dalmer, continued only to a depth of about 700 feet.

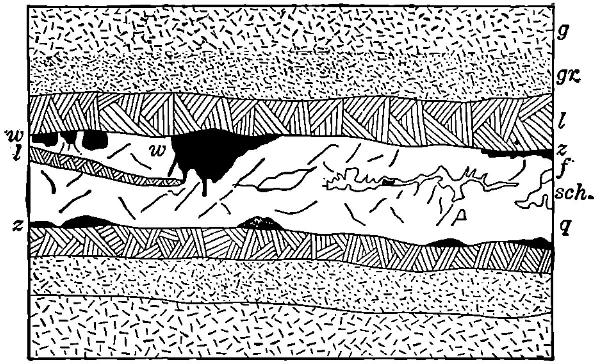

FIG. 268.—Diagrammatic section of vein at Zinnwald, Saxony. *g*, granite; *gr*, greisen; *q*, quartz; *l*, lithium mica; *z*, cassiterite; *w*, wolframite; *f*, fluorite; *sch*, scheelite. (*After R. Beck.*)

At Zinnwald the veins are likewise in granite, which with flat contact breaks through quartz porphyry. The numerous fissures are approximately parallel to the contact, and the veins are formed mainly by filling, sometimes with excellent banded structure by the development of mica

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[1] The older literature includes the papers by B. v. Cotta, H. Müller, E. Reyer, and A. W. Stelzner.

R. Beck, Einige Beobachtungen im Gebiete der Altenberg-Zinnwalder Zinnerzlagerstätten, *Zeitschr. prakt. Geol.*, 1896, pp. 148–150.

K. Dalmer, Der Altenberg-Graupener Zinnerzlagerstättendistrikt, *Zeitschr. prakt. Geol.*, 1894, pp. 313–322; see also *idem*, 1895, p. 228; 1896, p. 1; 1897, p. 265; 1900, p. 297.

J. T. Singewald, Jr., The Erzgebirge tin deposits, *Econ. Geol.*, 5, 1910, pp. 166–177; 265–272.

next to the wall; they contain a considerable amount of wolframite (hübnerite), besides the usual accompanying minerals (Fig. 268).

## Bolivia[1]

An important tin belt passes through Bolivia from southeast to northwest, along the high, eastern Cordillera Real. This imposing range, which towards the south turns into a deeply dissected plateau, is built of Silurian and Devonian clay slates with some quartzite; these are folded and faulted but not highly altered but are intruded by stocks of quartz monzonite (p. 580) in the Illimani and Quimsa Cruz sections. A little farther south extends the silver-tin area centering around Potosi, Oruro, Lake Poopo, Chocaya, and other districts, in which the deposits are connected with intrusive quartz monzonite porphyries and rhyolitic or dacitic flows (p. 580). The veins which carry tin only have not been worked until comparatively recently, while the silver mines have been worked since about 1550.

The relations of the tin veins are best illustrated in the Cordillera Quimsa Cruz, southward from La Paz. At Caracoles the quartz monzonite contains a great number of replacement veins carrying much tourmaline, with quartz and cassiterite and scant sulphides. The width of the veins ranges up to 15 feet, usually it is less. Filling is subordinate. The ore-shoots appear to extend horizontally. The paragenesis is chlorite and sericite (oldest), tourmaline, apatite, quartz, cassiterite (wolframite), pyrite, sphalerite, bismuthinite, quartz, ankerite, and pyrite.

The country rock is altered to a varying distance from the vein, and the minerals formed in it are chlorite, sericite, tourmaline, and a little cassiterite (for analyses see p. 648). The altered rock is not exactly a greisen, though the chemical changes tend in the direction of greisenization, with introduction of magnesia and iron, a process which, as explained above, is normal for many hypothermal veins.

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[1] W. R. Rumbold, The origin of the Bolivian tin deposits, *Econ. Geol.*, 4, 1909, pp. 321–364.

M. Armas, *Ann. des Mines*, 10th ser., 20, 1911, pp. 149–213.

W. Myron Davy, Ore deposition in the Bolivian tin-silver deposits, *Econ. Geol.*, 15, 1920, p. 463.

W. Lindgren, Replacement in the tin-bearing veins of Caracoles, *Econ. Geol.* 21, 1926, pp. 135–144.

F. R. Koeberlin, *Eng. and Min. Jour.-Press*, April 17 and Nov. 20, 1926 (secondary enrichment).

E. Kittl, La Cordillera de Quimsa Cruz y sus yacimientos estanniferos, *Revista minera*, Oruro, 2, 1927, pp. 65–80, 97–120.

R. Pilz and M. Donath, Betrachtungen über die Entstehung der Eruptivgesteine und Erzlagerstätten des andinen Bolivien, *Zeitschr. prakt. Geol.*, 37, 1929, pp. 125–138.

F. Ahlfeld, The tin ores of Uncia-Llallagua, *Econ. Geol.*, 26, 1931, pp. 241–527.

E. S. Turneaure (Llallagua), *Econ. Geol.* 30, 1935, pp. 14–66, 170–190.

In the Paleozoic slates, which surround this intrusive mass, is found an aureole of other tin deposits in which there is much less tourmaline and in which filling becomes more prominent. In none of these ores is there any notable amount of gold and silver. Bismuthinite is almost always present. Among these are the celebrated Araca veins which contain quartz, cassiterite, and a certain amount of sulphides. The cassiterite is often transparent and beautifully crystallized. Other notable veins occur at Chojñacota and at Monte Blanco; in the latter place cassiterite replaces quartzite. At Uncia and Llallagua, 250 kilometers southeast of Caracoles, are at present the most productive lode deposits of tin in the world, yielding about 20,000 tons of tin per annum. The veins occur in slate cut by a small stock of dacite-porphyry. The principal minerals are tourmaline, quartz, arsenopyrite, wolframite, pyrrhotite, siderite, and pyrite. There is said to be a tourmaline zone in depth and an upper zone of sulphides and sulphosalts. The zone of oxidation is about 1,000 feet deep. Below this there are much vivianite, wavellite, and other phosphates. It has been stated that 2,000 feet seems to be the depth limit for commercial ore. The grade is said to be about 4 per cent Sn.

At Chacaltaya, near La Paz, are veins in slate containing some tourmaline, lepidolite, fluorite, cassiterite, and pyrite. Stannite is generally absent from the veins described above.

For the silver-tin veins, see page 579.

### Tin Deposits in Other Countries

**United States.**—Few cassiterite veins are found in the United States and none of them are of great value. A vein in granite has been worked near El Paso, Texas, on the east side of the Franklin Mountains.[1] Some low-grade veins occur in the Temescal Mountains, near Riverside, California.[2] The tin-bearing deposits of the Black Hills of South Dakota are pegmatite dikes. Pegmatites and also a few quartz veins with cassiterite occur in the Appalachian region.[3] The tin deposits of Alaska are described on page 727.

**Tasmania.**[4]—The principal tin deposits of Tasmania are those of Mount Bischoff, where schists are intruded by dikes of granite porphyry,

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[1] G. B. Richardson, Tin in the Franklin Mountains, *Bull.* 235, U. S. Geol. Survey, 1906, pp. 146–149.

[2] H. W. Fairbanks, The tin deposits of Temescal, *Am. Jour. Sci.*, 4th ser., 4, 1897, pp. 39–42.

[3] L. C. Graton, Reconnaissance of some gold and tin deposits of the southern Appalachians, *Bull.* 293, U. S. Geol. Survey, 1906.

[4] W. von Fircks, Die Zinnerzlagerstätten des Mount Bischoff, *Zeitschr.* deutsch. geol. Gesell., Bd. 51, Heft 3, 1899, pp. 431–465.

J. G. Weston-Dunn, *Econ. Geol.*, 17, 1922, pp. 153–193 (Mt. Bischoff).

W. H. Twelvetrees and L. K. Ward, *Bull.* 8, Geol. Survey Tasmania, 1910.

H. Couder, *Bull.* 26, *idem*, 1918, pp. 1-96.

both rocks being rich in metasomatic tourmaline and topaz. The whole forms a weathered mass of rock traversed by cassiterite veins—a stockwork, large portions of which averaged 2 to 3 per cent tin. Magnetite, fluorite, pyrrhotite, sphalerite, wolframite, and siderite are mentioned as accompanying minerals.

The tin-bearing district of Zeehan, in northwestern Tasmania, contains deposits of various kinds which appear to show an unusually complete series of transitions. Silurian sediments are intruded by granite. There is a gradation from cassiterite veins, with tourmaline, in granite, to contact-metamorphic deposits, containing copper, lead, and zinc, in which cassiterite has been found, and from these to normal banded veins containing pyrite, chalcopyrite, galena, and stannite, and finally to galena-siderite veins. In other words, the gradation is one from high-temperature deposits to those formed in the cooler zone with a corresponding change of minerals deposited.

**Malaya.**[1]—A tin-bearing province of the greatest importance extends from the Dutch islands off Sumatra (Banka, Billiton and Singkep) through the Malayan states up into Siam and Burma. While the production is chiefly from placer deposits (p. 645), there are also primary deposits of great value, which range from disseminations in granite to pegmatites and pipes and finally to normal cassiterite veins. Much geological work has been done in this province, and at times opinions have clashed severely. Scrivenor's pioneer work is the outstanding feature.

From the somewhat conflicting evidence we may pick out the following as the most probable geologic history: A thick series of cliff-forming Permo-Carboniferous limestone is overlain by Triassic schistose sediments; both are intruded by Mesozoic granite; the placers occur in recent and high level alluvial beds. The lode deposits which yield but little production in proportion to the alluvials are veins, apparently of normal type. Most prominent among these are the Pahang Consolidated Mines, which, as W. R. Jones states, is the fourth largest tin mine of the world and which yields about 1,600 tons of tin annually.

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[1] R. A. F. Penrose, Jr., The tin deposits of the Malay Peninsula, *Jour. Geol.*, 2, 1903, pp. 135–154.

J. B. Scrivenor, The geology and mining industry of the Kinta district, Kuala Lumpur, 1913. The Lahat pipe in Perak, *Quart. Jour.* Geol. Soc. (London), 65, 1909, pp. 382–389. The topaz-bearing rocks of Gunong Bakau, *idem*, 70, 1914, pp. 363–381.

J. B. Scrivenor and W. R. Jones, The geology of South Perak, Kuala Lumpur, 1919.

W. R. Jones, Tin fields of the world, London, 1925.

E. S. Willbourn, The Beatrice mine, *Mining Mag.*, December, 1931; January, 1932.

R. H. Rastall, The geology of the Kinta Valley, *Mining Mag.*, June, 1927.

R. D. M. Verbeek, Ueber die Zinnerzlagerstätten von Bangka und Billiton, *Zeitschr. prakt. Geol.*, 1899, pp. 134–136.

C. T. Groothoff, De primaire Tinersafsettingen van Billiton, Dissert., Delft, 1916.

There is also more or less contact metamorphism at the granite contacts, but these pyrometasomatic deposits contain little tin.

**Japan.**—Takeo Kato[1] describes tin veins from Akenobe, where Paleozoic schists are intruded by diorite. The veins are argentiferous copper ores with small crystals of cassiterite and wolframite. There is a considerable tin production. The structure of the veins is crustified, and the paragenesis is as follows: (1) cassiterite; (2) wolframite and cassiterite, chalcopyrite, sphalerite. Colloform deposition of quartz, chalcedony, and cassiterite is noted. The country rock is chloritized, sideritized, and silicified. Fluorite is present. The similarity to Chocaya, Bolivia, is striking.

## The Cassiterite Pipes

It remains to discuss the peculiar forms of chimneys and pipes which are sometimes assumed by cassiterite deposits. It must be confessed that their place in the system of classification is uncertain. They are in places connected with pegmatites; in other places they recall contact-metamorphic mineralization.

Such roughly cylindrical pipes are described from New South Wales.[2] They occur in granite and are essentially highly altered rock with disseminated cassiterite. Similar veins or pipes of cassiterite, accompanied by specularite, tourmaline, sulphides, and siderite are found in the quartzite of the Rooiberg district in the Transvaal.[3] The pipes in the Kinta Valley, Federated Malay States (p. 657) are contained in limestone. J. B. Scrivenor and E. S. Willbourn have described many of these, the latest publication being that discussing the Beatrice mine (*cf.* footnotes, p. 657). The pipes occur in crystalline Permo-Carboniferous dolomitic limestone, close to granite. They are vertical or steep, and have no definite walls. Their mode of formation is compared to that of a jet of steam projected against a block of ice. They have been mined to depths of many hundred feet and the ore is usually rich. They contain much arsenopyrite, tremolite, cassiterite, fluorite, calcite, and a hydrated borate mineral. There is not much quartz.

These pipes bring to mind the similar, though smaller pipes or tubes of Orijärvi, Finland, described by Trüstedt[4] and Liesegang.[5]

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[1] A contribution to the knowledge of cassiterite veins, *Publ.* Imp. Univ. Tokyo, 43, Oct. 5, 1926.

[2] L. A. Cotton, The tin deposits of New England, N. S. W., *Proc.*, Linnean Soc. N. S. W., 34, 1909, pp. 733–781.

[3] H. Kynaston and E. T. Mellor, The geology of the Waterberg tin field, *Mem.* 4, Transvaal Geol. Survey, 1909.

[4] O. Trüstedt, Die Erzlagerstätten von Pitkäranta, *Bull.* Comm. Geol. de Finlande, 19, 1907.

[5] R. E. Liesegang, Geologische Diffusionen, 1913, pp. 186.

A. Knopf, *Bull.* 358, U. S. Geol. Survey, 1908.

### Cassiterite in Rhyolite Flows

Some of the Bolivian tin-silver veins occur in acidic effusive rocks (Chocaya), but these are of normal epithermal type. The deposits here to be described are found in rhyolite flows and have certain very peculiar characteristics, but so far have no economic importance. They are, it is believed, a sort of exudation deposit, having no deep connections, and have been formed directly from the residual solutions in local shear zones immediately after the consolidation of the flows. They evidently develop at high temperatures and low pressure. It may be recalled that topaz is sometimes found in lithophysae in rhyolite, indicating a retention of certain volatile constituents until consolidation of the flows.

The mineral association is wood-tin, and concretionary cassiterite with opal, chalcedony, and specularite, rarely as in Mexico, with some wolframite and bismuth minerals; the veinlets have finely developed colloform structure.[1] Occurrences of this kind have been described from Mexico,[2] Nevada,[3] and New Mexico.[4] To judge from specimens received, they are also found occasionally in the acidic flow rocks of Bolivia.

## WOLFRAMITE VEINS

Wolframite, including the tungstate of iron (ferberite, $FeWO_4$) and the tungstate of manganese (hübnerite, $MnWO_4$), has a field of occurrence similar to that of cassiterite. It appears in igneous rocks, in pegmatites, in cassiterite veins, and sometimes, with quartz and bismuth minerals, in veins which are evidently of the deep-seated type and allied to the tin veins. When cassiterite and wolframite occur together the latter is apparently always the later mineral. But wolframite also appears abundantly in veins formed under much more moderate temperature and pressure—for instance, in those of Boulder County, Colorado (p. 596). Small quantities of hübnerite are found in epithermal veins, as in those of Tonopah, Nevada, and Cripple Creek, Colorado. The principal production in the United States formerly came from Boulder County, Colorado, but now comes mainly from contact-metamorphic deposits (p. 728) in Nevada and California.

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[1] W. H. Newhouse and M. J. Buerger, Observations on wood-tin nodules, *Econ. Geol.*, 23, 1928, pp. 185–192.

[2] W. R. Ingalls, The tin deposits of Durango, Mexico, *Trans.*, Am. Inst. Min. Eng., 25, 1896, pp. 146–163; 27, 1898; E. Halse, *idem*, 29, 1900, pp. 502–511; E. Wittich, Zinnerze in der Sierra von Guanajuato, *Zeitschr. prakt. Geol.*, 1910, pp. 121–123.

[3] A. Knopf, Tin ore in northern Lander Co., Nevada, *Bull.* 640, U. S. Geol. Survey, 1916, pp. 125–138.

[4] J. M. Hill, The Mount Taylor tin deposits, New Mexico, *Bull.* 725, U. S. Geol. Survey, 1922, pp. 347–359.

Wolframite veins of the deep-seated type have been described from the Deer Park district,[1] in Washington, and from Sauce[2] in the Sierra de Cordova, Argentina. They also occur in Portugal. Wolframite lodes of great importance have been developed in the Tavoy district, Lower Burma,[3] where they occur in granite and metamorphic schist and contain in addition quartz, mica, (tourmaline), pyrrhotite, arsenopyrite, pyrite, bismuthinite, and galena. At the same place wolframite is also recovered from eluvial deposits. In 1916, the mines of Lower Burma produced about 3,000 tons of wolframite concentrates, or about one-third of the annual world production. In 1928, production had decreased to 843 tons. The crude ore is said to average 1.3 per cent $WO_3$ per ton.

Scheelite ($CaWO_4$) is the second important tungsten mineral. It is rather widespread in mesothermal gold-quartz veins, and is present in hypothermal tin-tungsten deposits. It is most common in certain contact-metamorphic ores.

Tungsten is used mainly for high-speed tool steel, the alloy ferrotungsten being first produced. The price for concentrates with 60 per cent $WO_3$ was about $400 in 1910; in 1918, under war conditions, $1,500; in 1932, about $500; all per short ton. After the World War, the prices crumbled and for several years there was scarcely any demand for the ores. During war times Bolivia[4] (from tin-tungsten veins) and China (from placers) were the outstanding producers.

## GOLD-QUARTZ VEINS

Gold-bearing veins of a deep-seated type are found in many regions in the pre-Cambrian and earliest Paleozoic rocks of the American continent. They appear in the gold belt of the Appalachian states, mainly from Maryland to Alabama; at various places in the Western states, particularly in South Dakota and New Mexico; in Ontario and Quebec; and finally in Brazil, in India, and in Rhodesia.

### The Veins of the Southern Appalachians[5]

The placer deposits of the southern Appalachians have, since their discovery, about 1800, yielded gold estimated at $30,000,000. The veins

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[1] Howland Bancroft, *Bull.* 430, U. S. Geol. Survey, 1910, pp. 214–216.

[2] Bodenbender, *Zeitschr. prakt. Geol.*, 1894, pp. 409–414.

[3] A. W. G. Bleeck, *Records*, Geol. Survey India, 43, pt. 1, 1913, pp. 48–73.

J. Coggin Brown and A. M. Heron, *Records*, Geol. Survey India, 50, pt. 2, 1919, pp. 101–121; abs., *Mining Mag.*, 22, 1920, pp. 241–243.

W. R. Jones, Tin fields of the world, London, 1925. On the relative occurrence in depth of tin and tungsten ore. *Trans.*, Inst. Min. and Met. (London), 1920; *Mining Mag.*, 22, 1920, pp. 371–375; 376–378.

[4] G. F. J. Preumont, *Eng. and Min. Jour.-Press*, April 5, 1919.

[5] G. F. Becker, Gold fields of the southern Appalachians, *Sixteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1895, pp. 250–331.

from which the placers were derived proved less productive, though they have been profitably worked at many places in North and South Carolina, at Dahlonega and the Franklin mine in Georgia, and at the Hogback mine in Alabama. All the deposits are not of the deep-seated type; there are some which approach the mesothermal gold-quartz veins. Others, like those described by Taber from Virginia, seem to be related to pegmatite dikes. The veins are contained in crystalline rocks, usually more or less schistose, which prove to be granites and quartz monzonites, intrusive into mica schists, clay slates, altered volcanic tuffs, and amphibolites. The age of the intrusives and veins is probably early Paleozoic.

Structurally the deposits may be classed as fissure veins and replacement deposits in schists. The veins are in general of the so-called lenticular type, illustrated in Fig. 49, in which the quartz lenses, which collectively constitute the veins, lie parallel to the foliated structure. In detail the lenses often cut across the schistosity and are sometimes of irregular form. Sharply defined veins cutting across the schistosity also occur. The quartz is massive, without banded or drusy structure. The replacement deposits form irregular bodies of silicified and pyritic schist; the deposit worked by the Haile gold mine is the most prominent example of this class. The ores form more or less regular shoots, often also pockets, and are in general of low grade; pyritic ore containing $2 per ton has been successfully treated at the Haile mine; many shoots, however, average much higher, sometimes $15 or $20 per ton. Free gold is generally, but by no means always, present below the zone of oxidation.

Quartz, often glassy and semi-transparent, is the principal gangue mineral and may be accompanied by calcite, dolomite, apatite, chlorite, ilmenite, magnetite, tourmaline, albite, and sometimes zinc spinel (gahnite) and garnet. The ore minerals are free gold, pyrite, arsenopyrite, pyrrhotite, molybdenite, more rarely galena, sphalerite, and chalcopyrite. Enargite, tetradymite, altaite, and nagyagite are recorded, but are rare. The pyrite is always the oldest sulphide, and the gold fills minute fractures in it, or in the quartz.

The metasomatic alteration of the wall rock shows considerable variations. The most intense action is shown by some quartz-tourmaline veins; the adjoining amphibolite is altered to garnet, tourmaline, and

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H. B. C. Nitze, *Bull.* 10, North Carolina Geol. Survey, 1897.

L. C. Graton and W. Lindgren, Reconnaissance of some gold and tin deposits, etc., *Bull.* 293, U. S. Geol. Survey, 1906.

H. D. McCaskey, Gold, etc., in the Eastern States, *Mineral Resources*, U. S. Geol. Survey, 1908, pp. 645–681 (with literature).

F. B. Laney, The Gold Hill mining district, *Bull.* 21, North Carolina Geol. Survey, 1910.

Stephen Taber, Geology of the gold belt in the James River basin, *Bull.* 7, Virginia Geol. Survey, 1913.

magnetite.[1] In some of the Dahlonega veins the included amphibolite, as well as the adjacent wall rock, is altered to well-developed crystals of pale-red garnet and a dark-green mica. The garnets contain visible gold; the quartz itself contains pyrite, pyrrhotite, galena, and chalcopyrite.[2] This mode of alteration is much like that noted in the rocks adjacent to the pegmatite dikes of the same region.

In other veins a chestnut-brown biotite is the only mineral resulting from metasomatic alteration; in places both muscovite, in comparatively large foils, and biotite are present, sometimes with calcite or dolomite, besides more or less pyrite or pyrrhotite. Amphibolite is the most easily attacked of the various kinds of country rock. The alteration of granite is usually slight.

The replacement bodies are generally in the acidic schist derived from volcanic fragmental rocks; these are extensively silicified and contain also both sericite and biotite as products of alteration.

### The Gold-Quartz Veins of Ontario[3]

Since 1915 the gold production of the United States has diminished from a maximum of $100,000,000 to about half that amount. On the other hand a great increase has taken place in Canada since 1910, the output reaching 3,050,600 fine ounces in 1932, which is about 63 million dollars (United States currency).[4] Of this, about 80 per cent now comes from Ontario, principally from the Porcupine and Kirkland Lake districts.[5]

[1] L. C. Graton, *op. cit.*, p. 47.

[2] W. Lindgren, *op. cit.*, pp. 126–127.

[3] A. H. Means, Tourmaline-bearing gold-quartz veins of the Michipicoten district, Ontario, *Econ. Geol.*, 9, 1914, pp. 122–135.

H. S. Robinson, Geology of the Pearl Lake area, Porcupine district, Ontario, *Econ. Geol.*, 18, 1923, pp. 753–771.

A. G. Burrows, The Porcupine gold area, *Ann. Rept.*, Ontario Dept. Mines, 33, pt. 2, 1925.

E. Y. Dougherty, Mode of formation of Porcupine quartz veins, *Econ. Geol.*, 20, 1925, pp. 660–670. Explains the quartz as deposited by a combination of filling and replacement. Mining geology of the Vipond gold mine, Quebec meeting, Canadian Inst. Min. and Met., 1934, pp. 25.

A. G. Burrows and P. E. Hopkins, The Kirkland lake gold area, *Ann. Rept.* Ont. Dept. Mines, 32, pt. 4, 1925. Reports on many other districts are published by the Ontario Dept. of Mines.

J. B. Tyrrell and R. B. Hore, The Kirkland Lake fault, *Proc.*, Royal Soc. Canada, 20, Sec. 4, 1926.

T. L. Gledhill, *Ann. Rept.*, Ontario Dept. Mines, 36, pt. 2, 1927.

E. W. Todd, Kirkland Lake gold area, *Ann. Rept.*, Ontario Dept. Mines, 37, pt. 2, 1928.

L. C. Graton and H. E. McKinstry, Outstanding features of Hollinger geology, *Bull.*, Can. Min. Met., 249, 1933, pp. 1–20.

[4] There is a premium on gold in Canada because (since 1931) of the decline of Canadian dollars measured by United States currency; therefore, production is best stated in ounces.

[5] S. D. Strauss, The cost of producing gold, *Eng. Min. Jour.*, July, 1932, pp. 368–370.

The pre-Cambrian gold deposits, widely scattered in Ontario, seem to be most productive in the central part of the province. Beginning in Quebec, they range toward the west through Rouyn, Kirkland Lake, Larder Lake, Abitibi, Porcupine, Matachewan, Swayze township, Michipicoten, Goudreau, Lake of the Woods, Rainy Lake, Red Lake, and Rice Lake, the last in Manitoba.

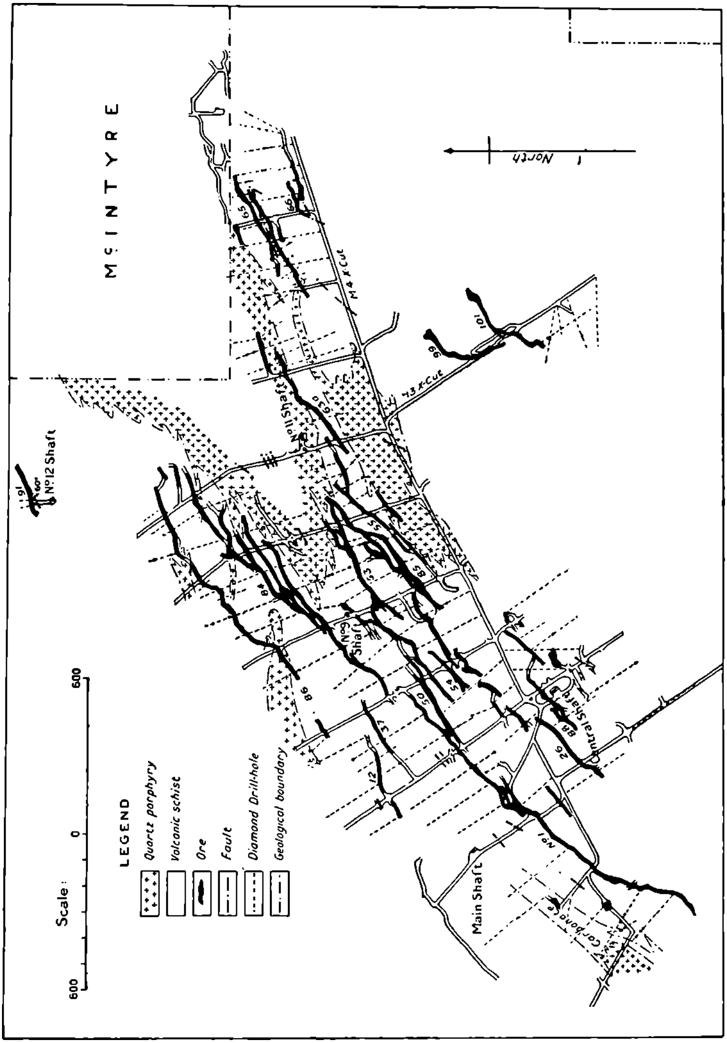


FIG. 269.—Plan of 550-foot level Hollinger mine. (After A. G. Burrows.)

The most common type is represented by quartz veins with little gold and scant sulphides, tourmaline, ankerite, pyrrhotite, and occasionally tellurides. Alteration of country rock is marked by development of sericite, carbonates, biotite, and tourmaline.

The second type is represented by the Kirkland Lake veins which carry quartz, native gold, no tourmaline, and scant sulphides. They are related to the mesothermal veins.

The third type, represented by the Rouyn deposits, shows lenses or veins with much pyrite, chalcopyrite, and pyrrhotite, sometimes magnetite and native gold (see p. 687 under Gold-Copper Deposits).

A distinct and peculiar magnesian metasomatism with minerals like chlorite, cordierite, gedrite, and other amphiboles accompanies this third type.

The fourth, more rare, type carries gold in arsenopyrite.

The most important deposits occur in folded synclines of Keewatin greenstone and lavas, and of Timiskaming sediments. Genetically, they are probably connected with post-Timiskaming and pre-Bruce intrusives of granite, quartz-monzonite, and syenites.

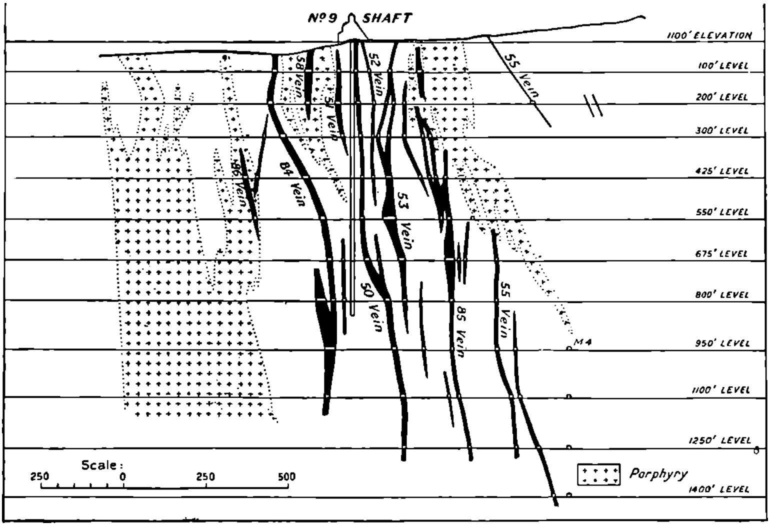


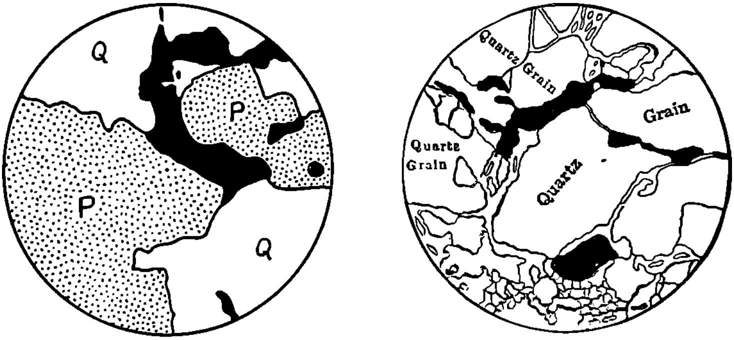
FIG. 270.—Section of Hollinger vein system looking northeast. (*After A. G. Burrows.*)

At various places there appear, also in the Keewatin greenstone, large hypothermal lenses of sulphide ores carrying copper and zinc with a little gold. Of these the Sherritt Gordon and the Mandy in Manitoba are well known examples (see p. 687 under Gold-Copper Deposits).

The Porcupine is the outstanding district. From 1910 to 1932 the production from this district amounted to over $300,000,000, the principal mines being the Hollinger, the Dome, the McIntyre, and the Vipond. Since 1910, the Hollinger has produced 60 per cent of the figure mentioned. In 1926, the mine yielded $14,800,000 from nearly 2,000,000 tons of ore averaging $7.64. Since then the annual production has declined to about $10,000,000. There is very little silver in the ore.

Most of the mines use the cyanide process preceded by fine grinding. Amalgamation is generally abandoned. Some of the new mills have introduced flotation with cyaniding of the re-ground concentrates.

The depth reached in the Hollinger mine is 3,800 feet. The Hollinger mine is working a complex vein system (Figs. 269 and 270) which is enclosed in a thick series of successive flows of Keewatin lavas, in part schistose, which vary from andesite to basalt, sometimes separated by graphitic (sedimentary?) bands. The whole series together with the overlying Timiskaming greywacke and conglomerate is folded in an east-pitching syncline. The Keewatin is intruded by "quartz-porphyry," which L. C. Graton thinks is intruded earlier than the Timiskaming sedimentation. The veins generally avoid the quartz porphyry.



*A* *B*

Fig. 271.—*A*, drawing of polished section of ore from the 800-foot level, Lake Shore mine, Kirkland Lake, showing gold (black) deposited on pyrite (*P*) and quartz (*Q*). Magnified 87 diameters. (*After E. W. Todd.*) *B*, Drawing of thin section showing native gold deposited in crushed gold-quartz, Rea vein, Porcupine. Black spots are native gold. Magnified 30 diameters. (*After A. G. Burrows.*)

The outlying intrusions are of quartz monzonite, and the veins are considered post-Timiskaming and later than the general folding. The quartz monzonite may have yielded the magmatic emanations which produced the veins.

The veins contain, besides predominating quartz, ankerite, albite, tourmaline, and clinozoisite. Metallic minerals are gold (in both quartz and metasomatic country rock), pyrite, arsenopyrite, pyrrhotite, galena and sphalerite, scheelite, molybdenite, and some tellurides (sylvanite, petzite, hessite, and altaite). The sulphides are present in scant amounts.

The altered country rock contains much pyrite, carbonates, and sericite; an analysis gives, for example, in per cent, $SiO_2$, 28.96; $Al_2O_3$, 16.89; FeO, 4.83; CaO, 7.97; MgO, 3.34; $K_2O$, 3.62; $Na_2O$, 1.12; $CO_2$, 11.04; $H_2O$, 1.79; $FeS_2$, 20.36 (Burrows). The leaching of $Na_2O$ is not so complete as is usual in the California districts, but otherwise the metasomatic processes are similar. The gold values are fairly evenly divided between quartz and altered rock. Much of it is so fine that it can not be recovered by panning or amalgamation. According to some authors, the coarser gold is deposited in the crushed quartz and seems to be distinctly later than the earlier quartz mineralization (Fig. 271). Graton says, however, that the distribution of the gold is not controlled by late fracturing or by deformation of the vein walls, but by the distribution of pyrite and ankerite.

The prevailing structure is lenticular (Fig. 272); sometimes, indeed, the veins appear corrugated and folded.

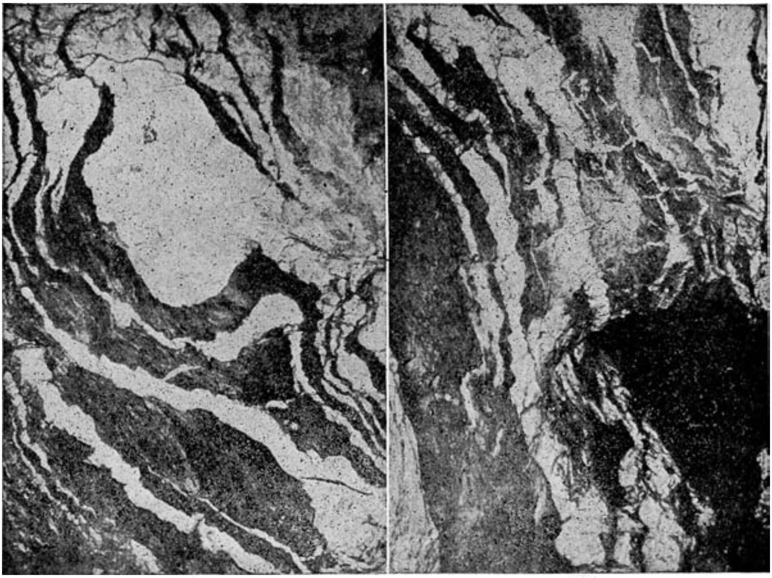


FIG. 272.—*A* and *B*, photographs of veins in Hollinger mine, Ontario. (*After A. G. Burrows.*)

The development of the quartz by a combination of filling and replacement was advocated by Dougherty (1925). Graton and McKinstry say that

> the striking complexity of vein shapes is ascribed to the effect of fracturing stresses on rock material already highly complex structurally, and to selective replacement of contorted beds rather than to important deformation of the quartz veins after they were formed. Minor faulting is the only post-vein disturbance.

They appear to believe that much of the quartz replaces an early ankerite which originally replaced the folded country rock. So proceeds the dispute between replacement and filling, already referred to in the description of the California quartz veins, with, as yet, no quite convincing conclusions. It must be pointed out, however, that it is a little difficult to account for a carbonate-pyrite-sericite alteration of the country rock and a simultaneous silicification of other parts of the country rock, as the two are separated by sharp boundaries.

The other outstanding district is Kirkland Lake. Since 1913, it has produced about $120,000,000, United States currency, the output rising rapidly. In 1932, the production reached about $22,000,000. The principal mines are the Lake Shore (which now exceeds the Hollinger in production), the Teck-Hughes, the Wright-Hargreaves, the Sylvanite, and the Kirkland Lake Gold.

The metallurgical process used is cyanidation, followed by fine grinding of tailings, flotation, and cyaniding. The presence of tellurides complicates the process; free gold is often visible in the ore. The ores of the district contain from $12 to $15 per ton.

FIG. 273.—Gold-quartz vein, Goudreau, Ontario, with dark bands of tourmaline. (*After T. L. Gledhill.*)

The Kirkland Lake shaft has reached 5,000 feet (Fig. 274). In the Lake Shore mine the deepest shaft is 3,800 feet, but 2,800 feet is the present deepest working level. In 1932, the Lake Shore produced about 50,000 fine ounces per month.

The Kirkland Lake deposits are contained in a synclinal belt of Timiskaming conglomerates, tuffs, and graywacke, intruded by syenitic rocks of varying acidity; the more basic facies is called "lamprophyre." The syenites are not schistose, but the Timiskaming rocks are slightly schistose in places. There are no Keewatin rocks in the district.

The veins, entirely different from those of the Porcupine district, are simply bodies of quartz filling fractures in and partly replacing the country rock along a faulted zone—the "main break." The faults follow a remarkably straight line for 2¼ miles striking N.65°W. The best ore is associated with the north fault, along which a reverse displacement of about 2,000 feet has taken place (Fig. 274). This shows in the Kirkland Lake, the Teck-Hughes, the Lake Shore, and the Wright-Hargreaves mines. The dip is steep to the south. There are also barren cross faults some of which carry calcite and barite. The main fault planes are generally surrounded by subsidiary fractures and brecciated structures over a width from 5 to 100 feet which are occupied by quartz, in part replacing, in part filling. There is often a gouge along "the main break." Ore-shoots are found in all rocks traversed by the veins.

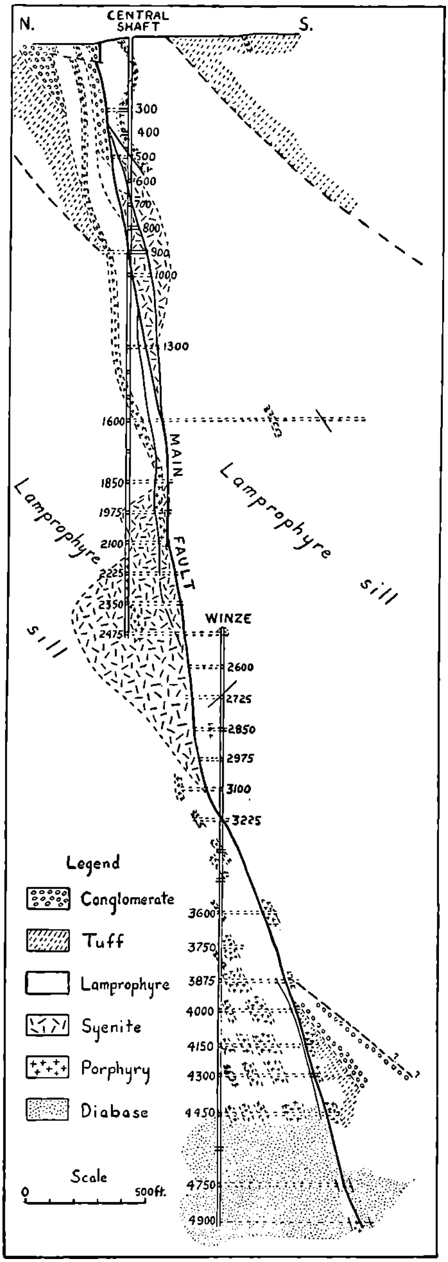


FIG. 274.—Vertical section of Kirkland Lake gold mine. (*After J. B. Tyrrell.*)

The alteration of the country rock, though widespread, is not intense, the new minerals consisting of quartz, calcite and other carbonates, (chlorite), albite, and sericite, probably also some adularia. Tourmaline is reported but is very scarce. The ore consists mainly of altered country rock with quartz stringers. Fragments and masses of unaltered orthoclase are often seen in the ore which occurs in syenite or porphyry.

The quartz was the first gangue mineral to form. Most of the metallic minerals are confined to fractures in quartz and country rock and consist of scant molybdenite, pyrite, chalcopyrite, tellurides, and gold. The gold is the latest mineral. There is not more than 2 per cent of sulphides in the ore. Altaite and calaverite are the most common tellurides, but there is also coloradoite and petzite, irregularly distributed. Todd presents many analyses of altered country rock, which, in general, indicate a strong introduction of carbonates. The relation between soda and potash is not consistently or strongly changed.

The veins are undoubtedly of mesothermal type. There is no trace of zoning. The veins will probably continue to much greater depths.

In some districts, *e.g.*, Larder Lake, the carbonatization of basic rocks has often resulted in large masses of coarsely crystalline products consisting of ankerite and bright-green mariposite (chromium mica). This altered rock, which appears to occupy considerable areas, is often cut by quartz veins and is remarkably similar to certain rocks along the Mother Lode in California, which have resulted from the alteration of serpentine.

In places, for instance, at the Rice Lake veins, Manitoba, biotite has developed in the country rock.

## The Pre-Cambrian Gold Veins of the Cordilleran Region

It has already been explained that of the pre-Cambrian veins, many, though not all, suggest formation at high temperature. The veins occur in the pre-Cambrian rocks of southern Wyoming,[1] at Atlantic and South Pass; also at Cherry Creek, Yavapai County, Arizona;[2] many veins in the pre-Cambrian of southwestern Arizona belong to this type. They usually carry more or less tourmaline.

The Black Hills of South Dakota[3] contain many gold-bearing deposits in the pre-Cambrian rocks. They occur, as a rule, in clay slates of sedimentary origin, not far from massive intrusive granites, also of pre-Cambrian age, and, while some of them are true veins with glassy quartz and free gold, others are lenticular bodies of highly altered rock. The best known among the latter is the Homestake lode at Lead. For many years the Homestake mine has been a regular producer, yielding annually from $4,000,000 to $8,400,000 (in 1930) in gold. The total production from 1875 to 1929, inclusive, was $226,121,128. The mine has been worked to a depth of 2,600 feet. The recovery is about $4 per ton (in 1930, $6.18 per ton) in gold, with little silver (Ag: Au = 1:5 by weight). About 63 per cent of the gold is recovered by amalgamation; the remaining very fine gold is extracted by the cyanide process.

Paige was the first to show that the Homestake deposit was formed by the replacement of an intensely folded and compressed bed of pre-Cam-

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[1] A. C. Spencer, *Bull.* 626, U. S. Geol. Survey, 1916, pp. 9–45.

[2] W. Lindgren, *Bull.* 782, *idem*, 1926.

[3] J. D. Irving and S. F. Emmons, *Prof. Paper* 26, U. S. Geol. Survey, 1904.

W. J. Sharwood, *Econ. Geol.*, 6, 1911, pp. 729–786.

Sidney Paige, The geology of the Homestake mine, *Econ. Geol.*, 18, 1923, pp. 205–237; *Bull.* 765, U. S. Geol. Survey, 1924; Folio 219, *idem*, 1925.

D. H. McLaughlin, *Eng. and Min. Jour.*, Oct. 12, 1931, Homestake Number. Best description of mine geology.

brian dolomitic rock interbedded with slate, biotite schist, and thin quartzite. The Homestake ore-bearing bed, of ferruginous dolomite

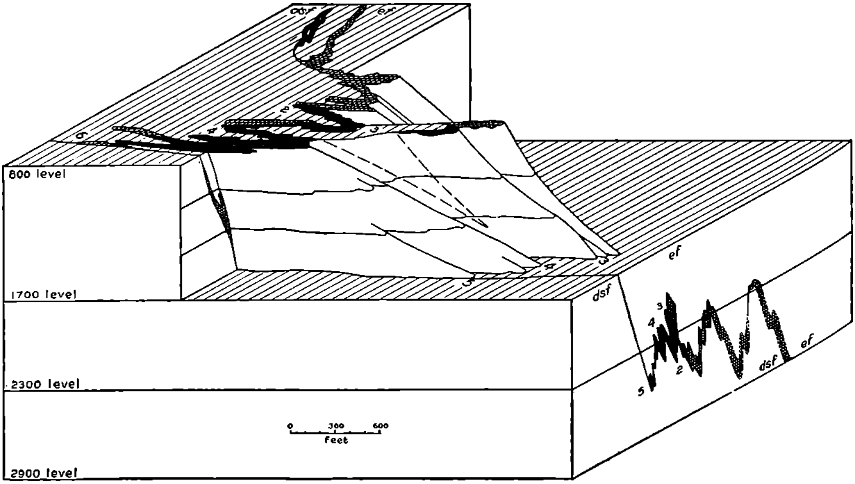

FIG. 275.—Isometric projection of the Homestake ore-bodies. Overlying Ellison formation (*ef*) removed over the plunging folds. Homestake formation; double cross-hatching; ore-bodies solid black. The underlying De Smet formation *dsf*. Ore-bodies: (1) Caledonia Ledge; (2) Incline Ledge; (3) Pierce Ledge; (4) No. 4 or Main Ore-bodies; (5) No. 5 Ledge; and (6) De Smet formation. (*After D. H. McLaughlin.*)

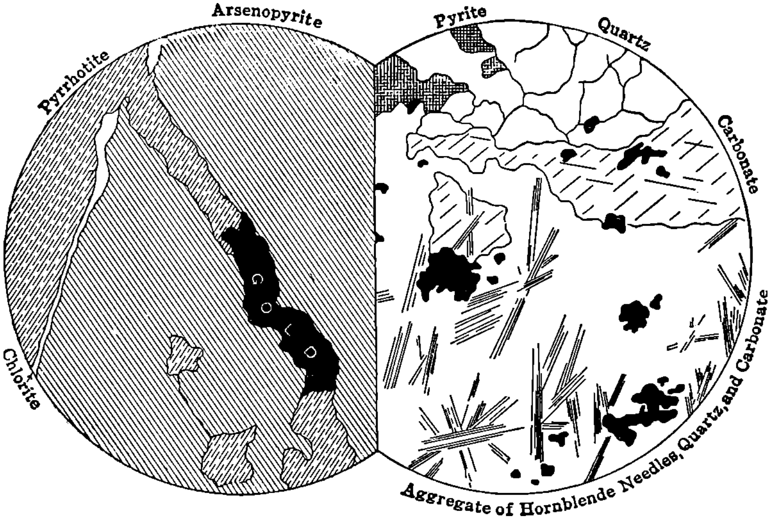

FIG. 276.—Thin sections of Homestake ores. Left: Gold and pyrrhotite (later), in arsenopyrite (earlier). Magnified 48 diameters. Right: Gold with quartz, iron-magnesium carbonate, pyrite, and cummingtonite. Magnified 32 diameters. (*After W. J. Sharwood.*)

with an original thickness of 40 to 60 feet, has in places been drawn out thin, in other places greatly re-folded and thickened. In general the dips are steep and the plunge is 40°S.E. (Fig. 275). The plunging ore-bodies lie along a group of minor anticlines that together form a large complex anticline. Each separate ore-body follows its own specific fold consistently to the southeast. In part, the dolomitic bed is replaced by cummingtonite (a monoclinic iron-magnesium amphibole) with quartz, ankerite, siderite, and calcite, also some garnet, iron-rich chlorite, and mica. The distribution of the sulphides conforms to the schistosity. The ore-bodies also include many small gash veins of coarse-grained, milky quartz which in places contains sulphides and gold. The coarse gold is chiefly associated with the sulphides. It often occurs in arsenopyrite or between pyrrhotite and pyrite (Fig. 276); the succession of ore minerals is arsenopyrite (oldest), pyrite, pyrrhotite, gold. Chalcopyrite is rare. Folding and shearing have in places affected the sulphides. McLaughlin thinks that gold was introduced with the sulphides, and that the cummingtonite and garnet were introduced earlier, during the general metamorphism of the beds.

Hosted and Wright hold that the mineralization is of Tertiary age and connected with the porphyries of that age occurring in the district. This view is probably untenable. The whole mineralization speaks emphatically in favor of a hypothermal, deep-seated mineralization.

### The Gold-Quartz Veins of Southeastern Alaska[1]

The gold-bearing veins of southern Alaska are closely allied to those of the Appalachian region and of Brazil, although they present some features that would rather connect them with the gold-quartz veins of California, which are believed to be formed under conditions of lower temperature or more moderate depth.

The veins occur mainly in the narrow strip of sharply folded Paleozoic slates and greenstones which form the western margin of the great batholithic mass of granodiorite of late Mesozoic age, 40 to 80 miles wide and continuous for many hundreds of miles parallel to the coast. The conditions are, therefore, essentially similar to those of the California gold belt, especially as the Paleozoic sediments farther west on Admiralty Island are adjoined by a belt of slates which are thought to correspond in age to the Mariposa slate of California.

In the long strip of coast country extending 300 miles there are numerous mining districts, among which are Windham Bay, Port Snettisham, Sheep Creek, Gold Creek, Douglas Island, Eagle River, and Berners Bay.

[1] A. C. Spencer, The Juneau gold belt, Alaska, *Bull.* 278, U. S. Geol. Survey, 1906.

A. Knopf, The Eagle River region, *Bull.* 502, *idem*, 1912.

A. F. Buddington and T. Chapin, Geology and mineral deposits of southeastern Alaska, *Bull.* 800, *idem*, 1929.

A. F. Buddington, Coincident variation of types of mineralization and of Coast Range intrusives, *Econ. Geol.*, 22, 1927, pp. 158–179.

The gold occurs in veins and lodes of various kinds, or more rarely, as on Douglas Island, in altered dikes of dioritic character that contain disseminated free gold and sulphides. The individual veins are rarely continuous for more than a few hundred feet, but often combine to form more extended stringer leads or lode systems. As the veins are later than the schistose structure of the rocks their tendency is to follow foliation planes, and in places they strongly resemble the lenticular veins of the Appalachian region, but continuous and cross-cutting veins also occur.

The gangue minerals are mainly milky quartz with some calcite or dolomite; tourmaline is occasionally reported, also magnetite. The ore contains free gold with more or less silver and associated with pyrite, pyrrhotite, sphalerite, chalcopyrite, galena, and arsenopyrite.

Few of the normal gold-quartz veins have yet been followed to great depth. Their width is from 1 to 8 or 10 feet at most, and the ores would probably range from $5 to $20 or more per ton.

The Treadwell ores, which have been mined on a large scale, are of low grade, containing about $3 in gold per ton, of which 60 to 75 per cent is free-milling, the concentrates yielding $30 to $50 per ton. The Treadwell deposits consist of a series of mineralized dikes of albite diorite in slates near the east shore of Douglas Island. The workings extend for 7,000 feet along the shore. The dikes dip about 50° northeast. The dimensions of the dikes are variable, the larger ones having a maximum width of over 200 feet. These ore-bearing dikes have been followed to a depth of 2,400 feet, and there appears to be no diminution of the average tenor of the ore at that depth. The average annual production of the Treadwell mines was, since 1910, $4,000,000. In 1917, an invasion of sea water filled most of the mines.

The ore-bodies are extensively fractured by a system of conjugated joints, along which there are irregular veinlets of quartz and calcite. The ore minerals are chiefly native gold, pyrite, and pyrrhotite, but chalcopyrite, galena, sphalerite, and molybdenite are also found. The important gangue minerals are albite, calcite, and quartz. The original diorite has been so thoroughly altered that it is difficult to establish its exact character. The metasomatic processes are described below in more detail.

On the mainland, opposite Douglas Island, the Alaska Juneau mine[1] is mining a great stringer lode from 100 to 300 feet wide in steeply dipping slate, amphibolite, and intruded gabbro, all of Mesozoic age. The minerals comprise quartz, ankerite, galena, pyrrhotite, and sphalerite. In the country rock, biotite and sericite have developed. The operations extend from sea level to 2,000 feet above. Though the ore contains slightly less than $1 per ton, profitable operations are now (1931) carried on in spite of many early difficulties. About 10,000 tons a day are treated

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[1] *Eng. and Min. Jour.*, September, 1932, Alaska Juneau number; Geology, by L. Wernecke, pp. 493–499.

by wet concentration. The concentrates which contain galena, pyrrhotite, pyrite, and native gold are amalgamated. The cost is about 55 cents per ton of ore.

The topographic features of this region permit the generalization that the vertical range of the deposits is over 5,000 feet. They have been followed 2,400 feet below sea level, and typical veins are found in the same regions at elevations of 3,000 feet or more above sea level. They were formed shortly after the great intrusion of granodiorite, and the vertical range now accessible must have been many thousands of feet below the surface of the earth at the time of the ore deposition.

### Metasomatic Processes in Veins of Southeastern Alaska

The facts that the Alaska veins contain abundant pyrrhotite and some tourmaline and magnetite and that the altered country rock contains biotite show that in many parts of the region the temperature of deposition was high. Albitization is a common process and appears to be independent of the amount of sodium in the country rock. It takes place not only in albite diorite but also in normal diorite, gabbro, and amphibolite.

Adjacent to the cross-cutting fissure veins of the Berners Bay district the metasomatic action is, as shown by Knopf, very similar to that in the California gold belt. Dolomite, sericite, albite, and pyrite are the principal new minerals formed in the rock.

The Treadwell mine worked large mineralized dikes of albite diorite in slates and greenstones. According to Spencer,[1] the original rock contained albite-oligoclase, microperthite, hornblende, and biotite, the latter two minerals in small amounts. The altered rock contains abundant albite, mostly developed by the replacement of microperthite, also quartz, calcite, muscovite, hornblende, rutile, epidote, magnetite, and pyrite. Albite is also found as narrow veinlets, although most of the veinlets consist of calcite and quartz. The composition of the altered rock differs considerably from place to place. Spencer holds that sodium has been added to the rock, together with carbon dioxide and sulphur. Calcium in the rock has been fixed by the carbon dioxide and suffered little leaching. The composition of one of the altered rocks is calculated as follows:

| | | | |
|---|---|---|---|
| Quartz | 2.34 | Magnesite | 0.11 |
| Albite | 84.36 | Siderite | 0.57 |
| Anorthite | 1.11 | Apatite | 0.13 |
| Zoisite | 0.91 | Rutile | 0.13 |
| Muscovite | 3.03 | Pyrite | 2.10 |
| Calcite | 3.80 | | |
| | | | 98.59 |

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[1] A. C. Spencer, The Juneau gold belt, Alaska, *Bull.* 287, U. S. Geol. Survey, 1906, p. 99.

Spencer and Knopf have shown that at several places on the mainland, near Juneau, dioritic rocks near the veins have been altered to products containing brown mica, probably biotite. Spencer describes the alteration in the Gold Creek district, which results in the development of biotite (I and II in the following table). The percentage mineral composition of the altered rock is calculated as follows: Quartz, 45; biotite, 22; carbonates, 20; titaniferous magnetite, 10.5; and sulphides, 2.5.

Knopf describes an altered and fresh amphibolite found near the Mendenhall Glacier (analyses III, IV, V, p. 675) and calculates the mineral composition approximately as follows:

| | Fresh amphibolite | Altered amphibolite |
|---|---|---|
| Orthoclase (mol.) | 6.7 | .... |
| Albite (mol.) | 18.3 | 39.3 |
| Actinolite | 43.7 | .... |
| Biotite[1] | 7.9 | 43.1 |
| Zoisite | 4.5 | 16.4 |
| Epidote | 18.3 | .... |
| Apatite | 0.6 | 1.2 |
| | 100.0 | 100.0 |

[1] By difference.

These changes differ greatly from those noted along fissure veins of the more ordinary type. In the first place, they include actual dehydration and distinct additions of aluminum, sodium, and potassium, the alkalies having doubled in quantity. In both localities ferric oxide is almost wholly removed, while there is some decrease in the ferrous oxide. Beyond this the two sets of analyses are dissimilar, for, while one indicates 20 per cent of carbonates, the other is entirely without carbon dioxide. As a consequence, the first has retained much more calcium and magnesium than the second. As to minerals, the rock rich in carbonate contains biotite, titaniferous magnetite, and sulphides; the one without carbonates yields albite, biotite, and zoisite, and Knopf considers that apatite has been formed in it.

In these deep-seated veins actinolite is unstable, while biotite, zoisite, and ilmenite were developed under the influence of the vein-forming solutions.

A similar development of biotite has been noted in the Kolar gold fields in Mysore, India,[1] which are very productive and are worked to a depth of 7,000 feet. The veins, which are probably of pre-Cambrian age, are contained in crystalline schists. The gangue is a glassy quartz

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[1] F. Hatch, The Kolar gold field, *Mem.*, Geol. Survey India, 33, pt. 1, 1901.
T. Pryor, *Bull.* 230, Inst. Min. and Met. (London), 1923.
The famous Kolar gold fields, *Mining Jour.* (London), Jan. 25, 1930.

ANALYSES OF FRESH AND ALTERED ROCKS FROM QUARTZ VEINS OF SOUTHEASTERN ALASKA
(Analysts, George Steiger, I, II, V; J. G. Fairchild, III, IV)

| | I | II | III | IV | V |
|---|---|---|---|---|---|
| $SiO_2$ | 47.76 | 44.69 | 48.30 | 52.92 | + 5.10 |
| $TiO_2$ | 1.48 | 2.25 | 1.01 | 0.99 | − 0.24 |
| $Al_2O_3$ | 13.98 | 14.97 | 13.59 | 20.53 | +17.70 |
| $Fe_2O_3$ | 1.99 | 0.60 | 3.12 | Trace | − 9.64 |
| FeO | 8.72 | 7.05 | 10.44 | 8.38 | − 7.80 |
| MnO | 0.14 | 0.14 | 0.25 | 0.28 | + 0.04 |
| MgO | 9.07 | 3.92 | 6.29 | 2.43 | −12.33 |
| CaO | 12.71 | 10.07 | 11.09 | 4.76 | −20.00 |
| BaO | Trace | 0.14 | ....... | ....... | ........ |
| $Na_2O$ | 1.65 | 2.36 | 2.16 | 4.67 | + 6.92 |
| $K_2O$ | 0.20 | 1.76 | 1.55 | 2.96 | + 3.83 |
| $H_2O-$ | 0.22 | 0.36 | 0.00 | 0.18 | + 0.34 |
| $H_2O+$ | 2.06 | 0.20 | 2.06 | 1.58 | − 2.25 |
| $ZrO_2$ | ....... | 0.02 | ....... | ....... | ........ |
| $CO_2$ | None | 8.47 | None | ....... | ........ |
| $P_2O_5$ | 0.12 | 0.26 | 0.26 | 0.57 | + 0.86 |
| S | 0.04 | ....... | ....... | ....... | ........ |
| $FeS_2$ | ....... | 0.27 | ....... | ....... | ........ |
| $Fe_7S_8$ | ....... | 2.25 | ....... | ....... | ........ |
| Less O | 0.02 | ....... | ....... | ....... | ........ |
| | 100.16 | 99.78 | 100.12 | 100.25 | −16.80 |

I. Green diorite, Gold Creek. Contains about 75 per cent green hornblende; remainder feldspar with some quartz.

II. "Brown diorite," Ebner mine, Gold Creek.

III. Amphibolite, Mendenhall Glacier. Sp. gr. 3.084. Dark olive-green rock.

IV. Altered amphibolite, Mendenhall Glacier. Sp. gr. 2.905. Dark-brown rock with pyrrhotite.

V. Gains and losses, in grams, in the alteration of 100 cc. of amphibolite to same volume of altered product.

with native gold and a little pyrite, pyrrhotite, arsenopyrite, and tourmaline. The veins contain pitching pay-shoots in which the ore is 5 feet wide, averaging $20 to $30 per ton in the deepest levels (Fig. 83) (p. 197). Total production was 16,300,000 ounces from 1880 to 1929.

## The Gold-bearing Veins of Brazil[1]

The great pre-Cambrian areas of Brazil contain auriferous lodes of great value. In many respects they are similar to those of the Atlantic coast of North America.

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[1] Orville A. Derby, Notes on Brazilian gold ores, *Trans.*, Am. Inst. Min. Eng., 33, 1903, pp. 282–287; *Eng. Min.-Jour.* 74, 1902, pp. 142–143. (*Footnote continued on following page.*)

The author owes the following notes to the kindness of Prof. L. C. Graton:[1]

The Morro Velho gold mine in Minas Geraes, Brazil, is remarkable for having produced in the neighborhood of $100,000,000 from a single ore-shoot of relatively small cross-section, mined down the dip for a distance of over 2 miles to a vertical depth of 7,526 feet. It is thus the second deepest mine in the world. The country rock, of earliest Paleozoic or pre-Cambrian age, is a great body of sericitic schist, which was originally a carbonaceous shale. Foliation and original bedding are at most places coincident; the southeasterly dip flattens toward the bottom of the mine from its normal value of about 45° to as low as 15°.

The attitude of the lode is unusual and has not been clearly made evident by earlier descriptions. The ore-body is a narrow band or "ribbon" averaging about 15 feet thick and some 500 feet wide. This ribbon extends diagonally downward from the surface along a direction of pitch that faithfully conforms with the plane, and (roughly) with the dip, of the bedding and regional foliation of the enclosing schist. But the ribbon stands edgewise rather than flat-wise to the plane of the rocks, so that its 500-foot dimension is (roughly) at right angles to the rock bedding. The lode is contained within the same 500-foot layers of beds from outcrop to mine bottom; in this sense it is conformable with its enclosing rocks. But the true plane of the lode strikes easterly and has almost vertical dip; therefore, it actually cuts strongly across the beds that enclose it. Decreasing dip in the bottom part of the mine has been attended by a squeezing and thinning of the beds; the ribbon itself is, therefore, folded and crumpled in the deeper levels contrasted with its relatively simple shape above. But this deformation affected the *locus* before the ore was introduced; the ore has not been disturbed since deposition.

The ore itself is a massive fine-granular sulphide, with pyrrhotite very strongly predominating over arsenopyrite and pyrite; but more than half its weight consists of ankerite, quartz, and albite. Minor to subordinate constituents include chalcopyrite, wolframite, chlorite, sericite, and scheelite. The gold occurs native in particles invisible except by magnification. The gold prefers the interior of arsenopyrite crystals and, as second choice, pyrrhotite or chalcopyrite close to arsenopyrite grains. The arsenic-gold ratio of the ore is notably constant. About the ore proper is an uneven envelope of *lapa seca*, a crystalline aggregate of ankerite, albite, and quartz with subordinate iron sulphides; and these same minerals penetrate the schist by replacement, as thin bands along certain bedding planes.

The ore is of surprising uniformity in mineral character and grade from top to bottom of the mine, averaging about $10 per ton and showing no dis-

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Georg Berg, Beiträge zur Kenntniss der Goldlagerstätten von Raposos, *Zeitschr. prakt. Geol.*, 1902, pp. 81–84.

E. Hussak, Der Quartzlagergang von Passagem, *Zeitschr. prakt. Geol.*, 1898, pp. 345–357.

Orville A. Derby, Gold-bearing lode of Passagem, *Am. Jour. Sci.*, 4th ser., 32, 1911, pp. 185–190.

[1] Extract from a paper in preparation by L. C. Graton and Guy N. Bjorge with permission of the St. John Del Rey Mining Company, Ltd.

coverable differences in kind, proportions, or textures of minerals throughout its great developed pitch length.

The lower half of the mine is dust-dry. Rock temperatures are unusually high, exceeding 130° F. at the bottom and with an almost constant increment of about 1° F. per 100 feet. Cooling and dehydration of the air by means of elaborate refrigeration plants, both on the surface and underground, suffice, however, to keep temperature-humidity conditions in a satisfactory state even in the deepest workings.

The Passagem lode, described by E. Hussak as a gold-bearing pegmatite dike, appears to be, according to O. A. Derby, a shattered pegmatite dike impregnated by gold, arsenopyrite, pyrrhotite, and tourmaline, with a little siderite and calcite (p. 769). The gold contains some bismuth. The appearance of cummingtonite allies the Passagem vein in an interesting manner to the Homestake deposit. In other parts of the Brazilian gold-bearing area palladium occurs in alloy with gold.

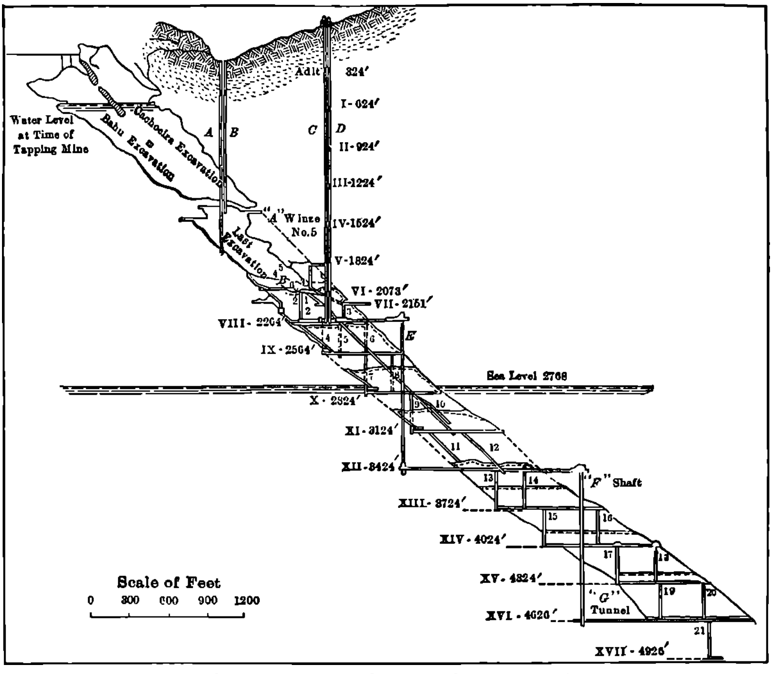

FIG. 277.—Vertical longitudinal section of the St. John del Rey Mining Company's Morro Velho Mine, Brazil.

### The Gold-telluride Veins of Western Australia[1]

Western Australia is an arid tableland of moderate elevation surmounted by short and low ridges (Fig. 278). Crystalline schists and

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[1] A. Gibb Maitland, *Bulls.* 4, 15, and 20, Geol. Survey Western Australia.
C. F. V. Jackson, *Bulls.* 13 and 18, *idem.* (*Footnote continued on following page.*)

granites are the principal rocks. The schists extend, with general north-south strike and vertical or steep dip, across the whole central part of the state and consist largely of amphibolites, massive or foliated, which have been derived by metamorphism from a basic rock—diorite, gabbro, or diabase. There are also highly altered sedimentary rocks such as quartzites and slates; more rarely limestones.

Granitic rocks, in part gneissoid, also occur extensively in the complex of crystalline schists. Lenticular masses of amphibolite are contained in the granitic rocks and vice versa, so that the sequence of the rocks is not always clearly apparent. Many of the granitic rocks are intrusive into the amphibolites.

The age of all these rocks is considered to be pre-Cambrian. Toward the northwest coast the old rocks disappear beneath transgressing horizontal Carboniferous limestones.

The gold deposits are contained chiefly in the amphibolites but also, though less commonly, in the granitic rocks.

The center of mining activity is at Kalgoorlie, about 350 miles east of Perth. From the mines within the so-called "Golden Mile" at that place, the larger part of the output of Western Australia has been derived, although other mines north and northwest of Kalgoorlie now contribute a considerable share. The most prominent among these outside mines are the Westralia-Mt. Morgan and Sons of Gwalia, in the Mt. Margaret gold field, and the Great Fingall, in the Murchison gold field, 300 miles northwest of Kalgoorlie. In this vicinity is also located the Wiluna mine, coming into production in 1932. The total gold production of Western Australia from 1886 to 1932, inclusive, was about $870,000,000; the annual yield, which is gradually decreasing, was $22,000,000 in 1917 and $10,000,000 in 1931. The yield of the Kalgoorlie mines from discovery to 1932 is about $430,000,000.

The geologists of Western Australia distinguish two principal modes of occurrence of gold-bearing lodes.

1. The normal quartz veins usually occur in the amphibolite or along the contact of granitic rocks and amphibolite. Most of the veins are short and conform in strike and dip with the steeply dipping schists, and

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E. S. Simpson, *Bull.* 6, *idem.*

T. A. Rickard, The telluride ores of Cripple Creek and Kalgoorlie, *Trans.*, Am. Inst. Min. Eng., 30, 1901, pp. 708–718.

P. Krusch, *Zeitschr. prakt.* Geol., 1903, pp. 321–331; 369–378.

C. O. G. Larcombe, The geology of Kalgoorlie, *Proc.*, Austral. Inst. Min. Eng., 5, 1910, pp. 1–312.

M. Maclaren and J. A. Thomson, *Min. and Sci. Press*, 107, 1913, pp. 45, 95, 187, 228, 374.

F. L. Stillwell, Geology and ore deposits of the Boulder belt, Kalgoorlie, *Bull.* 94, Geol. Survey Western Australia, 1929, 110 pp.

tend to form lenticular ore-bodies. The minerals accompanying the gold are galena, sphalerite, pyrrhotite, chalcopyrite, arsenopyrite, stibnite, bismuthinite, pyrite, scheelite, chlorite, calcite, and sericite. In addition tourmaline is reported from one mine, the Sons of Gwalia.

2. The quartz lenses are at many places surrounded by altered country rock; where this rock prevails, transitions are formed to the second class of composite replacement deposits or sheared zones, with no definite outlines, to which the name "lode formations" is given.

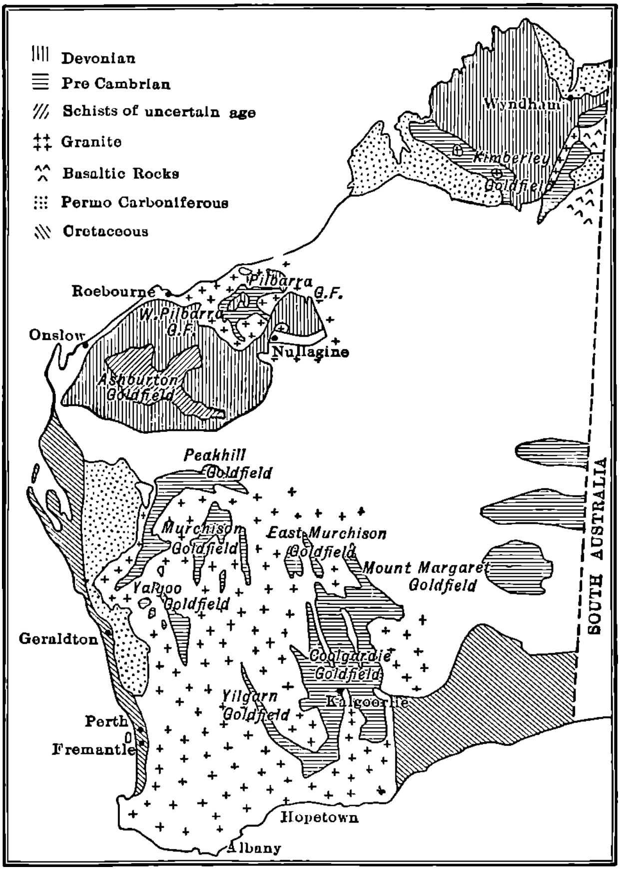

FIG. 278.—Geological map of Western Australia. Scale: 1 inch = 330 miles.

In the Boulder belt of the Kalgoorlie district the lodes intersect a quartz-dolerite-greenstone (Stillwell) derived by metamorphism and albitization of a quartz dolerite with which they seem genetically related. East and west of this are belts of amphibolite with sericite and carbonates; these are derived from basic lava flows and are known as "calcschists." They are considered unfavorable, and, as the scope of future mining in depth is limited by these rocks, hope for future production is said to lie in the southwestern part of the field and in lateral exploration. There are smaller bodies of sedimentary rocks, such as graphitic slate and quartzite. The lodes vary in width from 2 or 3 up to 80 feet.

The minerals of the lode formations include native gold and tellurides,[1] such as calaverite, krennerite, petzite, hessite, and coloradoite (telluride of mercury). Pyrite is abundant but is almost always finely divided, in contrast to the tellurides, which are often massive. Accessory minerals are chalcopyrite, sphalerite, galena, pyrargyrite, enargite, löllingite, fluorite, magnetite, specularite, rutile, calcite, dolomite, siderite, ankerite, tourmaline, albite, sericite, chlorite, and roscoelite. The ore-bodies, as shown by H. C. Hoover, form lenticular bodies. They have usually a rich core from which the gold content decreases outward, and the lenses are of large dimensions. Mining operations have attained a depth of 3,800 feet, and at this depth some of the mines are still in ore. The chief mines at Kalgoorlie were the Great Boulder, Ivanhoe, Horseshoe, Perseverance, Oroya-Brownhill, Associated, and Lake View Consolidated. The last is still in operation.

The ore varies from a dark-green, distinctly chloritic foliated schist, as in the Oroya-Brownhill mine, to pale-green sericite schists and to banded or massive dark rocks, flinty in places, and ranging from dark green to gray or brown. Small specks of pyrite are distributed through the ores, which contain bright-yellow gold associated with much pale-yellow calaverite and black, lustrous coloradoite. In places the tellurides and gold have developed as seams several centimeters thick in joint planes crossing the schistosity. The oxidized zone is from a few feet to 200 feet deep. At greater depths free gold is rarely visible and sulphides tend to take the place of the rich tellurides. The ores have gradually declined in value from $30 to $40 per ton near the surface to $7 to $10 in the deepest levels.

In Western Australia,[2] as in Alaska, there have been local differences in the processes of alteration. In the Pilbara gold field,[3] the granite next to the vein is altered to a greenish-gray rock, the calcium, magnesium, and sodium having been largely removed and the potassium considerably increased. It contains no carbonates. The course of alteration is entirely similar to that of many deposits found at intermediate depths in the Cordilleran region of the United States.

At Kalgoorlie, on the other hand, albite and carbonates are the principal products. The unaltered rock consists of an amphibolite contain-

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[1] F. L. Stillwell, The occurrence of telluride minerals at Kalgoorlie, *Proc.*, Austral. Inst. Min. Met., 84, 1931, pp. 115–190.

[2] W. Lindgren, Metasomatic processes in the gold deposits of Western Australia, *Econ. Geol.*, 1, 1906, pp. 530–544.

[3] A. Gibb Maitland, *Bull.* 15, Geol. Survey Western Australia, 1904, p. 12.

ing amphibole, chlorite, zoisite, and albite. The altered rocks forming the gold-bearing lode contain the minerals before enumerated and also nests and lenses of fine-grained quartz. The carbonate grains inclose irregular masses of tellurides and coarse gold, but the larger masses of calaverite also contain rhombohedrons of carbonates. Crystals of

ANALYSES OF FRESH AND ALTERED AMPHIBOLITES FROM KALGOORLIE

| | 3231 | 1936 | 206 | 1753 | 1751 |
|---|---|---|---|---|---|
| $SiO_2$ | 48.86 | 57.72 | 51.27 | 46.94 | 42.01 |
| $Al_2O_3$ | 14.91 | 9.68 | 13.85 | 12.49 | 8.42 |
| $Fe_2O_3$ | ......... | 6.49 | 1.54 | 0.33 | 2.45 |
| FeO | 11.13 | 9.17 | 2.63 | 9.20 | 15.76 |
| MgO | 7.65 | 1.63 | 4.18 | 3.56 | 1.67 |
| CaO | 12.19 | 5.05 | 6.40 | 6.43 | 7.07 |
| $Na_2O$ | 2.58 | 3.92 | 1.78 | 1.84 | 2.62 |
| $K_2O$ | 0.19 | 0.12 | 2.37 | 2.57 | 1.15 |
| $H_2O-$ | 0.04 | 0.16 | 0.40 | 0.09 | 0.23 |
| $H_2O+$ | 1.51 | 1.51 | 0.22 | 0.30 | 0.67 |
| $TiO_2$ | 0.22 | 1.13 | 0.23 | 0.14 | 0.81 |
| $CO_2$ | None | 1.84 | 8.02 | 13.41 | 15.65 |
| $P_2O_5$ | ......... | ......... | Trace | ......... | ......... |
| Te | ......... | ......... | Trace | ......... | ......... |
| $FeS_2$ | ......... | ......... | 8.41 | 2.25 | 0.30 |
| MnO | 0.90 | 0.09 | Trace | 0.32 | 0.41 |
| | 100.18 | 98.51 | 101.30 | 99.87 | 99.22 |

3231. Star of Colac, *Bull.* 6, Geol. Survey Western Australia, 1902, p. 67. "Rock consists of coarse-grained mixture of feldspar and chlorite. It contains colorless hornblende, saussurite with clear mosaic of albite, also ilmenite surrounded by leucoxene." Analyst, C. C. Williams.

1936. Hannans main shaft, at depth of 600 feet, *idem*, "Rather coarse-grained rock containing hornblende, chlorite, feldspar (albite?), ilmenite, and secondary quartz." Analyst, C. C. Williams.

206. Lake View Consols mine, 300-foot level, *idem*, pp. 23 and 67. "Strongly foliated gray lode stuff. Assay, 9 oz. 12 dwt. of gold and 6 oz. 7 dwt. of silver per ton. It shows chlorite and sericite on the cleavage planes." Analyst, E. S. Simpson.

1753. A foliated greenish-gray lode stuff from the 400-foot level, Ivanhoe mine, *idem*, pp. 23 and 67. "Contains chlorite, carbonates, a feldspathic material, ilmenite, and some quartz. Trace of gold and silver." Analyst, C. G. Gibson.

1751. Siderite rock, West cross-cut, 400-foot level, Ivanhoe mine, *idem*, p. 67. "Gray compact rock, containing carbonates, grains of quartz and of black iron ores, . . . altered feldspar, and some scaly green chlorite." Analyst, C. G. Gibson.

magnetite embedded in tellurides are reported. E. S. Simpson has shown that the ores are really derived from amphibolites by replacement. The replacement is irregular, albite, quartz, or carbonates alternately predominating.

The character of alteration is shown by the analyses on page 681. Analysis 3231 is probably fairly representative of many amphibolites of Western Australia. It contains no free quartz and is rich in lime and iron. There is very little potash, and only 2.58 per cent of soda.

The last three analyses may be roughly calculated as shown in the accompanying table:

PROBABLE MINERAL COMPOSITION OF ORES FROM KALGOORLIE

| | 206 | 1753 | 1751 | Remarks |
|---|---|---|---|---|
| Quartz | 29.52 | 25.20 | 21.44 | |
| Chlorite | 6.86 | 2.76 | 2.94 | $FeO.4MgO.Al_2O_3.3SiO_2.4H_2O$. |
| Albite | 15.12 | 15.70 | 22.12 | |
| Sericite | 10.54 | 21.52 | 9.58 | $K_2O.3Al_2O_3.6SiO_2.2H_2O$. |
| $CaCO_3$ | 11.42 | 11.50 | 12.61 | |
| $MgCO_3$ | 4.20 | 5.38 | 1.76 | |
| $FeCO_3$ | 2.32 | 14.60 | 23.20 | |
| $MnCO_3$ | ....... | ....... | 0.67 | |
| Pyrite | 8.41 | 2.25 | 0.30 | |
| Magnetite (?) | ....... | 0.47 | 3.53 | |
| $TiO_2$ | 0.23 | 0.14 | 0.81 | |
| $Fe_2O_3$ | 1.54 | ....... | ....... | |
| $Al_2O_3$ | 2.14 | 0.61 | ....... | |
| Hygroscopic water | ....... | ....... | 0.23 | |
| | 101.30 | 100.13 | 99.19 | |

At first glance the analyses of the altered rocks do not show very great changes so far as bases and silica are concerned. The principal differences are in the added carbon dioxide of the altered rocks, ranging from 8 to nearly 16 per cent.

Magnesium, calcium, and iron have been fixed as carbonates, the last also as pyrite. The combined water has decreased owing to replacement of chlorite by sericite. The silica set free has been deposited as quartz. Soda has decreased slightly but potash has increased, though not very greatly. The abundant development of albite and carbonates recalls the processes of replacement in the deposits near Angels Camp in California, and at the Alaska-Treadwell mine.

## Gold-Quartz Veins in South Africa

The pre-Cambrian areas of South Africa, particularly in the Transvaal and Rhodesia, contain many gold-quartz veins; although their production of about $10,000,000 a year seems small compared to the vast quantities of gold produced from the Rand conglomerates.

The Shamva[1] mine in South Rhodesia was the most productive, yielding nearly $2,000,000 per annum. It is now closed. This, probably hypothermal, deposit is a broad shear zone in quartzite; granite porphyry occurs on both sides of it. The shear zones are impregnated with calcite, quartz, and pyrite; the ore-shoots, which are separated only by assay walls from the adjoining quartzite, are long and wide, the maximum width being about 250 feet. The lowest-grade ore that can be mined contains $2.75 per ton. Many of the Rhodesian gold-quartz veins contain stibnite.

## Gold-Quartz Veins in Europe

It seems remarkable that the Fenno-Scandian shield should be almost barren of gold-quartz veins. Some of these deposits are found in northern Finland but have proved of little importance.

Gold-quartz veins, apparently of hypothermal type, occur in the Hohen Tauern of Austria, in gneiss and schist. They have been described by many eminent geologists,[2] beginning with von Cotta and Posepny. They contain quartz, pyrite, arsenopyrite, and native gold.

Other gold veins in the Austrian and Tyrolean Alps[3] are lenticular and some of them contain tourmaline.

In Bohemia, gold-quartz veins have been worked at Roudny,[4] and at Kasejovic.[5] At Roudny, the small but numerous veins occur in gneiss and aplite near a granite batholith and carry quartz, tourmaline, siderite, calcite, barite, fluorite; at Kasejovic, the veins contain pyrite, arsenopyrite, sphalerite, and native gold.

The gold-quartz veins in the Ural Mountains have likewise been described by many geologists.[6] The so-called ladder veins in the "beresite" dikes at Beresowsk contain quartz, tourmaline, and gold. A description by Purington differs somewhat from the older accounts.

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[1] J. M. McLaren, *Mining Mag.*, December, 1924, pp. 329–339.

[2] P. Krusch, *Zeitschr. prakt. Geol.*, 1897, p. 77.
F. Beyschlag, *idem*, p. 210.

[3] C. F. Behrend, *idem*, 1924, p. 9.

[4] F. Slavik, *Sitz-ber.* k. Böhm. Ges. Wiss., 12, 1912, pp. 1–28.
J. T. Singewald, Jr., *Econ. Geol.*, 5, 1910, pp. 257–264.

[5] A. Hofman and F. Slavik, *Bull.* Acad. Sci. de Bohême, 17 and 18, 1913.

[6] R. Beck, 1, 1909, p. 450.
C. W. Purington, *Eng. Min. Jour.-Press*, June 13, 1903.

Other gold-quartz veins are found in many Siberian mining districts, *e.g.*, in the Alatau and the Amur regions.

## COPPER DEPOSITS

### The Copper-tourmaline Deposits

In several parts of the world the association of chalcopyrite with tourmaline is fairly common. In addition the ores may contain some gold and silver. Other minerals present are magnetite, specularite, rutile, pyrite, pyrrhotite, molybdenite, bismuthinite, wolframite, scheelite, tetrahedrite, quartz, siderite, fluorite, and biotite. Anhydrite is often present but barite is rare.

The copper-tourmaline veins of Cornwall, which also carry cassiterite, establish the transition from these deposits to the tin-bearing veins. Other occurrences, like those of Meadow Lake, California, and the Passagem lode, Brazil, emphasize the transition to the gold-bearing quartz veins.

The deposits are in part fissure veins, in part replacements in brecciated or sheared zones. In both modes of occurrence the country rock is subject to intense metasomatic changes, and tourmaline is developed by replacement, often for a considerable distance from the solution ducts. The normal form of alteration in feldspathic rocks is sericitization, sometimes accompanied near the veins by silicification. The feldspars are replaced by sericite, and both feldspar and quartz are penetrated by acicular tourmaline prisms, usually of bluish-gray color. Chalcopyrite, pyrite, and sometimes other sulphides also develop in the altered rock. The final product, unless the tourmalinization is unusually intense, lacks the coarsely crystalline structure of the typical greisen; the mica foils are usually small. In some deposits biotite develops.

**Chile.**—Many tourmaline-copper deposits are found in Chile. They appear to be connected with basic rocks—gabbro, diabase, porphyrite, diorites, etc.[1]

Von Groddeck has described the formerly important Tamaya mines, where veins containing copper ores cut diabase and porphyrite. The tourmaline is not only present in the filling of the principal vein, which dips 35° and is from 3 to 6 feet thick, but is also abundantly developed in the calcitic, chloritic, and micaceous altered country rock. Asbestos and tremolite (cummingtonite) are also mentioned.

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[1] A. von Groddeck, *Zeitschr.* deutsch. geol. Gesell., 39, 1887, pp. 237–266.
A. W. Stelzner, *Zeitschr. prakt. Geol.*, 1897, pp. 41–53.
Moericke, *Tschermak's min. u. petr. Mitt.*, 12, 1891, pp. 186–198.
L. Darapsky, Das Department Taltal, Berlin, 1900, pp. 167–172.

Similar veins at Las Condes, 90 miles east of Santiago, have been described by A. W. Stelzner. The rocks are granite and altered andesites. The vein filling consists of pyrite, chalcopyrite, quartz, and a loose mass of tourmaline needles and minute crystals of zircon, octahedrite, and specularite. The country rock is bleached and impregnated with pyrite and tourmaline. At Peralillo, 31 kilometers from Santiago, a similar pyrite-chalcopyrite vein in diorite carries tourmaline, molybdenite, scheelite, and cupro-scheelite.

The most prominent representative of this type in the world is the Teniente, or Braden,[1] deposit, situated in the Western Cordillera, 50 miles S.S.E. of Santiago, at an elevation of 8,000 feet. Very large ore-bodies have been developed by tunnels over a vertical interval of 2,500 feet, and the production is now about 15,000 tons per day of 2 to 3 per cent ore.

An intrusive mass of andesite porphyry and monzonite porphyry is injected into Tertiary andesitic lavas and has been extensively mineralized. The greatest ore-bodies surround a volcanic explosion vent about 3,600 feet in diameter. This is filled with rudely stratified tuff and the tuff is again intruded by masses of andesitic breccias of several kinds. The period of greatest mineralization followed the intrusion of the breccias and resulted in the replacement of the rocks surrounding the "crater" by quartz, specularite, tourmaline, chalcopyrite, pyrite and occasionally other sulphides. A still later epoch of mineralization yielded richer ores of chalcopyrite, bornite, tennantite, siderite, rhodochrosite and anhydrite; while during the last commercially unimportant epoch were deposited chalcopyrite, bornite, quartz, barite, and gypsum. A crystal of gypsum 10 feet long and 2 feet in diameter was found in a cave coated by these products of the last mineralization.

The deposit has been somewhat enriched by chalcocitization effected by descending waters.

The mineralization probably took place at a depth below the surface of about 4,000 feet and was effected by solutions containing much boric acid. They were probably very hot and may have reached the surface in gaseous form like the "soffioni" of Tuscany. The abundance of boron emanations connected with volcanic action in the southern Andes is remarkable. The numerous borax deposits in Chile, Bolivia, and the Argentine bear witness that some of these emanations reached the surface.

A. Locke suggests that the Braden "crater" could be interpreted as an area of collapse caused by removal of material underneath. Many features, however, speak against such an interpretation.

**United States.**—In the Cordilleran region of the United States there are a number of similar smaller deposits. They have been described from

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[1] W. Lindgren and E. S. Bastin, *Econ. Geol.*, 17, 1922, pp. 75–99.

Meadow Lake, California,[1] from the Blue Mountains of Oregon,[2] and from the upper Pecos River in New Mexico.[3]

The most productive deposit of this type occurred at the Cactus mine,[4] in the San Francisco Range, southern Utah. This deposit is contained in a brecciated zone in post-Paleozoic monzonite; it is in places 200 feet wide and at least 900 feet long. The ores have been followed to a depth of 800 feet. Brown tourmaline, with quartz, pyrite, and chalcopyrite, coats the fragments of brecciated rock, which is sericitized and contains some metasomatically developed tourmaline. Butler contributes a detailed study of this alteration. Other minerals, formed somewhat later than the tourmaline, are siderite, anhydrite, specularite, and tetrahedrite. The development of chalcopyrite continued during the whole epoch of mineralization. In 1908, 177,000 tons of ore were mined which yielded copper, 2 per cent; silver, 0.2 ounce, and gold, 0.01 ounce per ton. The ore was concentrated and smelted. The mine is now closed, poorer ores having been found in depth.

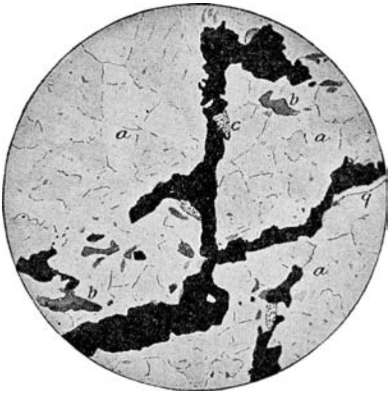

FIG. 279.—Replacement veinlet, War Eagle Mine, Rossland, B. C. *a*, granular aggregate of orthoclase with a little sericite; *b*, biotite; *q*, quartz; *c*, chlorite; black, pyrrhotite. Magnified 40 diameters.

### The Gold-copper Deposits

**British Columbia.**—In some of these copper deposits tourmaline is absent or rare. The Rossland district[5] is situated near the boundary line of the State of Washington. It has produced smelting ores since 1890 and has yielded a total of $62,300,000 in gold, copper, and silver. The ores contain about $5 to $10 in gold and 0.3 ounce silver, per ton, as well as 0.5 to 3 per cent copper. The mines are opened to a greatest depth of 2,200 feet; there was no production in recent years. The deposits are steeply dipping replacement veins along shear zones in monzonite and augite porphyrite. Small areas of granodiorite may represent upward extensions of the Trail batholith, the emanations from which are

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[1] W. Lindgren, *Am. Jour. Sci.*, 3d ser., 46, 1893, p. 201.

[2] W. Lindgren, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 551–776.

[3] P. Krieger, *Econ. Geol.*, 27, 1932, Nos. 4 and 5. W. Lindgren, *Prof. Paper* 68, *idem*, 1910, p. 113.

[4] B. S. Butler, Geology and ore deposits of the San Francisco and adjacent districts, Utah, *Prof. Paper* 80, *idem*, 1913, pp. 172–178.

[5] C. W. Drysdale, Geology and ore deposits of Rossland, B. C., *Mem.* 77, Canada Geol. Survey, 1915.

believed to have formed the deposits. The ore minerals are chalcopyrite and pyrrhotite (Fig. 279) with some pyrite, arsenopyrite, molybdenite, and bismuthinite. The gangue minerals comprise quartz, magnetite, calcite, and biotite with some garnet, arsenopyrite, actinolite, and wollastonite. The metasomatic action on the country rock has resulted in much secondary biotite. The veins were formed within the epoch of intrusion for they are intersected by many basic dikes related to camptonite. Apophyllite and other zeolites occur in druses. There is little evidence of secondary enrichment as perhaps is natural in a recently glaciated country.

**Manitoba.**—Lenticular replacement bodies of pyritic ores are found in the pre-Cambrian of Manitoba, Ontario, and Quebec. The sulphides, often massive, consist of pyrite, pyrrhotite, sometimes magnetite, with chalcopyrite and sphalerite. They usually approach the hypothermal class; the gangue is quartz, sericite, chlorite, calcite; garnet, amphibole, and scapolite are said to occur in some deposits. They occupy shear zones in gneiss or in Keewatin volcanics and are held to be later than the schistosity and shearing. They may be genetically connected with pre-Cambrian granitic intrusions.

In Manitoba,[1] the long pyritic deposit of the Sherritt-Gordon mine has been developed in late years. The Flin-Flon[2] deposit in the same region is actively being worked (in 1932) at the rate of 4,000 tons a day, producing annually 18,000 tons of copper, $2,000,000 in gold, and $200,000 in silver, besides much zinc. The ore contains, in per cent: copper, 1.71; zinc, 3.45; gold, 0.074 ounce per ton; and silver, 1.06 ounces per ton. The ore-body is traceable for 2,600 feet and is from 35 to 200 feet wide. It lies in ellipsoidal greenstone of the Amisk formation.

**Quebec.**—The Rouyn gold-copper district[3] has lately attracted much attention; the Noranda being the outstanding producing mine. It adjoins the Kirkland Lake district on the east. The deposits form replacements along shear zones in Keewatin rhyolitic lavas; more rarely, well-defined veins. The Keewatin, as well as the overlying Timiskaming beds, is intruded by many massive bodies ranging from granite to granodiorite and to syenitic rocks. There is also a diabase earlier than the deposits, and later diabase dikes probably later than the ore. The minerals are pyrite, magnetite, pyrrhotite, sphalerite, and chalcopyrite, the magnetite apparently later than the pyrite; the gangue consists of quartz, chlorite, sericite, amphibole (in part rhombic), and cordierite. It is distinctly an

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[1] E. L. Bruce, *Econ. Geol.*, 15, 1920, p. 386; *idem*, 24, 1929, p. 457.

[2] George Hanson, *idem*, 15, 1920, p. 574.

[3] H. C. Cooke, W. F. James, and J. B. Mawdsley, Geology and ore deposits of the Rouyn-Harricanaw region, Quebec, *Mem.* 166, Canada Geol. Survey, 1931, 314 pp.

Rogers Peale, The geology of the Waite-Ackerman-Montgomery ore deposit, *Bull.* 233, Can. Min. Met., 1931, pp. 1069–1086.

iron-magnesium, high-temperature metasomatism, allied to the later described cordierite-anthophyllite mineralization (p. 743). The Noranda (Horne) mine has become one of the more important of the world's gold-copper deposits with a present production of 4,500 tons of ore per day. Of this, 3,300 tons are smelted directly or used as flux and 1,200 tons are concentrated. The annual gold production is about $6,000,000. The mine is opened to a depth of 2,700 feet, though the present lowest stoping level is at 1,225 feet. It makes little water. The two main ore-bodies seem to follow tuff beds and stand vertically. There are smaller flat ore-bodies between rhyolite and andesite, ore replacing both. Other smaller ore-bodies are controlled by shear zones. Possibly the whole deposit is guided by the intersection of two major shear zones. The gold is irregularly distributed along smaller veins and along walls. According to Peter Price (oral information), it occurs in part native, in part also as tellurides. The average gold content is about $4 per ton. One gains the impression that a gold mineralization, perhaps similar to that of Kirkland Lake, has been superimposed on the pyritic mineralization.

**New South Wales.**—The copper lodes of Cobar[1] lie in an eroded desert range of older Paleozoic sediments; they are replacement veins from 10 to 120 feet wide, cutting sandstone and slate. The ore carries chalcopyrite, magnetite, and pyrrhotite in big lenses. The average contents are 2.5 per cent copper with $1 to $2 in gold and 2 to 3 ounces of silver per ton. Among the gangue minerals are quartz and an iron silicate, probably ekmannite. The greatest depth attained is 1,600 feet.

There was strong enrichment of copper in the upper levels as oxidized ores. A short distance below water level, secondary sulphides were found, comprising, according to Andrews, both chalcocite and chalcopyrite. There was little enrichment of gold and silver. No intrusive rocks are found within long distance of these deposits, but they were doubtless formed at great depth and at high temperature.

**Alaska.**—Somewhat similar are the copper ores of Prince William Sound.[2] The Beatson mine, now closed, was the largest copper mine in Alaska as to tonnage. The ores form elongated replacement bodies along shear zones in graywacke and slate of Jurassic (?) age, and contain pyrite, pyrrhotite, chalcopyrite, and cubanite, also a little sphalerite. The gangue is quartz with some ankerite and siderite.

**Sweden.**—About 1923, a rich sulphide deposit containing gold, copper, and arsenic was located by electric prospecting methods at

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[1] E. C. Andrews, Report on the Cobar copper and gold field, *Mineral Resources* 17, Geol. Survey N. S. W., 1913.

[2] A. M. Bateman, Geology of the Beatson Copper mine, Alaska, *Econ. Geol.*, 19, 1924, pp. 338–368.

Boliden,[1] on the Baltic shore of northern Sweden, an area heretofore markedly barren of gold-bearing ores. In 1931, 800 tons of ore were mined per day and smelted in local works. The smelting ore contains in per cent: 26 sulphur, 2.5 copper, and 12.5 arsenic; further about $12 in gold and 80 grams of silver per metric ton. It promises heavy production of gold and arsenic.

The steep lenses attain a length of 600 meters and a thickness of 30 meters and are enclosed in sericite schist, considered to be altered leptite; the leptite is an Archean supracrustal formation of acidic flows and tuffs (p. 740) overlain by phyllite and quartzite. The mineral succession is quartz, pyrite, fine-grained arsenopyrite, pyrrhotite, sphalerite, and chalcopyrite in coarse-grained intergrowth. Bismuth minerals, tetrahedrite, and galena also occur. The gold is chiefly connected with arsenopyrite occurring in minute grains. It is also found in chalcopyrite and as stringers in quartz. The ore shows pressure effects.

### The Copper-bearing Veins Allied to Contact-metamorphic Deposits and Pegmatites

In regions containing contact-metamorphic copper deposits it is not altogether unusual to find pyritic veins which exert an alteration on adjoining limestone similar to contact metamorphism, indicating that the vein-forming solutions possessed a high temperature.

The veins of cupriferous pyrite at Clifton, Arizona,[2] which intersect porphyry and contact-metamorphic limestone, are probably in part of this kind, for it was observed in many places that where they cut across limestone, tremolite and magnetite have developed adjacent to the veins. The primary deposits contain little copper but are enriched by surface waters.

One of the best instances is that of Massa Marittima, in Tuscany, described by B. Lotti[3] and V. Novarese. The great veins carry chalcopyrite, pyrite, galena, and sphalerite, and cut across Eocene limestone and clay shales. The limestone, but not the shale, is replaced near the vein by pyroxene, epidote, quartz, and sulphides. Some bismuth and a little tin are present in the ore. The mineralization is believed to be due to the intrusion of a granite of Tertiary age which on the surface does not come within several miles of the deposit.

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[1] S. Mörtsell, Gediget guld i Bolidenmalmen, *Geol. För. Förh.*, Stockholm, 53, 1931, pp. 394–414.

Alvar Högbom, Nya iakttagelser inom Norr-och Västerbottens urberg, *idem*, pp. 415–438 (Geology).

R. Schreiter, *Zeitschr.* deutsch. geol. Gesell., 83, 1931, pp. 233–243.

[2] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905.

[3] B. Lotti, Descrizione, etc., di Massa Marittima in Toscana, *Mem. descritt. Carta Geol. d'Italia*, 8, 1893.

K. Ermisch, *Zeitschr. prakt. Geol.*, 1905, pp. 206–241.

**Copper-titanium Veins.**—The small but interesting group of the chalcopyrite veins associated with titanium minerals is of uncertain affiliations. In some respects they are very closely allied to the pegmatites.

Such deposits have been described from Hereroland in South Africa,[1] at Rehoboth and Otjizongati. They are continuous quartz veins in mica schist and carry pyrite, chalcopyrite, bornite, and molybdenite with orthoclase, albite-oligoclase, rutile, ilmenite, apatite, and tourmaline. They contain also a little gold.

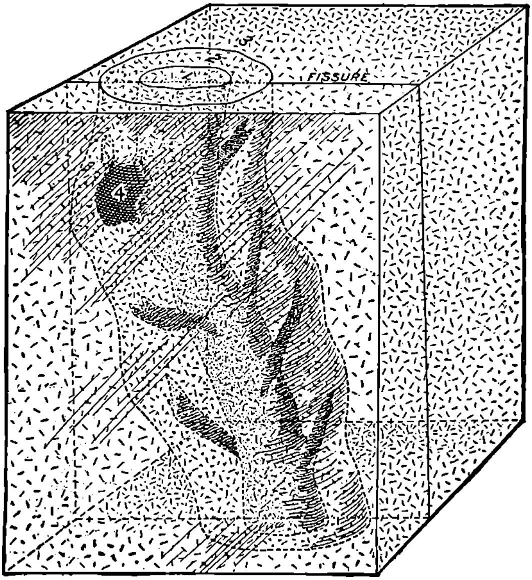

FIG. 280.—Stereogram showing relation of quartz pipe and mineralized quartz monzonite in the O.K. Mine, Beaver Lake district, Utah. 1, quartz; 2, altered monzonite; 3, monzonite; 4, high-grade ore. (*After B. S. Butler, U. S. Geol. Survey.*)

**Copper-molybdenum Veins.**—The association of copper and molybdenum is not uncommon but many other ore minerals are usually also present. Butler[2] describes the O.K. deposit, Beaver Lake district, Utah, in which the characteristics of pegmatites are curiously mixed with those of high-temperature veins. The deposit, worked to a depth of 400 feet, consists of a cylindrical body of extremely coarse and drusy pegmatitic quartz connected with a steep fissure and surrounded by a zone of sericitized quartz monzonite. The quartz has many offshoots of minor veins, which carry quartz, chalcopyrite, and molybdenite (Fig. 280). The ore in the upper levels is oxidized with secondary sulphides.

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[1] Eberhard Rimann, *Zeitschr. prakt. Geol.*, 22, 1914, pp. 223–226.

[2] B. S. Butler, *Prof. Paper* 80, U. S. Geol. Survey, 1913, p. 125.

## THE LEAD-SILVER-ZINC DEPOSITS

**Veins with Tourmaline.**—The combination of galena and tourmaline is rare, galena being generally found in deposits formed at lower temperatures. A. Knopf,[1] has shown that many of the veins in the contact zone and in the igneous rock of the Boulder batholith of quartz monzonite, in Montana, belong to this unusual group. The Alta vein is the best known and the richest of these deposits; it is supposed to have yielded over $32,000,000 in lead and silver, and was thus one of the greatest lead-silver deposits of the world. The monzonite contains a little tourmaline, its aplite dikes somewhat more, and the quartz veins are rich in this mineral. In the same district, H. V. and A. N. Winchell[2] observed a pyrite-tourmaline vein, the ore of which contains mainly silver with some copper and lead minerals. P. Billingsley and J. A. Grimes[3] have also examined these veins and conclude that they have been formed in or near the flat roof of that batholith.

**Veins with Garnet.**—The great Broken Hill lode[4] in the desert region of western New South Wales is representative of this rare class. This lode, which is said to have yielded lead, silver, and zinc to the value of about $688,000,000 since its discovery in 1883, has been described in great detail by Andrews in a notable monograph, and the intergrowth of the sulphides has been investigated by Stillwell. The lode has been opened over a length of 3 miles and the deepest shaft has attained 2,200 feet, ore continuing in depth. Output of zinc in 1931 was 54,000 tons.

The lode is contained in a thick series of pre-Cambrian sediments, which now appear as mica schists, sillimanite schists, and quartzite. Intrusive rocks were injected into the sediments, forcing them apart. The intrusive forms are pressure lenses and phacoliths, also *lit par lit* injections. The igneous rocks now appear as "augen-gneiss" (footwall gneiss) also pegmatites, diorite, and igneous quartz lenses. After the intrusion followed close folding and dynamic metamorphism. The folds are drag folds of weaker rocks against stronger buttresses. Garnet,

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[1] The tourmaline-silver-lead type of ore deposit, *Econ. Geol.*, 8, 1913, pp. 105–118; also *Bull.* 527, U. S. Geol. Survey, 1913.

[2] *Econ. Geol.*, 7, 1912, pp. 287–294.

[3] *Trans.*, Am. Inst. Min. Eng., 58, 1918, pp. 284–368.

[4] E. C. Andrews, Geology of the Broken Hill district, *Mem.* 8, Geol. Survey N. S. W., 1922.

F. L. Stillwell, Mineral constitution of the Broken Hill Lode, *Proc.*, Austral. Inst. Min. Eng., New Series, 64, 1926, pp. 97–172; *idem*, 67, 1927, pp. 187–220.

See also Report of subcommittee, *Trans.*, Austral. Inst. Min. Eng., 15, 1911, pp. 160–236.

W. E. Wainright and P. H. Warren, The Broken Hill South mine, *Mining Mag.*, January, 1918, pp. 12–19.

E. J. Kenny, The Broken Hill lode—its geological structure, *Proc.*, Austral. Inst. Min. Met., 87, 1932, pp. 217–245.

staurolite, sillimanite, micas, and other minerals developed along planes of schistosity. The ore-body developed in fault zones and drag faults and occurs sandwiched in between sedimentary sillimanite schist and intrusive gneisses. In the stress zone the lode takes the form of pressure lenses and replacement lenses comparable to those of the Homestake lode.

A fault zone 6 to 10 feet wide occupies the footwall of the deposit (Fig. 281). On the hanging-wall side the ore bulges in places to great

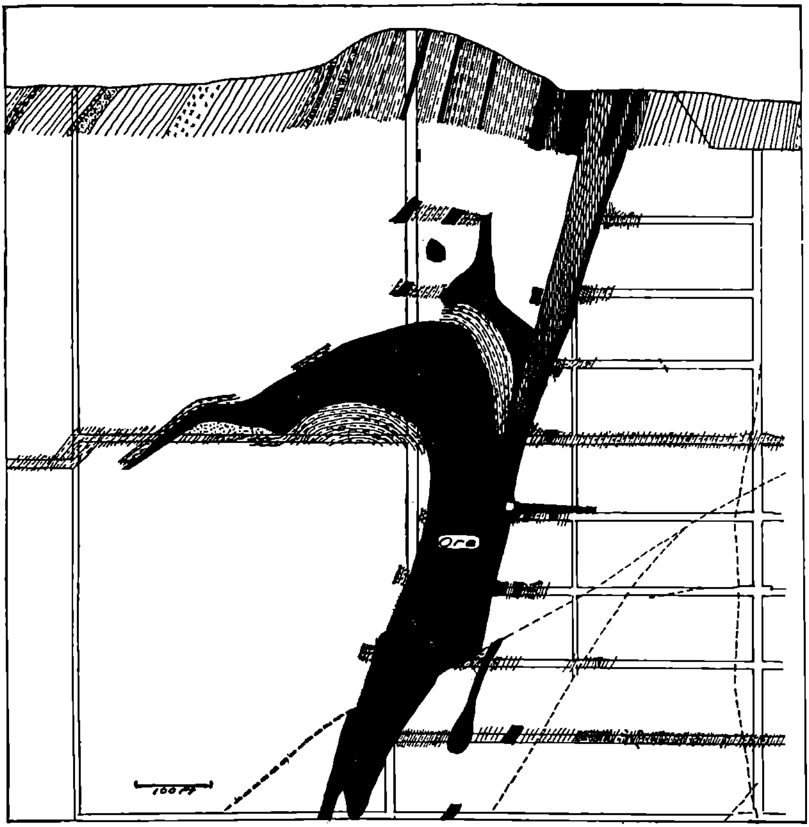

FIG. 281.—Vertical transverse section, Main Broken Hill Lode, Central mine, looking north. Ore replacing anticlinal drag fold. Black is ore. Full- and broken-line patterns indicate sillimanite gneiss. Minor intrusions of pegmatite and amphibolite. Scale: 0.4 inch = 100 feet. *(After E. C. Andrews. Reproduction from Trans. Am. Inst. Min. Met. Eng.)*

masses, which seem to follow the folded schists and form saddle-like bodies probably formed by replacement of the schist.

According to Andrews, the lode material rose as emanations under enormous pressure and forced its way along zones of strain. Its first appearance was as pegmatites with green feldspar; then followed the development of manganese garnet, rhodonite, pyroxene, magnetite, gahnite, apatite, calcite, and a little quartz. The quartz in the deposit has been formed at temperatures below 575° C. Later than the silicates are the sulphide minerals. Predominating are galena and sphalerite. The ore contains from 3 to 14 ounces of silver per ton, 14 to 16 per cent lead, and 8 to 18 per cent zinc.

The small amounts of pyrite, arsenopyrite, and a nickel-cobalt mineral are thought to be the oldest, followed by sphalerite, chalcopyrite, and cubanite, the latter two probably unmixing products of sphalerite. Then follow galena and tetrahedrite, overlapping in part with the sphalerite group. Later came recrystallization so that the chalcopyrite was partly converted to chalcostibite ($Cu_2S.Sb_2S_3$) and tetrahedrite partly dissociated into chalcostibite and chalcopyrite or pyrrhotite. The silver of the tetrahedrite appeared as pyrargyrite or dyscrasite. In some cases lead sulphide took part in the reaction and meneghinite ($4PbS.Sb_2S_3$) developed. See page 750 for a case where jamesonite forms by a similar reaction.

The surface gave little indication of the character of the deposit. Down to a depth of 300 feet there was a gossan, 20 to 100 feet wide, of quartz, limonite, manganese dioxide, hematite, and kaolin. Below this were found great masses of cerussite, anglesite, cuprite, and malachite, with abundant cerargyrite, embolite, and iodyrite. There was but little smithsonite, the zinc having been removed by leaching.

Where the oxidized ores changed to primary sulphides, there was a thin deposit of sooty chalcocite, rich in silver and copper; the slight thickness of these secondary sulphides is remarkable.

**The Sullivan Mine.**[1]—The important deposits in the Kootenay district of British Columbia must have been formed under similar conditions; they may be considered as high-temperature equivalents of the Coeur d'Alene lead deposits. The Sullivan mine is now probably the greatest producer of lead and zinc in the Americas. In 1930, nearly 2,000,000 tons of ore were mined. The workings reach 700 feet above the main adit and an incline is now being sunk below it.

The ores form replacement deposits in the Aldridge argillaceous quartzite of pre-Cambrian (Beltian) age. This formation is about 8,000 feet thick. The rocks are probably underlain by intrusive (Cretaceous?) granite.

The ore-bodies conform roughly to the strike and dip (10° to 30°) of the quartzites and grade into them; the greatest horizontal dimensions

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[1] S. J. Schofield, *Econ. Geol.*, 7, 1912, pp. 351–363; Geology of the Cranbrook area, B. C., *Mem.* 76, Canada Geol. Survey, 1915.

G. M. Schwartz, Microscopic features of Sullivan ore, *Eng. Min. Jour.-Press*, Sep. 4, 1926.

Engineers of Sullivan mine, *Bull.* 146, Canadian Inst. Min. Eng., 1924, pp. 401–465.

are 825 by 120 feet. The ore is an intimate, fine-grained mixture of galena and sphalerite with minor amounts of pyrite, pyrrhotite, magnetite, and jamesonite; the scant gangue contains manganese garnet, garnet, diopside, actinolite, and biotite with subordinate calcite. Banding by differential replacement is marked. The gangue minerals are earlier than the sulphides. Some cassiterite is present. The ore is said to contain 11 per cent lead and 13 per cent zinc.

**The Mexican High-temperature Lead Deposits.**—The close association of hypothermal vein deposits with contact-metamorphic deposits is not surprising when one considers that both are caused by highly heated solutions and fluids. When these reach a temperature at which lime silicates will be formed (400° to 500° C.) and when the solutions instead of permeating limestone beds follow fissures in limestone, the general mineralization will be very similar, for garnets and like silicates will appear. In veins, the mineralization is derived from deep-seated sources; in contact-metamorphic deposits, from the general body of intrusives adjacent. In many cases it may indeed be difficult to decide where to draw the line between these two kinds of deposits.

In 1906, Lindgren[1] described lead-bearing veins with garnet gangue from near Hachita, New Mexico. A little later Spurr and Garrey[2] gave excellent and detailed accounts of the deposit at Dolores, Mexico, where hedenbergite, andradite, actinolite, fluorite, and quartz occur in fissure veins, accompanied by pyrite and chalcopyrite. Zoisite, prehnite, and apophyllite occur in the later phases. Later, Basil Prescott brought attention to the chimneys of Santa Eulalia, Chihuahua, where pyrrhotite, pyrite, galena, and sphalerite are associated with iron and iron-manganese silicates, such as knebelite, ilvaite, hedenbergite, fayalite, and chlorite. In the upper parts, in the horizontal pipes, no silicates are found (p. 599).

A still more remarkable case is that of the Santa Barbara district,[3] where lead-zinc veins intersect Jurassic (?) shales with a limited amount of intruded rhyolite porphyry. These veins are in part at least characterized by the same silicates, particularly ilvaite, hedenbergite, and fayalite. There is apparently no limestone in the district.

## THE COBALT-TOURMALINE VEINS

The association of tourmaline with nickel and cobalt minerals in San Juan, Department Freirina, Chile, has been described by O. Stutzer.[4] In the same paper he gives a general review of the tourmaline veins.

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[1] W. Lindgren, *Prof. Paper* 68, U. S. Geol. Survey, 1910, pp. 335–337.

[2] *Econ. Geol.*, 3, 1908, pp. 688–725.

[3] Harrison Schmitt, Geologic notes on the Santa Barbara area, *Eng. Min. Jour.*, Sept. 15, 1928, p. 407.

[4] *Zeitschr. prakt. Geol.*, 1906, pp. 294–298.

# CHAPTER XXVIII

## PYROMETASOMATIC DEPOSITS

### INTRODUCTION

**General Features.**—In many geological provinces and during all ages molten magmas have invaded older rocks without reaching the surface. The intrusive magma cooled slowly and crystallized either as rocks with coarsely granular texture, such as granite, diorite, syenite, monzonite, gabbro, or diabase, or as the corresponding porphyries with holocrystalline groundmass. By means of uplift and subsequent erosion, these igneous rocks become exposed at the surface. If the rocks bordering the intrusives are crystalline schists or older igneous rocks, they seldom show much alteration along the contacts, but where they are of sedimentary

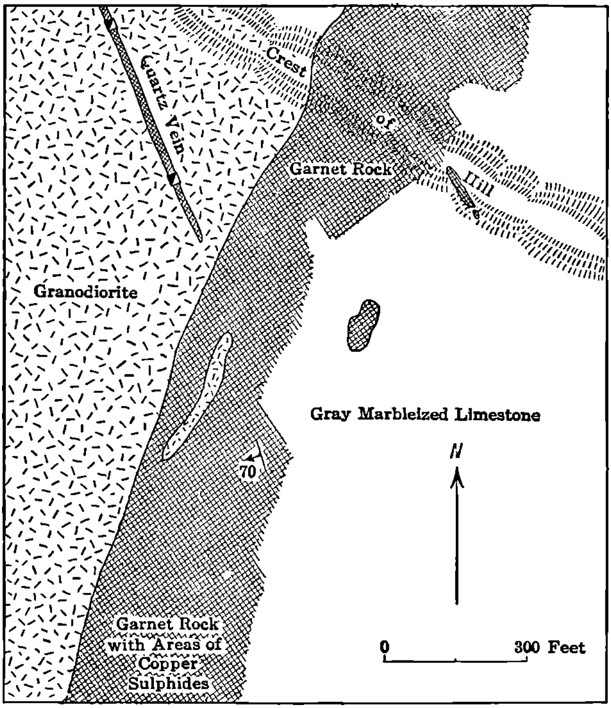

FIG. 282.—Sketch showing relation of ore zone to granodiorite and limestone, Bullion district, Nevada. (*After W. H. Emmons.*)

origin, like sandstone, shale, and limestone, considerable metamorphism is effected in them for a varying distance from the contact. In many places deposits of metallic ores or other useful minerals occur at these contacts, particularly where the older rock consists of limestone.

The form of such deposits is irregular and bunchy, but many of them are tabular by reason of following the contact (Fig. 282), or certain strata in the intruded rocks favorable for deposition (Fig. 283). Their mode of occurrence is metasomatic—that is, they are formed by replacement of the enclosing rock with addition of substance. They are formed at high temperature by emanations issuing from the intrusive. They are, therefore, suitably termed *pyrometasomatic* deposits.

The mineral association is characteristic: Chalcopyrite, pyrite, pyrrhotite, sphalerite, and molybdenite are the most common sulphides; magnetite and specularite the most common oxides. The heavy metals have been introduced from the outside. The most prominent gangue minerals are various silicates of calcium, magnesium, iron, and aluminum, the elements of which have in part been furnished by the carbonate rocks and shales. Among these so-called contact-metamorphic silicates are garnet, epidote, vesuvianite, diopside, tremolite, and wollastonite. Recrystallized, sometimes exceedingly coarse calcite is abundant; quartz is rarely present in large amounts. The ore minerals are usually later than the silicates.

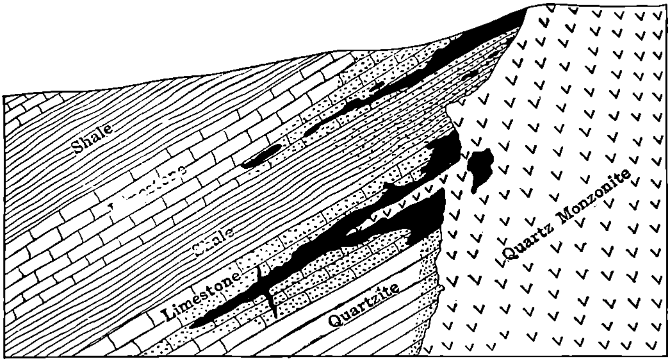

FIG. 283.—Diagram of contact-metamorphic deposit in vertical section. Ore shown in black. Contact-metamorphic rocks beyond ore stippled.

Some of these deposits contain valuable non-metallic minerals, like graphite or corundum, but ordinarily they are mined for the base metals. In the main the minerals are of simple composition and formulas. The association indicates an origin at high temperature, perhaps from 400° to 600° C. Close to the igneous rock the temperature may have been materially higher.

The igneous rock itself may be wholly fresh, or it may contain minerals closely allied to those in the deposit itself, such as garnet and epidote, or it may show sericitization with veins somewhat later than the alteration at the contact.

**History.**—In 1865 Bernard von Cotta described the iron deposits of the Banat province of Hungary and expressed the opinion that they were due to the action of intrusive rocks on the adjoining Mesozoic limestone. He also correlated these ores with those of Bogoslowsk, in the Ural Mountains, of Kristiania, in Norway, and of other districts. Von Groddeck,[1] however, first recognized them as a definite group to which he gave the name "Kristiania type." He stated that they were produced by contact metamorphism.

In 1894 Vogt[2] revived the interest in this class by describing the contact-metamorphic deposits of Kristiania. A little later the deposits at Seven Devils, Idaho[3]—the first of this type to be noted in the United States—were described, and, in a paper on the character and genesis of certain contact deposits,[4] the type was redefined and a number of examples from the United States were cited. W. P. Blake[5] mentioned the frequent occurrence of this type in Arizona. W. H. Weed[6] described a number of additional sub-types; a little later J. F. Kemp,[7] O. Stutzer,[8] and J. E. Spurr[9] discussed the subject again from a general standpoint.

Intrusive masses were found to be normally accompanied by contact-metamorphic deposits which, in some cases, were connected by transitions with the swarm of veins that usually surround these igneous bodies as an aureole of metallic treasure. The great importance of this type for the solution of problems related to the genesis of ore deposits became clear to the minds of many investigators. In Europe, many geologists have of late made detailed studies of contact-metamorphic deposits—among them B. Lotti, R. Beck, Loewinson-Lessing, E. Weinschenk, A. Bergeat, and V. M. Goldschmidt.

The views expressed in the above-mentioned papers, involving accession of material from the magma have not been allowed to pass unchallenged. F. Klockmann[10] expressed the opinion that these deposits were older accumulations of iron ore, altered at the intrusive contact; W. O.

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[1] A. von Groddeck, Die Lehre von den Lagerstätten der Erze, Leipzig, 1879, p. 260.

[2] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1894, pp. 177, 464; 1895, p. 154.

[3] W. Lindgren, *Min. and Sci. Press*, 78, 1899, p. 125.
W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, 1900, pt. 3, pp. 249, 253.

[4] W. Lindgren, *Trans.*, Am. Inst. Min. Eng., 31, 1901, p. 23.

[5] W. P. Blake, *idem*, 34, 1904, pp. 886–890.

[6] W. H. Weed, Ore deposits near igneous contacts, *idem*, 33, 1903, pp. 719 *et seq.*

[7] J. F. Kemp, Ore deposits at the contact of intrusive rocks and limestone, *Econ. Geol.*, 2, 1907, pp. 1–13.

[8] O. Stutzer, Kontaktmetamorphe Erzlagerstätten, *Zeitschr. prakt. Geol.*, 17, 1909, pp. 145–155.

[9] A theory of ore deposition, *Econ. Geol.*, 7, 1912, pp. 485–492.

[10] F. Klockmann, *Zeitschr. prakt. Geol.*, 1904, pp. 73–85.

Crosby[1] and A. C. Lawson[2] held that such bodies were simply the result of the ordinary circulation of meteoric character. As a general explanation neither of these views appears to be tenable.

In 1914, the discussion[3] regarding the origin of the garnet zones flared up again and was participated in by W. L. Uglow, W. Lindgren, J. F. Kemp, C. K. Leith, A. C. Lawson, and C. A. Stewart.

## CONTACT METAMORPHISM

**General Features.**—It will first be necessary to enter a little more deeply into the problem of contact metamorphism. This peculiar action of intrusive igneous bodies upon adjacent sedimentary rocks has been a well-known fact in geology since the days of Durocher (1846), and the processes have been described in much detail. Effusive rocks—that is, lava flows—rarely exert intense metamorphism beyond a baking or hardening of the sediments at the contact or an alteration of included rock fragments. The magmas intruded in sedimentary rocks, on the other hand, are in most cases surrounded by a halo of gradually fading metamorphism which may extend over a width of 1 or 2 miles, although usually much narrower. The immediate contact is ordinarily sharp, with no evidence of melting; only at contacts which were deeply submerged is there evidence of assimilation and extensive injection or feldspathization of the sediments. Slates and shales are ordinarily converted to hard, compact "hornfels"—that is, fine-grained holocrystalline rocks containing biotite, cordierite, andalusite, staurolite, scapolite, garnet, and feldspar; in extreme cases gneissoid rocks result. This metamorphism gradually diminishes, and at some distance the only evidence of change is a knotty texture of the rocks. Sandstones change to quartzite at the contact. Calcareous rocks become highly crystalline marbles and usually develop the contact-metamorphic minerals garnet, epidote, diopside, tremolite, vesuvianite, etc. According to the older view expressed by Rosenbusch, Zirkel, Brögger, and others, there is little change in composition aside from the expulsion of carbon dioxide from the limestones. The silicates are held to be formed under the influence of the heat of the magma from the impurities contained in the limestone.

It is well known that Rosenbusch[4] proposed a way of calculating the original character of a metamorphic rock from its present composition, and in regional metamorphism this is undoubtedly often justified. This

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[1] W. O. Crosby, *Trans.*, Am. Inst. Min. Eng., 36, 1906, pp. 626–646.

[2] *Min. and Sci. Press*, Feb. 3, 1912.

[3] See *Econ. Geol.*, 8, 1913 and 9, 1914; *Trans.*, Am. Inst. Min. Eng., 48, 1915, and *Min. and Sci. Press*, Oct. 17, 1914.

[4] Elemente der Gesteinslehre, 2d ed., 1901, p. 484.

thought has been followed by J. Barrell[1] in more direct application to contact metamorphism.

G. W. Hawes,[2] however, many years ago pointed out that emanations of boron and silica entered the sediments from the magma; and the introduction of tourmaline, for instance, has often been proved since then; it is in fact admitted even in the older textbooks of petrography.[3]

Petrographers of the present day, represented, for instance, by Alfred Harker and A. Lacroix, lean strongly toward the view that water and other gases containing $CO_2$, $H_2S$, S, Cl, B, and F must pass into the neighboring rocks while these are still heated by the intrusion.

The surrounding rocks of some intrusive bodies are, for instance, permeated for long distances by chloride solutions, this action resulting in the development of scapolite ($Na_4Al_3Si_9O_{24}Cl$).

The escape of these substances, as pointed out by Harker, may not have been uniform. A large part was doubtless given off while the magma was still fluid. Another part may have been liberated at the time of consolidation; still another part may have been retained until cooling had advanced considerably. Finally, fissures and shattering of the partly consolidated or congealed mass may have permitted gases from the still fluid interior or basal part to reach the outside of the intrusions.

It does not seem possible that atmospheric waters could have gained access to the contacts during intrusion. Both the heat of the magma and the pressure of the volatile compounds striving to free themselves would prevent such access.

The magmatic solutions enter into the rocks adjacent to the magma and produce a series of metasomatic changes the character of which will depend upon the composition of the solutions and their temperature, as well as upon the kind of country rock exposed to these hot extracts.

Ore deposits of this kind are rarely formed in argillaceous shales, sandstones and quartzite. Limestone, dolomite and calcareous shale, on the other hand, are easily permeable and appear to soak up the solutions like a sponge. It will be recalled that even at ordinary temperature limestone will absorb oil and similar substances. This permeation is facilitated by the fissuring and crushing along the contact. If the solutions are dilute, or if no opportunity is afforded for the carbon dioxide to escape, little replacement occurs but the limestone usually becomes coarsely crystalline, or scattered silicates will be produced by reaction between the calcite and the silica usually contained in most limestones. When the solutions are more concentrated and contain much silica,

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[1] Joseph Barrell, The physical effects of contact metamorphism, *Am. Jour. Sci.*, 4th ser., 13, 1902, pp. 279–296.

[2] G. W. Hawes, *Am. Jour. Sci.*, 3d ser., 21, 1881, p. 21.

[3] F. Zirkel, Lehrbuch der Petrographie, Leipzig, 2, 1894, p. 97.

sulphur, iron, and other metallic constituents, the replacement of carbon dioxide by silica will proceed with energy and the calcareous rock may be converted to a mass of ore and gangue minerals. In places whole beds of pure limestone or dolomite may be converted to garnet, diopside, and other silicates associated with recrystallized calcite and with magnetite, specularite, and simple metallic sulphides.

**Form and Texture.**—The frequency of rudely tabular bodies dependent upon parallelism with contacts or beds has already been emphasized. Sometimes the outlines are entirely irregular, but a selective action is very common by which the development of the best ore takes place along certain horizons in which the limestone is especially susceptible to metamorphism. This is doubtless caused by the physical condition of a particular bed, which may be a pure or an impure limestone; but the reason for this selective action is not always clearly apparent. While the deposits rarely extend far from the contacts, cases are known where they reach out along certain beds for a distance of 2,000 feet or more. The contact-metamorphic minerals in limestone often cease suddenly and form sharp contacts with the unaltered rock. Fissures may guide the solutions, and transitions to veins may be formed by deposition along them.

Favorite points of maximum ore development are limestone fragments included in the igneous rock or points where the calcareous rock projects into the intrusive. In most places the ore-bodies are of comparatively small size, but occasionally we find ore-bodies containing millions of tons. In case of copper deposits, subsequent oxidation and enrichment has often, as at Bisbee, Arizona, greatly enlarged the available ore-bodies.

The texture of the ore is commonly coarse, characteristic of replacement at high temperature. The absence of banding by crustification or deposition in open space is notable. But replacement processes sometimes result in a rude banding by a rhythmic metasomatic action or in orbicular or pipe-like arrangement of minerals. Such orbicular structures have been described by Trüstedt[1] in tin deposits at Pitkäranta, Finland, and by Knopf[2] in similar deposits in Alaska.

The calcite is often recrystallized to extremely coarse masses. Garnet and other silicates frequently show crystal outlines, but the sulphides, excepting pyrite, are rarely crystallized in distinctive form. Drusy cavities into which quartz crystals, garnets, etc., project are found in some deposits.

**Mineralogy.**—Among the ore minerals those of simple composition prevail. In order of abundance we have pyrite, chalcopyrite, bornite, pyrrhotite, sphalerite, molybdenite, arsenopyrite, and galena. Tetra-

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[1] O. Trüstedt, Die Erzlagerstätten von Pitkäranta, *Bull.* Com. géol. de Finlande, 19, 1907.

[2] Adolph Knopf, *Bull.* 358, U. S. Geol. Survey, 1908.

hedrite, jamesonite, and other complex sulphosalts are rare; most of those common in deposits formed at lower temperature are absent. Tellurides are rare, though tetradymite is reported from a few localities. Native gold is seldom present though most of the ores contain a trace of that metal. Graphite is not uncommon and platinum has been reported from one or two localities. The oxides are represented by magnetite, ilmenite, hematite, quartz, corundum, and spinels of various kinds. In all of the deposits of magmatic affiliations described above, magnetite and specularite are absent or scarce. Here they are plentiful. Magnetite is easily converted to specularite along the octahedral partings, and specularite crystals are often reduced to magnetite, wholly or in part.[1]

The silicates, especially those containing lime, magnesia, and iron, are abundant. We enumerate garnet (grossularite and andradite), epidote, zoisite, diopside, hedenbergite, tremolite, vesuvianite, forsterite, fayalite, anorthite, ilvaite, and wollastonite. Some of these contain the hydroxyl molecule.

Among the minerals containing chlorine, fluorine, or boron are scapolite, chondrodite, axinite, ludwigite, topaz, and tourmaline. The last two are not abundant.

Other minerals occasionally present are orthoclase, albite, muscovite, biotite, wolframite, scheelite, and fluorite while barite and anhydrite are absent. Andalusite, sillimanite, and staurolite, so common in contact-metamorphosed shales, are not usually connected with ore deposits. Chlorite is sometimes found in large crystals; serpentine is a product of decomposition of other silicates. Calcite, ankerite, and dolomite are usually present.

Many zeolites are also found but always crystallize during the last phase of mineralization.

On the whole, iron and copper deposits are the most common. Zinc, lead, gold, silver, and tin are much less abundantly represented.

**Intensity of Metamorphism.**—There are all degrees of metamorphism at the contacts of intrusive masses. Some magmas are evidently poor in volatile constituents and may exert only a slight recrystallization on adjacent limestone, while along the contact of others large bodies of ore may form. Along the same contact there are often great irregularities in mineralization. The degree of alteration of non-calcareous shales is a good indication of the intensity of the metamorphism, though these rocks rarely contain ore deposits.

**Influence of Composition of Igneous Rock.**—Highly acidic rocks, such as normal granites, are not always accompanied by ore deposits of the contact-metamorphic type, although they may produce widespread

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[1] F. F. Grout, Magnetite-martite-hematite, *Econ. Geol.*, 21, 1926, pp. 375–393.

P. Ramdohr, Beobachtungen an Magnetit, etc., Festschrift z. 150 Jahrfeier d. Bergakademie Clausthal, 1925.

effects of metamorphism and a later mineralization of quartz veins. The rocks accompanying the contact-metamorphic deposits of the Cordilleran type are generally monzonites or quartz monzonites or granodiorites.

Many examples show, however, that more basic rocks also may produce metallization of adjoining limestones—diabases, for instance, at Cornwall, Pennsylvania, and gabbro at the Nickel Plate mine, British Columbia.

**Alteration of the Intrusive Rock.**—Fresh granitic rock often adjoins the contact and is then separated sharply from the altered limestone. This is important to bear in mind. In many cases the intrusive rock is, however, more or less altered by sericitization or the development of other silicates. The alteration may have taken place simultaneously with the metamorphism of the sedimentary rock, or afterward.

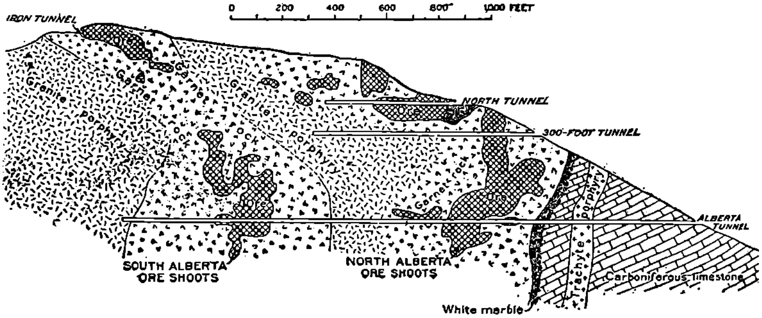

FIG. 284.—Longitudinal section, Empire mine, White Knob, Idaho. (*After J. B. Umpleby, U. S. Geol. Survey.*)

The hot intrusive, whether molten or just consolidated, and the adjacent cooler limestone form a system in which there will take place a vigorous exchange of material. Most of the changes will occur at the cooler side but the intrusive will also receive material, perhaps mainly carbon dioxide, and oxides of calcium and magnesium, from the sediments. Thus we find frequently epidote in large masses, more rarely garnet or diopside, developing in the igneous rock by replacement through the mass or in replacement veinlets. Such phenomena have been described by Spurr and Garrey from Dolores[1] and Velardeña,[2] by Bergeat[3] from Concepcion del Oro, all in Mexico; and by Umpleby[4] from White Knob, Idaho. At the latter place it is often impossible to define the contact between garnetized limestone and granite-porphyry (Fig. 284).

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[1] J. E. Spurr, G. H. Garrey, C. N. Fenner, *Econ. Geol.*, 7, 1912, pp. 444–492.

[2] J. E. Spurr and G. H. Garrey, *idem*, 3, 1908, pp. 688–725.

[3] A. Bergeat, *Boletin* 27, Inst. Geologico de Mexico, 1910.

[4] J. B. Umpleby, *Prof. Paper* 97, U. S. Geol. Survey, 1917; The genesis of the Mackay copper deposits, Idaho, *Econ. Geol.*, 9, 1914, pp. 307–358.

In other cases the alteration observed consists in sericitization, which is believed to be later than the contact metamorphism proper and largely caused by hot waters similar to those which form ordinary fissure veins; these waters may ascend from deeper portions of the intrusive from which emanations continue to be given off for considerable time after the irruption.

In the ordinary course of events an intrusive suffers various changes as to composition and texture at a contact. Basic facies may appear, or fine-grained texture, or a laminated or schistose structure.

The history of any igneous rock is marked by a successive development of minerals. Flett[1] states that in the Cornwall granites tourmaline, white mica, and topaz may be primary minerals and have a long period of formation, continuing to develop until the quartz separates out, and even long afterward.

Spurr and Garrey state that at Dolores and Velardeña many changes took place after consolidation in the small stocks of diorite, alaskite and monzonite breaking through Cretaceous limestone, and that a transition between contact-metamorphic deposits and normal veins is found at the former place. Pegmatitic, vuggy segregations are observed, and the older minerals have been altered to an early phase characterized by orthoclase, apatite, titanite, chlorite, quartz, diopside, and pyrite. At Velardeña grossularite garnet, and vesuvianite also appear.

Attention has lately been called to these so-called deuteric minerals[2] which also include such minerals as magnetite, chlorite, epidote, and zeolites (see p. 88).

**Succession of Events.**—The idea of successive epochs during the intrusion is frequently advanced. The general metamorphism by heat and dilute vapors is commonly considered to have taken place first, while foreign substances were added later. If we conceive magma rising to regions of lessening pressure, however, it is difficult to see why the escape of gases should not have begun at once when the melt was brought into contact with the cooler, surrounding sediments. The frequency of contact-metamorphic deposits in roof pendants of intrusive masses and along dikes in the roof shows that the volatile substances were concentrated in the upper parts, or cupolas, of the intrusives and in dikes radiating from these "gassed" parts of the magma. We observe also the scant metamorphism along flat intrusions, like laccoliths, and its relative strength in cross-cutting bodies where gases from below could more easily gain access.

The description of a few typical localities will show that the contact phenomena may vary greatly in different places. At times the general

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[1] J. S. Flett, *Mem.*, Geol. Survey England, Explan. Sheet 347, Bodmin and St. Austell, 1909, p. 58.

[2] R. J. Colony, *Jour. Geol.*, 31, 1923, pp. 169–178.

metamorphism is very weak while the metasomatic phases are strong and vice versa. Undoubtedly emanations along fissures continued after the general consolidation in many places, and if they had a high temperature the later alteration would be of the same type as that produced during the intrusive act.

Influenced by his experience in Mexico, where contact-metamorphic ores seem to grade into high-temperature veins, J. E. Spurr[1] has sought to minimize the importance of pyrometasomatic ore deposits, holding that they are generally the result of ascending emanations (ore magmas) rising along fissures from very deep-seated places of differentiation. This view is assuredly wrong. The phenomena of contact metamorphism are well established, and the metasomatic rocks having notable addition of substance cannot be separated from those of pure metamorphism. The development of hornfels, the recrystallization of limestone, the introduction of silica, the formation of lime-iron silicates, the growth of the iron oxides, and the sulphide phase, all form a continuous series and are caused by one and the same agency, *i.e.*, transfer of heat and substance from the intrusive.

The best proof that the mineralization came from the intrusive mass itself is the great tendency of the ores to form in salients and inclusions of the limestone.

In an admirable study of contact metamorphism at Marysville, Montana, Barrell[2] describes the metamorphic zone, ½ to 1 mile wide surrounding a small stock of quartz monzonite which widens in depth. The intensity of the metamorphism is thus increased.

Barrell distinguishes between (1) contact metamorphism, which results in recrystallization of the sediments to hornfels, marble and lime-silicate rock; and this was produced by the first wave of metamorphism; and (2) contact metasomatism in which magmatic emanations added some constituents to the altered rocks. This latter zone is at most 1,000 feet wide. Silica, iron, and sulphur were added. Diopside and hornblende, with a little apatite, tourmaline, garnet, and pyrite were the minerals formed. There are no contact-metamorphic deposits of economic importance at this place.

A very different state of affairs is described by Calkins[3] from Philipsburg, Montana, where a larger batholith invades Algonkian sediments of all kinds. The contact zones are half a mile or more in width. The author draws no sharp line between metamorphism and metasomatism, and evidently considers both to progress at the same time. The metamorphism is strong; close to the contact are masses of magnetite, in part carrying gold, with garnet, vesuvianite, humite, and forsterite. Scapolite

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[1] Ore magmas, New York, 1922, pp. 99, 588.

[2] Joseph Barrell, *Prof. Paper* 57, U. S. Geol. Survey, 1907.

[3] W. H. Emmons and F. C. Calkins, *Prof. Paper* 78, *idem*, 1913.

and tourmaline are distributed in the sediments for a distance of a mile or more from the contacts showing a widespread diffusion of magmatic chlorine and boron. Fluorine has also been introduced, likewise sodium, silica, and iron, the last only close to the contact.

A monograph by V. M. Goldschmidt[1] describes in great detail the type locality of the Norwegian contact-metamorphic deposits near Kristiania. They are of little economic importance. On a basement of Archean rocks rest Paleozoic sediments; these are broken by laccoliths of gradually more acidic composition, beginning with essexite, which is followed by syenitic rocks. Near the contacts of the essexite the metamorphism is exceedingly strong but takes place without addition of substance.

Along the syenite contacts Goldschmidt observed both an older normal contact metamorphism by recrystallization without addition and a younger "pneumatolytic" metamorphism by recrystallization under the influence of magmatic gases, resulting in "skarn rocks,"[2] in which the copper deposits are contained. The characteristic minerals of the hornfels or altered slates were formed before the consolidation of the magma and probably without the aid of other water than that normally contained in the rock. Though the "skarn rocks" and their metallic sulphides are later than the general metamorphism they were formed shortly before the crystallization of the magma, though the immediate contact may have been congealed. Pyroxene occurs in the inner and amphibole in the outer contact zone; according to Becke, the transition point between the stability fields of the two minerals is about 550° C. at 200 atmospheres.

The "skarn rocks" are coarsely crystalline and consist of andradite, hedenbergite, wollastonite, scapolite, axinite, adularia, albite, calcite, fluorite, zeolites, specularite, magnetite, bismuthinite, galena, chalcopyrite, primary chalcocite, primary willemite, sphalerite, pyrrhotite, molybdenite, and bornite. Many of these also occur as primary minerals in miarolitic cavities in the syenite. Magnetite forms nearer to the igneous rock than specularite. The scapolite becomes unstable at lower temperatures and is transformed to albite, epidote, microperthite and zeolites. The metallic ores are somewhat later than the skarn minerals.

From Ely, Nevada, Spencer[3] describes contact zones along small stocks of quartz monzonite. These zones extend a few hundred feet to half a mile from the contact. The monzonite porphyry is greatly altered with development of sericite, biotite, and pyrite but no garnet or diopside. The process involved loss of sodium, calcium, and some

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[1] Die Contactmetamorphose im Kristianiagebiet, Kristiania, 1911.

[2] An old Swedish mining term signifying the garnet-pyroxene-epidote rocks accompanying many Scandinavian magnetite deposits.

[3] A. C. Spencer, *Prof. Paper* 96, U. S. Geol. Survey, 1917.

alumina; gain of potassium and sulphur. The shales are converted to strongly pyritic hornfels involving addition of sulphur and iron. The limestones have developed white and brown mica, tremolite, pyroxene, garnet, epidote, scapolite, pyrite, pyrrhotite, chalcopyrite, galena, sphalerite, molybdenite, magnetite and hematite. In places very heavy masses of jasperoid have been formed from limestone. In general, silicon, sulphur, potassium, iron, copper, and other metals have been added.

Spencer again draws no sharp line between metamorphism and metasomatism, conceiving that the whole process took place by emanations from deeper parts of the magma after the parts now at the surface had consolidated. The areas exposed he considers too small to have effected the changes shown. It seems probable that in this case the alteration of the porphyry and the silicification of the limestone resulted from a somewhat later mineralization at lower temperature though practically continuous with the earlier phase.

Different conditions exist at White Knob, Idaho, described by Umpleby.[1] Here the Carboniferous limestone is very little altered in contact with fresh granite porphyry. Marble has developed close to the contact and more extensively in a roof segment of the batholith with scattered crystals of wollastonite, tremolite and diopside. Engulfed blocks of limestone are, however, more extensively garnetized with development of considerable ore-shoots (Fig. 284). The process shows great additions of iron, alumina and silica. To some extent the granite porphyry is also garnetized so that in places the contacts are indistinct, and it contains also diopside which replaces biotite and hornblende. The feldspars are the last minerals of the intrusive to be affected. Two stages of metamorphism are recognized: (1) Contact metamorphism at the time of intrusion; (2) contact metasomatism after the consolidation of the magma. The garnet and ore were developed during the last stage. That the last process took place after the intrusive had solidified is proved by garnetization following joint planes in the rock.

In the San Francisco district, Utah,[2] Butler finds silicate zones one-fourth mile wide and recrystallization of limestone over much wider areas. The contacts of garnetized limestone and fresh monzonite are often exceedingly sharp, the transition zone being less than 1 inch in width.

All this shows that the process varies considerably in different magmas, and that there are considerable differences of opinion as to the various stages of the process.

**Succession of Minerals.**—The introduction of sulphides and other metallic ores into the limestone is too obvious to be disputed. Many

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[1] J. B. Umpleby, *Prof. Paper* 97, U. S. Geol. Survey, 1917.

[2] B. S. Butler, *Prof. Paper* 78, *idem*, 1913.

observers have noted a certain succession of developments. Generally, the silicates and the magnetite are earlier than the sulphides, but the periods of deposition overlap.

In the deposits at White Horse, Northwest Territory, Stutzer[1] reports the succession: wollastonite, pyroxene, magnetite, garnet, calcite, and sulphides; at Berggiesshübel, Saxony, pyroxene, garnet, pyritic ores, sphalerite, arsenopyrite. At the Holgol Mine, Korea, Koto[2] found the succession to be ilvaite,[3] diopside, garnet, while the sulphides fill the interstices between the diopsides (Fig. 290). Frequently, chalcopyrite replacing calcite cements the garnet and other silicates.

Other authors admit that the sulphides are in part later but point out that to a considerable extent they have crystallized together with the silicates. This seems to be the condition in the White Knob, San Francisco, Ely, Clifton,[4] and Camp Hedley[5] districts. At the Imperial mine, Utah, Butler finds magnetite in part later than chalcopyrite.

Takeo Kato[6] found that at the Okufo mine, Japan, the succession began by wollastonite, which is replaced by andradite. The sulphides are deposited contemporaneously with the andradite or at the very close of its deposition.

C. H. Clapp[7] describes contact-metamorphic deposits on Vancouver Island, in which the order is diopside, epidote, garnet, magnetite, pyrrhotite, pyrite and chalcopyrite, but the silicates continued to form after some metallization had taken place.

The succession of the sulphides generally begins with arsenopyrite and pyrite; then follow pyrrhotite, sphalerite, chalcopyrite, galena; the rarer sulphosalts were formed last.

The early appearance of pyrite is explained by Spencer[8] by the accession of additional iron in the solutions from the sediments and igneous rock which would bring about precipitation of iron compounds before those of copper. He regards the magmatic solutions as containing much $SiO_2$, $H_2S$, KSH, $CO_2$, $HCO_3$, F, Fe, and Cu. The relative abundance of metals in the deposits at Ely is Fe, Cu, Pb, Zn, Mo and in the main that is also the sequence of deposition. This leads to the suggestion that the principal control in determining the order of deposition of the sulphides was the relative concentration of metal radicals in the mineralizing solutions.

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[1] *Zeitschr. prakt. Geol.*, 17, 1909, pp. 116–120.

[2] *Jour. Coll. Sci.*, Imp. Univ. Tokyo, May 28, 1910.

[3] Hedenbergite according to D. F. Higgins, *Econ. Geol.*, 13, 1918, p. 19.

[4] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905.

[5] C. Camsell, *Mem.* 2, Canada Geol. Survey, 1910.

[6] *Jour.* Geol. Soc., Tokyo, 20, 1913, pp. 13–32.

[7] *Mem.* 13, Canada Geol. Survey, 1912, p. 158.

[8] A. C. Spencer, *Prof. Paper* 96, U. S. Geol. Survey, 1917, pp. 64–72.

Detailed studies supported by many analyses have been made of the contact-metamorphic changes in limestone at Bingham, Utah.[1] The metasomatic processes were followed in detail, showing introduction of garnet, epidote, wollastonite, tremolite, and diopside with pyrite, magnetite, siderite, and some chalcopyrite. Much Fe, $SiO_2$, Mg, and some Na was added, while carbon, $CO_2$, and CaO were carried away. There was no important change in volume.

**Volume Relations.**—The replacement of limestone by sulphides, oxides, and silicates liberated a large volume of carbon dioxide; a portion may have been resorbed in the magma, but a larger part was doubtless dissipated in the fractures surrounding the intrusive mass and gradually escaped or mingled with the escaping magmatic water and some distance away with the ground-waters, thus adding to the load of ascending thermal springs.

In the Morenci district, Arizona, the clearest evidence is given by the transformation effected along dikes of unaltered quartz monzonite porphyry, which come from near the apex of an intrusive stock and which seem to have carried unusual quantities of magmatic emanations. One of these, 20 to 50 feet wide, crosses the successive Paleozoic formations with no evidence of fractures that could have admitted solutions from the depths after the consolidation. In the lower limestones the contact zones are only a few feet wide, consisting of epidote next to the intrusive rock, followed by garnet, which adjoins the unaltered limestone. The addition of iron and silica to this narrow zone, which shows no evidence whatever of contraction of volume, is so clear that it hardly admits of discussion. Farther up, the same dike cuts across a pure limestone about 80 feet in thickness. This has been changed to massive andradite garnet, with some epidote, for a distance of about 100 feet from the dike. Stains of malachite are present, but this particular rock is poor in copper (Fig. 285).

If all of the lime has been used in the garnetization and only $CO_2$ has escaped, the volume of the rock would have increased about one-half. If, on the other hand, as seems probable, the volume has remained approximately constant, then 460 kilograms CaO and 1,190 kilograms $CO_2$ per cubic meter have been carried away, while 1,330 kilograms $SiO_2$ and 1,180 kilograms $Fe_2O_3$ have been added. These are astonishing figures and give an idea of the vigorous transfer of material which took place during metamorphism. The fresh limestone contains 94 to 96 per cent $CaCO_3$, less than 1 per cent $MgCO_3$, 1 per cent $SiO_2$, and 1 to 3 per cent $Al_2O_3$ and $Fe_2O_3$. The andradite garnet contains 1.53 per cent

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[1] W. Lindgren, Contact metamorphism at Bingham, *Bull.* Geol. Soc. Am., 35, 1924, pp. 507–514.

A. N. Winchell, Petrographic studies of limestone alterations at Bingham, *Trans.*, Am. Inst. Min. Met. Eng., 70, 1924, pp. 884–899.

$Al_2O_3$, 31.41 $Fe_2O_3$, 42.63 $SiO_2$, and 23.37 CaO. The transfer has been mutual, for at some places the intrusive rock next to the contact has been strongly epidotized by lime derived from the calcareous rock.

That the volume has remained practically constant even in the most intense metasomatism may be considered proved and confirmed by the observations in the Ely, San Francisco and White Knob districts referred to above. Preservation of structure, like stratification planes, joints and fossils, has repeatedly been observed in the silicate rock and in the sulphides.[1]

C. K. Leith and E. C. Harder[2] have attempted to account for the silicate rock at the contact of the Iron Springs intrusive in southern Utah by assuming that the silica and alumina have remained constant while the volume of the rock has been decreased as much as 60 or 80 per cent by reason of abstraction of lime. No definite field evidence, however, was found in favor of this view.

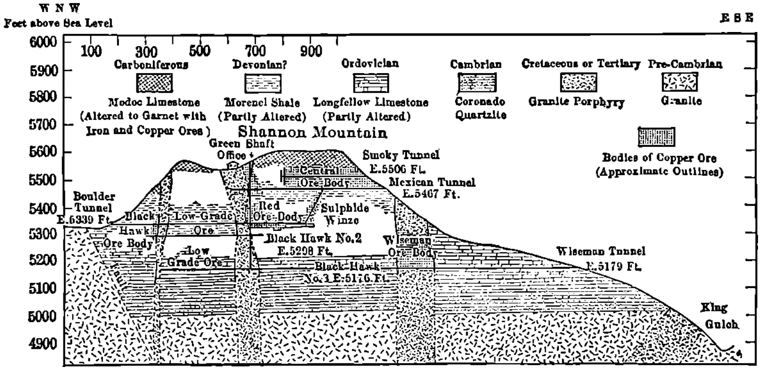

FIG. 285.—Vertical section showing flat ore-bodies at Shannon mine, Clifton, Arizona Ore-body in porphyry dike consists of secondary chalcocite

On the whole, it may be asserted that contact metamorphism takes place without change in bulk volume, though in places some vuggy or drusy structures may appear.

**Mode of Transfer.**—During the alteration of the carbonate rock much of the carbon dioxide and lime was carried away. In various proportions this has been compensated by additions from magmatic emanations of silica, iron, alumina, sodium and a number of other useful metals. Mineralizers like sulphur, chlorine, boron, fluorine, and arsenic have also been introduced. Many writers, like Leith and Harder, Calkins and Goldschmidt, believe that the heavy metals were introduced as

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[1] Adolph Knopf, *Bull.* 580, U. S. Geol. Survey, 1915, pp. 1–18.
W. Lindgren, *Prof. Paper* 68, *idem*, 1910, p. 294.

[2] *Bull.* 388, U. S. Geol. Survey, 1908. See also review by Kemp, *Econ. Geol.*, 4, 1909, p. 782, and answer by Leith, *idem*, 5, 1910, p. 188.

chlorides, fluorides, etc.; and this view is very likely correct. The mode of transfer of silica from the intrusive to the limestone is a problem not yet fully solved.

The equations roughly representing the transformations would be as follows:

$$\begin{aligned} 2FeCl_3 + 3H_2O &= Fe_2O_3 + 6HCl \\ 2FeF_3 + 3CaCO_3 &= Fe_2O_3 + 3CaF_2 + 3CO_2 \\ 2FeCl_3 + 3CaCO_3 &= Fe_2O_3 + 3CaCl_2 + 3CO_2 \\ CaCO_3 + SiO_2 &= CaSiO_3 + CO_2 \end{aligned}$$

H. von Eckerman[1] has contributed valuable information regarding the progress of silicatization in large limestone masses included in granite magma, and has also pointed out that $CO_2$ may exert an oxidizing action on iron, resulting in $Fe_2O_3$ being formed from FeO of the igneous rock. On the other hand the CO thus generated would exert a reducing influence.

**Physical Conditions at the Contact.**—The temperature at the contact, according to the composition of the magma, may have been as much as 1500° C.; the siliceous rocks consolidating at a temperature of 500° to 1100° C. Where there is no chemical action involved $CaCO_3$ begins to lose carbon dioxide at 550° C., but the reaction would probably begin at a lower temperature for $MgCO_3$. Magmatic gases acting chemically on the calcite might still further reduce the temperature necessary for silicatization. Under abyssal conditions no opportunity would be afforded for the liberated carbon dioxide to escape through the fissured rocks.

Even where the carbon dioxide can not escape there may be intense action between the igneous rock and the limestone. The two rocks will form a chemical system with great difference of temperature, and it may be assumed that there will be intense transfer of material between the two. Possibly lime and carbon dioxide will be absorbed by the magma in exchange for metallic constituents exhaled from the igneous rock. At great depths the action will be sustained over a longer period and the results may be different from those obtained within the cooler and more brittle upper zone. In its earlier stages the process is surely pneumatolytic, that is, effected by magmatic gases.

A gradual lowering of the temperature from the maximum obtaining at the contact will take place (1) during the diffusion of the gases outward into cooler sediments; (2) during the gradual cooling of the intrusive itself. The presence of wollastonite indicates that the temperature at this particular place could not have exceeded 1300° C., above which point this mineral is unstable. It may form at 825° C. but not at 660° C.[2]

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[1] H. von Eckermann, *Geol. För. Förh.*, 45, 1923, pp. 465–537.

[2] H. Ehrenberg, *Zeitschr. phys. Chemie*, 14, 1931, pp. 421–434.

Lindgren and Whitehead[1] attempted to determine the temperature by the solubility curve of sodium chloride, which salt is present as sharp cubes in fluid inclusions in quartz in a contact-metamorphic deposit at Zimapan, Mexico. They concluded that the temperature of formation was about 400° to 500° C.

Wright and Larsen[2] have shown that the quartz in contact-metamorphic deposits was formed below 575° C. Spencer[3] examined fluid inclusions in jasperoid at Ely, Nevada, and by calculations based on relative volume of gas and liquid found probable temperatures of 200° to 350° C. The jasperoids were, however, probably formed at lower temperatures than the contact-metamorphic silicates.

The garnet in pyrometasomatic deposits commonly shows double refraction. H. E. Merwin[4] found that such birefringent garnet became isotropic upon heating to 800° C. and hence concluded that the temperature of formation must have been below the figure indicated.

**Depth of Formation.**—In many cases it is difficult to ascertain the depth below the surface at which contact-metamorphic deposits were formed. In the Cordilleran region of America, which contains the most abundant and characteristic examples of this type, the conditions of sedimentation and intrusion were such that approximately correct measurements are feasible. Brögger, many years ago, pointed out that granular intrusive rocks by no means always crystallized at abyssal depths and that some intrusions in the Kristiania region had congealed less than 1,000 meters below the surface.

In the central Cordilleran region sedimentation was almost continuous from the Cambrian to the late Cretaceous; the intrusive rocks now exposed were injected into these sediments during Cretaceous or early Tertiary time, well up in the zone of fracture and far above normal "anamorphic" conditions. In 1907, Barrell[5] showed that at Marysville, Montana, the batholith reached within 4,000 feet of the surface, and Leith and Harder gave the same figure for the iron deposit at Iron Springs, Utah. In New Mexico[6] similar conditions existed. At the close of the Cretaceous period practically the whole state was covered by a mantle of sedimentary rocks from 6,000 to 9,000 feet thick. The Cretaceous part of this section, into which most of the numerous intrusive masses reached, was between 3,000 and 5,000 feet thick; much of it consisted of tough but

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[1] *Econ. Geol.*, 9, 1914, pp. 435–462; see also N. L. Bowen, Geologic thermometry, in Fairbanks, Laboratory investigation of ores, New York, 1928, pp. 172–199.

[2] *Am. Jour. Sci.*, 4th ser., 27, 1909, pp. 421–447.

[3] *Prof. Paper* 96, U. S. Geol. Survey, 1917, p. 63.

[4] In *Prof. Paper* 87, U. S. Geol. Survey, 1915, p. 108.

[5] *Prof. Paper* 57, U. S. Geol. Survey, 1907.

[6] W. Lindgren, L. C. Graton, and C. H. Gordon, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 41.

pliable shales not easily broken through by the intrusions. At many places contact-metamorphic deposits were formed less than 3,000 feet below the surface.

The intrusive "trap" sheets of Triassic age in Connecticut, New Jersey, and Pennsylvania have exerted some contact-metamorphic action and produced small copper deposits; at Cornwall, Pennsylvania,[1] important magnetite deposits were formed in calcareous Carboniferous rocks. The depth below the surface was probably less than 1,000 feet.

For contact-metamorphic deposits in pre-Cambrian areas and in general where periods of dynamic metamorphism have intervened exact data of this kind can rarely be obtained. Some deposits of this class were formed at great depth and under distinctly abyssal conditions.

**Distribution.**—In the northern part of the Cordilleran region the deposits seem to be less abundant, though several representatives may be found in British Columbia, Alaska, Montana, Idaho, Nevada, and Utah. There are relatively few of them in Colorado, but they occur in much greater abundance in New Mexico, Arizona, and Mexico.

A reconnaissance of the metal deposits in New Mexico permits a good review of the frequency and relationships of these ores. Along a belt extending from the northern boundary down to the southwestern part of the state the Paleozoic and Mesozoic strata are intruded by at least 20 stocks of early Tertiary quartz monzonite or monzonite, usually of moderate size. The major part of the commercial mineral deposits cluster around these intrusions. Contact-metamorphic deposits were found in 14 districts and at least 25 mines have been worked on a commercial scale. At San Pedro and Jarilla, primary chalcopyrite ores are smelted; at Magdalena, the deposits yield zinc, copper, and lead; at Hanover, magnetite, sphalerite, and chalcopyrite. In the minor deposits the ores may simply form irregular masses at the contact, rarely extending more than 200 feet away from it.

In southern Arizona the deposits are equally common. Among them are the copper deposits of Clifton, Bisbee, Saddle Mountain, Twin Buttes, Washington, Silver Bell, Ajo, Casa Grande, and Vekol. At Clifton and Bisbee the ores have been greatly enriched by oxidation; at Saddle Mountain, Twin Buttes, Washington, and Silver Bell primary chalcopyrite ores are worked.

In eastern Mexico cupriferous contact-metamorphic deposits are common where monzonites break through Mesozoic limestones, as first mentioned by Ordoñez and Aguilera. Since then detailed descriptions have been given of the districts of Santa Fe, in Chiapas; Velardeña, in Durango; San Jose, in Tamaulipas; Concepcion del Oro, in Zacatecas; Dolores, in San Luis Potosi; and Cananea, in Sonora.

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[1] A. C. Spencer, *Bull.* 359, U. S. Geol. Survey, 1908.

## PRINCIPAL TYPES OF CONTACT-METAMORPHIC DEPOSITS

The contact-metamorphic deposits may be classified as follows:

1. Magnetite deposits.
2. Chalcopyrite deposits. Principal ore minerals are chalcopyrite, pyrite, pyrrhotite, sphalerite, molybdenite, magnetite, and specularite.
3. Galena and sphalerite deposits.
4. Arsenopyrite-gold deposits. Principal minerals are arsenopyrite and pyrrhotite.
5. Gold deposits.
6. Cassiterite deposits.
7. Scheelite deposits.
8. Graphite deposits.

The chalcopyrite and magnetite deposits present the most common type; the ores containing galena, arsenopyrite, gold, or cassiterite are distinctly rare.

### Magnetite Deposits

**General Character.**—The magnetite deposits of this class are of common occurrence, though rarely very large. Associated with the magnetite is more or less specularite, almost always a little pyrite and chalcopyrite, and the ordinary contact silicates including andradite, ilvaite, olivine, and hedenbergite—all four rich in iron (Fig. 286). The magnetite is sometimes crystallized and often developed in coarsely granular masses (Fig. 288). The ore-bodies are of irregular form, unless, as often happens, they follow the stratification for some distance.

**Foreign Occurrences.**—Among the European deposits those of Berggiesshübel, in Saxony; Schmiedeberg, in Silesia; and Gora Magnitnaja and Wyssokaja Gora, in the Ural Mountains, are usually described in the textbooks. Regarding the latter two occurrences, the opinions seem to be somewhat divided.

The classical deposits of the Banat province, in southeastern Hungary, first described by von Cotta, deserve special mention. In this region early Tertiary intrusive rocks, designated as diorite, syenite, and their porphyries, break through Mesozoic limestones. Along the contacts the limestones become coarsely crystalline, and the usual metamorphic silicates,[1] together with irregular masses of magnetite and some sulphides, develop in them. A banded structure, sometimes apparent, is caused by alternating layers of garnet and magnetite of contemporaneous origin. Masses of garnet from seventy to several hundred feet thick occur. The best-known mines of this region are at Moravicza, Dognacska, and Oravicza. The present annual production is small. According to

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[1] A ferromagnesian borate, ludwigite, is recorded from Moravicza.

Bergeat, there can be no doubt that the ores are of contact-metamorphic origin.

The celebrated mineral deposits of Elba,[1] Italy, with their beautifully crystallized hematite, are likewise of contact-metamorphic origin and were formed under the influence of post-Eocene granite.

**Fierro, New Mexico.**—Many magnetite deposits of this kind are known in the United States, particularly in the Western States, but most of them are comparatively small. A deposit at Fierro,[2] in southwestern New Mexico, is actively worked at present, the ore being shipped to Pueblo, Colorado. The ore occurs at the contact of quartz monzonite porphyry, probably of early Tertiary age, with Paleozoic limestone; it outcrops in bold masses and is mined in open cuts. The ore-bodies are mainly irregular, lenticular masses of magnetite with a little chalcopyrite; in part they are pure magnetite containing from 60 to 70 per cent of iron. Those parts which contain a notable quantity of chalcopyrite are left as pillars. Small bunches of garnet and epidote are present in the ore, and in places there are horses of more or less metamorphosed limestones; the phosphorus is rarely above 0.07 per cent; the sulphur averages 0.02 per cent. Similar deposits, richer in copper, have been mined for flux, the ores being used in the copper furnaces at Douglas, Arizona.

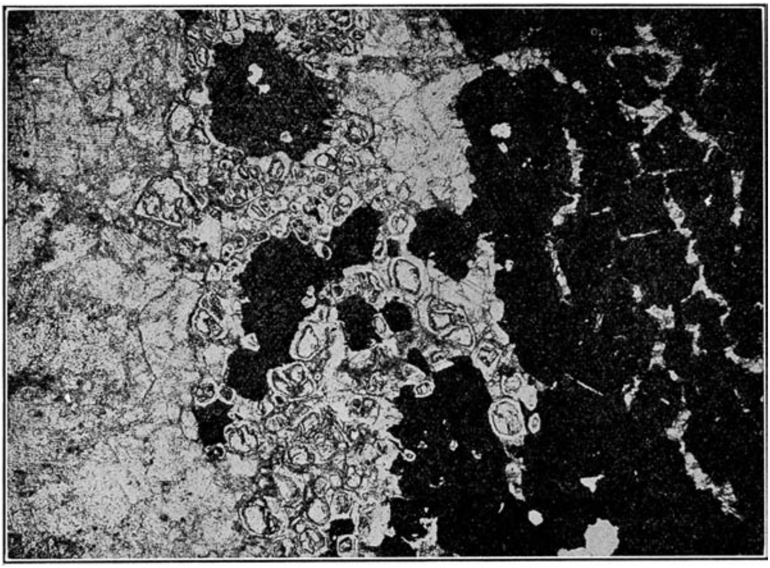

FIG. 286.—Thin section showing magnetite replacing limestone in contact-metamorphic zone, Philipsburg, Montana. Intermediate zone rich in olivine. (*After F. C. Calkins, U. S. Geol. Survey.*)

[1] B. Lotti, *Zeitschr. prakt. Geol.*, 1905, p. 141.

[2] L. C. Graton, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 313.
Sidney Paige, *Bull.* 380, *idem*, 1909, pp. 199–214; Folio 199, *idem*, 1916.

**Heroult, California.**—Another deposit is situated in Shasta County, California.[1] The ore was formerly smelted locally at Heroult, in an electric furnace. The ore-bodies are found mainly at the contacts of diorite and Triassic limestone, and also, to a minor extent, at the contacts of the same diorite with Permian shale and with granodiorite. Magnetite, garnet and hedenbergite, ilvaite and quartz, pyrite and chalcopyrite are the minerals present. The limestone is practically pure, and that material has been transferred from the intrusive seems to be the unavoidable conclusion. The ore is low in phosphorus and sulphur.

**Iron Springs, Utah.**—The important deposits of Iron Springs, in southern Utah, have been described by C. K. Leith and E. C. Harder.[2]

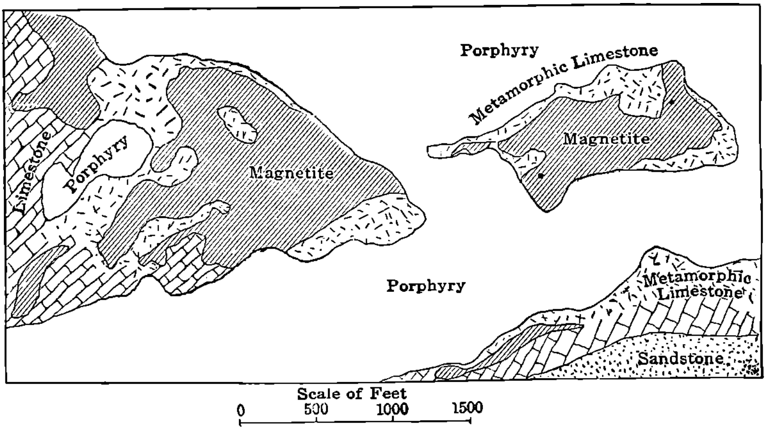

FIG. 287.—Plan showing magnetite and limestone in projecting point of limestone, Iron Mountain, Iron Springs, Utah. (*After Leith and Harder.*)

Ores from this district, mined in open cuts, and averaging about 53 per cent iron, are now smelted in blast furnaces at Provo in the same state. A laccolith of quartz syenite porphyry (andesite according to the nomenclature of the authors) breaks through sediments of Carboniferous, Cretaceous, and Tertiary age (Fig. 287). The magnetite appears in fissure deposits and replacements along the contact with the Carboniferous limestone. Quartz, garnet, diopside, apatite, and hornblende are minor constituents of the ore. According to the authors only a part of the ore is associated with contact metamorphism, for the probably gaseous emanations of iron compounds continued after the consolidation, and the resulting magnetite, sometimes associated with apatite, garnet, etc., filled contraction fissures in the intrusive and replaced the limestone near the contact.

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[1] Basil Prescott, *Econ. Geol.*, 3, 1908, pp. 465–480.

[2] *Bull.* 338, U. S. Geol. Survey, 1908.

D. McVichie, *Trans.*, Am. Inst. Min. Met. Eng., 74, 1926, pp. 163–173.

**Cornwall, Pennsylvania.**[1]—In Pennsylvania and also in Virginia intrusions of Triassic diabases come in contact with Paleozoic calcareous shales; at many of these places deposits of magnetite have been formed.

The principal deposit is at Cornwall, and has been extensively worked as an open cut and later by underground workings. About 31,000,000 tons have been mined and concentrated. Chalcopyrite is also recovered. Thick sills of diabase are in contact with Triassic shales and with a 90-foot thick bed of Paleozoic calcareous shale. The former have been indurated at the contact but contain no deposits. The calcareous shale has been extensively contact-metamorphosed with the development of tremolite and magnetite (Fig. 288) and a small amount of specularite, garnet, epidote, and diopside. The order of crystallization is tremolite, magnetite, pyrite, and chalcopyrite. The ores continue in places for several hundred feet from the contact.

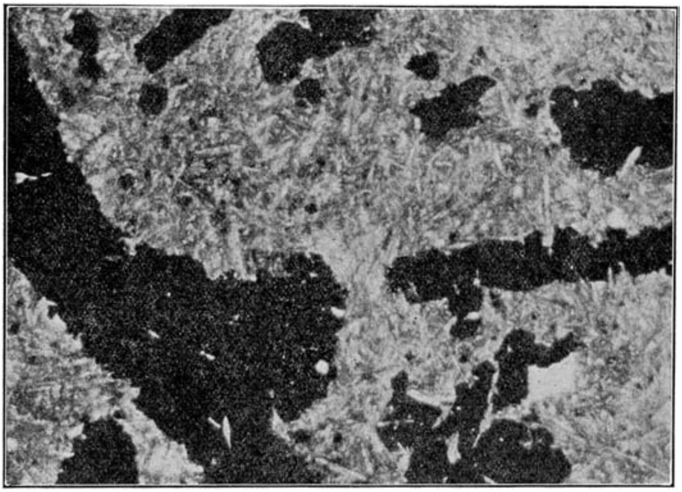

FIG. 288.—Thin section showing magnetite (black) replacing tremolite, Cornwall, Pennsylvania. Magnified 35 diameters. (*After W. H. Callahan.*)

All those who have examined the deposit agree that the iron and copper were derived from the solidifying diabase magma, and not from any other direction. Callahan shows that specularite, magnetite, pyrite, chalcopyrite, apophyllite and albite occur on the joints in the diabase. The evidence adduced by Shannon from Virginia is very important and conclusive.

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[1] A. C. Spencer, Magnetite deposits of the Cornwall type in Pennsylvania, *Bull.* 359, U. S. Geol. Survey, 1908, pp. 74–76; *Bull.* 430, *idem*, 1910, pp. 247–249.

E. C. Harder, Structure and origin of the magnetite deposits near Dillsburg, Pennsylvania, *Econ. Geol.*, 5, 1910, pp. 599–622.

E. V. Shannon, *Proc.*, U. S. Nat. Mus. 66, 1924, pp. 1–86.

W. H. Callahan and W. H. Newhouse, *Econ. Geol.*, 24, 1929, pp. 403–411.

## Chalcopyrite Deposits

**General Character.**—The contact-metamorphic deposits that carry chalcopyrite as the predominating ore mineral are not abundantly represented in Europe, but are the most common type in North America, particularly in New Mexico, Arizona, and Mexico, and are sparingly present in the Cordilleran region of South America. Similar deposits have been found in Australia, Japan, and Korea. They occur, as a rule, at the contacts of smaller intrusive masses of monzonite or quartz monzonite against limestone. Their form is irregular or tabular. The tabular deposits follow certain beds in the limestone formations, and their hanging and footwalls may consist of little-altered or unaltered limestone. The structure of the ore is massive and coarse granular (Figs. 289 and 290). The ore minerals consist of chalcopyrite, bornite, pyrite, more rarely pyrrhotite and sphalerite, often also molybdenite and other sulphides; galena is rare. The ore frequently contains much magnetite and specularite. The gangue minerals are andradite, grossularite, epidote, diopside, tremolite, vesuvianite, wollastonite, ilvaite, and calcite. The deposits are poor in gold and silver and are frequently enriched in copper by oxidation, but in many occurrences the primary ore is rich enough to be utilized.

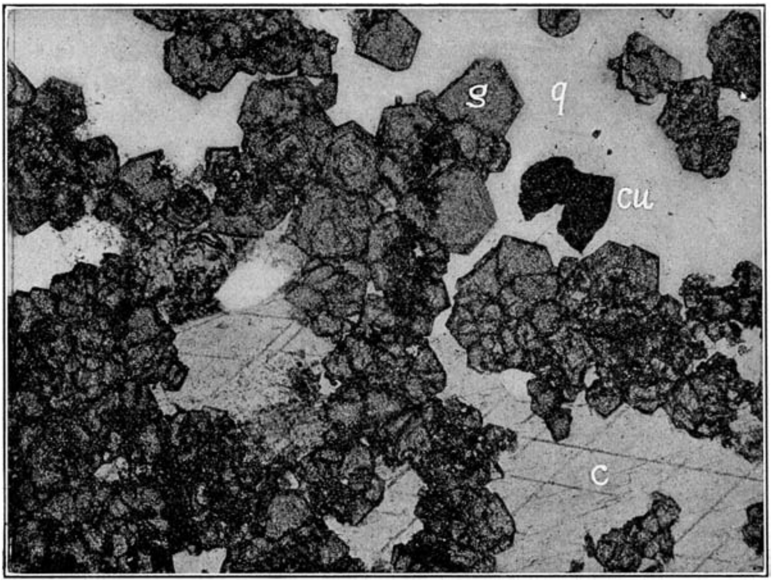

FIG. 289.—Thin section showing contact-metamorphic ore, Clifton, Arizona. *c*, calcite; *g*, garnet; *q*, quartz, *cu*, chalcopyrite. Magnified 15 diameters.

**New Mexico.**[1]—In the fourteen districts of New Mexico distinguished by contact-metamorphic deposits, the copper ores are by far the most common.

The most important of these deposits, economically, is that of the San Pedro mine, in the laccolithic mountain group of the same name. Beds of Upper Carboniferous rocks, over 700 feet thick, have been metamorphosed by the underlying laccolith of granodiorite porphyry and by dikes extending upward from it. The lower 200 feet of shaly limestone is only partly altered, with local development of garnet and tremolite and a little chalcopyrite and pyrrhotite, but along a certain bed of purer limestone garnetization has taken place for half a mile, the thickness of this strongly metamorphosed stratum being about 50 feet. Bunches of chalcopyrite are irregularly distributed in it. Within this zone beds of pure crystalline limestone adjoin wholly garnetized beds. In places the rock consists of a mixture of garnet and coarsely crystalline limestone. On the dip, the gently inclined ore beds have been followed for 300 feet. The upper beds of the series consist mainly of somewhat metamorphosed and baked shale and sandstone.

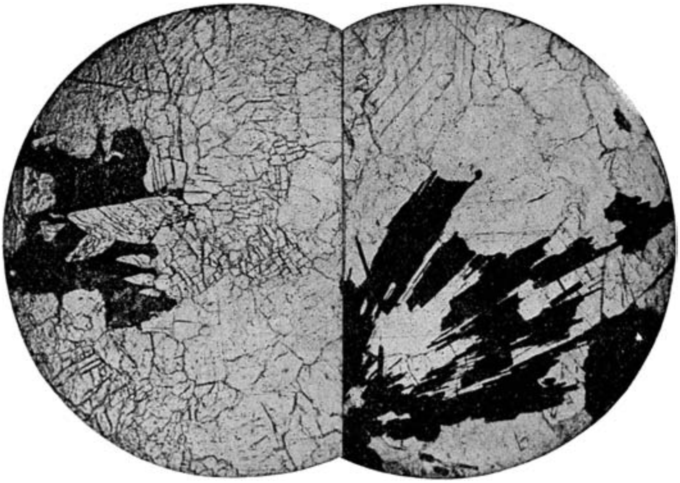

FIG. 290.—Thin sections showing contact-metamorphic ores at Holgol, Korea. *A*, chalcopyrite (black) in diopside. *B*, radiating crystals of ilvaite (black) in granular limestone. Ilvaite incloses crystals of diopside. Magnified 30 diameters. (*After B. Koto.*)

**Clifton, Arizona.**—In Arizona almost all the contact-metamorphic deposits yield copper as the principal metal. Near Clifton[2] a stock of

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[1] Lindgren, Graton, and Gordon, *Prof. Paper* 68, U. S. Geol. Survey, 1910.

[2] W. Lindgren, *Prof. Paper* 43, *idem*, 1905.

L. E. Reber, Jr., The mineralization at Clifton-Morenci, *Econ. Geol.*, 11, 1916, pp. 528–573.

granite porphyry and quartz monzonite porphyry breaks across pre-Cambrian granite, a Paleozoic series about 1,000 feet thick, and Cretaceous sediments about 400 feet thick. The Paleozoic limestones and shales, as well as the Cretaceous sandstones, are contact-metamorphosed. The ore deposits lie mainly near Morenci and Metcalf; at both places the beds are cut by an unusual number of dikes, which have exerted a specially strong contact-metamorphic action on the sediments.

The ore deposits form a complicated series very similar to those observed at Cananea, Mexico, at Ely, Nevada, and at Bingham, Utah. The oldest ores are contact-metamorphosed limestones; these, as well as the adjoining porphyry, are cut by a series of pyritic veins, poor in copper, which in the sericitized porphyry spread out into disseminations of pyrite. Widespread oxidation has altered all the deposits and enriched them. Well-defined chalcocite zones (p. 832) have formed by replacement of the pyrite by descending cupric sulphate solutions, and the present importance of the district is due wholly to the exploitation of these chalcocite ores, which contain from 1 to 2 per cent of copper.

The primary contact-metamorphic deposits lie in limestone and form irregular bunches or tabular deposits parallel to dikes or following the stratification. Wherever the character is not masked by oxidation, these primary ores consist of garnet, epidote, diopside, calcite, chalcopyrite, pyrite, magnetite, and sphalerite, occasionally also molybdenite. In the earlier days of the district, from 1875 to 1900, these oxidized ore-bodies were mined; they were easily reduced and comparatively rich in copper, containing mainly malachite, azurite, and limonite. The rich Longfellow ore-body formed a funnel-shaped mass in Ordovician limestone, between two porphyry dikes.

Farther west, in the Manganese Blue and Detroit mines, were several tabular ore-bodies, following the stratification in the Ordovician, Devonian, and Carboniferous limestones; these also owed their richness to several porphyry dikes, a few hundred feet from the main contact. Along the main contact were many irregular bunches of oxidized contact-metamorphic ores. At Metcalf, the Shannon Mountain contained several similar ore-bodies (Fig. 285), lying in an isolated mass of Paleozoic limestones extensively cut by porphyry dikes.

**Bisbee, Arizona.**[1]—At Bisbee, Arizona, pre-Cambrian rocks are overlain by about 5,000 feet of Paleozoic limestones, which after deposition

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[1] F. L. Ransome, *Prof. Paper* 21, U. S. Geol. Survey, 1904.

W. L. Tovote, *Min. and Sci. Press*, Feb. 4, 1911.

Arthur Notman, The Copper Queen mine and works, *Trans.*, Inst. Min. and Met. (London), 22, 1913, pp. 550–562.

Y. S. Bonillas, J. B. Tenney, and L. Feuchère, Geology of the Warren mining district, *Trans.*, Am. Inst. Min. Eng., 55, 1917, pp. 284–355.

J. B. Tenney, The Bisbee mining district, *Eng. Min. Jour.-Press*, May 21, 1927.

were deformed by folding and faulting. After a period of erosion a transgression of the Cretaceous sea deposited a thick series of beds on the older series. The Paleozoic series is cut by intrusions of granitic porphyry of doubtful (Tertiary ?) age, which is intimately connected with the origin of the copper deposits (Fig. 291). The principal mass, of which the most prominent point is Sacramento Hill, close to Bisbee, is about 1 mile long and 1½ miles wide.

The copper deposits lie in the limestones, surrounding crescent-like the east side of the intrusive mass, and appear as irregular and ill-defined or rudely tabular masses, which on the whole follow the dip of the stratification. They are almost entirely oxidized, even down to depths of 1,400 feet below the surface. The oxidized ore, containing copper carbonates, cuprite, and sometimes also chalcocite, with limonite and some siderite, passes gradually on its peripheries into "ledge matter" or limonitic clays, which in turn grade into altered and contact-metamorphosed limestones. The contact metamorphism is unusually inconspicuous, but the limestone surrounding the intrusive mass contains fine-grained tremolite, diopside, garnet, vesuvianite, and quartz, associated with magnetite, pyrite, bornite, and a little chalcopyrite, galena and sphalerite. In the Oliver shaft, close to the contact, on the 1,200-foot level, the limestone is more intensely metamorphosed than elsewhere and contains bodies of pyrite, mixed with chalcopyrite and bornite.

The porphyry dips underneath the limestones, and the contact is thus found farther west in the mines than on the surface. The limestones are also cut by a considerable number of dikes. A heavy mantle of pyritic quartzose ore, with some chalcopyrite and chalcocite, surrounds the porphyry, pitching parallel to the contact underneath the limestone of the surface.

Lead ores occur towards the east of Sacramento Hill suggesting a zonal arrangement of metals.[1] Later work by the Calumet and Arizona Copper Company has developed extensive replacement ores towards the southwest, at the Campbell shaft, with pyrite, chalcopyrite, bornite, sphalerite, chert, and specularite. The last is said to be a border phase following the chalcopyrite. In the same direction are widespread chert replacements in the favorable limestones (Naco and Escabrosa). While the ores adjoining the main mass of porphyry and associated dikes are probably of pyrometasomatic origin, the metallization continued over a considerable period and takes the form of hypothermal or meso-

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G. M. Schwartz and C. F. Park, Jr., *Econ. Geol.*, 27, 1932, pp. 39–51 (Ores from Campbell mine).

Ore deposits of the Southwest, *Guidebook* 14, Sixteenth Intern. Geol. Congress, 1933: Papers by F. L. Ransome and J. B. Tenney.

[1] W. H. Emmons, Relations of metalliferous lode systems to igneous intrusives, *Trans.*, Am. Inst. Min. Eng., 74, 1926, pp. 29–70.

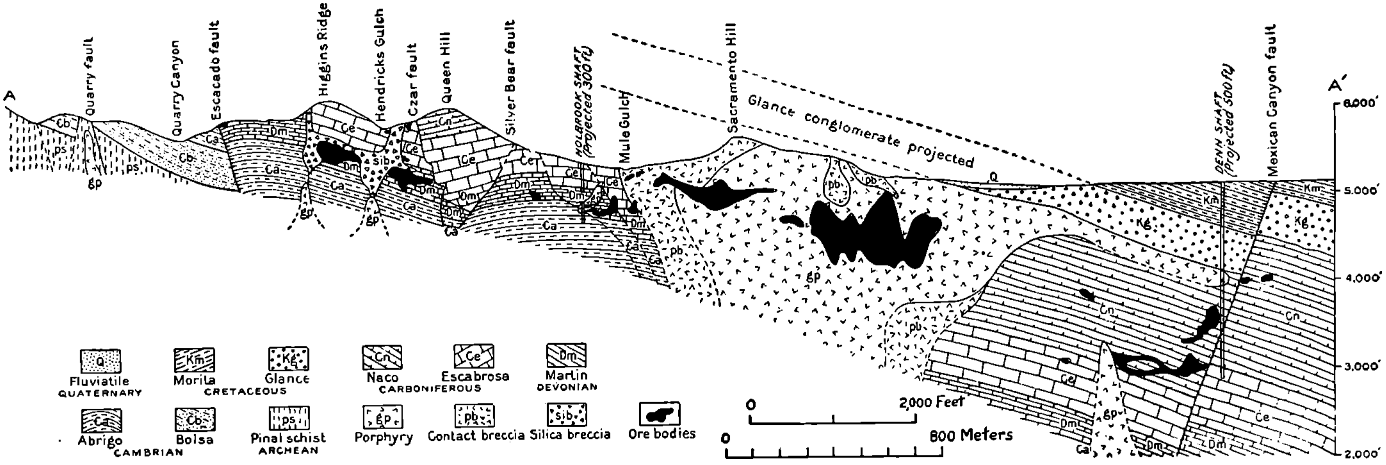

FIG. 291.—East-West section across Bisbee mining district. (*After J. B. Tenney.*)

thermal replacements in the areas removed from the center of the intrusions.

The deep oxidization of the ores certainly dates back to the Tertiary period and took place under conditions of water level different from those of to-day. The sericitized and chloritized porphyry, particularly that of Sacramento Hill, contains some large bodies of low-grade chalcocite ore, this mineral replacing pyrite.

The Bisbee (Warren) district produced in 1916 about 2,000,000 tons of 5 to 6 per cent ore yielding about 100,000 tons of copper with some lead, gold, and silver; the total metallic value was $52,000,000. The principal production is derived from the Copper Queen and Calumet & Arizona mines. In 1929, the output was 2,796,000 tons of ore yielding $1,410,400 in gold; 2,264,000 ounces of silver; 93,100 tons of copper; 1,020 tons of lead; valued in all at $35,505,000 or $11.90 per ton. The total value of the production amounts to about seven hundred million dollars.

**Silver Bell, Arizona.**—At Silver Bell,[1] northwest of Tucson, extensive primary chalcopyrite deposits have been worked. The mines are near the summit of one of the numerous desert ranges of that region; the ores were smelted, without concentration, at the Sasco plant. Several small masses of Paleozoic limestone are engulfed in a large mass of granite porphyry, and along their contacts metamorphism is irregularly developed —in part by marmorization, in part by garnetization. The ore consists of chalcopyrite and light-brown garnet, said to be andradite, with a little magnetite, sphalerite, galena, and molybdenite. Much of the ore averaged 7 per cent copper, but that smelted would average about 4 per cent. About 800 tons were mined per day in 1909, and some production is still maintained. A trace of gold and 1 to 2 ounces of silver per ton are present. The oxidation is shallow, wholly fresh rock being encountered at the 200-foot level. The porphyry is locally silicified, but otherwise not greatly altered, except for some disseminated pyrite and chalcopyrite. No extensive chalcocite zone has been found in the porphyry.

**Cananea, Mexico.**—The mines at Cananea lie a short distance south of the Arizona-Sonora boundary line, in one of the short ranges that rise out of the gently sloping desert plains. Since 1900, these deposits have yielded large quantities of copper from ores enriched by oxidation and development of secondary chalcocite. The district was described by S. F. Emmons.[2]

The deposits show some similarity to those of Clifton, Arizona, but the geological history is much more complicated. Three successive

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[1] C. F. Tolman, Jr., *Min. and Sci. Press*, Nov. 27, 1909.
C. A. Stewart, *Trans.*, Am. Inst. Min. Eng., 43, 1913, pp. 240–290.

[2] S. F. Emmons, *Econ. Geol.*, 4, 1909, pp. 312–356.
W. H. Emmons, *op. cit.*, p. 720.

irruptions, termed diorite porphyry, granodiorite, and quartz porphyry, have caused contact metamorphism in relatively small areas of Paleozoic limestone. Among the primary minerals are chalcopyrite, bornite, sphalerite, magnetite, and specularite; the limestones are garnetized, marmorized, and silicified.

A second epoch of mineralization by aqueous solutions resulted in veins and disseminations of pyrite, chalcopyrite, and quartz. Both classes have been enriched by oxidizing solutions resulting in valuable deposits of chalcocite.

**Bingham, Utah.**[1]—The Bingham mining district near Salt Lake City, Utah, is remarkable for its enormous ore production amounting to a value of $915,693,000 since 1865. The geological features comprise about 10,000 feet of Carboniferous quartzite with numerous lenses of limestone, all folded into an open syncline. In this has been injected an irregular, spreading mass of quartz monzonite porphyry occupying only about 3 square miles. The metallization took place in and near the intrusives and formed a great variety of ore deposits the development of which was spread over a time interval of gradual cooling following the intrusion. A low-grade "protore" in the intrusive mass has been enriched by supergene processes and now forms the ore-body worked by the Utah Copper Company; it is one of the largest low-grade copper mines of the world (p. 847). In many other parts of the district supergene processes have been at work enriching the ore deposits.

Some of the limestone lenses near the porphyry have been extensively contact-metamorphosed and near the intrusives chalcopyrite and pyrite have been introduced, as in the Highland Boy mine where the ores for a long time averaged from 2.5 to 3 per cent copper. In depth, lower grade pyritic ores were met. It seems probable that these ores are of early contact-metamorphic origin following the silicatization. But there are also many replacement deposits and veins, partly in the altered white limestone, partly in the unaltered black limestone, which carry mainly lead and zinc without any high-temperature minerals and which are considered to be formed at a lower temperature and during the later stages of metallization.

**Ketchikan, Alaska.**—Several contact-metamorphic copper deposits in southeastern Alaska are described by F. E. and C. W. Wright.[2]

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M. J. Elsing, Secondary enrichment at Cananea, *Eng. and Min. Jour.*, Sept. 25, 1930.

[1] J. M. Boutwell, *Prof. Paper* 38, U. S. Geol. Survey, 1905.

A. N. Winchell, Petrographic studies of limestone alteration at Bingham, *Trans.*, Am. Inst. Min. Met. Eng., 70, 1924, pp. 884–899.

W. Lindgren, Contact metamorphism at Bingham, *Bull.* Geol. Soc. Am., 35, 1924, pp. 507–514.

R. N. Hunt, The ores in the limestones at Bingham, *Trans.*, Am. Inst. Min. Met. Eng., 70, 1924, pp. 856–883.

[2] *Econ. Geol.*, 3, 1908, pp. 410–417; *Bull.* 347, U. S. Geol. Survey, 1908.

Those of Copper Mountain, Prince of Wales Island, present an unusually excellent illustration of deposits occurring at intervals along the contacts of an isolated granite intrusion. On the Kasaan Peninsula are several magnetite-chalcopyrite deposits, also containing pyrrhotite and pyrite, in a gangue of amphibole, epidote, orthoclase, garnet, and calcite. Wright believes that the ores were formed after the consolidation of the last intrusions of syenite. At both places shear zones and vein deposits containing copper accompany the contact deposits.

### Zinc and Lead Deposits

Almost all contact-metamorphic sulphide deposits contain some sphalerite, and often also a little galena, but only a few deposits are known in which these metals constitute the principal value of the ore. Where they occur, the amphiboles and epidote appear to be more prominent than garnet.

One of the best examples is furnished by the Magdalena mines,[1] in New Mexico, which in the oxidized zone, 200 to 300 feet deep, were worked for their lead, silver, and zinc. In depth large bodies of sphalerite were found, together with a little galena and chalcopyrite. According to Gordon, the Magdalena Range consists of faulted blocks of Paleozoic (Mississippian and Pennsylvanian) limestone, resting on pre-Cambrian crystalline rocks. The limestones are cut by dikes of granite porphyry, which are exposed near the Graphic mine and which are believed to have caused the mineralization. In the limestones, which dip westward, toward a hidden contact with the granite porphyry, mineralization has taken place at five horizons, of which only one, just below the "Silver Pipe" limestone, is of great importance. The orebodies are roughly lenticular and may be as much as 40 feet in thickness. Besides the sulphides mentioned, they contain magnetite and specularite, with much epidote, pyroxene, and tremolite, but little if any garnet. The distance along the dip of the strata to the intrusive rock is probably not less than 2,000 feet.

Important deposits of sphalerite with ilvaite and hedenbergite are found at Hanover,[2] New Mexico, not far from Fierro (p. 714); they are contained in the outer part of the metamorphic aureole, surrounding a stock of granodiorite porphyry.

Knopf[3] describes lead deposits from Darwin, Inyo County, California, which present a succession ranging from contact-metamorphic types to fissure veins. Over large areas the Carboniferous calcareous rocks are altered to wollastonite, diopside, and grossularite rocks, with perfect

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[1] C. H. Gordon, *Prof. Paper* 68, U. S. Geol. Survey, 1910, pp. 241–258.

[2] Ore deposits of the Southwest, *Guidebook* 14, Sixteenth Intern. Geol. Congress, 1933, p. 34 (Sidney Paige).

[3] Adolph Knopf, *Bull.* 580, U. S. Geol. Survey, 1915, pp. 1–18.

preservation of structure. An enormous quantity of material, chiefly silica, has been added during metamorphism. The contact-metamorphic ores lie between quartz-diorite and lime-silicate rocks. The minerals are galena, andradite, calcite, and fluorite. Orthoclase and apatite are present in other similar deposits. Numerous veins of galena and fluorite break through lime-silicate rock in other parts of the district. The contact-metamorphic deposits are considered by Knopf to be later than the general metamorphism, but the argument is not wholly convincing.

Contact-metamorphic deposits have been described from eastern Mexico, where the thick Cretaceous limestone is broken through by many small intrusives. Most of these are copper deposits, but sometimes they contain lead as at La Sirena mine,[1] near Zimapan, Hidalgo, where dikes and masses of monzonitic character intruded into the pure limestone have caused mineralization along the contacts. The ore formation began by development of quartz, albite, danburite ($CaB_2Si_2O_8$), apatite, garnet, actinolite, and fluorite in the order given. Then followed deposition of arsenopyrite and pyrrhotite, and lastly galena, sphalerite, and jamesonite.

## Gold Deposits

Gold is present in traces in almost all sulphide deposits of the pyrometasomatic type, and a few ounces of silver to the ton is likewise not unusual, but it is rare to find in such deposits ores which are valuable chiefly on account of their precious metals.

One of the best examples of a gold-bearing contact-metamorphic deposit is that of the Cable mine, Montana, described by W. H. Emmons.[2] The ores are included in a long, thin block of limestone, in contact on both sides with quartz monzonite. The principal minerals are calcite, quartz, pyrrhotite, pyrite, magnetite, and chalcopyrite, with actinolite, garnet, and green mica. The gold, in part coarse, is disseminated in calcite, quartz, and sulphides. Altaite, the lead telluride, is reported by Sharwood. This deposit has yielded about $3,000,000.

Platinum is rarely found. One occurrence in Sumatra is mentioned by L. Hundeshagen;[3] the metal occurs in wollastonite. It also appears in the contact-metamorphic zone of the Bushveld complex (Transvaal) (p. 782).

**Gold-Arsenopyrite Type.**—A deposit of this rare type, described by C. Camsell,[4] is that worked by the Nickel Plate mine, British

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[1] W. Lindgren and W. L. Whitehead, *Econ. Geol.*, 9, 1914, pp. 435–462.

[2] *Prof. Paper* 78, U. S. Geol. Survey, 1913, pp. 221–231.

[3] L. Hundeshagen, *Trans.*, Inst. Min. and Met. (London), 1904, pp. 1–3. See also under Platinum, Chap. XXXI.

[4] Geology and ore deposits of the Hedley district, B. C., *Mem.* 2, Canada Dept. Mines, Geol. Survey Branch, 1910.

Columbia. Gently folded Carboniferous limestones associated with shale, quartzite, and volcanic tuffs are intruded by sheets of gabbro and diorite. Along the contacts of these sheets, and particularly of their apophyses, the impure limestones are converted into contact-metamorphic minerals with arsenopyrite. The commercial ore-bodies, which have yielded several million dollars in gold, are tabular and follow the dipping contacts of the basic rock, which are not exactly parallel with the inclination of the strata. The principal ore-body has been followed 350 feet along the dip and has a width parallel to the contact of 125 feet.

The ore minerals are, in order of abundance, arsenopyrite, pyrrhotite, chalcopyrite, pyrite, sphalerite, tetradymite ($Bi_2Te_3$), and molybdenite. The depth of oxidation is slight, but in the upper levels free gold occurred associated with tetradymite, while at the greater depth now attained it seems to be intimately bound up with the arsenopyrite and is not amenable to amalgamation. The pure arsenopyrite may contain as much as 12 ounces of gold per ton. The gold tenor varies from $6 to $14 per ton, but beyond the ore-body minor quantities of gold are widely disseminated in the contact-metamorphic rocks. There is very little silver; traces of platinum (as sperrylite?) and nickel are present. The gangue minerals are andradite, pyroxene, epidote, calcite, and axinite, and the sulphides are closely intergrown with them; the chalcopyrite is somewhat later than the arsenopyrite. Although the rocks are faulted and fissured by post-intrusive stresses, these fractures contain practically no valuable ores.

The only similar deposit described in the literature is that of Reichenstein, Silesia, where auriferous arsenopyrite has been worked on a small scale, probably since the thirteenth century. According to C. Wienecke[1] and H. Reh,[2] the ore-producing intrusive is probably a neighboring granite, and the altered rock a dolomitic limestone.

Contact-metamorphic deposits carrying telluride ores are rare. W. H. Weed[3] describes such an occurrence at the Dolcoath mine at Elkhorn, Montana, where auriferous tetradymite is found in a 15 to 18-inch bed of garnet, diopside, and calcite. Weed[4] also mentions a deposit at Bannock, Montana, where tellurides occur with specularite, garnet, pyrite, and free gold at a contact between diorite and limestone.

It seems well established, then, that tellurides may sometimes crystallize at high temperatures. They are not known as products of igneous consolidation.

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[1] *Zeitschr. prakt. Geol.*, 1907, p. 273.

[2] Herbert Reh, Beitrag zur Kenntniss der erzgebirgischen Erzlager, *Neues Jahrbuch*, Beil. Bd., 65-A, 1932, pp. 1–86.

[3] W. H. Weed and J. Barrell, Elkhorn mining district, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, p. 506.

[4] *Trans.*, Am. Inst. Min. Eng., 33, 1903, p. 732.

### Cassiterite Deposits

Contact-metamorphic deposits with the assemblage of fluorine and boron minerals characteristic of cassiterite veins are rare. The tin-bearing magmas, which are generally normal granites, seem to retain the tin and associated substances until a later stage, after consolidation of the rock.

Cassiterite occurs, in part in connection with fissures, in the contact-metamorphic deposits of Pitkäranta, in Finland; of Schwarzenberg and Berggiesshübel, in Saxony; and of Campiglia Marittima, in Tuscany. Other examples of more typical character have been noted at Dartmoor, in Devonshire, England;[1] by A. Knopf[2] on Lost River, Seward Peninsula, Alaska; and by Steidtmann.[3] At the latter place the granite has produced a narrow but long contact zone of lime-silicate rocks rich in tourmaline, axinite, ludwigite, hulsite and paigeite (the last two are ferromagnesian stannoborates), vesuvianite, fluorite, scapolite, chondrodite, galena, sphalerite, arsenopyrite, pyrrhotite, scheelite, magnetite, pyroxene, and cassiterite. In the banded limestone the argillaceous layers are converted to tourmaline, with tremolite and vesuvianite, while the purer calcareous layers are marmorized. The orbicular structure of the contact minerals of this district has been mentioned on page 700. None of these deposits are of great economic importance.

At the last named locality cassiterite also occurs in tourmalinized granitic masses or dikes, in quartz veins cutting granite and developing greisen, in quartz porphyry dikes, and in quartz stringers cutting limestone and slate.

The dikes of quartz porphyry, which pierce the limestone, contain cassiterite, pyrite, arsenopyrite, wolframite, and fluorite, with mica and topaz. The adjoining limestone is reticulated by veins which carry cassiterite, and around these veinlets, hornblende, vesuvianite, fluorite, plagioclase, mica, and topaz have formed.

In the so-called Dolcoath lode a narrow dike is transformed into danburite (borosilicate of calcium) and tourmaline, with some arsenopyrite and cassiterite.

In the offshoots from the main granite mass are found fluorite, cassiterite, muscovite, tourmaline, and topaz, the last two having crystallized after the feldspar and quartz.

These interesting observations clearly show the intimate connection and, in fact, the transition between contact-metamorphic deposits and hypothermal veins.

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[1] K. Busz, *Neues Jahrbuch*, Beil. Bd. 13, 1899, p. 100.

[2] *Bull.* 358, U. S. Geol. Survey, 1908.

[3] E. Steidtmann and S. H. Cathcart, Geology of the York tin deposits, Alaska, *Bull.* 733, U. S. Geol. Survey, 1922.

## SCHEELITE DEPOSITS

Scheelite ($CaWO_4$), a heavy white mineral of non-metallic luster, occurs in many contact-metamorphic deposits. Such scheelite deposits have been discovered in Inyo County,[1] California, apparently of economic importance. The scheelite is associated with garnet, calcite, hornblende, and pyroxene. Metallic minerals are rare. In 1917 and 1918 several such deposits were discovered and worked in the Humboldt Range, Nevada, particularly near Mill City.[2] The association is calcite, scheelite, garnet, epidote, and pyrite; the deposit occurs in limestone close to granite and is intersected by a dike of aplite. Possibly many such deposits have been overlooked. The intrusive rock is usually a granodiorite; the replaced rock always a limestone or a dolomite. The succession is in general garnet, diopside, epidote; scheelite was apparently formed with the silicates; the scant sulphides were later. The granitic rock adjoining the contact had solidified at the time of ore formation. Hess and Larsen give the name of "tactite" to the contact-metamorphic

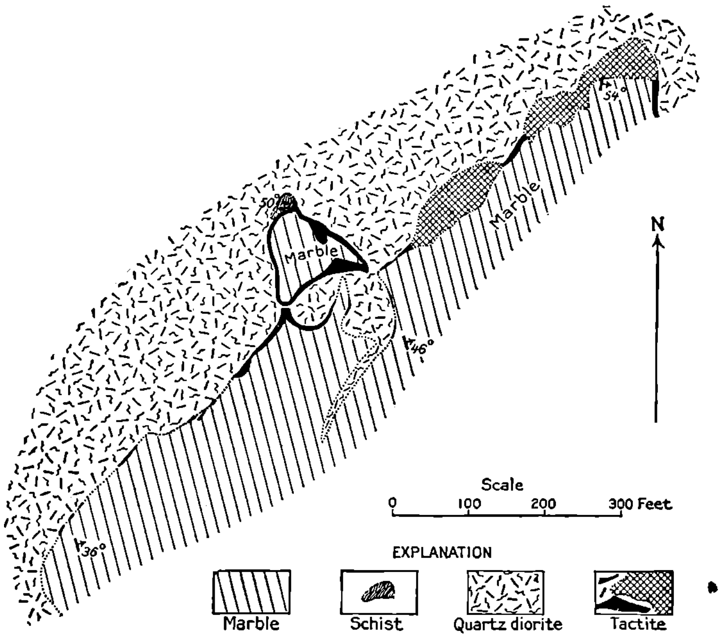

FIG. 292.—Map of scheelite contact zone near Golconda, Nevada. (*After Hess and Larsen, U. S. Geol. Survey.*)

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[1] Adolph Knopf, *Bull.* 640, U. S. Geol. Survey, 1917, pp. 229–249.

F. L. Hess and E. S. Larsen, Contact-metamorphic tungsten deposits of the United States, *Bull.* 725, *idem*, 1922, pp. 245–311.

[2] *Information Circ.* 6280, U. S. Bur. Mines, 1930.

rock. Many interesting illustrations show the tendency of the tungsten ore to form along the periphery of limestone inclusions in the intrusive (Fig. 292).

## Graphite[1]

**Properties.**—Graphite is a form of carbon crystallizing in the rhombohedral system; it is soft, steel-gray to black, and has a greasy feel. Even in its purest form it contains a little volatile matter and ash, usually less than 1 per cent. Many varieties are impure, and for some purposes, like paint-making, material with as little as 35 per cent graphitic carbon is utilized. Analyses quoted by Cirkel show that the commercial grades of graphite, even those considered as of high quality, contain several per cent of volatile matter and may be high in ash. It is stated that graphite always contains hydrogen and nitrogen.

The question whether graphite really exists in some of the varieties of "graphitic slate" yielding "amorphous graphite" is debatable; the minuteness of the particles renders it difficult to determine whether they are crystalline or not. The term "graphitoid" has been proposed for such substances. The best test for graphite is said to be its characteristic property of yielding "graphitic acid" ($C_{11}H_4O_5$) with strong oxidizing reagents such as nitric acid. The amorphous carbons do not respond to this test.

According to H. Moissan, graphite begins to oxidize at 650° to 700° C. In texture graphite is flaky or scaly or, when in veins, often fibrous perpendicular to the walls; these varieties are called "crystalline" in the trade. "Plumbago" and "black lead" are trade names for the mineral.

**General Occurrence and Origin.**—Graphite appears mainly in rocks which have suffered intense regional or igneous metamorphism. The literature on its occurrence and origin is very extensive and shows plainly that the mineral may have originated in several ways:

1. It may form an integral part of rock magmas and crystallize together with the rock. This possibility is indicated by its presence in meteorites, in the terrestrial iron of Ovifak, Greenland, in nepheline syenite,[2] and in pegmatites.[3] In some of the occurrences in pegmatite dikes the graphite has, however, clearly been absorbed from the surrounding crystalline limestone. This is the origin of one of the occurrences

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[1] F. Cirkel, Graphite, its properties, occurrence, refining, and uses, Dept. Mines, Canada, 1907, p. 307.

B. L. Miller, *Mineral Industry*, Annual issues.

E. Donath, Der Graphit, Leipzig and Vienna, 1904.

E. Weinschenk, Der Graphit, etc., Leipzig, 1904.

T. H. Clark, The origin of graphite, *Econ. Geol.*, 16, 1921, pp. 167–183; A. N. Winchell, *idem*, p. 492.

[2] T. H. Holland, *Mem.*, Geol. Survey India, 30, 1901, p. 201.

[3] G. O. Smith, *Bull.* 285, U. S. Geol. Survey, 1906, pp. 480–483.

described by George Otis Smith, as well as of the graphite in a dike near Franklin Furnace, New Jersey, described by A. C. Spencer.[1] These deposits are rarely of economic importance.

2. Graphite forms by the recrystallization of carbonaceous matter in metamorphic sedimentary rocks resulting from sandstone, shale, limestone, or coal. This transformation is well established and can evidently be effected under conditions of intense regional or igneous metamorphism, but it probably requires a relatively high degree of heat, certainly well above 200° C. The development of graphite in the zone of contact metamorphism is assumed by some authors, like E. Weinschenk, to mean that the carbon has been supplied by emanations from the magma. Weinschenk also applies this theory to its occurrence in many areas of regional metamorphism, but this view is probably not justified.

In studying the contact-metamorphic graphite from Ticonderoga, New York, E. S. Bastin showed by experiments that the contemporaneous quartz crystals had not been exposed to a temperature of 575° C. While a very high temperature is necessary for the manufacture of artificial graphite, the transformation can evidently be effected in nature at a much lower degree of heat.

Winchell[2] has shown that two reactions are possible by which the carbon in limestone can be expelled in gaseous form, as, for instance, has evidently happened at Bingham, Utah, during contact metamorphism; they are: $C + 2H_2O = CO_2 + 2H_2$ and $C + CO_2 = 2CO$. If these reactions are reversible and move towards the deposition of free carbon upon cooling, this would afford a satisfactory explanation of the formation of graphite. The first reaction does not occur below 550° and proceeds very slowly below 600° while the second reaction does not begin below 600° but proceeds freely at 800° to 900° C. The free carbon was probably driven out of the limestone while the temperature was at least 700° C. The same solutions produced tremolite which can not be formed above 1000 to 1100° C.[3] If both processes went on simultaneously, therefore, the temperature was probably about 800° C.

3. Lastly, graphite occurs in veins, sometimes 2 or 3 feet wide, having the appearance of resulting from the filling of open fissures; and in this form the mineral usually possesses a marked transverse fibrous structure. Such veins are found in igneous rocks like pegmatites and granites, and also in the surrounding metamorphosed sediments. Fine examples are seen in the graphite regions of New York, Canada, and Ceylon.

The origin of this type is less easy to explain. As the veins are usually found near intrusive contacts where high heat prevailed, it may be conjectured that they were formed by deposition from gaseous carbon

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[1] Franklin Furnace folio, 161 *Geol. Atlas*, U. S. Geol. Survey, 1908.

[2] *Trans.*, Am. Inst. Min. Eng., 70, 1924, pp. 884–899.

[3] E. T. Allen and J. K. Clement, *Am. Jour. Sci.*, 4th ser., 26, 1908, p. 101.

compounds, such as carbon monoxide or cyanogen compounds, perhaps with metals; in some of these graphites the ash contains much iron. The prevailing opinion is that the carbon is derived from surrounding sediments and was deposited shortly after the intrusion, but E. Weinschenk[1] and others consider it as originating from exhalations of igneous origin. Types 2 and 3 form many valuable graphite deposits.

**Occurrences.**—Deposits of graphite have been found in a number of states in the Union, but few are of economic importance; many of them are graphite slates or clays which are utilized as pigments or as lubricants.

A part of the domestic supply of "crystalline" graphite is obtained from New York; the mines are located in Essex, Warren, Washington, and Saratoga counties, in the Adirondack region,[2] and the largest mine, that of the American Graphite Company, has been worked for 30 years. The rocks are pre-Cambrian crystalline schists of sedimentary origin. The principal bed worked is a dark silver-gray quartz-graphite schist and is said to average about 6 per cent graphitic carbon. Elongated quartz grains, muscovite, apatite, pyrite, and graphite, the last in thin and ragged flakes, averaging about 1 millimeter in length, are the constituents. Two beds are known, one about 8 feet thick, the other from 3 to 20 feet. Excavations have extended for 2,000 feet along the gentle dip of the thicker bed, the greatest depth below the surface being 250 feet. The associated rocks are garnetiferous gneisses and limestones of the Grenville series. The sediments are metamorphosed by intrusion and injection of granite and gabbro of Laurentian and possibly Algoman age.

Three miles northwest of Ticonderoga, in the same region, coarse graphite plates are irregularly distributed throughout the contact zone between pegmatite and pegmatitic granite and the schists and limestones which these rocks intrude. Contact-metamorphic minerals, like scapolite, pyroxene, and vesuvianite, occur in this zone. The graphite also forms veins, 1 to 2 inches wide, which cut across both the schist and granite. The deposits at this locality have been worked for a number of years.

The production of flake graphite from a belt of Paleozoic mica schist in Clay County, Alabama, and the adjacent region has acquired considerable importance. The ore contains about 3 per cent graphite.

A deposit containing graphite in veins similar to those of Ceylon has been found near Dillon, Montana.[3] The veins occur along a contact

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[1] *Abh.* Bayer. Akad. Wiss., 21, 1901, pp. 218–335.

[2] E. S. Bastin, *Econ. Geol.*, 5, 1910, pp. 134–157.

D. H. Newland, Annual reports of the graphite industry, *Bulls.*, 1905 to 1916, N. Y. State Mus.

Ida H. Ogilvie, *Bull.* 96, *idem* (*Geological map*).

H. L. Alling, The Adirondack graphite deposits, *Bull.* 199, N. Y. State Mus., 1918.

[3] A. N. Winchell, *Econ. Geol.*, 6, 1911, p. 218.

E. S. Bastin, *idem*, 7, 1912, p. 435.

zone of granites and pegmatites, intrusive in pre-Cambrian schists and calcareous rocks which have been contact-metamorphosed.

At several places in New Mexico[1] intrusions of basic igneous rocks have altered the coal beds of the Tertiary or Cretaceous formations. At Madrid the coal was converted to anthracite. Near Raton the intrusions have turned the coal into a coke-like material, but at one place 7 miles southwest of Raton a number of sills produced exceptionally intense metamorphism, converting the coal to graphite. Graphite also occurs in irregular masses in the diabase and has a more or less columnar texture normal to the faces of the igneous rock.

Similar conditions produced the important deposit of amorphous graphite of Santa Maria, in central Sonora, Mexico. According to F. L. Hess,[2] several coal beds, attaining a maximum thickness of 24 feet, have been subjected to contact metamorphism and folding by intruding granite and are converted into amorphous graphite. The main vein averages 86 per cent graphitic carbon and furnishes a good material for the manufacture of lead pencils.

The graphite deposits of Ceylon[3] are among the most productive in the world, yielding about 14,000 metric tons a year of high-grade product, principally from the Ragadera mine. The mineral is said to occur as filled fissure veins, from 12 to 22 centimeters wide. They show little or no replacement but consist almost entirely of graphite with radial fibers at the selvages, and platy masses in the middle. According to F. D. Adams, the earlier minerals are pyroxene, calcite, apatite, rutile, and pyrite. The calcite seems to be contemporaneous with the graphite. The mines are from 100 to 1,200 feet deep. The rough material often contains up to 50 per cent impurities and is hand picked and sorted.

According to Bastin, the veins are found in a fine-grained acidic or basic gneiss to which he applies the name granulite. Some crystalline limestone is also present. The gneisses are intruded by granites and pegmatites. In the last few years Madagascar is rivaling Ceylon. In 1929, 16,050 metric tons are said to have been produced, the quality of flake graphite being about the same as that of the domestic output.

The Siberian deposits, in the Batagol Mountains near Irkutsk, yield material of great purity, which formerly supplied the lead-pencil industry. L. Jaczewski,[4] describing the Alibert mines in this region, states that the graphite occurs in a nepheline syenite close to the contact of a schist that also contains graphite, the latter, as well as the inclusions in the igneous rock, being considered of organic origin.

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[1] W. T. Lee, *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1908, p. 733.

[2] *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1908, p. 734.

[3] E. S. Bastin, *Econ. Geol.*, 7, 1912, pp. 419–443 (with literature).
F. D. Adams, *Trans.*, Can. Inst. Min. Eng., 1926, pp. 496–503.

[4] *Neues Jahrbuch*, 1901, ref. 2, p. 74. (Original in Russian.)

The deposits at Passau, in Bavaria, comprise few veins; the graphite occurs in a crushed, schistose rock; and Weinschenk[1] regards the deposits as caused by volcanic emanations. Passau is discussed again by E. Ryschkewitsch and W. Peinert,[2] both authors[3] apparently upholding Weinschenk's theory of magmatic origin.

The graphite deposits of Canada are contained chiefly in Buckingham and Grenville townships, Quebec, near Ottawa. The production in 1929 was about 1,325 metric tons. These deposits, which have been described by A. Osann,[4] show particularly clear relations to contact metamorphism. The rocks are largely gneiss, quartzite, and crystalline limestone of Grenville age cut by granite, pegmatite, and diorite. Graphite is widely distributed in fissure veins or lenticular masses in these intrusions or near their contacts, also as disseminated flakes in limestone or gneiss (Figs. 293 and 294). Associated minerals are apatite and scapolite, often appearing in the wall rocks of the veins, also biotite, titanite, wollastonite, and pyrite. The analogy of these deposits with the apatite veins is striking and the conclusion seems justified that they were developed by igneous emanations shortly after the close of the intrusive activity.

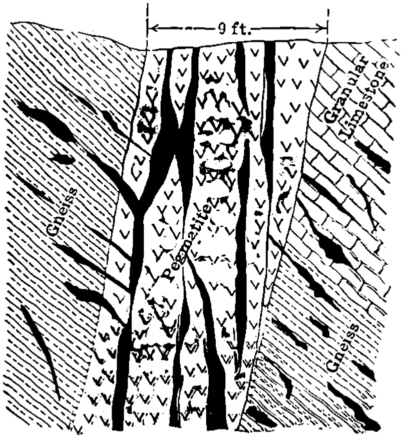

FIG. 293.—Vertical section of graphite veins, Buckingham Township, Quebec. Black, graphite. (*After A. Osann.*)

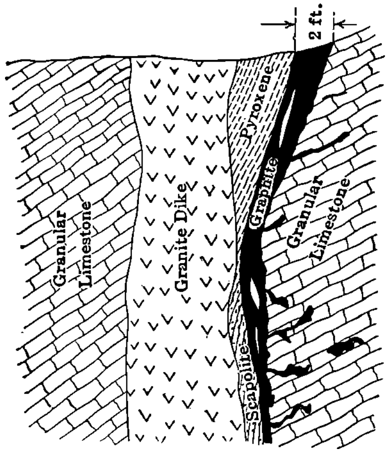

FIG. 294.—Vertical section of graphite vein in limestone, Grenville district, Quebec. (*After A. Osann.*)

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[1] E. Weinschenk, Zur Kenntniss der Graphitlagerstätten., *Abh.* Bayer. Akad. Wiss., 19, 1899, pp. 511–563.

[2] *Zeitschr. prakt. Geol.*, 1924, pp. 70–76.

[3] *Idem*, 1925, pp. 12–15.

[4] *Ann. Rept.*, Canada Geol. Survey, 1899, pp. 660–820. See also Cirkel's report quoted above.

Deposits from South Australia similar to those of the Adirondacks have been described by C. E. Tilley.[1] Tilley shows inclusions of graphite crystals in garnet.

**Production and Uses.**—The production of natural graphite in the United States has varied considerably, owing to the large quantities of low-grade material used for paints and fertilizers. The output of flake graphite from New York State was in 1918 about 1,600 tons per annum. Alabama, in 1924, produced 292 tons; the total domestic output of natural graphite in 1929 was 6,458 short tons, of which 2,900 tons were flake graphite, mainly from Alabama. There was no domestic output in 1931. Artificial graphite is manufactured in electric furnaces at Niagara Falls at the rate of 6,000 tons per annum, from anthracite coal mixed with a small percentage of ash. In addition about 20,000 to 42,000 tons of graphite are imported from the highly productive mines in Ceylon, Mexico, Korea, and Madagascar. Ceylon graphite sold in New York (1929) for 3 to 9 cents a pound. Domestic crystalline graphite brought an average price of 4.6 cents per pound in 1929.

There is a great demand for graphite from many branches of industry. The inert and heat-resisting nature of the "crystalline" graphite makes it particularly valuable for crucibles, the fibrous Ceylon product being most suitable for this purpose.

Graphite is extensively used as a lubricant, with oil, and for this purpose the artificial mineral, which is "deflocculated," causing it to remain indefinitely in suspension in oil, is especially employed. Other uses are for pencils, foundry facings, polishing powder, paint, electrodes, and, strange to say, as an adulterant for fertilizers; it is claimed that it prevents absorption of moisture and caking.

The low-grade material from New York State is concentrated at the mines by crushing, washing on buddles or other appliances, and settling, but the details of the process have not been made public. The practice for amorphous graphite in Clay County, Alabama, includes dry crushing, water flotation and drying.[2]

## Garnet

Some varieties of garnet, especially almandite, are mined and used as abrasive material. In the state of New York there are several deposits of this kind.[3] The garnets occur in highly altered rocks of somewhat uncertain history but probably the result of contact metamorphism.

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[1] *Econ. Geol.*, 16, 1921, pp. 185–198.

[2] Irving Herr, *Eng. and Min. Jour.*, April 21, 1917.

[3] W. J. Miller, Garnet deposits of Warren County, N. Y., *Econ. Geol.*, 7, 1912, pp. 493–501.

W. M. Myers and C. O. Anderson, Garnet, its mining, milling and utilization, *Bull.* 256, U. S. Bur. Mines, 1926.

W. M. Myers, Mineralogy of commercial garnets, *Am. Jour. Sci.*, 5th ser., 12, 1926, p. 115.

## PYROMETASOMATIC DEPOSITS NOT RELATED TO CONTACTS

**General Features.**—The deposits thus far described lie close to the well-defined contact of an intrusive rock with a sedimentary series. There are deposits, however, in which the mineral association points to the same mode of origin, but which are not clearly related to any given contact. This may result from a sloping or irregular contact of a large batholith, so that a point on the surface that is several miles from the contact horizontally may be only a few thousand feet from it in a vertical direction. General metamorphism, without special development of mineral deposits, appears to have been effected by such conditions at the northern end of the great batholith of Idaho between the Clearwater and St. Joe rivers.[1] During a long and deep immersion into the abyssal zone, metallic gases given off by magmas may have penetrated farther from the intrusion than they have near the surface. Obscure fissures may have guided the emanations. Erosion may have cut away the metallizing dike or mass, so that its relation to the deposit is no longer apparent.

Ores of copper, zinc, lead, and iron are included in this class. Many representatives are found among the obscure deposits in the pre-Cambrian of Scandinavia.

**Boundary District.**—At Phoenix[2] and Greenwood, in British Columbia near the international boundary, are a number of ore-bodies which have yielded several million tons of ore. The geology of the region is complex. A thick series of volcanic rocks (porphyrites), both clastic and massive, crystalline limestones, and argillites, all of upper Paleozoic age, is intruded by a granitic batholith of probable Jurassic age and smaller masses of syenite.

The large ore deposit of the Granby Company, now exhausted, lies in a mineralized zone which represents a part of the limestone replaced by epidote, garnet, etc. The ore-bodies are lenses or large masses, one of which is 2,500 feet long and 900 feet wide and has a maximum thickness of 125 feet. The dip becomes flat in depth and the ore ceases at a vertical depth of 675 feet. The ore consists of chalcopyrite, pyrite, hematite, and magnetite, with andradite, actinolite, and epidote. Calcite and quartz fill the interstices between the lime-iron silicates. The ore as smelted contained from 1.2 to 1.6 per cent copper with 0.04 ounce of gold and 0.3 ounce of silver per ton. The original limestone, which appears in some remnants near the ore-body, is comparatively pure and contains from 1 to 10 per cent of silica and little or no iron. The limestone is in large part converted to jasperoid, the alteration appearing to have taken place before the development of the ore.

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[1] F. C. Calkins and E. L. Jones, *Bull.* 530, U. S. Geol. Survey, 1913, pp. 75–86.

[2] O. E. LeRoy, *Mem.* 21, Canada Geol. Survey, 1912.

No large bodies of igneous rocks appear in or near the deposits, and the nearest—small—outcrops of granodiorite are 1 to 2 miles away; one of these outcrops has been locally replaced by garnet, epidote, and actinolite. Deep drilling below the deposits failed to disclose intrusive rocks. It is held that the ores were formed by igneous emanations of iron, silica, and copper which traversed the limestone laterally from some unit of the intrusive series that is now eroded.

**Ducktown, Tennessee.**—The copper ores at Ducktown have been worked since about 1848 and still maintain an output of about 8,000 tons of copper a year. In addition, about 1,000 tons of sulphuric acid are now obtained daily from these ores. The district, which lies in the mountainous area of the southern Appalachians, has been the subject of repeated geologic investigation by C. Heinrich, J. F. Kemp, and W. H. Weed. W. H. Emmons and F. B. Laney[1] reported on the deposits in 1926. The last paper is by C. S. Ross, 1935.

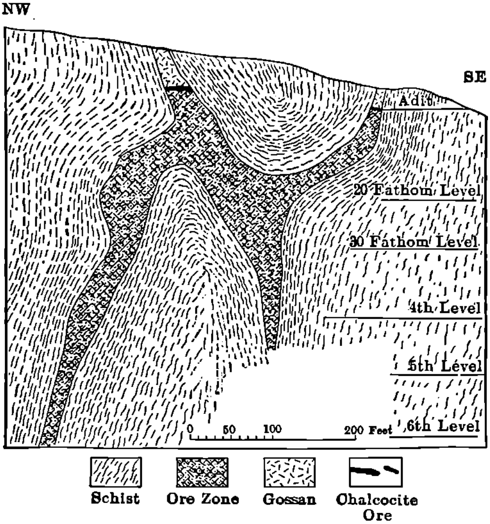

FIG. 295.—Cross-section of Mary mine, Ducktown, Tennessee. (*After W. H. Emmons, U. S. Geol. Survey.*)

The deposits are contained in a highly compressed, metamorphosed and schistose series of arkosic sediments of Cambrian age, consisting of poorly sorted conglomerate, sandstone, and shale. Garnet and staurolite have developed abundantly in the rocks, the staurolite following certain horizons persistently. Thin lenses of limestone were contained in the series and are exposed in some places in the mines; they are now crystalline and contain layers of biotite and muscovite. Here and there

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[1] W. H. Emmons and F. B. Laney, Geology and ore deposits of the Ducktown mining district, Tenn., *Prof. Paper* 139, U. S. Geol. Survey, 1926. C. S. Ross, *Prof. Paper* 179, 1935. Ross denies the presence of limestone lenses and holds that the carbonates were deposited by hydrothermal solutions.

are small ill-defined lenses of a highly metamorphic rock looking like a diorite-pegmatite and consisting of quartz, feldspar, hornblende, and garnet. These peculiar phases are now believed to be the result of strong metamorphism of the arkosic sediments. Dikes of gabbro, later than the mineralization, are intruded in the sediments.

The deposits are elongated, roughly tabular masses, some of them curved, lens-shaped, or folded, striking northeast and mostly dipping southeast (Fig. 295). The ore beds are parallel to the strike of the schists and average 60 feet in width. The primary ore is a coarsely crystalline mass of pyrrhotite, pyrite, chalcopyrite, sphalerite, specularite, magnetite, actinolite, calcite, tremolite, quartz, pyroxene, garnet, zoisite, chlorite, mica, graphite, titanite, and feldspars, all of practically contemporaneous crystallization.

Much of the ore is almost massive pyrrhotite and pyrite. Along the strike and dip the ore may grade into a lime-silicate rock of gangue minerals, and these in places grade into crystalline limestone. The contact between schist and ore is sharp or gradational within a few inches. The beds have been worked to a maximum depth of 1,400 feet. A thin but rich chalcocite zone due to enrichment by surface waters was found at a depth of 50 feet (p. 835), but below this the ores contain 1.5 to 3.0 per cent copper, a small amount of silver, and a trace of gold. The ores from the Mary mine average 2.5 per cent. It is held that the ores are formed by the replacement of thin limestone beds; all the abundant gangue minerals are in fact rich in lime. The sequence is believed to have been as follows: (1) Development of lime silicates and pyrite; (2) crushing by continued dynamo-metamorphism; (3) introduction of pyrrhotite and, later, sulphides of copper, zinc, and lead.

**Franklin Furnace, New Jersey.**[1]—The great zinc-manganese deposits of northern New Jersey are of exceptional complexity and interest. Known since 1650, and actively worked since 1860, they now yield annually about 773,000 short tons of ore containing about 98,000 tons of zinc. The treatment of the crude ore by magnetic concentration yields franklinite, "half and half," and willemite; the first is used for the manufacture of zinc oxide for paints and leaves a manganiferous residue which goes to the blast furnace to make spiegeleisen; the second is also used for zinc white; and the third after further concentration yields a product of willemite from which a high-grade spelter (zinc) is made.

[1] A. C. Spencer, H. B. Kümmel, J. E. Wolff, and Charles Palache, Franklin Furnace folio, *Geol. Atlas* 161, U. S. Geol. Survey, 1908; review by C. K. Leith, *Econ. Geol.*, 4, 1909, p. 265.

H. Ries and W. C. Bowen, Origin of the zinc ores of Sussex County, N. J., *Econ. Geol.*, 17, 1922, pp. 517–571.

J. E. Spurr and J. Volney Lewis, *Eng. Min. Jour.-Press*, Feb. 21, 1925.

Charles Palache, The minerals of Franklin and Sterling Hill, *Prof. Paper* 180 U. S. Geol. Survey, 1935.

The two ore deposits of Mine Hill and Sterling Hill, 3 miles apart, are situated along a belt of pre-Cambrian crystalline limestone adjoined on the west by coarse gneisses of igneous origin. Cambrian limestone covers these rocks to the east and west. Both deposits form canoe-shaped beds in the pre-Cambrian limestone. The Mine Hill ore bed (Fig. 296) is closely adjoined along its west flank by the gneiss, the contact of which is parallel to the ore-body. The ore mass is thus a layer varying from 12 to 100 feet in thickness, which, bent upon itself, forms a long trough or one-half of a canoe with sides of unequal height, the keel pitching north at a gentle angle.

The mines are opened by a vertical shaft 965 feet deep and an incline 1,500 feet long. The ore forms transitions into the limestone, and at Sterling Hill the limestone between the flanks also contains lean ore. Pegmatite dikes cut ore, limestone, and gneiss. The ore is a coarse aggregate of franklinite, 50 per cent; willemite, 20 to 30 per cent; zincite, 2 to 6 per cent; and calcite, 3 to 11 per cent. Franklinite ((Fe, Mn, Zn) O. (Fe, Mn)$_2$O$_3$), contains about 42 per cent iron, 15 per cent manganese and 12 per cent zinc; willemite, ($Zn_2SiO_4$), 58 per cent zinc; zincite, (ZnO), 77 per cent zinc. The four minerals mentioned are held to constitute the original ore. Besides, there are a great number of rarer minerals such as tephroite ($Mn_2SiO_4$), zinc pyroxene (schefferite), zinc amphibole, zinc spinel (gahnite), manganese garnet (polyadelphite), axinite (borosilicate of Al, Ca, Fe, Mn), apatite and scapolite (containing chlorine), rhodochrosite, fluorite, sphalerite, galena, arsenopyrite, chalcopyrite, and löllingite. In all, about 100 minerals have been identified in the deposit. Most of these are regarded as products of secondary metamorphism due to the pegmatite dikes. Many veins cut the deposits,

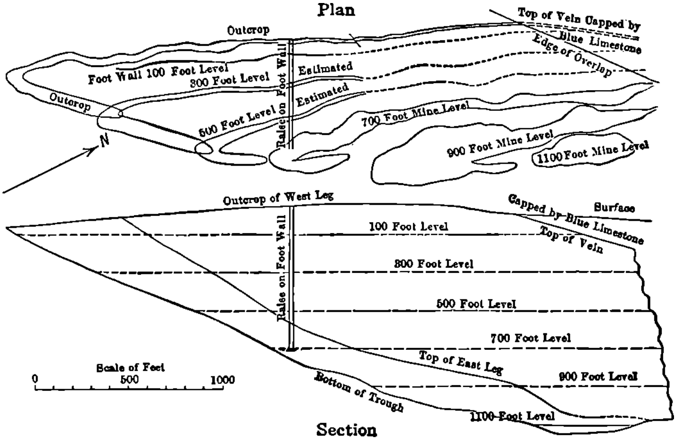

FIG. 296.—Plan of outcrop and levels, and vertical section of Mine Hill ore-body, Franklin Furnace, New Jersey. (*After A. C. Spencer, U. S. Geol. Survey.*)

some of them contain the normal recrystallized ore minerals, others are distinctly later with sulphides associated with calcite, albite, bornite, quartz, dolomite, etc.

In the older literature the deposits were considered of sedimentary origin. The question of genesis was reopened in 1889 by F. L. Nason, who admitted the possibility of igneous origin. Spencer believes that the original deposit was formed by the injection of magmatic emanations from the gneiss intrusions into the limestone. Participation in the general deep metamorphism which affected this region in pre-Cambrian time has further complicated the relations. It is certain that the texture of the ore and the universal rounding or corroding of the ore minerals point distinctly to igneous metasomatic action. The abundance of the spinel minerals is indicative of high temperature.

Ries and Bowen believe that the ore minerals were formed by replacement of a limestone bed which subsequently was folded. Spurr and Lewis suppose that the deposit was originally a magmatic ore-body of sphalerite, pyrite, and rhodochrosite which later was metamorphosed by an injection of magma. Still another theory which has something to recommend it, is that the deposit was originally an epigenetic sulphide mass in limestone, which was thoroughly oxidized by weathering processes and subsequently intruded and metamorphosed by granite.

**Långban, Sweden.**[1]—This remarkable manganese deposit, from which 108 minerals have been described, in some respects forms a close analogy to Franklin Furnace. Magnetite, hematite, braunite, and hausmannite are accompanied by andradite, diopside, and tremolite. There are also lead silicates, arsenates, arsenites, antimonites, and native lead. It is held that a primary Fe-Mn ore, probably sedimentary, was exposed to granitic intrusions which gave rise to the extremely complicated mineralogical development.

**Metasomatic Magnetite Deposits of Sweden.**[2]—Many of the earliest known and longest worked of the Swedish iron deposits form irregular

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[1] N. H. Magnusson, Långbans malmfält, Geol. Survey Sweden, *Mon.* 23, 1930, 111 pp.

[2] Hj. Sjögren, The geological relations of the Scandinavian iron ores, *Trans.*, Am. Inst. Min. Eng., 38, 1908, pp. 766–835.

Hj. Sjögren, The question of the origin of the iron ores in the older pre-Cambrian series of Sweden, *Geol. För. Förh.*, 30, 1908, pp. 115–155.

H. Johansson, The question of the origin of the middle-Swedish iron ores (Swedish), *Geol. För. Förh.*, 28, 1906, pp. 516–538; 29, 1907, pp. 143–186, 232–255; 30, 1908, pp. 232–235.

Review, *Econ. Geol.*, 5, 1910, pp. 494–498.

See also a summary of literature by A. Bergeat in *Fortschritte der Mineralogie*, etc., 2, Jena, 1911, pp. 43–44.

Excellent descriptions of individual districts are found in the guide to the excursions of the Internat. Geol. Congress, Stockholm, 1910.

P. Geijer, Some problems in iron ore geology in Sweden and America, *Econ. Geol.*, masses or lenses in rocks of Archean age. They are either directly associated with crystalline limestone, or they occur near limestone but surrounded by masses of silicates, like pyroxene, garnet, and epidote, to which the term "skarn" is usually applied. Though not so large as some of the more recently discovered deposits of certain or probable magmatic origin, the deposits have in the aggregate furnished much ore of exceptional purity and as yet are far from being exhausted. Formerly these deposits were considered by the Swedish geologists as of sedimentary origin, like bog iron ores, but subsequently metamorphosed. At present few observers hold to this view. There is a strong similarity between the Swedish ores and the pyrometasomatic deposits of the Banat. Unmistakable also is their similarity to the pyrometasomatic deposits of North America, many of which contain much magnetite and which at some places are worked for iron.

The Swedish deposits are, however, not so simply explained, for while in the districts mentioned the ores unquestionably adjoin igneous intrusions, the granitic rocks of Sweden are generally later than the deposits, which normally are contained in a peculiar fine-grained rock with gneissoid texture that is widely distributed in the iron region and has been variously designated "hälleflinta," eurite, leptite, or granulite. These rocks, which form wide zones in the pre-Cambrian of Sweden and are locally associated or interbedded with amphibolites and masses of more distinctly sedimentary quartz-mica slates and also with beds or lenses of crystalline limestone or dolomite, are salic rocks, generally with at least 67 per cent silica, and consist largely of albite, orthoclase, and quartz. The most prevalent opinion is that these rocks are in part effusive, perhaps originally tuffaceous, and in part intrusive, generally supracrustal, and that the limestone and mica schist are of sedimentary origin.

The bodies of magnetite are in general associated with masses of crystalline limestone in this leptite formation. The ores form stock-like masses with greatest extension in a vertical direction and border directly against granulite, limestone, or "skarn." The bodies have been followed to a depth of about 1,000 feet; some of them cease distinctly at various depths, but other stocks still continue below the greatest depth reached. Many of them, but not all, conform with the banding of the leptite (Fig. 298).

The "limestone ores" are more directly embedded in limestone, but here too skarn minerals may be present. In such an ore-body at Klack-

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10, 1915, pp. 209–239.

P. J. Holmquist, Structure and metamorphism of Swedish iron ores (Swedish), *Geol. För. Förh.*, 35, 1913, pp. 233–272.

N. H. Magnusson, Persbergs Malmtrakt, *Bull.* 2, Dept. Comm., Stockholm, 1925; Nordmarks Malmotrakt, *Mon.* 13, Geol. Survey Sweden, 1929, 98 pp.

berg a narrow zone of dark-brown garnet was noted along the contact of limestone and ore, and in the limestone itself was disseminated a dark-brown amphibole. The "skarn" consists mainly of pyroxene and amphibole. At Persberg and other places, there are also some cordierite-magnetite ores poor in lime. The limestone ores often carry manganese and some of them constitute manganese deposits, like that of Långban. Stretching and schistosity were superimposed upon the deposits in places, and sometimes the direction of the stretching indicates the pitch of the ore-body. The magnetite is fine-grained; it contains in places a little specularite. Some deposits contain small quantities of pyrite, pyrrhotite, chalcopyrite, and arsenopyrite.

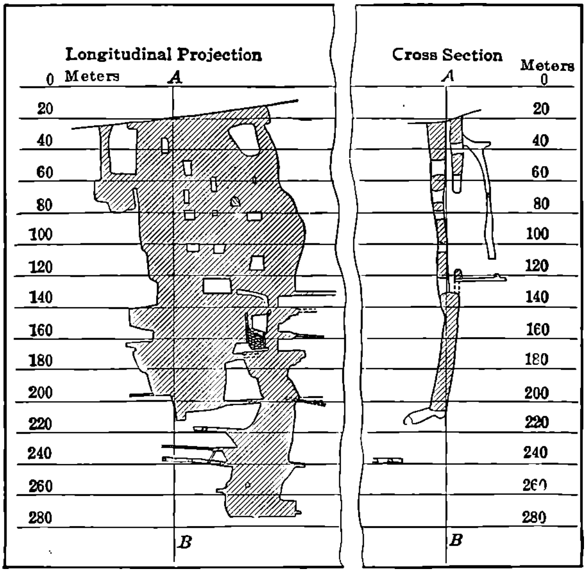

FIG. 297.—Vertical sections of the Kran mine, Persberg, Sweden. Shaded areas indicate stopes. (*After Hj. Sjögren.*)

The composition of one of the Persberg ores is as follows:

| | | | |
|---|---|---|---|
| $Fe_3O_4$ | 71.56 | CaO | 4.85 |
| FeO | 5.11 | $Al_2O_3$ | 0.77 |
| Fe | 55.79 | $SiO_2$ | 12.76 |
| MnO | 0.17 | $P_2O_5$ | 0.005 |
| MgO | 4.18 | S | 0.031 |

Secondary changes have resulted in crushing along certain zones (skölar) and a great development of chlorite and other minerals of dynamometamorphic affiliations. Among the celebrated deposits of this type should be mentioned those of Persberg, Taberg (in Wermland), Nordmark, Norberg, and Dannemora.

The field relations indicate beyond doubt that the ores and skarn are metasomatic replacements of limestone or dolomite similar to contact-metamorphic deposits, probably effected by very hot solutions, containing iron, manganese, silica, etc., derived from intrusive magmas.

Holmquist holds that the bedded, supracrustal leptites with accompanying sedimentary iron ores (p. 313) subsided into underlying granitic magma, which effected igneous metamorphism of the bedded ores and developed magnetite and lime silicates in the limestone. The later events included a regional metamorphism which affected the ores to some degree, and the final intrusions of granite, pegmatite, and diabase which have exerted very slight influence on the deposits.

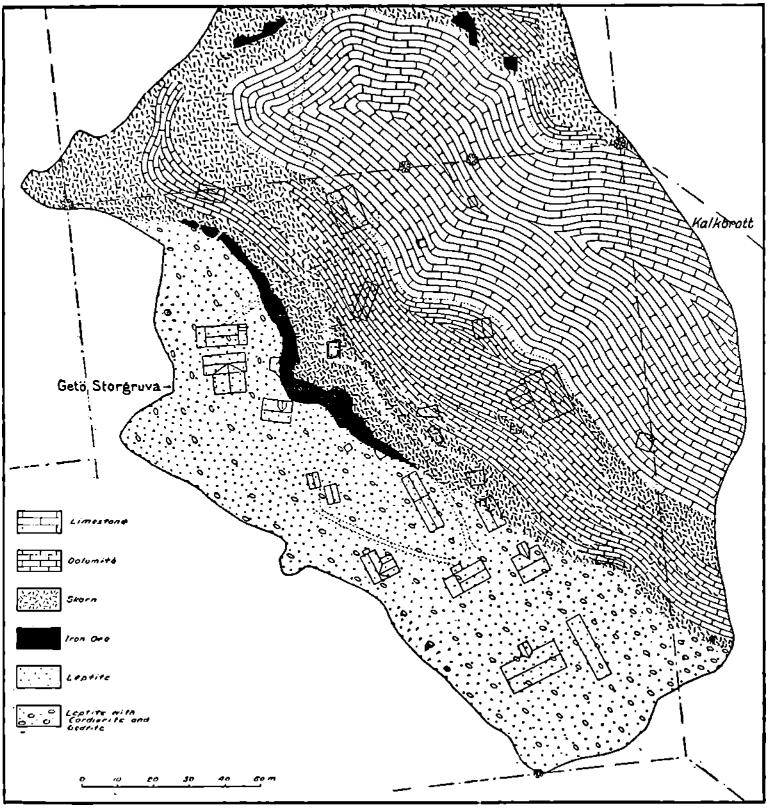

FIG. 298.—Geological map of Getön, near Persberg, Sweden. Showing relations of magnetite to the leptite-limestone series. (*After N. H. Magnusson.*)

During the last years opinions have again shifted somewhat. Sundius, Lindroth and Magnusson[1] appear to believe that the iron ores

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[1] See discussion of N. H. Magnusson and Per Geijer in *Geol. För. Förh.*, 47, 1925, pp. 357–361. (Footnote continued on following page.)

stand in closest connection with the effusive leptites; and, the leptites being supracrustal, it is suggested that one must choose, therefore, between a sedimentary origin and metasomatism near the surface at lower temperatures. The formation of the iron ores was followed by the intrusion of the ancient granites which metamorphosed the skarn ores and the leptite with development of silica, fluorite, pyrite, scheelite, anthophyllite, gedrite, and cordierite, probably also sulphide ores (p. 744). Still later came the younger granites which produced very slight changes. P. Geijer holds to the view of deep-seated pyrometasomatism, and his arguments have considerable weight. The relations are exceptionally well illustrated by Fig. 298.

**Magnetite Deposits in the United States.**—Deposits of magnetite, which are similar to the Swedish ores just described, are found in the United States at few places. The Tilly Foster mine,[1] in New York State, contained a steep lenticular body of ore embedded in gneiss; the magnetite was associated with calcite, dolomite, chondrodite, enstatite, epidote, chlorite, garnet, and scant sulphides. The ore-body was followed to a depth of about 600 feet.

**The Iron Ores of the Lahn Syncline, Germany.**—This district of folded Devonian slate, "schalstein," and effusive, probably submarine, diabase flows (Fig. 16) has attracted much attention.[2] The ores lie between diabase and Devonian slates or tuffaceous beds and contain magnetite, specularite, pyrite, siderite, and quartz; magnetite often replaces crystals of specularite. It is held probable by the authors cited that the ores were formed by submarine volcanic exhalations, the products of which were deposited on the sea bottom. This explanation is not universally accepted in Germany.

## PYROMETASOMATISM OF THE CORDIERITE TYPE

Cordierite is a mineral typical of the contact-metamorphic zones; it is common in hornfels but does not ordinarily develop in limestone. In certain kinds of copper deposits, which occur only in regions of strong metamorphism and very deeply buried rocks, we find the association of cordierite, anthophyllite, or gedrite, quartz, chlorite, biotite, and mus-

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N. H. Magnusson, Persbergs Malmtrakt, *Bull.* 2, Dept. Comm., Stockholm, 1925.

G. Lindroth, *Geol. För. Förh.*, 48, 1926, pp. 357–409.

N. Sundius, *Tschermak's min. u. petr. Mitt.*, 38, 1925, pp. 175–194.

[1] F. R. Koeberlin, The Brewster iron-bearing district of New York, *Econ. Geol.*, 4, 1909, pp. 713–754.

R. J. Colony, Magnetite iron deposits of southeastern New York, N. Y. State Mus., *Bulls.* 249, 250, 1923.

[2] H. Schneiderhöhn, *Geol. Rundschau*, 14, 1923, pp. 60–68.

A. Cissarz, *Mitt.* Kaiser-Wilhelm Inst. Eisenforsch., 5, 1924, pp. 109–126.

E. Dannecker, *Neues Jahrbuch*, Beil. Bd. 53-*A*, 1925, pp. 197–208.

covite, in places chondrodite (also a magnesian mineral), with pyrite, pyrrhotite, chalcopyrite, more rarely other sulphides, sometimes magnetite. These deposits have been variously interpreted. In the second edition of this book some of them were referred to as "injected pyritic deposits," following Weinschenk and Bergeat. Although truly magmatic injections of pyritic ores occur, some of the examples mentioned belong to a different class. They have been formed by replacement, and have received important magmatic additions of magnesia, iron, copper, and sulphur. We know the origin of these deposits chiefly by the work of P. Eskola and P. Geijer. The former has demonstrated their existence at Orijärvi, Finland, in a notable publication; the latter has shown that many representatives of the class are found in Sweden. In the United States the type is very rare.

In brief, these deposits are pyrometasomatic in origin and have developed in slates, schists, or quartzite, not in limestone. They are of irregular, roughly tabular, or stocklike form. Many important copper mines have been working ores of this origin.

**Bavaria.**—At Bodenmais,[1] granite intersects gneisses. The ore deposits lie in cordierite gneiss. The ores contain pyrrhotite and pyrite, with some sphalerite rich in cadmium and galena rich in silver; the bodies lie in general parallel to the dip of the gneiss. The contact between ore and gneiss is sharp, though there are some disseminated sulphides in the surrounding rock. Many of the gangue minerals are rounded or corroded. Recent papers show widely varying opinions of origin.

**Sweden.**—The renowned copper deposit at Falun,[2] in Sweden, forms a huge inverted cone enclosed in gray quartzose and gneissoid rocks and extending to a depth of 1,200 feet. The ore-body is really composed of the same rock, impregnated to greater or less extent with pyrite, pyrrhotite, and chalcopyrite. The gangue minerals accompanying the ore are cordierite, magnetite, andalusite, spinel, and garnet. It is difficult to arrive at a definite conclusion regarding the origin of this deposit; at any rate it was formed at high temperature. According to Vogt, the total production of copper from Falun from 1300 to the present time is about 500,000 metric tons.

The copper deposit at Bersbo,[3] in Sweden, is of a similar type. The ores are quartzose and are embedded in gray fine-grained "granulite"

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[1] E. Weinschenk, *Zeitschr. prakt. Geol.*, 1900, pp. 65–71.

For different interpretations, see papers by Hegemann and Maucher and by Dessau, Annotated Bibliography of Economic Geology, VI (1): 243; and *idem*, VI (2): 214, both 1934.

[2] Hj. Sjögren, The Falun copper mine, *Guide exc.*, XI$^e$ Cong. geol. internat., Stockholm, 31, 1910.

A. E. Törnebohm, *Geol. För. Förh.*, 15, 1893, pp. 609–690.

[3] A. Bergeat, Die Erzlagerstätten, 2, 1904, p. 978.

N. Sundius, Åtvidabergstraktens geologi, Geol. Survey Sweden, ser. C, 306, 1921.

or "leptite" (p. 740). In thin section the ores show a texture resembling that of contact-metamorphosed schist and contain as gangue minerals quartz, cordierite, spinel, biotite, hornblende, and garnet. On the whole, the succession is magnetite (oldest), pyrite, pyrrhotite, sphalerite, and chalcopyrite.

More recently P. Geijer has contributed valuable data on this group. He has re-examined Falun and finds that it is practically a mass of contact-metasomatic quartzite and schist. At Riddarhyttan[1] are many examples of this type.

**Finland.**—At Orijärvi, P. Eskola[2] has described the very extensive contact metamorphism produced in a series of limestones and leptites and ore-bodies of pyritic copper ores replacing the latter. Much cordierite is present.

**United States.**—It is significant that no deposits of this type have been found in the Cordilleran region where contact metamorphism is very common but usually of a kind which developed relatively close to the surface. It seems that very high temperature and pressure are necessary for the formation of the cordierite-chalcopyrite deposits.

At Blue Hill, Maine,[3] copper ores have been worked intermittently, last during the World War. The Ellsworth schists (Cambrian?) have been intruded by granite. Near the contact irregular or tabular bodies of pyrite and chalcopyrite have developed in the chlorite-sericite schist. A considerable amount of quartz and cordierite accompanies the ores.

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[1] P. Geijer, Riddarhyttans Malmfält, *Bull.* 1, Dept. Comm., Stockholm, 1923.

[2] *Bull.* 40, Comm. géol. de Finlande, 1914.

[3] J. L. Gillson and R. M. Williams, *Econ. Geol.*, 24, 1929, pp. 182–194.

# CHAPTER XXIX

## REGIONALLY METAMORPHOSED SULPHIDE DEPOSITS

Rocks subjected to stress at moderate depths within the zone of fracture may rupture in closely spaced breaks, producing the appearance of a schistose structure. In such rocks no great chemical changes would occur, except perhaps by subsequent deposition along the tight fissures. At greater depth deformation may take place by granulation and recrystallization, accompanied by chemical changes that are effected by the aid of the scant rock moisture. Uralite and chlorite may form from pyroxene, the soda-lime feldspars may recrystallize to zoisite and albite, the quartz crystals may be crushed and elongated, new mica, particularly muscovite, may develop; also crystals of aluminum garnet. The chemical composition of the rock will, however, change but little; although the various transformations involve transportation of substance, this movement is not free, but is limited and hindered in all directions.

Under these circumstances it is improbable that processes of concentration could have much opportunity to assert themselves; the minute quantities of useful metals contained in the original rock could not easily assemble to form larger masses.

Nevertheless certain deposits may develop by recrystallization of the substances contained in the original rocks. Disseminated graphite deposits (p. 730) may result from the crystallization of carbonaceous matter. Talc and soapstone may form from basic rocks. The crystallization of garnet may lead to deposits which can be utilized for abrasive purposes. Aluminum silicates, like cyanite and sillimanite, when plentifully present, may be useful for refractory material.

In amphibolite schist small grains of chalcopyrite, often intergrown with pyrrhotite and magnetite, appear to be more common than in the primary igneous rock from which the schist was derived. If even the slightest and slowest circulation of water were established during the deformation, some concentration of chalcopyrite could well take place, as it does in fissures traversing similar rocks.

When the deformation takes place at higher temperatures a number of minerals are developed which are similar to or identical with those of contact metamorphism. It is often difficult, indeed, to draw the line between regional and igneous metamorphism, especially in intensely metamorphosed regions where intrusive masses are abundant. There is reason to believe that at temperatures of, say, several hundred degrees some of the rocks, particularly limestones, become permeable to the gaseous emanations of water and metallic compounds yielded by intrusive masses, and thus an opportunity is afforded for the introduction of new substances, which in places may become concentrated into ore deposits. To such a permeation in the deep zone, many of the most enigmatic ore deposits of the crystalline schists may owe their origin. These deposits would then differ in some respects from the ordinary contact-metamorphic ores, which have, as a rule, developed only close to intrusive contacts, in most cases also actually within the zone of fracture.

Dissemination of sulphides is a phenomenon often encountered in almost any area of crystalline schists. In the majority of occurrences pyrite, pyrrhotite, and chalcopyrite are prominent; the sulphides of lead and zinc are far less common. Such disseminations are particularly connected with amphibolitic or chloritic rocks. As indicated above, these ore minerals may have various modes of origin. In the first place, the dissemination may be caused by mineralization along both sides of a fissure, parallel with the schistosity—that is, by the formation of a "bedded vein." Such mineralization is later than metamorphism, and the metamorphic minerals will probably be found to be altered—sericitized, carbonatized, or more rarely silicified.

If, on the other hand, the sulphide minerals were contained in the rock previous to metamorphism, or if they were developed during that process, they will be found intergrown with the metamorphic minerals, such as amphibole, epidote, chlorite, garnet, and albite, and usually accompanied by some magnetite or ilmenite.

Larger pyritic masses of this kind are, in most cases, probably original products of magmatic concentration; or they may be old fissure veins or replacement veins which have been rendered unrecognizable by deformation; or, finally, they may be of contact-metamorphic origin.

Sparser disseminations, often following certain lines along the strike of the schist, are often called "fahlbands" (the German "fahl" meaning rusty brown and referring to the oxidized outcrops). Such fahlbands, first noted in Kongsberg, Norway,[1] where they enrich the silver veins, may be several miles long and vary in thickness between a fraction of a foot and several hundred feet. The enclosing rocks vary from gneiss to mica schist, diorite, and amphibolite. The ore minerals are pyrite, pyrrhotite, sphalerite, chalcopyrite, molybdenite, and sometimes cobalt minerals. They are often intergrown with amphibole or garnet. The fahlbands are rarely of economic importance, but many of them characteristically enrich intersecting veins, causing native silver and gold

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[1] C. A. Münster, Ref. in *Zeitschr. prakt. Geol.*, 1896, p. 93.
J. H. L. Vogt, *idem*, 1899, pp. 177–181.

as well as cobalt and nickel ores to appear at the intersections. This is probably only a special case of the general law that veins are enriched where they cut across belts of pyritic impregnation. Fahlbands rich in cobaltite, with pyrite, chalcopyrite, pyrrhotite, and molybdenite, were worked at Skutterud and Snarum, in the Modum parish, Norway, from 1776 to 1899. For a long time these deposits were among the principal sources of cobalt oxide, which is used to impart a deep blue color to glass and porcelain. According to the older literature quoted by Stelzner and Bergeat,[1] the fahlbands at Skutterud lie between gneiss or quartz schist and amphibolite. Other minerals mentioned are malacolite, anthophyllite, and rarely graphite and tourmaline. The ores were poor, containing, even when sorted, less than 1 per cent cobalt. A parallel belt at Snarum is said to be enclosed in amphibolite and contains more copper.

The fahlbands have been variously interpreted. At a time when the crystalline schists were generally considered as altered sediments, they were held to be sedimentary deposits. Dall and Kjerulf,[2] in 1880, held them to be impregnations related to gabbro intrusions. Vogt considered the gray gneiss of Kongsberg as a pressed granite and held that it had been impregnated with sulphides at the same time as the surrounding schists.

That the dissemination of sulphides in its present form is dependent upon dynamo-chemical metamorphism is clearly shown by the minerals with which the sulphides are now intergrown. Sulphide emanations from intrusive magmas at a considerable distance from their source do not usually crystallize with amphibole, pyroxene, and garnet, but rather with calcite, sericite, and quartz as gangue minerals. Still, the recrystallization under pressure does not necessarily explain the ultimate origin of the minerals, and it is probably hopeless to speculate on this subject until the metamorphic series at the location of typical fahlbands has been more carefully examined as to the original character of its rocks.

Somewhat similar fahlbands in amphibolite and gneiss are reported in the older literature from Schladming, in Styria, where they enrich intersecting cobalt-nickel veins, and from Les Challanches, in France, where similar relations exist.[3] Descriptions from both places show that the so-called fahlbands are in reality narrow veins accompanied by alteration of the wall rocks.[4]

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[1] Die Erzlagerstätten, 1, 1904, pp. 269–271.

[2] Die Geologie des südlichen und mittleren Norwegens, 1880.

[3] Stelzner and Bergeat, Die Erzlagerstätten, 1, 1904, pp. 268–269.

[4] C. Schmidt and J. H. Verloop (Schladming), *Zeitschr. prakt. Geol.*, 17, 1909, pp. 271–276.

T. A. Rickard (Challanches), *Trans.*, Am. Inst. Min. Eng., 24, 1894, pp. 689–705.

## THE PLASTIC DEFORMATION OF MINERALS

In rocks which have been exposed to strong pressure, practically all minerals have been crushed and deformed; even the hardest of minerals are not exempt if the pressure is strong enough. Naturally the deformation takes place most easily in the softer minerals; it is expressed by crushing, sheeting, and transformations of grains and crystals into lenticular masses. It may also find expression in translation gliding or in twinning. The latter mode is exemplified by the frequent twinning in calcite. The softer ore minerals are easily affected, particularly galena, sphalerite, and chalcopyrite.[1]

Galena deforms by translation along cube planes; and, by twinning, along other planes; evidences of deformation are often seen in lead deposits (p. 567). Sphalerite deforms by gliding on octahedral planes resulting in twinning. Chalcopyrite deforms by gliding on the pyramid and sometimes becomes twinned. Pyrrhotite also deforms in a plastic way. Pyrite resists vigorously and is usually crushed or sheeted.

Ore deposits contained in schists and gneisses usually show a more or less perfect alignment of the ore minerals with the schistose structure. This banded structure may have resulted from the dynamic metamorphism of a massive ore-body, but it is also possible that the ore minerals may have been introduced after the metamorphism and that the banded structure may have resulted from a preferential replacement of certain bands, or minerals, in the schist. It is often of importance to distinguish between these two cases. The subject has been well treated by Newhouse and Flaherty.[2] In the case of dynamic metamorphism of the ore, two generations of each mineral will generally be found, and there will be evidence of elongation and straining of the ore minerals.

It is clear that fracturing, crushing, and plastic deformation are not the only effects of pressure on sulphide ores. They are accompanied by more or less extensive recrystallization. Newly formed aggregates heal and cement the material; in part they are gangue, in part sulphides or similar compounds. It has been supposed that the recrystallization takes place by the aid of very small amounts of water or gas which permit solution and re-deposition of the original material. Many of the ores affected must, however, be nearly dry. For such conditions it is well to bear in mind the results obtained by metallographic research. In

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[1] M. J. Buerger, The plastic deformation of ore minerals, *Am. Mineralogist*, 13, 1928, pp. 1–7, 35–51 (with literature and account of older data); Translation gliding in crystals, *idem*, 15, 1930, pp. 45–64.

F. D. Adams, Differential pressure on minerals and rocks, *Jour. Geol.*, 18, 1910, pp. 489–525.

[2] W. H. Newhouse and G. F. Flaherty, The texture and origin of some banded or schistose sulphide ores, *Econ. Geol.*, 25, 1930, pp. 600–620 (with literature).

metals and alloys, recrystallization often takes place upon relatively slight heating in the absence of a fluid phase, so that it must be concluded that a limited rearrangement and migration of molecules can also occur within the solid phase itself.

The Mandy ore-body at Flin-Flon, Manitoba, although occurring in schist, shows the microscopic texture of a normal replacement ore.

The copper deposit at Rammelsberg,[1] in the Harz Mountains (p. 626), well illustrates the effects of local dynamo-metamorphism at no great depth. Under strong pressure the softer minerals, like galena, chalcopyrite, and sphalerite, are easily deformed along gliding planes and pressed out to plastic streaky masses. Pyrite, being harder, is crushed without plastic deformation and subsequently cemented.

Pyritic deposits, with chalcopyrite, forming lenses or bedlike masses in probably Paleozoic schists are found at several places in New England and Quebec. Among them are the Milan, New Hampshire, the Ely, Vermont, and the Eustis, Quebec. The ores appear to be later than the schistosity and replace the schist under the influence of Paleozoic intrusions. Some slight later deformation may be observed.

At the Milan mine,[2] cordierite and quartz are among the gangue minerals, but neither they nor the sulphides antedated the metamorphism of the surrounding schists.

## POST-MINERAL REACTIONS

It is well understood that dynamic metamorphism may change the structure, texture, and even the mineral composition of ore deposits. It is likewise well known that many changes take place under the influence of descending, oxidizing surface waters; these we call *supergene* processes, and they will be described in detail later. But there are other changes that may occur after the process of mineralization by ascending waters has ceased. Little attention has been given to these, but it is almost self-evident that complex ores may not always be in a state of equilibrium with the residual solutions which still linger in the deposit. The temperature has been reduced, but it may still be fairly elevated, say from 50° to 100° C. Under such conditions some of the minerals may prove unstable or may react together. Such minerals as marcasite and complex sulpho-salts may form. Waldschmidt[3] was probably the first to call attention to the development of boulangerite or jamesonite

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[1] W. Lindgren and J. D. Irving, *Econ. Geol.*, 6, 1911, pp. 303–313.

F. D. Adams and J. A. Bancroft, On the amount of friction developed in rocks during deformation, etc., *Jour. Geol.*, 25, 1917, pp. 597–637.

Paul Ramdohr, Ueber den Mineralbestand und die Struckturen der Erze des Rammelsbergs, *Neues Jahrbuch*, Beil. Bd. 57-*A*, 1928, pp. 1013–1068.

[2] W. H. Newhouse and G. F. Flaherty, *op. cit.*, p. 609.

[3] W. A. Waldschmidt, Deformation in ores, *Econ. Geol.*, 20, 1925, pp. 573–586.

in the ores from the Coeur d'Alene district as a distinctly post-mineral process. Later investigations (unpublished) have shown that in the deep levels of the Morning mine of that district tetrahedrite reacts with galena to form a lead sulphantimonide. Pressure seems to favor the reactions. Observations made by Stillwell[1] indicated that meneghinite forms from a similar reaction.

In the Porvenir silver mine at Huanuni, Bolivia, franckeite (a complex tin-lead sulphantimonide), also argentiferous jamesonite, marcasite, and pyrite develop from galena, sphalerite, arsenopyrite, and cylindrite, likewise apparently under pressure. To what extent descending waters may contribute to these reactions is not yet fully known. In the Morning mine at least, the process went on far beyond the reach of surface waters.

### THE ZINC ORES OF ÅMMEBERG, SWEDEN

One of the most mysterious of ore deposits is that of Åmmeberg,[2] in Sweden, where the zinc ore is disseminated in banded and contorted gray gneissoid "leptite." Some layers of gabbro, crystalline limestone, and lime-silicate rocks are intercalated in the steeply dipping leptite. Sphalerite, with very little galena, is widely disseminated in the leptite, seemingly taking the place of magnetite, and along certain zones has accumulated as long, lenticular folded bands, some of which are 30 to 50 feet in width and have been followed to depths of 1,000 feet. The ores are rich in sphalerite but contain few other minerals. A fahlband of disseminated pyrrhotite and arsenopyrite lies in the leptite of the footwall.

The deposit certainly seems to be of syngenetic origin and the mineral association indicates that it has been subjected to high temperatures.

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[1] F. L. Stillwell, Mineral constitution of the Broken Hill lode, *Proc.*, Austral. Inst. Min. Eng., 64, 1926, pp. 97–172.

[2] H. E. Johansson, The Åmmeberg zinc ore field, *Geol. För. Förh.*, 32, 1910, *Guide exc.*, XIème Congrès géol. internat., Stockholm, 35, 1910.

## CHAPTER XXX

# MINERAL DEPOSITS OF THE PEGMATITE DIKES

### INTRODUCTION

Each large intrusive mass is usually accompanied by a series of later dikes. These "complementary" dikes have, as a rule, a composition similar to that of the prevailing rock, but differ from it in showing an enrichment of certain constituents and a reduction of others. They are generally regarded as products of magmatic differentiation, forming residual parts of the dominant magma after its consolidation has begun. Some of them are basic, like kersantite, minette, or camptonite; others are acidic, like granite porphyry, aplite, or pegmatite.

Under the name of pegmatite are grouped the coarse granular dike rocks, often with well-developed idiomorphic texture, which accompany intrusive rocks, each group being characterized by pegmatites of special types.

Gabbros are sometimes accompanied by basic pegmatites of feldspar and pyroxene, and diorite by similar dikes of a basic feldspar and hornblende. The anorthosites are followed by pegmatitic dikes containing labradorite, hypersthene, and ilmenite; the nepheline syenites by pegmatites of soda feldspars, nephelite, sodalite, lepidomelane mica, ægirine, arfvedsonite, and minerals containing zirconium and titanium.

Most abundant are the granitic pegmatites, which consist mainly of coarsely crystallized orthoclase and quartz with muscovite; they often contain tourmaline, cassiterite, beryl, orthite, topaz, and a host of other rare minerals.

### MINERALIZERS AND THE NATURE OF THEIR ACTION[1]

The processes of intrusion and crystallization bring about an increasing concentration of the volatile constituents in the magma, if no avenue of escape is open to such substances. In subaerial eruptions they are given off into the atmosphere. These volatile substances, which of course formed an integral part of the original magma, consist of water and compounds of boron, fluorine, chlorine, phosphorus, sulphur, carbon, arsenic, and other rarer elements. They exert a peculiarly favorable

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[1] A. Harker, Natural history of igneous rocks, 1909, pp. 282–302.

J. F. Kemp, The pegmatites, *Econ. Geol.*, 19, 1924, pp. 697–723.

K. K. Landes, Origin and classification of pegmatites, *Am. Mineralogist*, 18, 1933, pp. 33–55; 95–103.

action upon the crystallization of magmas and minerals by decreasing their viscosity, lowering their freezing point, and furthering the development of minerals which otherwise do not crystallize from dry magmas. Harker says:

> The action is doubtless partly physical, partly chemical. The nature of the chemical effect, where the agent does not enter the crystallized product, is sometimes designated as a *catalytic* action, signifying a peculiar property possessed by certain bodies of inducing chemical changes in other bodies without themselves entering into the composition of the final product. In other instances the "mineralizer" forms part of the crystallized material.

In the acidic rocks, which are known to contain fluorine and boron, the action of mineralizers is particularly clear, but they are doubtless present also in basic rocks, in which chlorine, phosphorus, and sulphur take the place of fluorine and boron. More or less water is always present.

The presence of inclusions containing water in quartz crystals of acidic intrusive rocks shows plainly enough that the magma contained some water, as do also the rare transitions from pegmatite dikes to deep-seated ore-bearing veins.

The residual magma contains, besides these volatile mineralizers, the principal elements of the igneous rock crystallizing as quartz, feldspar, ferromagnesian minerals, and muscovite, and a number of rarer elements, such as tin, tungsten, zircon, tantalum, columbium, cerium, beryllium, molybdenum, lead, copper, lithium, and cesium. These rare elements appear to have been carried along in the process of differentiation by the mineralizers, which in many cases have also carried large quantities of iron differentiated from the main igneous body.

It is possible that these fluid extracts may have been pressed out of the semi-consolidated magma by mechanical action. The residual magma has a high vapor tension, and both factors facilitate its separation.

The mineralizing agents do not confine their action to the later stages of differentiation, but doubtless play a part in the crystallization of the main body of every magma. This is shown by the occurrence of molybdenite, pyrite, bismuthinite, sphalerite, titanite, and zeolites in the druses of granitic rocks; among such are the occurrences at Striegau, described by A. Schwantke,[1] and those in the syenitic rocks in the vicinity of Kristiania, mentioned by Goldschmidt.[2] The granites of the Island of Elba contain in druses such minerals as albite, tourmaline, beryl, garnet, pyrite, arsenopyrite, cassiterite, and zeolites.[3]

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[1] Drusenmineralien des Striegauer Granits, Leipzig, 1890.

[2] V. M. Goldschmidt, Die Contactmetamorphose im Kristiania Gebiet, Kristiania, 1911.

[3] G. Vom Rath, Die Insel Elba, *Zeitschr.* deutsch, geol. Gesell., 22, 1870, p. 466.

Many of the silicate minerals, formed by the aid of mineralizers as the last stage of intrusive action, are of remarkably complex chemical nature. To many minerals of this class no formula can confidently be assigned. Other minerals, especially the sulphides, are characteristically of simple formula and composition.

A distinct paragenesis or succession of minerals is noted in many pegmatites. With successively lower temperatures new sets of minerals were formed and many of those stable at a higher degree of heat became subject to alteration as the temperature became lower. Thus in the Norwegian pegmatite dikes, Brögger distinguishes three epochs of crystallization ending with the zeolites. Later investigators have emphasized this feature.

The residual magma is injected into the earlier consolidated portion and also into the encasing rocks; its fluidity and low melting-point are factors of great importance, allowing it completely to soak and penetrate schistose and fissile rocks encountered in its way. The pegmatites are essentially residual magmas, but they may become so admixed with water and dissolved gases that we may speak of them as in aqueo-igneous fusion at a temperature of around 575° C.

A praiseworthy attempt to carry this problem forward by the construction of diagrams showing the temperature of formation during magmatic and post-magmatic epochs has been made by Fersmann,[1] in regard to the pegmatites of Trans-baikalia and Chibina (Kola).

## OCCURRENCE AND GENERAL CHARACTER

The pegmatites form dikes, sheets, pipes, and irregular masses; where appearing as dikes or sheets no great regularity or extended continuation in depth can be counted upon, and this is important to consider in the exploitation of such bodies. Probably this irregularity is explained by the sudden and explosive action by which they make room for themselves and hold the cavities open until their substance is crystallized. Very different is this action from the slowly applied compressive stresses by which the fissures of most veins are opened.

The pegmatites are essentially coarsely crystalline rocks. Under some circumstances the dimensions of the crystals may be enormous. In the Ural Mountains a quarry was opened in a single orthoclase crystal; in India muscovite plates 10 feet in diameter have been found; at the Etta mine (Fig. 299), in the Black Hills of South Dakota, spodumene occurs in crystals resembling tree trunks and as much as 42 feet in length; quartz crystals several feet in length are not uncommon. At Albany, Maine, beryl crystals 18 feet long and 4 feet in diameter have been

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[1] A. Fersmann, Geochemische Diagramme, *Neues Jahrbuch,* Beil. Bd. 64, 1931, pp. 663–680. On classification of granite pegmatites, see same author, Tschermak's *min. u. petr., Mitt.,* 41 (1), 1931, pp. 64–94.

found. Feldspar and quartz often crystallize together as graphic granite, but in other pegmatites there is a distinct succession, with muscovite, for instance, at the walls and quartz and feldspar in the center, or with feldspar crystals along the walls and a central filling of quartz. The rarer minerals usually form the later generations and probably crystallized below 575° C., the crystallographic inversion point for quartz. The pegmatites are generally not eutectics. The rarer minerals are accessory, as a rule, for there are enormous masses of pegmatites which contain little but quartz and feldspar.[1]

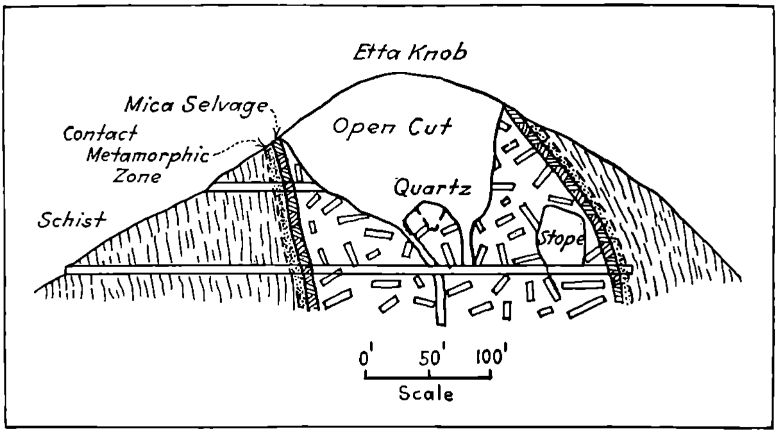

FIG. 299.—Cross-section of the Etta pegmatite dike showing crystals of spodumene. (*After G. M. Schwartz, Econ. Geol.*)

In their present condition there is little evidence of water as a constituent of their magma, but facts already referred to force us to the belief that some water was present as well as some carbon dioxide.

Liquid inclusions in pegmatitic quartz from Branchville, Connecticut,[2] were found to consist of 98.33 per cent $CO_2$, 1.67 per cent nitrogen, and traces of hydrogen sulphide, ammonia, fluorine, and possibly chlorine.

A marked contact-metamorphic action, sometimes stronger than that of the original magma, characterizes many pegmatites. H. B. Patton[3] describes, for instance, a pegmatite dike in Colorado which is 10 feet wide and which contains but little tourmaline, but which strongly impregnates the surrounding rock with this mineral for a distance of 2 or 3 feet from the contact. However, no sulphide impregnations analogous to the normal contact-metamorphic deposits have been found at the contacts of pegmatite and limestone. The quartz monzonites

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[1] E. S. Bastin, Origin of the pegmatites of Maine, *Jour. Geol.*, 18, 1910, pp. 297–320; *Bull.* 445, U. S. Geol. Survey, 1911.

[2] G. W. Hawes and A. W. Wright, *Am. Jour. Sci.*, 3d ser., 21, 1881, pp. 203 and 209.

[3] *Bull.* Geol. Soc. Am., 10, 1898, p. 21.

of the western states, along whose contacts most of the deposits mentioned occur, are rarely accompanied by pegmatitic dikes.

The pegmatites often absorb material from their walls, and near them minerals, otherwise foreign, are likely to appear; andalusite, garnet, and staurolite are among these minerals.

## TYPES OF PEGMATITES

**Acidic Pegmatites.**—The most common type consists of the granitic pegmatites, which always contain orthoclase, albite, and quartz, usually also microcline and muscovite. Among the accessory minerals magnetite, often in crystals, is perhaps most common. Other rarer minerals are tourmaline, topaz, fluorite, cassiterite, apatite, ilmenite, rutile, orthite, monazite, beryl, samarskite, spodumene, amblygonite, and many more. The typical mineralizers are boron and fluorine, together with a little phosphorus and sulphur. Lithium and the metals of the cerium and thorium groups are also characteristic. Among sulphides, molybdenite and bismuthinite are the most common, but pyrite, arsenopyrite, pyrrhotite, chalcopyrite, bornite, and sphalerite are occasionally found. In the south Norwegian granitic pegmatites, lithium and tin are absent.[1]

Schaller[2] holds that the development of pegmatites involves an original injection of quartz-orthoclase magma, as graphic granite, and a long series of high-temperature replacements during which the rarer and well-crystallized minerals were formed. At Pala, California, the gently dipping pegmatites are contained in granodiorite or in gabbro. The lower part of the dike is banded by quartz and albite, which upwards goes over into graphic granite. Albitization comprises the first stage of the replacements; then follow muscovitization and the development of tourmaline and garnet. Lastly, the lithium minerals replace all others, mainly along the middle seam of the dike; the solutions ascend through the middle and hotter part of the mass.

Hess[3] discusses the lithium pegmatites of the Black Hills independently and arrives at similar conclusions.

Lacroix[4] describes the pegmatites of Madagascar which contain many rare minerals such as columbates and tantalates of uranium; also danburite, gold, specularite, bismuthinite, etc.

The pegmatites of central Maine[5] contain original microcline, microperthite, and quartz, with subordinate black tourmaline, beryl, lepidolite,

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[1] W. C. Brögger, Die Mineralien der südnorwegischen Granit-pegmatit Gänge, *Videnskabsselskabets Skrifter*, Math.-naturw. Klasse, Kristiania, 1909, 6, 159 pp.

[2] W. T. Schaller, The genesis of lithium pegmatites, *Am. Jour. Sci.*, 5th ser., 10, 1925, pp. 269–279.

[3] F. L. Hess, Natural history of pegmatites, *Eng. Min. Jour.-Press*, Aug. 22, 1925.

[4] A. Lacroix, Les minéraux de Madagascar, Paris, 1922.

[5] K. K. Landes, The paragenesis of the granite pegmatites of central Maine, *Am. Mineralogist*, 10, 1925, pp. 355–411.

garnet, biotite, amblygonite, and arsenopyrite. Residual solutions moving up through the dikes replaced the original constituents and deposited (1) quartz, cleavelandite (variety of albite), lepidolite, beryl, gem tourmalines, spodumene, and apatite; (2) amblygonite and rare phosphates, rhodochrosite; (3) latest quartz, apatite, and cookeite (related to lepidolite).

**Syenitic Pegmatites.**—A second group is formed by the syenitic pegmatites, rich in alkalies and especially in sodium. These contain soda orthoclase, ægirine, acmite, arfvedsonite, biotite, nephelite, sodalite, låvenite, and a number of rare titanium and zirconium minerals, also fluosilicates. There is little or no quartz. The characteristic mineralizers are fluorine and chlorine. Brögger[1] distinguishes in south Norwegian syenite pegmatites four phases of crystallization. In the earliest phase he places feldspars, nephelite, sodalite, ægirine, lepidomelane, barkevikite, and magnetite, followed by fluorite, rosenbuschite, låvenite, and woehlerite (containing fluorine); by sodalite (containing chlorine); by helvite (containing sulphur); by löllingite (containing arsenic); and by homilite and melanocerite (containing boron). There is no tourmaline, topaz, or quartz.

The second phase consists in the filling of drusy cavities in part by destruction of the older minerals; these druse minerals consist of leucophane and fluorite, representing the fluorine group; of homilite and datolite, representing the boron group; and of the simple sulphides, such as molybdenite, sphalerite, and galena.

The third phase, at a considerably lower temperature but still probably above 100° C., comprises the zeolites, which are followed by a fourth phase of low-temperature carbonates and fluocarbonates.

Pegmatite pipes in the riebeckite granite of Quincy, Massachusetts, have been described by Warren and Palache.[2] A zone of a coarse granitic aggregate of quartz, feldspar, riebeckite, and ægirine graduates into a central mass of almost pure massive quartz, sometimes containing molybdenite, sphalerite, galena, and chalcopyrite and, in its miarolitic cavities, fluorite, octahedrite, ilmenite, and parisite, the last a fluocarbonate of the cerium metals.

**Basic Pegmatites.**—The basic pegmatites are less common. Boron and fluorine are not usually present, but phosphorus and chlorine, probably also sulphur, play important parts. Soda-lime feldspars, amphibole, pyroxene, quartz, apatite, rutile, scapolite, and brown mica are the most abundant minerals.

## ECONOMIC FEATURES OF PEGMATITE DIKES

The pegmatites, on one hand, contain many of the common minerals in exceptional size of grain and purity, and, on the other hand, they are a

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[1] W. C. Brögger, *Zeitschr. Kryst. Mineralog.*, 16, 1890.

[2] *Bull.* Geol. Soc. Am., 21, 1910, p. 784.

storehouse for a great number of the rarest minerals, many of which are not found elsewhere. These deposits are therefore of considerable economic importance and their valuable products are of manifold kind.

**Feldspar and Quartz.**[1]—The orthoclase and quartz of granitic pegmatites are mined or quarried at numerous places, particularly in Maine, Connecticut, and North Carolina, in Norway, and in many other countries. The total value of the quartz and feldspar obtained from pegmatite dikes in the United States amounts annually to several hundred thousand dollars. Crude feldspars bring about $6.50 per ton. Very fine crushing raises the cost greatly. The feldspars are used for pottery, for soft abrasives, etc. Potassium feldspars are preferred on account of the higher melting point; sodium feldspars are used for glazes. The quartz is also utilized as an abrasive, in paints, and for the coating of tarred roofing. A minor quantity of quartz is also cut as a semiprecious stone under the names rock crystal, smoky quartz, rose quartz, and rutilated quartz. In the Red Rose mine in the Black Hills unusually large masses of rose quartz are found.

**Mica.**[2]—White mica, more commonly known as muscovite, is also an important product of the granitic pegmatites. It occurs as irregularly disseminated bunches of foils, or "books," in pegmatite dikes, sometimes crystallizing along the walls. The mica-bearing pegmatites are worked in three belts in the North Carolina mountain region, where they break into the pre-Cambrian crystalline schists and gneisses. The dikes, which also carry orthoclase, perthite, oligoclase, and quartz, are of varying thickness and persistence, at some places lenticular, following the schistosity, at others cutting across the country rock (Fig. 300). Accessory minerals are biotite and several rare or gem minerals, among them beryl and its variety aquamarine. Some of the occurrences constitute transitions to quartz veins, which are assumed to have been formed by more distinctly aqueous solutions.

Muscovite is also mined in New Hampshire, Virginia, South Dakota, Alabama, and Georgia.

Muscovite owes its usefulness to its transparency, elasticity, great resistance to heat and weathering, and applicability as a non-conductor of electricity. The essential use of sheet mica is for electric insulation.

[1] *Information Circ.* 6044, U. S. Bur. Mines, 1927; also 6381, *idem,* 1930.

E. S. Bastin, Economic geology of the feldspar deposits of the United States, *Bull.* 420, U. S. Geol. Survey, 1910, 85 pp.

[2] D. B. Sterrett, Mica deposits of South Dakota, *Bull.* 380, U. S. Geol. Survey, 1909, pp. 382–397.

D. B. Sterrett, Mica deposits of North Carolina, *Bull.* 315, *idem,* 1907, pp. 400–422.

D. B. Sterrett, Mica deposits in the United States, *Bull.* 740, *idem,* 1923.

H. S. deSchmid, Mica, *Publ.* 118, Canada Dept. Mines, 1912.

Ground mica is chiefly used for roofing. There are many minor uses. Crystals of mica may be several feet in diameter, but this is exceptional, sheets of 1 foot in diameter being considered large. The largest on record are 12 feet in diameter.[1] Smaller sheets a few inches square find ready use and are split into thin lamellæ and cut into proper shapes for stove doors and for various electrical insulating purposes. The scrap from the trimming is often ground and compressed into mica board or "micanite" for use in insulating. It is also used for the manufacture of wall papers and roofing materials. The quality of mica is best judged by the transparency of sheets about 2 millimeters thick; it is graded as "wine" or "rum" or smoky and spotted mica, the latter being undesirable for insulation. The price paid for sheet mica varies greatly according to the size of the sheets. The average price ranges up to $1 per pound for sheets 2 by 3 inches; larger sheets are worth several dollars per pound. In 1929, the domestic output was 1,018 tons of sheet mica and 6,253 tons of scrap. Considerable quantities are imported, mainly from Canada and India.

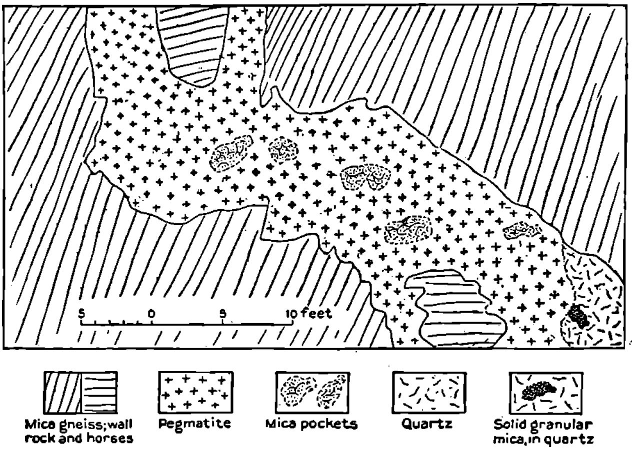

FIG. 300.—Vertical section across pegmatite dike, Thorn Mountain mine, North Carolina. (*After D. B. Sterrett, U. S. Geol. Survey.*)

The Indian mica, which is mined on a large but primitive scale, is generally muscovite contained in pegmatite dikes cutting gneissoid rocks. The Canadian mica, of which much is also exported, is mainly a phlogopite, or brown magnesium mica, and is better adapted to electrical uses than the muscovite. It occurs with apatite, in dikes or veins of pyroxene in gneiss or limestone, the principal localities being north of Ottawa. Associated minerals are calcite, scapolite, titanite, various metallic sulphides, among which molybdenite is mentioned, and one or two

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[1] A. L. Hall, Mica in eastern Transvaal, *Mem.* 13, Union of South Africa Geol. Survey, 1920, p. 21.

zeolites. These peculiar deposits are undoubtedly analogous to the Norwegian apatite-scapolite veins; in part they are certainly derived from limestone by contact metamorphism. Many of these dikes have been worked to a depth of several hundred feet. The quantity of trimmed mica obtained from the rock mined is small, often less than 1 per cent. Occasionally plates 5 feet in diameter are found.

**Oxide Ores.**—Specularite, magnetite, and ilmenite are of common occurrence in pegmatites, but scarcely ever of economic importance. Cassiterite, or oxide of tin, is also very common in many granitic pegmatites.

Pegmatites containing cassiterite, with phosphates and lithium minerals, have been mined near Gaffney, South Carolina, and about 50 tons of tin have been obtained from the detrital deposits.[1] The average tenor is low, but the mineral is concentrated along certain lines in the dike not unlike a shoot in a metalliferous vein. Tin-bearing pegmatites occur also in the Black Hills of South Dakota, where attempts to mine them have shown that they carried a very low percentage of the metal.[2] At Tinton, in the northern Black Hills, mining operations have been carried on and some cassiterite recovered. At the Etta or Harney Peak mine, in the southern Black Hills, the percentage of tin appears to be too small for successful recovery.

In the New England district in New South Wales, as described by E. C. Andrews and L. A. Cotton,[3] there are pipes of greisen with transitions into pegmatite containing cassiterite associated with wolframite, molybdenite, bismuth, arsenopyrite, tourmaline, fluorite, and beryl.

Remarkable rutile deposits have been discovered in Virginia, in Amherst and Nelson counties.[4] They are probably pegmatitic developments of pre-Cambrian magmas of gabbro or anorthosite, which, in other parts of the world, are also characterized by the concentration of titanium and phosphorus.

It is apparently a case where it is difficult to draw the line between ordinary rock differentiation and pegmatitization, but the features of the deposits clearly recall the latter process. The districts mentioned contain a predominant rock of quartz monzonite gneiss with an unusually

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[1] L. C. Graton, Gold and tin deposits of the southern Appalachians, *Bull.* 293, U. S. Geol. Survey, 1906.

[2] F. L. Hess, Tin, tungsten, and tantalum deposits of South Dakota, *Bull.* 380, U. S. Geol. Survey, 1909, pp. 131–163.

[3] E. C. Andrews, The geology of the New England Plateau, *Records*, Geol. Survey N. S. W., 8, pt. 2, 1905, pp. 131–136.

L. A. Cotton, The tin deposits of New England, *Proc.*, Linnean Soc. N. S. W., 34, pt. 4, Nov. 24, 1909.

[4] T. L. Watson and S. Taber, The Virginia rutile deposits, *Bull.* 3-A, Virginia Geol. Survey, 1913.

C. S. Ross, *Trans.*, Am. Geophys. Union, 1, 1934, p. 245.

large percentage of titanium and phosphorus. Besides, there are dikes of gabbro still richer in titanium. The pegmatitic facies consist essentially of a bluish quartz with plagioclase, orthoclase, and pyroxene, the last converted into hornblende with much rutile and accessory apatite and ilmenite. The rock has an even granular texture and contains as much as 59 per cent titanium dioxide and 12 per cent phosphoric pentoxide. Fluorine is present in quantities of 1 per cent or more and sulphur to the same amount, but there is very little chlorine. The rutile as well as the ilmenite is recovered by concentration and is used mainly for the manufacture of arc-lamp electrodes.

**Wolframite.**—As noted above, wolframite usually accompanies cassiterite in pegmatites, but only a small amount of the world's supply of tungsten is derived from these sources.

**Columbite and Tantalite.**[1]—These minerals are columbates and tantalates of iron and manganese ($(Fe, Mn)(Cb, Ta)_2O_6$). Their home is in the granitic pegmatites, from which the small quantities needed for incandescent lamps, electrodes and surgical instruments are derived. Large masses of columbite in black tabular crystals have been found in the pegmatites of the Black Hills, especially at the Etta mine. One ton of columbite was mined there in 1926. Mangano-tantalite, richer in tantalum, is mined from similar sources in Western Australia. Columbite is not uncommon in many regions characterized by pegmatite dikes, such as Connecticut and Virginia. Strüverite ($FeO.(Ta,Cb)_2O_5.4TiO_2$), isomorphous with rutile, has been found in abundance at the Etta mine, South Dakota and, in places, in the Federated Malay States.

Metallic tantalum is one of the most useful of the rare metals owing to its hardness, ductility, and resistance to acids. Its density is 16.9 and its melting point 2770° C. It is said to be more resistant to corrosion than any other known metal or alloy. It is used for many scientific and industrial purposes and has been called a rival of platinum.[2] Its price ranges from $90 to $160 per kilogram or about $4 to $7 per ounce.

**Yttrium, Thorium, Cerium, and Uranium Minerals.**[3]—Among the many rare earth minerals the following are the most important: thorite, ($ThSiO_4$); monazite [$(Ce, La, Y, Th) PO_4$]; gadolinite (beryllium-iron-yttrium silicate); allanite (cerium epidote); yttrialite (silicate of yttrium and thorium); euxenite (columbate and titanate of cerium metals and uranium); samarskite (columbate and tantalate of cerium metals and uranium). Some of these minerals, mainly monazite, are used extensively as a source of thorium salts in the manufacture of incandescent

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[1] F. L. Hess, *Bull.* 380, U. S. Geol. Survey, 1909, pp. 157–161.
*Information Circ.* 6348, U. S. Bur. Mines, 1930.

[2] *Eng. Min. Jour.-Press*, May 28, 1927; *Trans.*, Am. Inst. Min. Met. Eng., 71, 1925, pp. 696–697.

[3] *Information Circ.* 6321, Thorium, U. S. Bur. Mines, 1930.

mantles; the yttrium minerals, like fergusonite and gadolinite, were used in the manufacture of Nernst lamps in which the incandescent part consisted of 25 per cent yttria and 75 per cent zirconia. The cerium minerals have a limited use for chemicals, etc., as well as for the manufacture of ferrocerium, an alloy emitting sparks when rubbed against a hard substance. The Welsbach incandescent mantles are coated with a substance containing 60 per cent zirconia, 20 per cent yttria, and 20 per cent oxide of lanthanum.

At the present time incandescent mantles are said to contain 99 per cent $ThO_2$ and 1 per cent $CeCO_3$. The principal compound manufactured from the thorium minerals is thorium nitrate, which is worth $1 to $2 per pound. Most of the thorium nitrate used in the United States was imported from Germany where it was manufactured from Brazilian monazite. In 1929, 10,000 pounds were imported (see p. 249). In 1929, about 400 metric tons of monazite sands were concentrated from marine sands in India and Brazil. The small amount of cerium used is derived from monazite. Mesothorium, an element similar to radium and used for curative purposes and luminous paint, is also contained in thorium minerals and is extracted as a by-product.

All these minerals find their home in the granitic pegmatites and in the placers derived from them (p. 248). To some extent they are also primary constituents of igneous rocks. In Scandinavia there are some celebrated occurrences, like those of Hitterö, in southern Norway, and of Ytterby, Korarfvet, Brodbo, and Finbo, in Sweden. One of the most renowned localities in the United States is Baringer Hill, 100 miles northwest of Austin, Texas; few other localities have yielded as large amounts of rare-earth minerals as this place.[1]

Baringer Hill is a low mound, about 100 feet wide and 250 feet long, preserved from erosion by its relative hardness. The country rock is a coarse, porphyritic granite of pre-Cambrian age, and the dike itself an unsymmetrical body or pipe. At the edge of the dike is pegmatite of the "graphic" variety 1 to 6 feet wide. The central part is made up of large individuals of quartz and feldspar, the latter being microcline and albite. In the center of the dike the quartz appears to be concentrated. Some of the feldspar crystals are several feet long. Vugs are lined with smoky quartz. One smoky quartz crystal weighed 687 pounds. The rarer minerals, some of which occur in large amounts, are fluorite, ilmenite, gadolinite, allanite, fergusonite, and polycrase—in short, a

[1] W. E. Hidden and C. H. Warren, *Am. Jour. Sci.*, 4th ser., 22, 1906, p. 515.

F. L. Hess, Minerals of the rare-earth metals at Baringer Hill, Llano County, Texas, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 286–294.

K. L. Kithil, Monazite, thorium and mesothorium, *Tech. Paper* 110, U. S. Bur. Mines, 1915. See also *idem*, *Information Circ.* 6321, U. S. Bur. Mines, 1930.

S. J. Johnstone, The rare earth industry, London, 1918, 136 pp.

series of silicates, columbates, titanates, and uranates of cerium, yttrium, and other rare metals. There are also sulphides, particularly chalcopyrite, pyrite, sphalerite, and molybdenite, the last named being the most abundant. The rock contains no tourmaline, beryl, zircon, garnet, or cassiterite. The deposit is worked intermittently for the yttrium which its minerals contain. Some of the minerals show a marked radioactivity.

Uraninite ($UO_2$) in isometric crystals occurs in many pegmatites and has been mined for uranium and radium at Wilberforce, Ontario, and other places. Regarding radium, see page 415.

**Zircon.**[1]—Zircon ($ZrSiO_4$) is an accessory mineral of granitic and monzonitic rocks; it also occurs in pegmatites. It is absent from veins formed nearer the surface or under conditions of lessened temperature and pressure. The metal hafnium occurs in notable amounts in zirconium minerals.

Zircon occurs in considerable amounts in many placer deposits derived from the disintegration of granitic and pegmatitic rocks. In the miner's pan it is concentrated, with the gold, as a string of minute crystals of brilliant white, almost metallic luster. The best-known deposits are at Zirconia, near Green River, in Henderson County, North Carolina; from the decomposed outcrops of a pegmatite dike at this locality many tons of zircon have been obtained. Formerly zircon was concentrated from beach sand at San Pablo, Florida, where 624 tons were produced in 1925. The value of the concentrated zircon sand is about 7 cents per pound.

A zirconium mineral,[2] at first thought to be baddeleyite ($ZrO_2$), has been found in Brazil, in the Caldas district on the border of the states of Minas Geraes and São Paulo. It occurs in large quantities and is probably connected with pegmatite dikes in nephelite-syenite, and appears to consist of brazilite, a fibrous variety of $ZrO_2$, zircon and an unknown zirconium mineral with 75 per cent $ZrO_2$, possibly a silicate but not identical with zircon. In 1930, 1,500 tons of this material were imported.

The principal use of zircon is as an excellent refractory material; it has also a very low coefficient of expansion. Zirconia is also used for incandescent mantles and zirconium as a steel-hardening metal. Nickel-tungsten-zirconium (cooperite) is an alloy used for high-speed tools. The value of zircon varies: $115 to $140 per ton have been paid.

**Apatite.**—Apatite associated with pyroxene (malacolite), hornblende, phlogopite, titanite, and much calcite occurs in many deposits in southern

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[1] J. H. Pratt, Zircon, monazite, etc., *Bull.* 25, North Carolina Geol. Econ. Survey, 1916.

E. P. Youngman, Zirconium, *Information Circ.* 6456, U. S. Bur. Mines, 1931.

[2] W. T. Schaller, *Mineral Resources*, U. S. Geol. Survey, 1916, pp. 377–379.

Ontario, near Kingston.[1] The deposits are said to be dikes of pyroxenite with segregations of apatite and calcite intruded in gneiss and limestone of pre-Cambrian (Grenville) age. It seems certain now that some of these pyroxene rocks are contact-metamorphosed limestone (Fig. 294); but there are also dikes, though with what intrusions these dikes are associated is not certain. In part the apatite may be regarded as of pegmatitic origin. The mineral is well crystallized and of greenish color. In small quantities apatite also occurs here in veins in gneiss, and is then associated with pyroxene and scapolite. The output of Canadian apatite has ceased.

The Norwegian apatite veins,[2] described in detail by Vogt, now yield a small production. In some features they stand between the high-temperature veins and the pegmatites. The minerals are chlor-apatite, rutile, ilmenite, pyrrhotite, hornblende, enstatite, malacolite, and specularite. The feldspars in the rocks adjacent to the veins are altered to scapolite indicating the presence of chlorine in the emanations.

Very rich apatite deposits have recently been discovered in the alkaline syenites of the Kola Peninsula, Russia.

Apatite is used for the manufacture of phosphatic fertilizers, but comparatively little is now mined.

**Lithium Minerals.**[3]—Among the alkaline metals lithium accompanies potassium in the pegmatites and appears in a series of minerals, the most common of which are lepidolite, or lithium mica (4 per cent lithia); spodumene, a lithium-aluminum silicate allied to pyroxene (8 per cent lithia); petalite, lithium-aluminum disilicate (5 per cent lithia); triphylite, a lithium-iron-manganese phosphate (9 per cent lithia); and amblygonite, a fluophosphate of aluminum and lithium (10 per cent lithia). Spodumene and amblygonite are the principal raw materials from which lithia salts are manufactured. These minerals have been mined in the Ingersol, Peerless and Etta pegmatite dikes, in the Black Hills, South Dakota. In 1928, 4,600 tons of lithium minerals valued at $94,750 were mined. Other lithium pegmatites are found at Pala, San Diego County, California; in New Mexico, and in Manitoba. The use of lithium salts is small, chiefly for fireworks, medicine, and certain storage batteries.

At the Etta mine attempts have also been made to mine the pegmatite for tin and columbite. The Etta deposit is a roughly circular mass of coarse pegmatite about 150 by 200 feet in extent (Fig. 299). Spodumene crystals as much as 42 feet in length and having a cross-section

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[1] F. D. Adams and A. E. Barlow, Geology of the Haliburton and Bancroft areas, *Mem.* 6, Canada Geol. Survey, 1910, p. 383.

[2] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1895, pp. 367–370, 444–459, 465–740.

[3] *Information Circ.* 6215, U. S. Bur. Mines, 1930.

Lithium, The mineral industry of the British Empire, London, 1932, 27 pp.

of 3 by 6 feet are found here. Hess mentions a mass of amblygonite 27 feet in diameter. The list of primary minerals found at this remarkable locality is given below.[1] No topaz or axinite is present.

MINERALS FOUND AT THE ETTA MINE

Orthoclase
Albite
Microcline
Almandite
Grossularite
Andalusite
Muscovite
Biotite
Lepidolite
Petalite
Spodumene
Tourmaline
Epidote
Beryl
Titanite
Quartz
Zircon
Rutile
Spinel
Cassiterite
Corundum
Ilmenite
Columbite
Tantalite
Wolframite
Monazite
Amblygonite
Apatite
Triplite
Triphylite
Molybdenite
Arsenopyrite
Löllingite
Leucopyrite
Bismuth
Galena
Stannite

**Beryl.**[2]—The common greenish variety of beryl [$Be_3Al_2(SiO_3)_6$ with 14 per cent BeO] is found in small amounts in pegmatites all over the world. It has been mined in Maine (Albany, crystals up to 18 feet long); New Hampshire; at the Etta mine, South Dakota; Namaqualand, South Africa; and other places. Commercial ore should contain from 8 to 10 per cent BeO and brings about $50 per ton. Beryllium added to copper, aluminum and magnesium alloys greatly increases their strength and corrosion resistance. About 100 tons are absorbed by the markets per annum. The price of the metal is about $100 per pound. Concerning emerald, see page 766.

**Cryolite.**[3]—Cryolite ($3NaF.AlF_3$, with 12.8 per cent aluminum) is a white to brown or even black mineral of which only one large deposit is known. The locality is Ivigtut, in west Greenland, close to the sea, where it occurs as a large mass having surface dimensions of 160 by 600 feet; it has been worked to a depth of 150 feet. The cryolite occurs in a coarse granite enclosed in gneiss and is undoubtedly to be classed as an unusual pegmatite mass. The coarsely crystalline mineral is associated with siderite, argentiferous galena, auriferous chalcopyrite, pyrite, fluorite, topaz, and ivigtite; also druses of many rare fluorides. About 17,000 tons are produced annually, 7,718 long tons being imported into the United States in 1929.

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[1] F. L. Hess, *Bull.* 380, U. S. Geol. Survey, 1909, p. 149.

F. C. Lincoln, Pegmatite mining in the Black Hills, *Eng. Min. Jour.-Press*, June 18, 1927.

[2] *Information Circ.* 6190, U. S. Bur. Mines, 1929.

Beryllium, The mineral industry of the British Empire, London, 1931, 26 pp.

[3] N. V. Ussing, Denmark Geol. Undersög., 2d ser., No. 12, 1902, pp. 97–102.

R. Baldauf (and R. Beck), Ueber das Kryolith-Vorkommen in Grönland, *Zeitschr. prakt. Geol.*, 18, 1910, pp. 432–446.

S. H. Ball, The mineral resources of Greenland, Meddelelser om Grönland, 63, Copenhagen, 1922.

S. H. Gordon, Mining cryolite in Greenland, *Eng. Min. Jour.-Press*, Feb. 6, 1926.

A pegmatite mass adjoining the cryolite contains the same minerals and also cassiterite in a coarse-grained aggregate of microcline, albite, and quartz. It is thought by Gordon that the cryolite and the sulphides may be later replacements of an original pegmatite. The deposit is thus a most unusual magmatic concentration of fluorides.

Bauxite, the hydroxide of aluminum, is now used for the manufacture of the metal. Before the present methods of smelting aluminum were introduced the easily fusible cryolite was used for this purpose, and even now it is added to the charge to promote the melting. It is also used for the manufacture of soda; for enameling iron ware, and for making white Portland cement.

**Precious Stones.**[1]—The pegmatite dikes have always been famous as the source of gem minerals, which are valued for ornaments on account of their color, hardness, and brilliancy. Many of these beautiful crystals appear to belong to one of the later magmatic stages of consolidation and usually occur in druses of the rock. Among the most productive American regions are North Carolina, Maine, and San Diego County, California. The pink tourmalines of Pala,[2] and other places in San Diego County are famous. Accompanying this mineral are hiddenite and kunzite, the green and lilac-colored gem varieties of spodumene.

Green tourmaline comes from Maine; emerald and aquamarine ($Be_3Al_2(SiO_3)_6$), both varieties of beryl, are found in pegmatites in North Carolina, accompanied by quartz, albite, and tourmaline. Aquamarine of gem quality with much greenish beryl is found in pegmatite quartz of southern New Hampshire. The emeralds of Colombia,[3] at Muso, occur in carbonaceous Cretaceous limestone, in calcite veins which seem to belong to the high-temperature veins rather than to the pegmatites. The emeralds found near Ekaterinburg in the Ural Mountains are said by Fersmann[4] to occur in granitic pegmatites injected in biotitic and magnesian schists. An exchange of elements has taken place between these two rocks. The chromium coloring the green emeralds was derived from the basic magnesian rocks.

Rubies are also found in pegmatite dikes. Most of the supply is derived from a pre-Cambrian crystalline limestone of Mogok, in Burma,[5]

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[1] G. F. Kunz, The production of precious stones, *Mineral Industry*, annual issues.

S. H. Ball, The geologic and geographic distribution of precious stones, *Econ. Geol.*, 17, 1922, pp. 575–602. An excellent summary. See also *idem*, 26, 1931, pp. 681–738.

E. H. Kraus and E. L. Holden, Gems and gem materials, New York, 1931, 260 pp.

[2] W. T. Schaller, *Bull.* Univ. Cal. 3, 1903, p. 266.

[3] J. E. Pogue, *Trans.*, Am. Inst. Min. Eng., 55, 1917, pp. 910–934.

[4] A. Fersmann, Geochemische Migration der Elemente, *Abh. prakt. Geol.* (G. Berg), 18, 1929, 116 pp.; *idem*, 19, 1929, 86 pp.

[5] F. D. Adams, A visit to the gem districts of Ceylon and Burma, *Bull.* Canadian Inst. Min. Met., February, 1926.

### LIST OF GEM MINERALS BY ORIGIN

(*After S. H. Ball*)

| Pegmatite | Basic and alkaline igneous rocks | Contact-metamorphic |
|---|---|---|
| Tourmaline | Obsidian | Ruby |
| Beryl | Sodalite | Spinel |
| Emerald | Hauynite | Tourmaline |
| Quartz | Zircon | Grossularite |
| Spodumene | Sapphire | Vesuvianite |
| Topaz | Ruby | Sphene |
| Moonstone | Diamond | Epidote |
| Anatase | Pyrope | Willemite |
| Rutile | Olivine | Apatite |
| Cordierite | (Peridot) | Cordierite |
| Phenacite | Labradorite | Sapphire |
| Axinite | Enstatite | |
| Spessartite | | |
| Apatite | | |
| Cyanite | | |
| Danburite | | |
| Sphene | | |
| Sapphire | | |
| Ruby | | |
| Diopside | | |
| Zircon | | |

| Regional-metamorphic | Ascending waters (in part magmatic) | Descending waters |
|---|---|---|
| Nephrite | Opal | Turquoise |
| Jadeite | Variscite | Chrysoprase |
| Cyanite | Agate | Chrysocolla |
| Staurolite | Chalcedony | Malachite |
| Diopside | Jasper | Azurite |
| Andalusite | Amethyst | Dioptase |
| Almandine | Quartz | Smithsonite |
| Crocidolite | Gold-quartz | |
| (Tiger eye) | Thompsonite | |
| Sillimanite | Prehnite | |
| | Datolite | |
| | Rhodonite | |
| | Benitoite | |
| | Uvarovite | |
| | Chrysoprase | |

where also pink spinels and blue sapphire occur, but some good stones have been obtained from gravels near pegmatite dikes in Cowee Valley, North Carolina.

Topaz is found in pegmatite as well as in lithophysae of rhyolite and in high-temperature veins. Quartz of the clear, smoky, and rutilated varieties is common in pegmatites.

Ball has presented many interesting facts as to the value and distribution of precious stones. The world's production of rough stones has an annual value of about $81,000,000, of which the diamond (p. 777) accounts for 94.3 per cent; sapphire, amber (fossil resin), emerald, ruby, jadeite, turquoise, and opal together, another 4.6 per cent. Thus the value of the subordinate stones is only about $1,000,000. The sources of all gems, excepting the diamond, are as follows: The pegmatite dikes furnish precious stones to a value of about $700,000; other igneous rocks, $85,000; metamorphic and metasomatic rocks, $282,000; deposited by water (opal, turquoise, etc.), $595,000; fossil material (amber, etc.) $825,000; gravels and sands (secondary deposits) $2,096,500; all these items making $4,583,500.

**Native Metals, Sulphides, and Arsenides.**—The sulphides and allied minerals so abundant in fissure veins play a very subordinate part in the pegmatite dikes; nevertheless their occurrence is of great scientific interest, for the pegmatites form a transition between the magmas and many ore deposits connected with igneous rocks.

Gold in visible form is rare. Spurr states that gold occurs in pegmatite dikes or in quartz veins closely connected with them in the Yukon districts in Alaska and at Silver Peak, Nevada, but the quantities reported are, as a rule, small—at most 0.05 ounce per ton.

A. Lacroix[1] mentions several cases of gold in pegmatite from Madagascar. B. S. Butler[2] describes a pipe-like deposit from near Gold Hill, Utah, which contains enough gold to be worked. It carries quartz, microcline, actinolite, tourmaline, chalcopyrite, bismuth, scheelite, and molybdenite, also apatite and sphene. Evidently this is related to pegmatite.

A. Knopf[3] describes similar quartz-microcline dikes or veins from near Rochester, Nevada, which have yielded some ore containing native gold.

A very interesting occurrence is that of the Natas mine in Southwest Africa, investigated by E. Reuning.[4] An apex of a granitic stock is accompanied by pegmatite dikes which in places contain some gold,

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[1] Les minéraux de Madagascar, Paris, 1922.

[2] *Prof. Paper* 111, U. S. Geol. Survey, 1920, p. 476.

[3] *Bull.* 762, U. S. Geol. Survey, 1924, p. 52.

[4] E. Reuning, Natas mine, Southwest Africa, *Neues Jahrbuch*, Beil. Bd. 52, 1925, pp. 192–264.

chalcopyrite, and scheelite. In the same vicinity are quartz veins with more gold, in which the feldspars are generally absent. Reuning emphasized the difference existing between the quartz veins and the pegmatites though realizing their close relationship.

In 1898, E. Hussak[1] described the Passagem lode, in Brazil, and regarded it as a gold-bearing pegmatite dike. Orville A. Derby[2] has reviewed the evidence and arrived at the following conclusions, which are quoted because they very likely apply to many similar occurrences:

> The Passagem lode presents evidence of three distinct processes of filling . . . (1) An extensive fissure opened by stress . . . was closed by an invasion of pegmatite running off into clear quartz. At this stage the lode contained only the minerals characteristic of a granitic magma . . . (2) A subsequent stress . . . fractured this pegmatitic quartz, rendering it accessible to a pneumatolytic action . . . which filled its fissures with tourmaline and sericitized the feldspar of the pegmatite. (3) A third stress coming near the end of the second phase of the lode fractured the tourmaline filling and gave access to a pneumatolytic action characterized by sulphur, arsenic, metallic oxides, and metals (gold and silver), which filled the fissures of the lode, invading to some extent its pre-existing portions and probably also some of the adjacent and enclosed country rock.

Arsenopyrite, löllingite, galena, sphalerite, pyrite, and pyrrhotite have been reported from numerous localities in the granitic and syenitic pegmatites; there is not the slightest reason to doubt that they are here primary minerals, even if they belong to one of the later phases of magmatic consolidation. There are, however, no deposits known in which they are abundant enough to be mined.

Bismuth and bismuthinite are reported from many places, and are said to be so abundant in certain pipes of pegmatite in the New England district of New South Wales as to have some economic value. The occurrence of bismuth is interesting owing to its low melting point (271° C.).

**Molybdenite.**[3]—Molybdenite is an accessory mineral in certain granites, pegmatites, and aplites; it is usually of late origin, connected with quartz veinlets and sericitization. It is common in many veins of the deep-seated class, more or less closely connected with pegmatites. It is also of frequent occurrence in contact-metamorphic deposits and in ordinary fissure veins, both in those formed at great depth and in those deposited near the surface, but in the main is confined to deposits genetically allied to acidic igneous rocks. In the pegmatites and the

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[1] E. Hussak, *Zeitschr. prakt. Geol.*, 1898, pp. 345–357.

[2] *Am. Jour. Sci.*, 4th ser., 32, 1911, pp. 185–190.

[3] V. L. Eardley-Wilmot, Molybdenum, *Rept.* 592, Canada Dept. Mines, Mines Branch, 1925, 292 pp.

A. V. Petar, Molybdenum, *Econ. Paper* 15, U. S. Bur. Mines, 1932 (literature).

abyssal veins the individual particles of molybdenite are often large and sometimes well crystallized. In the deposits formed under conditions of less intense heat and pressure molybdenite usually appears as small or microscopic scales.

In Quebec, Canada, some pegmatitic dikes contain enough molybdenite to be of economic importance (Fig. 301). In Ontario, near Kingston,[1] molybdenite occurs with pyrrhotite in contact-metamorphic or pegmatitic deposits, similar to those which carry pyroxene, apatite, and brown mica (Fig. 293).

Pegmatites containing molybdenite occur in Washington and Hancock counties, Maine.[2] Some of them may have economic importance. One of the principal deposits occurs at Cooper, 22 miles southwest of Calais, associated with the more quartzose phases of the pegmatite.

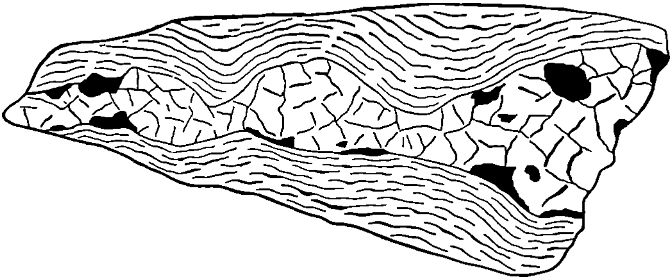

FIG. 301.—Molybdenite (black) along the borders of pegmatite dike in gneiss, Romaine, Quebec. One-half natural size. (*After T. L. Walker.*)

Similar deposits are found in many parts of the United States but difficulties of concentration long confined the product to hand-picked material. By aid of the flotation process, the production of molybdenite on a large scale has become of great importance. Formerly, most of the product came from Ontario and British Columbia; in 1916, 84 tons of pure sulphide were produced from concentrates containing 60 to 70 per cent of molybdenite.

The world's greatest supply is now obtained from Climax, Summit County, Colorado.[3] In 1930, about 1,000 tons of ore were milled daily. The deposit is contained in pre-Cambrian granite. Around a central silicified core is a zone up to 300 feet thick, which contains veinlets of quartz, orthoclase, and marginal grains of molybdenite. The grade of ore is about 0.5 per cent Mo. Another deposit of considerable importance is worked at Questa, Taos County, New Mexico. In 1930, the domestic production of molybdenum, in ores and concentrates,

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[1] E. Thomson, A pegmatitic origin for molybdenite ores, *Econ. Geol.*, 13, 1918, pp. 302–313.

[2] Geo. O. Smith, *Bull.* 260, U. S. Geol. Survey, 1905, pp. 197–199.

[3] B. S. Butler and J. W. Vanderwilt, The Climax molybdenum deposit of Colorado, *Proc.*, Colorado Sci. Soc., 12(10), 1931, pp. 309–353.

amounted to 3,759,269 pounds valued at $2,068,000. New South Wales, Queensland, and Norway also yield molybdenite.

Wulfenite ($PbMoO_4$), a yellow tabular mineral, is not uncommon in many oxidized deposits, the primary ore of which contained galena and molybdenite. Some of the output of molybdenum in the United States has been derived from this mineral, though it is less desirable than molybdenite.

Aside from many chemical and industrial uses, the principal value of molybdenum lies in its steel-hardening qualities. Molybdenum steel, containing a very small amount of the metal, is used for rifle barrels, propeller shafts, and especially for high-speed steel-cutting tools.

Ferromolybdenum with up to 80 per cent Mo is made in the electric furnace from molybdenite. The melting point of ferromolybdenum is very high, 2500° C. Usually the molybdenum is introduced in the steel in the form of calcium molybdate, which is made by roasting $MoS_2$ and combining the oxide with lime. The price of the $MoS_2$ in 1931 was from 40 to 45 cents per pound; of calcium molybdate 85 cents per pound; of ferromolybdenum $1 per pound. Mainly from the Climax deposit, there were produced in 1931 about 1,150 tons of molybdenite concentrates (about 80 per cent $MoS_2$). About 300 tons of such concentrates were produced in Norway.

# CHAPTER XXXI

# MINERAL DEPOSITS FORMED BY CONCENTRATION IN MOLTEN MAGMAS

## GENERAL FEATURES[1]

Certain kinds of mineral deposits form integral parts of igneous rock masses and permit the inference that they have originated, in their present form, by processes of differentiation and cooling in molten magmas. The minerals are of simple composition and few in number; most prominent among them are magnetite, ilmenite, spinel minerals, cassiterite, pyrrhotite, pentlandite, chalcopyrite, molybdenite, löllingite ($FeAs_2$), arsenopyrite, corundum, platinum, and diamond. At some places the resulting deposits are large and rich, but as a whole they are of much less importance than those formed by aqueous solutions.

The characteristic feature of a deposit of this class is that it is a part of a body of igneous rock; the crystals of its minerals formed in the magma solution from which the rock crystallized, or in one similar to it. The associated gangue minerals are those which make up igneous rocks. Structures other than those of purely igneous origin should be absent. If there is evidence of metamorphism or metasomatic replacement, with the development of minerals like sericite, carbonates, chlorite, uralite, garnets,[2] or epidote, or bleaching or kaolinization, we must conclude that processes other than those of purely igneous origin have been active. Many igneous deposits have, at a later period, been subjected to influences producing alteration and their original characteristics may then have become veiled.

Some igneous deposits are simply parts of the rock, which contains disseminations of the useful mineral, like diamond in certain peridotites, and have then the form of that rock mass itself—a dike or a volcanic neck, for instance. In other deposits the massive ore forms a dike, as in certain titaniferous magnetites. Or again, the ore minerals may have become concentrated in parts of the igneous rock and form rudely tabular or wholly irregular, usually ill-defined, masses in the rock. Unless the deposit is of large cross-section it can rarely be followed to great depth like a fissure vein, for the movements in a viscous magma facilitated the

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[1] J. H. L. Vogt, Magmas and igneous ore deposits, *Econ. Geol.* 21, 1926, pp. 207–233, 309–332, 469–497.

[2] Garnets are, however, also occasionally found in normal igneous rocks, but their primary or secondary nature is not difficult to establish.

formation of irregular streaky or pasty masses—often termed "schlieren," after the German usage—rather than bodies persistent for long distances in a given direction.

These deposits represent extreme conditions in mineral formation; the temperature of basic surface lavas is considered to have been about 1000° to 1250° C., but the deep-seated granular rocks in which most of the igneous mineral bodies occur crystallize more slowly than lavas and in general at lower temperatures, probably from about 575° to 1000° C. The temperature before emission and consolidation may have been hundreds of degrees higher than the various figures given above.

In order to understand these deposits, which are often designated as "liquid magmatic," and in order to distinguish them from other deposits of magmatic affiliations, it is necessary to take up first the law that governs the constitution, crystallization, and differentiation of magmas. For this purpose the reader is directed to Chapter X of this book.

We may recall the fact that mineral deposits of this type may be formed by the simple crystallization of a magma rich in some valuable constituent, as corundum in some syenites or diamonds in peridotite. The ore mineral will then be scattered in the rock mass from which it may be recovered by some process of concentration. On the other hand, differentiation by crystallization involves the early separating of minerals like magnetite, chromite, or ilmenite, the sinking of these crystals in the liquid magma and, consequently, their accumulation in the lower part of the magma body. Uplift and erosion may bring such masses to the surface, or they may be re-melted and injected as dikes of almost solid ore.

Finally, the magma may, at a high temperature, carry in solution a considerable amount of sulphides of iron, copper, and nickel; when the temperature drops part of the sulphides becomes immiscible with the silicate magma, and may sink to the lower part of the mass, coalescing to drops and accumulating to larger masses, which will remain liquid till all of the rock constituents have separated out. Then, again, these sulphides may become exposed at the surface by uplift or erosion or, they may conceivably be injected as liquid dike-like masses into cooler rocks.

Deposits formed by processes like these have been called *orthotectic*, by Graton and McLaughlin.[1]

The conditions are not always as simple as outlined, for there are transitions to hypothermal deposits. The ore minerals are no longer accompanied only by minerals of the igneous rock, but hydrothermal products begin to appear, such as sericite, epidote, etc. The oxide or sulphide melt contains mineralizers which will more or less deeply modify

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[1] *Econ. Geol.*, 13, 1918, p. 85.

the strictly magmatic influences. Such transitions are observed at the Sudbury nickel deposits and in the New Jersey magnetite deposits. For such transitional forms the authors mentioned suggested the term *pneumotectic*.

## PRINCIPAL TYPES OF DEPOSITS

Among the valuable minerals formed during the consolidation of magmas are diamond, platinum, chromite, ilmenite, magnetite, corundum, cassiterite, pyrrhotite, pentlandite, pyrite, chalcopyrite, molybdenite, sperrylite, and apatite. A much more complex series of minerals is contained in the pegmatite dikes, which are described separately. For each kind of rock certain minerals are characteristic, and most of the rocks are of the deep-seated type, crystallizing with granular structure.

Diamonds, chromite, platinum, and sometimes corundum are associated with peridotites; corundum also with certain nepheline syenites. Chalcopyrite, pyrite, pentlandite, and pyrrhotite follow the basic rocks, especially gabbros. Apatite and magnetite are connected with alkali-rich syenites; ilmenite and titaniferous magnetites with anorthosites (labradorite rocks) and gabbros; cassiterite with granite.

## DIAMONDS[1]

Diamond is pure carbon, crystallizing in the isometric system. It is the hardest of all minerals and has a specific gravity of 3.52. Usually it is white or yellowish, but other pale colors are not uncommon. Rounded gray forms with aggregate structure are called bort and occur in the Premier and other South African deep mines. The dark-gray or black carbonado, found in placers in Bahia State, Brazil, is granular and without visible cleavage. Its specific gravity is less than that of the

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[1] Gardner F. Williams, The genesis of the diamond, *Trans.*, Am. Inst. Min. Eng., 35, 1905, p. 440.

A. von Fersmann and V. C. Goldschmidt, Der Diamant, Heidelberg, 1911. A fundamental and important work.

C. Doelter, Handbuch der Mineralchemie, Dresden, 1, 1911, pp. 28–56.

Percy A. Wagner, The diamond fields of southern Africa, Johannesburg, 1914.

Percy A. Wagner and L. Reinecke, Mineral deposits of the Union of South Africa, Third Empire Min. and Met. Congress, Johannesburg, 1930, 310 pp. An excellent review.

A. F. Williams, The genesis of the diamond, 2 vols. London, 1932.

R. A. F. Penrose Jr., The Premier diamond mine, Transvaal, *Econ. Geol.*, 2, 1907, pp. 275–284.

W. N. P. Reed, The carbonado, *Eng. and Min. Jour.*, 1930, pp. 289–293.

diamond. Both varieties are used for drilling on account of their great hardness. About 1,000,000 carats of bort and 25,000 carats of carbonado is the world's present annual output of these abrasives. Diamond powder is used as an abrasive.

Diamonds are mined in placer deposits of detrital origin (p. 250), and this was the only source before 1871. Since that time one-half or more of the world's production has been obtained from primary deposits which consist of intrusions, usually pipes, of basic igneous rocks, more or less altered, belonging to the peridotite family. Practically the whole of this production comes from the Union of South Africa. It is estimated that since 1883, about 126,000,000 carats have been obtained from the placers and deep mines of the Union. The placer diamonds are of a better quality than those in the pipes.

The only American occurrence of note is near Murfreesboro, Pike County, Arkansas,[1] where stones of good quality are found in decomposed peridotite and breccias of late Cretaceous age, very similar to the South African occurrences. From 1906 to 1920, a total of 5,300 stones, averaging 0.4 carat, had been found. The largest stone found weighed 20¼ carats.

In 1871 the Kimberley diamond field in South Africa was discovered, the first known occurrence of diamonds in primary rock. The district lies in the northern part of Cape Colony and the adjacent part of the Orange Free State. Another district centers at Jägerfontein, in the Orange Free State; still another at the Premier mine, near Pretoria, in the Transvaal.

The diamonds in the South African field are disseminated in volcanic necks, commonly called "pipes," of "kimberlite," a serpentinized rock derived from peridotite. In places the pipes are connected with dikes of similar material. The age is Cretaceous and the intrusions have been deeply truncated by erosion. The kimberlite, first recognized in 1886 by Carvil Lewis, has been the subject of much investigation. The best description is that by Percy A. Wagner in the work noted. The kimberlite is a peridotite, often tuffaceous or brecciated, always more or less serpentinized and of a dull, greasy appearance. There are two varieties, one the so-called basaltic type; the other, a lamprophyric variety rich in phlogopite. The rocks are porphyritic; the ground mass is greatly altered but may contain olivine, apatite, perovskite, ilmenite, apatite, chromite, with grains of enstatite, garnet, and diopside. Calcite, iddingsite, serpentine, apophyllite, and chlorite are secondary minerals. An analysis of the Premier "blue ground" gave, in per cent: $SiO_2$, 38.29;

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[1] G. F. Kunz and H. S. Washington, Diamonds in Arkansas, *Trans.*, Am. Inst. Min. Eng., 39, 1908, pp. 169–176.

H. D. Miser and C. S. Ross, *Econ Geol.*, 17, 1922, pp. 662–674; *Bull.* 735, U. S. Geol. Survey, 1923, pp. 279–322.

$TiO_2$, 2.00; $Al_2O_3$, 2.66; $Fe_2O_3$, 5.77; FeO, 2.93; CaO, 2.42; MgO, 29.46; $K_2O$, 1.03; $Na_2O$, 0.30; $H_2O-$, 3.13; $H_2O+$, 10.19; $CO_2$, 0.02; and $P_2O_5$, 1.44; total, 99.82. Analyst, M. Dittrich. This rock breaks through the horizontal quartzitic sandstones, volcanic flows, and shales of the Karroo formation (Carboniferous to Triassic).

The Kimberley pipe has been worked down to a depth of about 3,500 feet, its diameter being about 500 feet. Near the surface the kimberlite was decomposed ("yellow ground") but at greater depth the fresher "blue ground" was met. The latter is now mined exclusively; and the tough rock is allowed to slack on the surface for many months before it can be washed in pans and jigs, the final concentration being effected on greased tables, the grease having the property of holding the diamonds while the other constituents of the rock are washed over it. Successive crushing is now often substituted for the slacking and treatment in pans. The pipes are probably to be regarded as explosion vents and the rock filling them is undoubtedly of igneous origin. There are hundreds of these vents scattered from Cape Colony to the Congo and many more are probably concealed by the covering deposits of the Kalahari Desert. Only a few of these pipes carry diamonds in economic quantities. The diamonds are often crystallized as octahedrons, with convex or concave faces, but most of the crystals as recovered are broken. The color is white, yellowish, greenish, lilac, and even deep yellow. Small rounded masses with concentric growth have also been noted and many dark-gray semi-transparent pieces are found.

The "blue ground" contains fragments of carbonaceous shale, which at first were thought to have furnished the material for the diamonds. Stelzner, Bonney, and others have shown that the gem crystallized as an integral part of the magma. Stelzner mentions intergrowths of pyrope garnet and diamonds, and Gardner F. Williams has collected specimens from Kimberley which show crystallized diamonds still partly enclosed by garnet and ferromagnesian minerals. Trübenbach and Bonney have also found that diamonds actually occur in fresh eclogite, a garnet-pyroxene rock closely allied to peridotite, from the Newlands mine, about 40 miles west of Kimberley. T. W. E. David described a ⅓-carat diamond from New South Wales, embedded in a solid matrix of hornblende diabase.

The production at Kimberley by the De Beers Company is about 1,000,000 carats per annum; the average price for the rough stone ranges from $16 to $26 per carat. The best ground is said to average about 1 metric carat (200 milligrams) per ton.

The deposit at the Premier mine, discovered about 1900, is a pipe ½ mile by ¼ mile in horizontal section penetrating the older Pretoria series of sediments. The mine is 325 feet deep. About 2,800,000 tons of material are treated annually, yielding about 733,000 carats. In this mine the largest diamond ever found was obtained; it is known as the Cullinan diamond and formed a broken octahedron 4 inches long and 2 inches wide; its weight was 3,025¾ carats.

Much of the product from the South African mines is sold in the United States, the imports reaching a value of about $60,000,000 a year. Good cut stones sell from $200 to $500 per carat.

Production, and imports into the United States, dropped sharply after 1929. The period of depression since 1929 has had a disastrous effect on the diamond market.

The world's production of diamonds for 1929 was estimated at about 7,150,000 metric carats, of which 3,661,000 came from the Union of South Africa, over 1,800,000 from the Congo, 860,000 from Angola and Southwest Africa, and 214,000 from British Guiana. In the three latter regions the gems occur in placer deposits (p. 250).

A government control board now regulates the production of diamonds from alluvials in the Union of South Africa. A private corporation controls the output of the deep mines to avoid overproduction.

Carbon is soluble in molten magmas and can crystallize from them. Small diamonds have been artificially produced in several ways, well summarized by F. W. Clarke in his book, "Data of Geochemistry." They have been obtained by dissolving carbon in molten iron, fused olivine, and lime-magnesia magmas. The discovery of small diamonds in meteorites of iron or peridotitic rock is another fact clearly pointing to a magmatic origin of this mineral. It has been shown by R. A. A. Johnson[1] that chromite, a mineral of the peridotite rocks, contains microscopic diamonds.

## OTHER PRECIOUS STONES

Other precious stones contained in igneous rocks are sapphire, garnet, and peridot (pp. 766 and 797).

Pyrope (magnesium-aluminum garnet) of the beautiful deep-red color which is necessary for gem quality is usually found in basic rocks of igneous origin. The garnets of Bohemia, obtained in washing a Cretaceous conglomerate, are probably derived from a serpentine. The diamond-bearing serpentine of South Africa contains pyrope of gem quality, called Cape ruby. Almandite (iron-aluminum garnet) is not so extensively used. It occurs in granite and aplite, and also, as a product of metamorphic action, in crystalline schists. In the Navajo Reservation, Arizona, pyrope and peridot (yellowish-green olivine) are obtained as disintegration products of a basaltic rock.

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[1] *Mem.* 22, Canada Geol. Survey, 1913, p. 83.

## PLATINUM AND PALLADIUM[1]

Nearly all the platinum of the world is derived from placers, mainly in the Ural Mountains in Russia (p. 246), but also in Colombia, California, Canada, South Africa, Abyssinia, and New South Wales (p. 246).

Native platinum occurs as an alloy with others of the platinum group—osmium, iridium, palladium, ruthenium, and rhodium. Native iridium, iridosmine, and other alloys are found with it.

An analysis of the crude platinum sand of California by Deville and Debray showed the following percentages: Platinum, 85.50; iridium, 1.05; palladium, 0.60; rhodium, 1.00; gold, 0.80; copper, 1.40; iron, 6.75; iridosmine, 1.10; sand, 2.95.

The platinum in the placers forms small rounded, also concretionary and knobby dark-gray pieces. Bright silvery scales of iridosmine occur with it. In the Ural, large pieces of platinum have been found, the largest weighing about 26 pounds.

Platinum and associated metals are siderophile elements and mainly occur as magmatic products in basic rocks, particularly in peridotite. To some extent they follow the high-temperature copper ores, though here there is usually more palladium than platinum present.

Platinum has been found in primary deposits, but until recently few of them have been of economic importance. The modes of occurrence of platinum are as follows: (1) in placers; (2) disseminated in peridotite and olivine gabbro, associated with chromite; (3) in magmatic deposits in basic rocks, associated with chalcopyrite and pyrrhotite (with palladium); (4) in small quantities in quartz veins (with palladium); (5) in contact-metamorphic deposits; (6) in traces in copper deposits of many kinds (with palladium); (7) concentrated by processes of oxidation in replacement ores of copper and gold in limestone (with palladium).

Daubrée showed, in 1875, that the Russian platinum is intergrown with olivine, pyroxene, and serpentine. R. Beck, on authority of S. Conradi, reports it in Paleozoic peridotite from Solowioff Mountain, in the Ural Mountains, and states that the metal forms zonal crystals of unquestionable magmatic origin lying between grains of chromite. Duparc has described the geological conditions relating to platinum in the Ural Mountains.[2] Kemp has found platinum in dunite from the

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[1] J. F. Kemp, *Bull.* 193, U. S. Geol. Survey, 1902.

C. W. Purington, *Trans.*, Am. Inst. Min. Eng., 29, 1899, p. 3.

D. T. Day, *Trans.*, Am. Inst. Min. Eng., 30, 1900, p. 702.

F. W. Clarke, Data of geochemistry, *Bull.* 770, U. S. Geol. Survey, 1924, pp. 717–722.

L. Duparc, and M. Tikonowitch, Le platine et les gîtes platinifères de l'Oural et du monde, Geneva, 1920.

[2] A. N. Zavaritsky, Primary platinum deposits of the Urals, Com. géol., Matér. géol., livr. 108, 1928, Leningrad, 56 pp.

Tulameen River, British Columbia. The serpentines of this region also yield traces of platinum.

In the nickel deposits at Sudbury, Ontario, which are considered of magmatic origin, platinum arsenide, sperrylite, probably accompanied by palladium arsenide, is found as small silvery-white cubes intergrown with pyrrhotite and chalcopyrite. The same mineral was discovered at the Rambler mine,[1] Wyoming, in copper ores, mainly chalcopyrite and covellite, probably of igneous origin and forming a lens in a dioritic rock. Much palladium is also present.

The Salt Chuck mine,[2] on Prince Edward Island, Alaska, near Ketchikan, carries disseminated bornite, also some chalcocite, covellite, and epidote in a dark-green pyroxenite. The ore-body was mined for several years before it was recognized that the bornite contained an appreciable amount of palladium. About 3,000 ounces of palladium were recovered in 1925. The concentrates contain copper 40 per cent, gold 1 ounce per ton, silver 5 ounces, palladium 3 ounces. Mertie thinks the deposit is epigenetic.

Another rare mode of occurrence of platinum is in quartz veins, described by Bell,[3] from the southern island of New Zealand, from northern Finland, and from Canada.

A peculiar occurrence of platinum with wollastonite and grossularite in a contact-metamorphic rock has been reported from Sumatra.[4]

Palladium occasionally occurs alloyed with gold. E. Hussak[5] found in Brazil such palladium-bearing gold in a limestone close to the contact of an igneous rock. Platinum and palladium are also recovered in the electrolytic refining of copper bullion.

A gold-platinum-palladium deposit concentrated by processes of oxidation has been described by Adolph Knopf.[6]

The ore occurs in a mesothermal lead-copper-gold replacement deposit in limestone in the Yellow Pine district, southern Nevada. The material which contains the precious metals is plumbojarosite, a sulphate of iron, lead and bismuth, and occurs in important amounts. It may average, in ounces to the ton: Gold, 3.46; silver, 6.4; platinum, 0.70; palladium, 3.38. All metals are in minute division. The gold is rough, black and spongy, the palladium-platinum appears as microscopic black grains. The platinum metals were probably contained in the primary sulphides and, most likely, have been concentrated by solutions in which they were present in colloidal suspension.

[1] S. F. Emmons, *Bull.* 213, U. S. Geol. Survey, 1903, pp. 94–97.

[2] J. B. Mertie, Jr., *Bull.* 714, U. S. Geol. Survey, 1921, pp. 121–126.

[3] J. M. Bell, *Econ. Geol.*, 1, 1906, p. 749.

[4] L. Hundeshagen, *Trans.*, Inst. Min. and Met. (London), 13, 1903–1904.

[5] *Sitz.-ber.* Akad. Wiss. Wien, 113, No. 1, July, 1904, pp. 1–88.

[6] *Bull.* 620, U. S. Geol. Survey, 1915, pp. 1–44.

**Production and Use.**—At best the world's annual output of platinum, mainly from the Ural Mountains, was 400,000 troy ounces. War conditions greatly decreased the Ural production, but it is now increasing slowly, with production of 100,000 ounces in 1929. The dredges in Colombia now turn out about 45,000 ounces per annum. The output of platinum metals from all domestic sources is now (1930) about 10,000 ounces, of which 527 ounces is from Pacific coast placers; the rest comes from electrolytic copper refineries and oxidized ores; more than one-half of this consists of palladium. The nickel mattes of Sudbury, Ontario, yielded in 1931 about 45,000 ounces each of platinum and palladium. The price of platinum per troy ounce rose from $20 per ounce to a maximum of $120 but has now (1932) dropped to $35 per ounce; palladium now costs $20 per ounce. Iridium is obtained in refining crude platinum or osmiridium. Some is produced by picking out iridium grains from the concentrate. The price is about $90 per ounce. The world's production of platinum in 1930 was about 230,000 ounces.

Platinum is used for jewelry, in dentistry, for chemical utensils, for spark devices, and in the contact process for making sulphuric acid. Attempts are now made to save platinum by using alloys of tungsten, molybdenum, chrome-nickel, and palladium.

Palladium, a silvery-white ductile metal, soluble in $HNO_3$, is used for silvering circles on surveying instruments, and other electroplating; also in dentistry and, in alloy with gold, as substitute for platinum; the last named use being the most important.

Iridium, exceedingly hard and resistant, is used for hardening platinum; 5 to 20 per cent are added. It is also employed for chemical and physical instruments, contact devices, etc.

Rhodium finds use with platinum for thermo-couples. There is little demand for it, or for ruthenium.

Osmium is available in considerable quantities from the refining of platinum sand. Formerly it was used in incandescent lamps. At present, it is little used for industrial purposes.

## LATE DISCOVERIES OF PLATINUM METALS

Needless to say the shortage of platinum caused by the shutdown of the Ural mines stimulated prospecting all over the world. The developments in Colombia were one result of this.

The recovery of osmiridium in the Rand gold field since 1921 is also to be noted. It is believed the mineral, like the gold, is of detrital origin.[1] About 6,000 ounces are now recovered per annum.

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[1] An analysis from Government Areas mine gave 35.78 Ir., 39.71 Os., 9.60 Ru., 6.12 Pt., and 0.16 Rh. Iridosmine comprises two varieties: (1) osmiridium (sisertskite), with less than 40 per cent iridium; and (2) iridosmium (nevyanskite), with more than 40 per cent iridium. Both occur as hexagonal tabular or flaky crystals. Iridosmium is silvery white, osmiridium gray.

It has also been found that platinum is widely distributed in the basic rocks of the Transvaal[1] and probably also in South Rhodesia (Fig. 302).

Much attention has been given to these developments and the literature is already extensive. Wagner calls attention to the extraordinary richness of South Africa in some of the siderophile elements, such as chromium, platinum, and diamond.

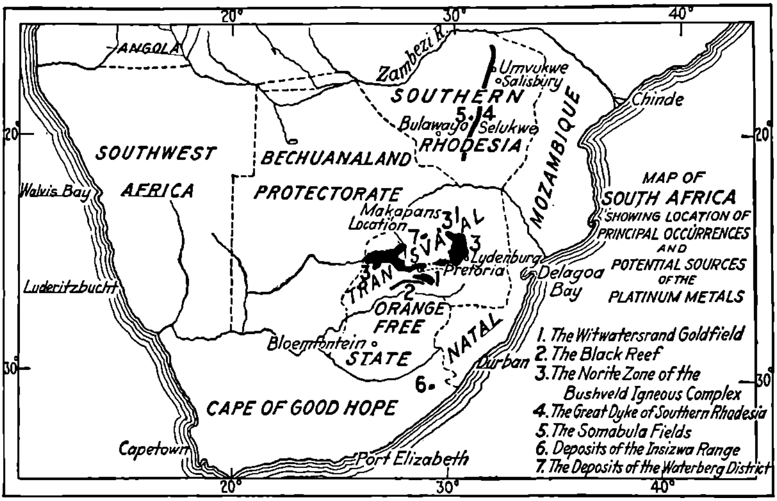

FIG. 302.—Map of South Africa showing location of principal occurrences and potential sources of the platinum metals.

There are several different kinds of platinum deposits in the Transvaal. The first group is connected with the lower norite zone of the great Bushveld igneous complex,[2] which breaks through the Transvaal system (probably pre-Cambrian) and which occupies an area of over 150,000 square miles. This great intrusive sheet or "lopolith" presents many analogies to the smaller intrusive sheet of Sudbury, Ontario (p. 804). A peculiar pseudostratification, or division into horizontal sheets of different composition seems to be characteristic of the norite zone, which comprises many rocks ranging from peridotite to anorthosite and norite. There are much assimilation and alteration of the underlying beds of

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[1] A. W. Newberry and J. F. Kemp, Platinum in the Transvaal and elsewhere, *Eng. Min. Jour.-Press*, 121, 1926, pp. 717–725; 763–768.

Percy A. Wagner, Occurrence of the platinum metals in South Africa, *Econ. Geol.*, 21, 1926, pp. 109–134; 243–270.

The platinum deposits and mines of South Africa, Edinburgh, 1929, 326 pp., with a chapter by H. Schneiderhöhn.

G. H. Stanley and P. A. Wagner, *Jour.* Chem. Met. Min. Soc., South Africa, March, 1925; also *Mining Mag.* (London), 32, 1925.

[2] R. A. Daly and G. A. F. Molengraaff, *Jour. Geol.*, 32, 1924, pp. 1–35.

quartzite, shale, and dolomite. Even the upper acidic differentiates of the complex carry some platinum in places.

1. **Occurrences in Hortonolite-dunite.**[1]—In the lower part of the norite are pipes, segregations, or veins of a peridotite of the above-named variety, extremely rich in iron and containing, besides olivine, some phlogopite, diallage, hornblende, chromite, and titaniferous magnetite. In some of these masses platinum occurs in important amounts, say 0.1 to 7.0 ounces per ton, as cubical crystals and irregular grains up to 4 millimeters in diameter. Small amounts of Fe, Os, Ir, Pd, and Rh enter the composition of the platinum.

2. **Chromite Deposits.**—Flat "beds" of chromite occur in the norite zone and some of these contain platinum (Fig. 303).

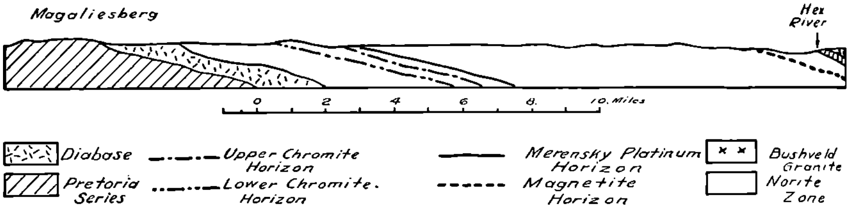

FIG. 303.—Section across Norite Zone of Bushveld Complex, east of Rustenburg showing position of Magnetite Horizon in relation to Platinum and Chromite Horizons. (*After P. A. Wagner.*)

3. **Platinum with Small Amounts of Ni-Co-Fe Sulphides.**[2]—These deposits occur in bronzitite and norite. Palladium and platinum are associated with pyrrhotite, chalcopyrite, and pentlandite.

The Merensky-horizon type forms important low-grade platinum deposits occurring in a persistent flat sheet of diallage-norite, 3 to 30 feet thick. Pyrrhotite, chalcopyrite, and pentlandite are present and are, according to Wagner, of purely magmatic origin; *orthotectic*, to use Graton's term.

The platinum is present as metallic grains, probably also as sperrylite. Over miles this rock contains platinum, about 0.1 to 0.15 ounce per ton. These deposits occur in the Lydenburg and Pretoria districts. Another occurrence and one of great importance appears near Potgietersrust, also in the norite horizon. It is near the rocks of the Transvaal system and includes both strictly magmatic and contact-metasomatic deposits in the basal dolomite; here too the platinum is associated with the sulphides. One horizon has been traced for 1,700 feet and averages 0.3 to 0.5 ounce platinum to the ton. Promising values have also been

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[1] P. A. Wagner and E. T. Mellor, *Trans.* Geol. Soc. South Africa, 28, 1925, pp. 1–18.

[2] P. A. Wagner, On magmatic nickel deposits of the Bushveld igneous complex, *Mem.* 21, Geol. Survey Union South Africa, 1924.

shown in crushed zones of banded iron stone (taconite) occurring at the base of the dolomite series.

The main horizon in the Potgietersrust field is probably the equivalent of the Merensky horizon in the Lydenburg district.

In this vicinity, the most important deposits in South Africa have been opened; they occur in a sheet of diallage-norite, which is 150 feet thick, covered and underlain by anorthosite-norite. Platinum occurs mainly in the upper 50 feet, associated with some sulphides and averaging about 0.1 to 0.2 ounce per ton. Crystals of sperrylite half an inch in diameter have been found. An important mineral in norite is said to be cooperite, $PtS_2$; there are also native platinum and stibiopalladinite ($Pd_3Sb$).[1]

**Platinum-palladium Deposits of the Waterberg District.**—These deposits are most interesting. They occur in central Transvaal and are brecciated quartz veins or composite lodes occupying faults of post-Karroo age in felsite or felsite tuff, belonging to the Bushveld complex.

The main lode can be traced for 2½ miles, is 6 to 50 feet thick, and dips steeply. The banded quartz shows comb structure. There are quartz, and chalcedony, also specularite, oxidized pyrite, and chromiferous chlorite.

Platinum[2] occurs in small gray specks, showing under the microscope a concentrically banded structure suggesting colloidal deposition. The platinum contains up to 37 per cent palladium, and it has been suggested that the colloform structure is caused by alternating bands richer in palladium. The distribution of the platinum is erratic. Assays up to 150 ounces per ton have been obtained. Some sections average 14 ounces over 60 inches. Operations were suspended in 1926.

These deposits present a remarkable similarity to the colloidal cassiterite deposits in rhyolite, described from Nevada, New Mexico, and Mexico. There are the same high-temperature minerals, deposited apparently near the surface. Traces to small amounts of platinum occur in the post-Karroo dolerite, in the great serpentine dike of Rhodesia, and in the kimberlite of the diamond pipes.

**Production.**—The recovery of platinum from these deposits formed a difficult subject. Concentration on tables, amalgamation, or flotation and smelting of the resulting low-grade concentrate are the processes adopted. The recent drop in the price of platinum may considerably retard the development of the deposits. At any rate, the production from the Potgietersrust, Lydenburg, and Rustenburg districts has

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[1] H. R. Adam, *Trans.*, Geol. Soc. South Africa, 33, 1930, pp. 103–109.

[2] P. A. Wagner and T. G. Trevor, Platinum in the Waterberg district, *Bull.* 101, Industries bulletin series, Pretoria, 1924.

G. H. Stanley and P. A. Wagner, Note on the microstructure of Transvaal platinum, *Jour.* Chem. Met. Min. Soc. South Africa, March, 1925.

increased rapidly since 1926 and in 1930 reached 47,021 ounces, including a notable percentage of palladium.

## IRON AND NICKEL

Native iron occurs sparingly in some basalts. Large masses have been found in a basalt at Ovifak, West Greenland, where the natives used the metal for their tools and weapons. The iron contains from 2 to 3 per cent of nickel and 3 per cent of carbon and was long thought of meteoric origin. The basalt breaks through Tertiary beds containing coal, and it is believed by some that the metal was reduced from the rock by means of the coal. The nickel it contains militates against this view; more likely it was carried up from some deep-seated source by the basalt.

Awaruite ($FeNi_2$) is disseminated in gravels and also as small grains in the serpentine and peridotite of Red Mountain, on the south island of New Zealand, and is found also in sluice boxes for gold washing at Hoole Canyon, Yukon Territory.[1] A similar mineral, josephinite ($FeNi_3$), has been found in detritus in areas of serpentine in southwestern Oregon and a few other localities.

## CHROMITE[2]

Chromite ($FeO.Cr_2O_3$),[3] a mineral of the spinel group and usually admixed with other spinel molecules, is an almost constant accessory in peridotites and the serpentines derived from them and is often found in them as accumulations large enough to be mined. The ore, more or less mixed with the rock, forms irregular bunches and masses or flat bodies along the contacts or in the interior of the intrusive masses; frequently also it forms ill-defined streaks or "schlieren." Investigations, of Vogt and others, have shown that chromite in large masses mainly represents purely magmatic separations in peridotite magmas. Vogt showed that the succession in the Norwegian deposits was chromite, olivine, and soda-lime feldspar, and in all cases the chromite appears to

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[1] J. Keele, *Summ. Rept.*, Canada Geol. Survey, 1910, p. 257.

[2] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1894, pp. 384–393.

E. Glasser, Les richesses minérales de la Nouvelle Calédonie, *Ann. des Mines* (10) 4, 1903, pp. 299–536.

W. Glenn, The chromites of Maryland, *Trans.*, Am. Inst. Min. Eng., 25, 1896, p. 481.

E. C. Harder, Some chromite deposits in western and central California, *Bull.* 430, U. S. Geol. Survey, 1910, pp. 167–183.

J. S. Diller, Recent studies of domestic chromite deposits, *Trans.*, Am. Inst. Min. Met. Eng., 63, 1920, pp. 105–149.

[3] Theoretically, chromite should contain 68 per cent $Cr_2O_3$ and 32 per cent FeO, but $Al_2O_3$ and MgO are always present and the actual content of $Cr_2O_3$ is rarely more than 60 per cent. It is one of the most difficultly fusible minerals, melting at 2180° C.

be the earliest consolidated constituent. Deposits in serpentine are often admixed with magnetite. It is noteworthy that during the weathering of peridotite few chromium silicates are formed, while nickel silicates often develop. A little chromiferous muscovite (mariposite), also ouvarovite or chrome garnet, as well as chloritic chromium minerals, in places accompany the chromite.

In the later literature[1] the opinion seems to be gaining ground that chromite may in part be early magmatic, in part late magmatic, and in part even hydrothermal. L. W. Fischer, C. S. Ross, and Edward Sampson have expressed such views. Sampson shows that in South Africa chromite occurs in pyroxenite and rocks rich in bronzite and anorthite, rather than in peridotite.

Chalcopyrite is occasionally found with the chromite. Niccolite (NiAs) has been found in serpentines and peridotites at Malaga, Spain.[2] The mineral is later than the chromite and, according to R. Beck,[3] cements crystals of augite.

Chromite ores should contain 45 to 50 per cent $Cr_2O_3$. Deposits occur in many countries. Until lately, the largest supplies were received from Asia Minor, near Antiochia, Smyrna, and Brussa. These deposits are of unusual size. The Daghardy mine,[4] near Brussa, yielded annually 12,000 to 15,000 metric tons of ore. One of the masses from this mine was 70 meters long, 25 meters wide, and 20 meters deep. Much chromite is also exported from New Caledonia, where the ore occurs in part as residual deposits, which are concentrated, in part as masses and "veinlike segregations" in serpentine. Chromite has been mined near Baltimore, Maryland, in North Carolina, in California, and in Oregon. Although there are many deposits, particularly in the great Serpentine Dike of California, in Eldorado and Placer Counties, the total production is small. Deposits of chromite in the serpentine areas of Quebec have been worked by concentration.

The great deposits of southern Rhodesia now supply much of the demand. Foremost among these is the Selukwe district,[5] where the ore occurs in talcose, chloritic, and serpentine schists, evidently representing dynamo-metamorphosed peridotites.

Other important deposits occur as successive flat sheets, about 10 inches thick, in the serpentine of the Great Dike of southern Rhodesia.

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[1] See papers in *Econ. Geol.* 24, 1929; also Edward Sampson, Varieties of chromite deposits, *Econ. Geol.* 26, 1931, pp. 833–839; *idem,* 27, 1932, pp. 113–144.

[2] F. Gilman, Notes on the ore deposits of the Malaga serpentines, *Trans.*, Inst. Min. and Met. (London), 1896, pp. 159–165.

[3] Erzlagerstätten, 1, 1909, p. 89.

[4] R. E. Weiss, *Zeitschr. prakt. Geol.*, 1901, pp. 249–262.

W. F. A. Thomae, Emery, chrome ore, etc., in Asia Minor, *Trans.*, Am. Inst. Eng., 28, 1898, pp. 208–225.

[5] A. E. V. Zeally, *Trans. and Proc.*, Geol. Survey South Africa, 17, 1914, pp. 60–74.

This remarkable intrusion, similar to the Serpentine Dike of the Sierra Nevada in California, extends north and south for 300 miles (p. 781) and is 3 to 4 miles wide. The exact mode of intrusion of this dike has not been studied. It seems analogous in some ways to the Bushveld intrusions. It is composed of norite, pyroxenite, and peridotite, the serpentine being derived from the latter rock. The total exports of chromite from southern Rhodesia in 1930 are reported as 205,631 metric tons.

Chromium is used as a steel-hardening metal, for stainless steel, and also for the preparation of bichromate of potassium, principally for tanning, also pigments. Chromite bricks are used as refractory material in the steel industry. Chromium is extensively used for electroplating of metals. The metal resists nitric acid; also fruit acids.

The maximum of domestic production, during the World War, was 82,430 long tons. In 1930, 326,617 tons of chromite were imported, mainly from Rhodesia, Greece, Cuba, New Caledonia, and the Union of South Africa. The price f.o.b. shipping ports is about $20 per ton.

## ILMENITE OR TITANIC IRON ORE

**General Features.**—At many places in the world large masses of ilmenite ($FeTiO_3$, containing oxygen 31.6, titanium 31.6, iron 36.8), are found associated with more or less magnetite, olivine, pyroxene, and soda-lime feldspars. Ilmenite, with magnetite, is one of the earlier products of consolidation in magmas and is contained in almost all diabases, basalts, and gabbros; it occurs also in other less basic rocks, but the real home of ilmenite is in the rocks poor in silica. The larger masses of ilmenite are simply facies of the rock itself produced by concentration from the same magma. Near such masses the dark constituents first increase; finally the feldspar disappears and the ore-body consists of a mixture of ilmenite with ferromagnesian silicates. Vogt[1] first called attention to this well-defined group of ore deposits and explained its origin.

The ilmenite deposits, though large, have thus far been little utilized on account of difficulties in the metallurgical treatment; but these do not seem to be insuperable, and, as it is known that titanium confers valuable qualities of hardening on steel, the ores may at some time become important in metallurgy.

Ferrotitanium is manufactured from rich titaniferous iron ores. Rutile is used for arc electrodes. Titanium white ($TiO_2$), an excellent pigment, is made from Ti rich ilmenite by treatment with $H_2SO_4$ and electrolysis of the titanium sulphate.[2] Most of the ilmenite used for

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[1] Die Bildung der Erzlagerstätten durch Differentiation in basischen Eruptivmagmata, *Zeitschr. prakt. Geol.*, 1893, 1894, 1900, and 1901.

[2] *Information Circ.* 6365, U. S. Bur. Mines, 1930.

pigment is obtained from beach sands near Travancore, Southern India. In 1930, 29,000 metric tons were exported from there.

**Microstructure of Ilmenite.**—The complex intergrowths of ilmenite with magnetite, rutile, and hematite have been described by J. T. Singewald;[1] by C. H. Warren;[2] and by Paul Ramdohr.[3]

Warren summarizes his results as follows: One type, represented by occurrences at Miask, Arendal, and Snarum, is homogeneous, though ilmenites from the two latter localities contain a great excess of $Fe_2O_3$ compelling the belief that there is a wide range of miscibility between the molecules $RTiO_3$ and $Fe_2O_3$. A second type presents an intergrowth of grains of homogeneous ilmenite and magnetites. The occurrences at Lake Sanford, New York, and Iron Mountain, Wyoming, belong here. Many so-called titanic iron ores are magnetic.

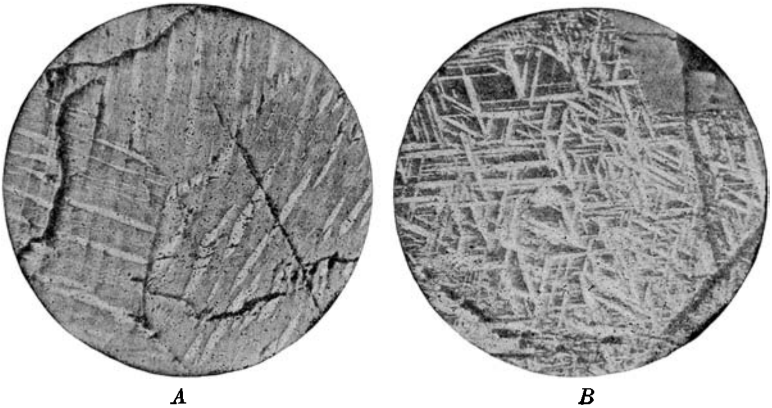

FIG. 304.—*A*. Photomicrograph of polished section showing intergrowth of hematite (light); ilmenite (dark), St. Urbain, Quebec. Magnified 100 diameters. *B*. Intergrowth of magnetite (dark); ilmenite (light); a grain of olivine in upper right corner, Cumberland, Rhode Island. Magnified 180 diameters. (*Both after C. H. Warren.*)

A third type, represented by St. Urbain, Quebec (Fig. 304*A*) and Egersund, Norway, consists of a well-defined crystallographic intergrowth of ilmenite and hematite, caused by the unmixing of an originally homogeneous solid solution.

A fourth type, illustrated by specimens from Cumberland, Rhode Island and Iron Mountain, Wyoming, shows an extraordinary minute regularly oriented intergrowth of magnetite and ilmenite (Fig. 304*B*). Warren believes that there exists a limited solid solution of the ilmenite and

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[1] J. T. Singewald, Jr., *Bull.* 64, U. S. Bur. Mines, 1913.

[2] C. H. Warren, *Am. Jour. Sci.*, 4th ser., 33, 1912, pp. 263–277. The microstructure of certain titanic iron ores, *Econ. Geol.*, 13, 1918, pp. 419–446.

[3] P. Ramdohr, Beobachtungen an magnetit, ilmenit, etc., Festschrift 150, zur Jahrfeier d. Berg. Akad. Clausthal, 1925, pp. 307–341.

P. Ramdohr and H. Schneiderhöhn, Erzmikroskopie, 1931, Berlin, pp. 539–550.

magnetite molecules, with a eutectic; and that ilmenite and hematite form a complete solid solution at higher temperatures, with an inversion interval and limited miscibility at lower temperatures.

Ramdohr also interprets the intergrowth of $Fe_2O_3$ and $FeTiO_3$ as a result of unmixing by slow cooling. Similar unmixing of the solid solution $Fe_3O_4$-$FeTiO_3$ produces the intergrowths of magnetite and ilmenite.

**Irregular Bodies.**—The titanic iron ores form irregular masses or rather sharply outlined streaks in the central parts of gabbro or norite intrusives. The transitions to the country rock indicate that these masses have been formed by differentiation in the magma after the irruption into its present place. In these differentiated magmas ilmenite and magnetite have, as a rule, crystallized after the silicates, though some titanium and iron are retained in the residual magma. The order of crystallization is ferromagnesian silicates, spinel, ilmenite, and magnetite. Probably little water was present and the temperature of consolidation was high, perhaps near 1450° C., the fusion point of ilmenite, according to Brun. Vogt has shown that during the differentiation of a gabbro or norite magma a concentration of ferric oxide takes place, as well as of titanium, chromium, and vanadium; the lime, magnesia, and particularly silica diminish greatly, the silica to such an extent that the alumina and magnesia are forced to crystallize as corundum and spinel, both of which occur frequently in these deposits. Little sulphur or phosphorus is present.

**Dikes.**—Separated during a deeper-seated differentiation and remelted, veritable dikes of almost pure ilmenite may be injected into the prevailing rock, which then is usually an anorthosite.

**Occurrences.**—Vogt and Kolderup have described the Norwegian occurrences in norite and anorthosite in the great intrusive region in Ekersund; the largest body is 3 kilometers long and from 30 to 70 meters thick. Its composition is about 21 per cent plagioclase, 41 per cent hypersthene, and 38 per cent ilmenite. At Routivare, in northern Sweden, a gigantic mass of titanic iron ore is included in a mass of highly altered gabbro, intruded in Cambro-Silurian strata. Some pyrrhotite is associated with the ore.

A large deposit at St. Urbain, in Quebec, is described by C. H. Warren[1] and J. B. Mawdsley.[2] Elongated, sometimes dike-like masses of ilmenite are inclosed in anorthosite. Specularite is intimately intergrown with ilmenite. Much rutile and blue grains of sapphirine ($Mg_5Al_{12}Si_2O_{27}$), also andesine, biotite, and spinel are contained in some of the ore. Other varieties in which no rutile is present contain only 5 to 6 per cent of other minerals.

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[1] *Am. Jour. Sci.*, 4th ser., 33, 1912, pp. 263–277.

[2] *Mem.* 152, *Canada Geol. Survey*, 1927, 58 pp.

J. F. Kemp[1] has described the large deposits in the Adirondacks of New York, near Elizabethtown and Lake Sanford. These ores are contained in a gabbro which is intrusive in a large "massif" of anorthosite. They are granular mixtures of magnetite and ilmenite with a maximum of 15 per cent $TiO_2$ and form irregular or tabular masses. Osborne[2] finds that the ore-bodies are in part intrusive into the country rock, in part concordant with the rock structures. As the ilmenite is later than the silicates, the second type is interpreted as concentrations by "filter pressing" or some such process. The ores contain plagioclase, pyroxene, olivine, hornblende, garnet, pyrite, apatite, spinel, and quartz, but are low in sulphur and phosphorus. At the principal locality, probably several million tons of ore are exposed above the level of the lake.

Other deposits are known to occur in Minnesota[3] and Ontario.

In eastern Wyoming, at Iron Mountain,[4] a dike of almost solid ilmenite, in places 300 feet wide, breaks through anorthosite containing but little pyroxene and scarcely any ilmenite. This is a most remarkable instance of complete deep-seated differentiation of magmas. An analysis of the ore is as follows:

| | | | |
|---|---|---|---|
| $SiO_2$ | 0.76 | MnO | 1.38 |
| $TiO_2$ | 23.49 | MgO | 1.56 |
| $Al_2O_3$ | 3.98 | CaO | 1.16 |
| $Cr_2O_3$ | 2.45 | $P_2O_5$ | trace |
| $Fe_2O_3$ | 45.03 | S | 1.44 |
| FeO | 17.96 | ZnO | 0.47 |
| | | | 99.68 |
| Total Fe | | | 45.49 |

Some olivine, spinel, and magnetite are present as inclusions in the ore.

Concentrations of ilmenite with prevailing magnetite are not uncommon in gabbros and diabases, though rarely of economic importance.

Taberg, in southern Sweden, is of interest as one of the first places in which the existence of magmatic ore deposits was demonstrated, by A. Sjögren, Törnebohm, and Igelström. Taberg is a prominent hill 400 feet high, composed of norite. Toward the center the ilmenite and magnetite are greatly concentrated and form a mass of ore with some olivine, biotite, and plagioclase. The ore has at times been smelted as iron ore; it contains about 6 per cent titanic dioxide.

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[1] Titaniferous iron ores of the Adirondacks, *Nineteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, 383–422.

[2] F. F. Osborne, Certain magmatic titaniferous iron ores, *Econ. Geol.*, 23, 1928, pp. 724–761; 895–922.

[3] F. F. Grout, *Econ. Geol.*, 20, 1925, pp. 424–430.

[4] S. H. Ball, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 206–212.

Flat sheets of titaniferous magnetite occur in the upper part of the Bushveld igneous complex, South Africa. They are up to 15 feet thick and can be traced for miles; they contain from 16 to 19 per cent titanium.

**Influence of Pressure.**—Kolderup has shown that, under dynamometamorphic influences, the ilmenite in deposits of this class changes to rutile, pyroxene is replaced by amphibole, garnet is developed, and new biotite and feldspar aggregates are formed.[1]

## MAGNETITE

As iron is contained in some basic igneous rocks in great quantities and as even the average of this metal in all igneous rocks is 4.46 per cent, it might be expected that magmatic concentrations of magnetite would be abundant. This does not seem to be the case, however. In the early stages of consolidation of igneous rocks some magnetite is crystallized, together with other accessories, but the tendency to form silicates is strong and sufficient silica is usually available to take care of all the iron in the form of ferromagnesian silicates. In basic rocks much iron remains in the residual magma and is separated at a late stage of the consolidation.

Commercially valuable deposits of magnetite as differentiation products of magmas have been found only in connection with syenites, syenite porphyries, and keratophyres, and here the magnetite is usually later than the ferromagnesian silicates and the feldspars. Fluorapatite frequently accompanies the magnetite, but sulphides are rare. Some basalts contain as much as 8 per cent FeO and 4 per cent $Fe_2O_3$; some gabbros and norites as much as 15 per cent FeO and 1 or 2 per cent $Fe_2O_3$. Magnetite requires 69 per cent ferric oxide and 31 per cent ferrous oxide. Any process of differentiation favoring the separation of magnetite thus requires a transfer of part of the iron to a ferric state.

By later processes the magnetite may develop hematite along cleavage planes or by pseudomorphism (martite). Magnetite deposits are rarely found in surface lavas, doubtless because the time before consolidation has not been long enough to permit of differentiation. Such concentrations in basalt have been described from Colorado[2] and a similar occurrence in andesite is on record from Nevada.[3]

**The Iron Ores of Northern Sweden.**[4]—The great magnetite deposits in the extreme northern part of Sweden, the largest in the world, have been studied by many geologists. For many years they remained

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[1] R. Beck, Erzlagerstätten, 3d ed., 1, 1909, p. 57.

[2] H. S. Washington and E. S. Larsen, *Jour.* Washington Acad. Sci., 3, 1913, pp. 449–452.

[3] J. C. Jones, *Econ. Geol.*, 8, 1913, pp. 247–263.

[4] O. Stutzer, Die Eisenerzlagerstätten bei Kiruna, *Zeitschr. prakt. Geol.*, 14, 1906, pp. 65–71; 137–142. Ref. *Econ. Geol.*, 2, 1907, pp. 88–91.

Per Geijer, Igneous rocks and iron ores of Kiirunavaara, etc., Stockholm, 1910, unworked on account of their high percentage of phosphorus, which, since the invention of the Thomas process, is no longer objectionable.

The deposit at Kiruna forms a north-south ridge which is, as exposed at the surface, about 2.8 kilometers in length and rises 248 meters above Lake Luossajärvi, or 748 meters above the sea. The magnetite forms the highest part of the ridge and is from 32 to 152 meters wide. The total amount of ore proved above the level of the lake is said to be 265,000,000 metric tons, and the total, to a depth of 300 meters, proved by borings, about 1,075,000,000 tons. The annual production from open-cut and underground workings reached a maximum of about 9,000,000 metric tons in 1929; the ore is exported to England and Germany.

The ore-body and surrounding formations are of pre-Cambrian age and dip steeply to the east (Fig. 305). In the footwall lies a syenite porphyry with fluidal structure and rich in soda (per cent: 61 $SiO_2$, 8 iron oxides, and 6 to 7 $Na_2O$), the pyroxene of which is largely altered to amphibole, chlorite, and epidote. Magnetite is present in two generations, the later of which may surround the feldspars and enter them along cleavage planes. The contact between the ore-body and footwall porphyry is apparently sharp, but shows in detail a narrow zone of transition due to small, sharply defined dikelets of magnetite in the porphyry. An ultimate product of this zone is a mixture of magnetite with green hornblende; it contains vugs filled with apatite, titanite, and magnetite. The deposit itself is free from vugs. The reddish quartz porphyry of the hanging wall is essentially similar in microscopic character, but contains, in per cent, about 71 silica, 5 iron oxides, and 5 to 6 $Na_2O$, and has been classified as a quartz keratophyre. It contains fragments of magnetite ore and occasionally of the footwall porphyry. Above this hanging wall porphyry lie quartzites, clay slates, and conglomerates, with water-worn fragments of iron ores and keratophyre.

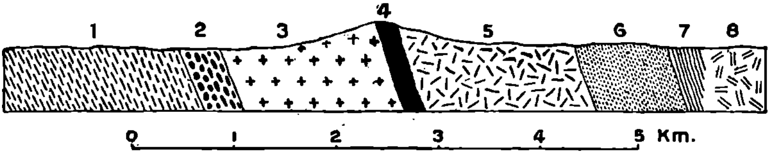

FIG. 305.—Schematic cross-section of iron deposit at Kiruna. 1, soda greenstone; 2, Kurravara conglomerate; 3, syenite porphyry; 4, magnetite deposit; 5, quartz porphyry; 5 and 6, Hauki Complex; 7, amphibolite; 8, quartz porphyry.

The ore is compact and fine-grained, consisting chiefly of magnetite and apatite in intimate intergrowth, apparently having crystallized

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278 pp. Author's abstract, *Econ. Geol.*, 5, 1910, pp. 699–718 (English).

Per Geijer, Berggrunden inom malmtrakton Kiruna-Gellivare (English summary), *Ann. Rept.* 24, Swedish Geol. Survey, Ser. C, No. 366, Stockholm, 1931, 225 pp.

Per Geijer, The iron ores of the Kiruna type, *idem*, No. 367 (English), 1931, 39 pp.

together. In places it contains pyroxene. The ore is said to average 68 per cent iron. The phosphorus is, as a rule, above 2 per cent, and some parts of the ore yield from 3 to 4 per cent or even more of this substance. The apatite seems to have crystallized first. In places magnetite and apatite appear to be contemporaneous (Fig. 306). The sulphur is not above 0.05 per cent; manganese not above 0.70 per cent; a similar amount of magnesia is recorded, about 1.5 per cent silica, 0.75 per cent alumina, about 3 per cent lime, and 0.3 per cent $TiO_2$. In places a fluidal structure of the magnetite and branching veinlets of apatite, suggesting a eutectic, are observed in the ore.

In 1898, H. Lundbohm published the first monograph on the deposit. A little later Högbom proved the magmatic origin of the deposit, though he believed that the differentiation had proceeded in place. The investigations of Stutzer and the later monographic works by Geijer have shown plainly that the ore was differentiated from magmas in depth and that it has been brought to its present position in a molten condition. The differences largely depend upon whether the porphyries are to be considered as effusive or intrusive. Stutzer held the latter and more probable view, regarding the syenite porphyry as the earlier rock followed by the intrusion of a dike of magnetite. On the east side the magnetite was later intruded by quartz porphyry which includes fragments of ore. Geijer, holding that the porphyries were extrusive, thought that the ore was intruded as a sheet of molten material between the porphyries while the flows were still horizontal, and that it was later brought into a vertical position.

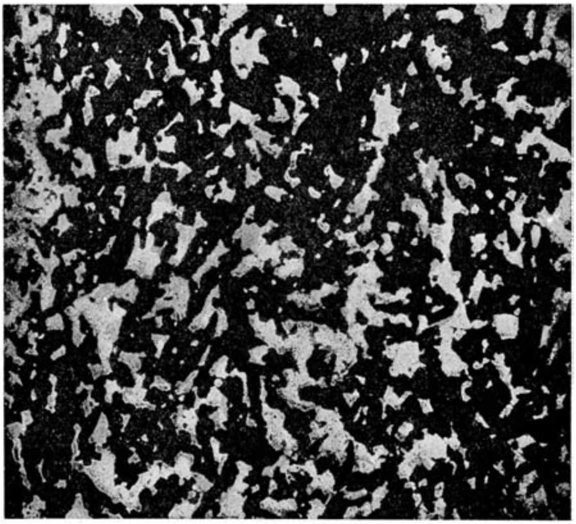

FIG. 306.—Thin section of apatite-rich ore, Kirunavaara. Apatite, white; magnetite, black. Magnified 14 diameters. (*After Per Geijer.*)

The injection theory necessitates a magmatic differentiation in depth and probably requires also a remelting before intrusion. On the whole, the deposit illustrates beautifully the extreme magmatic differentiation from an alkaline magma with concentration of iron, titanium, phosphorus, and fluorine. Undoubtedly the latter elements greatly depressed the melting point of the magnetite.

The great iron mines of Gellivare, a short distance south of Kiruna, produce about 1,500,000 metric tons of ore per annum.

The ore is principally mined in open workings and contains the same minerals as that of Kiruna—that is, magnetite and apatite—but it has a coarser grain. Locally it contains pyrite, chalcopyrite, fluorite, calcite, and zeolites. The ore is rudely schistose, conforming with the steep dip of the country rock, and forms large lenses, partly imbricating on parallel and curving strike lines.

The rocks are chiefly gneisses. The red gneiss is most common near the deposit and is traversed by many irregular veins of magnetite. It consists of albite with some quartz, chlorite, apatite, and biotite.

The reddish-gray gneiss is similar in composition but contains also soda-lime feldspar, microperthite, augite, and hornblende. Both rocks are rich in soda.

Dikes of diorite or quartz keratophyre cut across the ore-body.

The deposit at Gellivare[1] has probably a similar origin to that at Kiruna. This view is supported by the tectonic relationship; the ore-body is by no means confined to a single horizon in the gneiss. On the whole the analogy with Kiruna is very striking, though at Gellivare the rocks are clearly of intrusive origin. At both places the same genetic relations seem to exist; the earliest rock is rich in soda and of low to medium acidity, then follows an intrusion of magnetite-apatite rock, and lastly a quartzose soda-rich igneous rock was intruded.

Gellivare is, then, simply a dynamo-metamorphosed Kiruna.

The hematite deposits at Iron Mountain,[2] Missouri, are believed by Geijer[3] to be closely allied to the magnetite of Kiruna. They are indeed very similar to the magnetite injected in porphyry at Tuolluvaara[4] in northern Sweden. He thinks the hematite may have been derived by alteration from magnetite and points to the association with apatite and the occurrence as dike-shaped masses in the pre-Cambrian porphyry.

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[1] Per Geijer, Gällivare malmfält, *Paper* 22, Geol. Survey Sweden (Eighth summary), 1930, 115 pp., Stockholm.

[2] C. W. Crane, The iron ores of Missouri, Missouri Bur. Geol. and Mines, 2d ser., 10, 1912, pp. 107–144.

J. E. Spurr, *Eng. Min. Jour.-Press*, Feb. 26, 1927.

[3] *Econ. Geol.*, 10, 1915, p. 324.

[4] Per Geijer, Tuolluvaara malmfälts geologi, Geol. Survey Sweden, *ser. C*, 296, 1920.

**The Magnetites of the Ural Mountains.**—According to the investigations of Loewinson-Lessing and Jakowleff, the magnetite deposits of Wyssokaia Gora and Gorablagodat, in the Urals, are products of differentiation in magmas, although at the former locality contact-metamorphic deposits also appear to be present. A summary of the Russian literature has been given by Beck.[1] In both places the igneous rocks are augite syenites; at Gorablagodat[2] the ore forms streaks or "schlieren" in this rock; it has a brecciated structure, the magnetite cementing the augites and feldspar, but these are in part probably of contact-metamorphic origin. They contain very little apatite, and the ores are not injected dikes.

**The Magnetites of the Adirondacks.**[3]—The eastern part of the Adirondack Mountains, in northern New York, contains a number of deposits of magnetite which have been worked since the early part of the last century and which still possess considerable economic importance. The total output is estimated at 40,000,000 long tons. The annual mine production in the last 25 years has varied from 1,000,000 to 2,000,000 long tons. The latter figure was reached in 1917. The more important operations are carried on in the Mineville district, but the deposits are spread over a large area. As some of the ores contain much apatite, magnetic concentration is used. The concentrates contain 60 to 65 per cent of iron, and a by-product of impure apatite is obtained which is used as a fertilizer. The ores are extracted through shafts, the deepest of which, at Lyon Mountain, is 1,500 feet deep on the incline.

The associated rocks include syenitic, granitic, and dioritic gneisses, garnetiferous schists, amphibolites, and crystalline limestones. The deposits considered of magmatic type occur in a belt of syenitic gneisses, in part also massive syenites and their pegmatites. These rocks contain from 1.5 to 6.5 per cent magnetite.

In the Archean sedimentary rocks, known as the Grenville series, are a number of smaller deposits, many of which contain pyrite as well as magnetite; in the rocks graphite, sillimanite, garnet, and scapolite have been noted. The genesis of these deposits is in doubt; they may be of sedimentary origin and subsequently metamorphosed.

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[1] Erzlagerstätten, 1, 1909, pp. 29–34.

[2] Vera de Derwies, Recherches géol. dans la région du Blagodat, *Bull.* Com. géol., Leningrad, 1924, p. 1439.

[3] J. F. Kemp, Geology of the magnetites near Port Henry, *Trans.*, Am. Inst. Min. Eng., 27, 1898, pp. 146–203.

D. H. Newland and J. F. Kemp, Geology of the Adirondack magnetic iron ores, *Bull.* 119, N. Y. State Mus., 1908, p. 182.

D. H. Newland, Sedimentary phases of Adirondack magnetites, *Econ. Geol.*, 18, 1923, pp. 291–296.

W. J. Miller, Magnetite iron ores of Clinton County, New York, *idem*, 14, 1919, pp. 509–535.

The magnetites associated with undoubtedly igneous rocks appear as long lenses or pod-like bodies, often bent, curved, or folded, and have clearly participated in the general metamorphism of the country; at first they were probably tabular bodies. The ore lenses range in thickness from a few feet up to 25 feet or more, especially where curved or folded. In part the magnetite ore is pure, but more commonly it is mixed with the minerals that make up the wall-rocks, into which the ores often grade; these minerals are feldspar, quartz, pyroxene, and hornblende.

Apatite is usually present, and the non-Bessemer grade may contain as much as 10 per cent of this mineral. While much of the ore yields 60 per cent iron, there are large masses of ore with 50 per cent iron or less that are suitable for concentration. According to Newland the lowest grade of workable milling ore carries about 35 per cent iron. An average analysis of 65 car-loads from pit 21 of the Mineville group of mines gave, in per cent, iron, 60.03; silica, 4.48; phosphorus, 1.635; sulphur, 0.021; and titanium, 0.12. The concentration resulted in a first-class product with 67.34 Fe and 0.67 P and a "high-grade" apatite containing 3.55 Fe and 12.71 P.

The intimate association and intergrowth of the magnetite with the feldspar, augite, hypersthene, and hornblende of the augite syenite are considered by Kemp to prove its origin by magmatic differentiation. Syenitic pegmatites are also present, and the processes of pegmatitization are considered to have contributed to the formation of the ore; fluorite and titanite are often found in the ores.

Newland thinks that conditions in the Mineville region are complicated by the injection of the syenites in the Grenville Series, which is so metamorphosed that it is difficult to identify. The ores seem to present an extreme differentiation product of alkaline rocks, with concentration of iron, titanium, and fluorine.

Similar relations are noted by W. S. Bayley[1] in regard to the magnetite deposits of New Jersey, which have long been worked near the town of Dover. There are no deposits which can be shown to have separated from the liquid syenitic rock; they are all high-temperature replacements accompanied by amphibole and biotite, often coarsely crystallized. They are sometimes veinlike and the flat ore-shoots conform with the elongation of the minerals in the gneiss. On the other hand, they are not normal pegmatites; but pegmatites with some magnetite occur in the region.

While some magnetite is of early crystallization, these deposits were formed by late, iron-rich residual solutions probably after the pegmatites had developed and certainly after the consolidation of the rock. They are more nearly related to certain hypothermal deposits than to liquid-magmatic separations.

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[1] W. S. Bayley, Iron mines and mining in New Jersey, Geol. Survey New Jersey, *Final Report*, 7, 1910, pp. 131–193.

The magnetite "ore vein" of Cranberry, North Carolina, is, according to W. S. Bayley,[1] of similar origin. The deposit is continuous for over a mile and contains a large amount of concentrating ore, averaging about 45 per cent Fe. The enclosing rocks are gneisses. The ore contains much amphibole, epidote, and garnet. Bayley regards it as "an injected pneumotectic deposit, related to pegmatite." Basic, pyroxenic pegmatites, with magnetite, were followed "by magmas or solutions that produced pyroxene and magnetite, and finally mainly magnetite." Bayley's description suggests that the ores are in part formed by a high-temperature replacement.

It looks, therefore, as if the liquid-magmatic orthotectic iron ores were very scarce. We really only know them as injected dikes, but that they are derived by differentiation from alkaline magmas seems fairly certain.

## CORUNDUM[2]

**General Mode of Occurrence.**—Corundum ($Al_2O_3$) has long been known as a product of regional and contact metamorphism; that it may also result from direct crystallization from a molten magma has been established by late investigations. Alumina is contained in all magmas, usually forming about 14 to 17 per cent. Certain syenites, nepheline syenites, and anorthosites may contain as much as 30 per cent. The peridotites, on the other hand, contain only from a fraction of 1 per cent up to 10 per cent of alumina. The corundum of magmatic origin is chiefly associated with those rocks of exceptionally high or low content of alumina, in which at the same time the silica is low.

By some observers the corundum of igneous rocks is regarded as due to recrystallization of included shale fragments. On the whole the differentiation theory fits the facts better.

Corundum is fusible only at 2250° C. (Moissan), but it by no means follows that it crystallized from the magma at this temperature; Hautefeuille and Perrey showed, for instance, that alumina is soluble in a nepheline or leucite melt and that upon cooling the greater part of it crystallizes as corundum. Very likely it will be found that similar conditions would exist in melted olivine. Morozewicz has shown that when

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[1] *Bull.* 29, Geol. Survey Tennessee, 1923.

[2] The blue transparent corundum is called sapphire; the red transparent variety forms ruby; both varieties are valuable gems. Colorless, yellow, and green varieties also occur. The ordinary bluish or gray corundum is an inconspicuous mineral with good basal cleavage and great hardness, whence its principal use as an abrasive. Mixed with magnetite, mainly in metamorphic rocks, it is termed emery, the name being derived from Cape Emeri, on the island of Naxos.

J. H. Pratt, Corundum, *Bull.* 269, U. S. Geol. Survey, 1906.

A. E. Barlow, Corundum, *Mem.* 57, Canada Geol. Survey, 1915, p. 377.

G. P. Merrill, The non-metallic minerals, 1904, pp. 69–85; 2d ed., 1910, pp. 73-89.

J. Morozewicz, *Tschermak's min. u. petr. Mitt.*, 18, 1898, pp. 22–83.

in a silicate melt the alumina is in excess of the ratio $RO:Al_2O_3::1:1$, corundum, spinel, sillimanite, or cordierite will form.

**Corundum in Igneous Magnesian Rocks.**—A long belt of magnesian rocks of probable pre-Cambrian age, mainly peridotites, gabbros, and norites, extends along the Appalachian region from Alabama to Massachusetts, and in these rocks corundum has been found in commercial quantities at a number of places.

In North Carolina and Georgia the mineral occurs as vein-like bodies from a few inches up to 15 feet in width at the contact of peridotite with gneisses and schists, in part also in the peridotite itself. Chlorite, enstatite, and spinel are associated with the corundum. Among the principal localities are the Laurel Creek mine, in Rabun County, Georgia; Corundum Hill, Macon County, and Webster, Jackson County, North Carolina. None of the southern localities has been productive in late years.

Deposits of emery, an impure corundum mixed with magnetite, are worked by open cuts near Peekskill, Westchester County, New York. The emery here occurs in the intrusive Cortlandt series of rocks described by G. H. Williams[1] and consisting of peridotites, norites, and diorites. The corundum and magnetite are, according to Williams, simply segregations in the norite, the constituent minerals of which occur even in the present emery. Hercynite—an iron spinel—and garnet accompany the magnetite and the corundum.

T. L. Watson[2] has described the Virginian emery deposits. He finds that they form lenses and masses in a granite intrusive in highly altered Cambrian sediments. The emery consists of magnetite and spinel with more or less corundum and ilmenite. The deposits are regarded as produced by igneous contact metamorphism; they probably resulted from the absorption of Cambrian schist xenoliths by the magma.

Regional metamorphism easily changes many of these magnesian rocks to amphibolites and chloritic schists. During this process the corundum is apparently little affected. At Chester, Massachusetts, according to B. K. Emerson,[3] emery occurs for a long distance along a belt of amphibolite contained in sericite schists. The emery-bearing part of the schist is in places 12 feet wide and has been mined to a depth of 300 feet below the surface.

Corundum of gem quality is occasionally found in these deposits or in the gravels derived from them, but most of the sapphires obtained in the United States are derived from a different source. At Yogo Gulch, Fergus County, Montana, long dikes belonging to the monchiquite-camptonite series of lamprophyric rocks contain sharply crystallized

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[1] *Am. Jour. Sci.*, 3d ser., 33, 1887, p. 135.

[2] *Econ. Geol.*, 18, 1923, pp. 53–76. Probably the last contribution from his pen.

[3] *Mon.* 29, U. S. Geol. Survey, 1908, pp. 117–147.

rhombohedral sapphires of excellent quality. The decomposed rock is allowed to slack and is washed in sluice boxes. The deposits have been described by Weed,[1] Pirsson,[2] and Sterrett.[3]

**Corundum in Syenite.**—The most important deposits of corundum in Canada were discovered by W. F. Ferrier about 1896. The mineral occurs as an essential constituent of syenites and nepheline syenites and their pegmatites in Haliburton, Peterboro, Hastings, and Renfrew counties, Ontario.[4] The rocks form dikes or intrusive masses in gneisses and contain as much as 12 or 15 per cent of bluish or gray, often roughly crystallized corundum, many of the crystals being 2 or 3 inches in diameter. The rock is quarried, crushed, and concentrated on tables.[5] The production, formerly great, has now diminished. The value of corundum is about $100 per ton.

**Other Occurrences.**—Corundum has also been found in anorthosites—rocks consisting principally of labradorite, or anorthite feldspar. According to T. H. Holland[6] corundum is abundant in India and representatives of the various classes already described are present.

In a number of occurrences of corundum in "gneiss" we have probably to deal with igneous rocks, like syenite, made schistose by pressure. Corundum is, however, unquestionably also developed by the contact metamorphism of limestone, probably by transfer of alumina from the magma. The largest known deposits of emery occur on the island of Naxos, in the Greek Archipelago, and near Smyrna in Asia Minor; they are contained in metamorphosed limestone.

**Production in the United States.**—At present little or no corundum is produced in the United States. In 1929, 924 tons of emery at about $11 per ton represented the domestic output. Imports of corundum and emery from Canada, Greece, and Turkey, in 1929, had a total value of about $494,174. In response to a great demand, artificial corundum (alundum) is now manufactured at Niagara Falls by fusing bauxite in the electric furnace.

**Uses.**—Corundum finds its principal use as an abrasive, wheels and abrading tools of all kinds being manufactured from the crushed material.

## SULPHIDE ORES OF IGNEOUS ORIGIN

**General Principles.**—That sulphide minerals may crystallize from a magma has been ascertained beyond doubt, but the number of minerals

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[1] *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1899, pp. 454–460.

[2] *Am. Jour. Sci.*, 4th ser., 4, 1897, p. 421.

[3] *Mineral Resources*, U. S. Geol. Survey, 1907, pt. 2, p. 816.

[4] A. P. Coleman, *Eighth Rept.*, Ontario Bur. Mines, 1899, pp. 250–253.

[5] V. L. Eardley-Wilmot, *Publ.* 675, Canada Dept. Mines, 1927, 51 pp.

[6] A manual of the geology of India, Economic Geology, pt. 1, Corundum, Calcutta, 1898.

which have this origin is limited to a few species, mainly pyrrhotite, pyrite, chalcopyrite, molybdenite, sphalerite, and pentlandite; arsenides like niccolite and sperrylite are also known. This view of the igneous origin of certain ores has been gained largely by the labors of J. H. L. Vogt.[1]

While it is clear that sulphides are not freely miscible with silicate magmas,[2] Vogt has shown that the monosulphides are soluble to some extent in dry melts of basic character—that is, with much iron, calcium, and manganese. Under favorable circumstances, particularly at temperatures of about 1500° C., as much as 6 or 7 per cent of these sulphides may become dissolved. But it is also to be noted that Vogt found but little solubility for the sulphides of copper, nickel, lead, and silver. Upon crystallization the sulphides always separate out first, as oldhamite (CaS), alabandite (MnS), troilite (FeS), and sphalerite (ZnS). These experimental results do not correspond with the conditions found in nature, for of the sulphides mentioned sphalerite is the only one encountered at all in igneous rocks, and the sulphides of slight solubility, like those of copper and nickel, are the most abundant. Later investigations by W. Wanjukoff[3] have shown that the sulphides of copper, nickel, iron, zinc, and cadmium are soluble in basic slags to a notable degree and in the order of abundance indicated. Very likely the presence of mineralizers other than sulphur would increase this solubility as, in fact, already suggested by Vogt.

In the surface lavas, which correspond most closely to dry melts, small grains of primary pyrite, pyrrhotite, chalcopyrite, and bornite are occasionally found.[4] The sulphides of economic importance are almost wholly confined to the peridotites, norites, and gabbros, all rocks of deep-seated crystallization. The characteristic metallic association is iron, copper, nickel, platinum, and occasionally a little arsenic.

Many observers have stated that the pyrrhotite and chalcopyrite which often occurs in basic rocks, more or less intergrown with magnetite, are of primary magmatic origin. None has treated the subject better and more convincingly than E. Howe,[5] who described the gabbro-norite and pyroxenite of the Cortlandt series. His conclusions are as follows: The extremely fresh rocks contain small amounts of pyrrhotite, pentlandite, and chalcopyrite; there is no post-magmatic alteration. The sulphides are as essentially magmatic as the silicates. Although most of the sulphides separated from solutions at an early stage in the cooling of the magma, small quantities continued to separate or to redissolve until

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[1] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1893.

[2] J. H. L. Vogt, Die Sulfid-Silikatschmelzlösungen, Kristiania, 1917.

[3] *Metallurgie*, 9, 1912, pp. 1–48.

[4] W. H. Newhouse, *Bull.* Geol. Soc. Am. 47 (1), 1936, pp. 1–52.

[5] Sulphide-bearing rocks from Litchfield, Conn., *Econ. Geol.*, 10, 1915, pp. 330–347.

the magma had nearly crystallized. The form and the interstitial relations of the sulphides seem to show that, although they may have separated early from the silicates, they remained liquid until the silicates had crystallized. The sulphide-bearing rocks are regarded as products of differentiation from magmas poorer in these substances.

The melting points of some of the sulphides are indicated on Fig. 307, which also shows the lowering of the melting points and the eutectic points for several binary sulphide systems. The molten sulphides mix in all proportions to a uniform fluid phase. They are extremely liquid with low viscosity and, perhaps excepting some systems in which ZnS enters, they crystallize later than the rock minerals. They crystallize without notable under-cooling. Silicates are not soluble in a sulphide melt but, according to Vogt, magnetite is slightly soluble and may separate from the melt upon cooling.

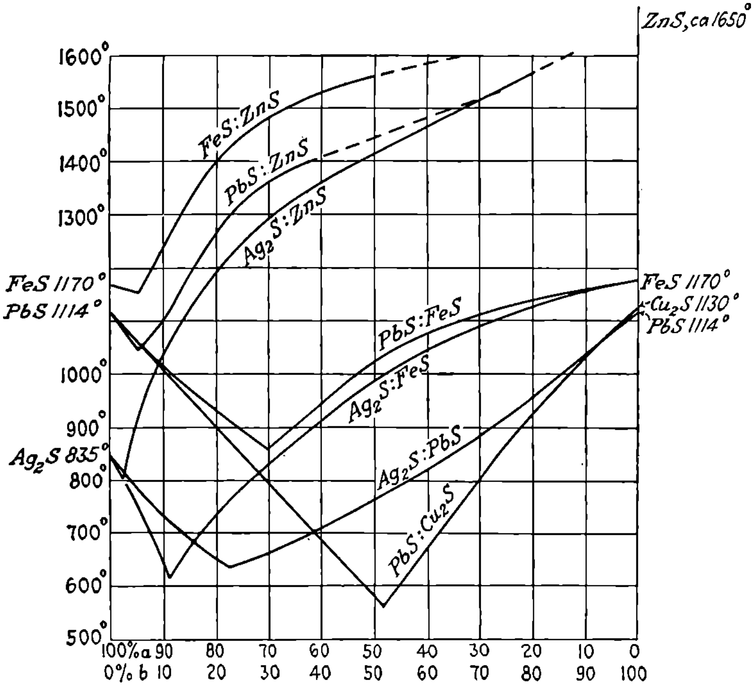

FIG. 307.—Temperature-composition diagram for some binary sulphide systems. (*After J. H. L. Vogt.*)

The liquid-magmatic sulphide ores contain magnetite, pyrrhotite, pentlandite, cubanite, chalcopyrite, and bornite, but no gangue minerals other than the primary rock-forming minerals. They contain no sericite, chlorite, garnet, or epidote. The sulphides often corrode and embay the older silicates but without any indication of secondary substances. One often wonders what became of these dissolved portions. There is a certain succession among the sulphides. Where pyrite is present it is often octahedral, and etched depressions on its faces are sometimes filled with chalcopyrite. Pyrrhotite is the most abundant mineral. The usual succession of minerals, as established by Tolman and Rogers, and other authors, is as follows: Silicates, magnetite, hematite, pyrrhotite, pentlandite, chalcopyrite, and bornite. Nevertheless, it seems certain that, in these rocks, there is also magnetite which is earlier than the silicates. It will be remembered that Vogt in his studies on silicate melts frequently obtained two generations of this mineral. Sperrylite ($PtAs_2$) and millerite ($NiS$) have also been reported.

The determination of the succession is often made difficult by the occurrence of solid solutions and unmixing. Cubanite and pentlandite both appear to separate from pyrrhotite by slow cooling. According to Schwartz[1] the cubanite is taken up in solid solution when the unmixed material is heated to 450° C. According to W. H. Newhouse,[2] similar relations apply to pentlandite and pyrrhotite.

At times it becomes extremely difficult to hold closely to the above-mentioned definitions and rules. Most deposits of magmatic sulphides are not uniform disseminations but rather marginal concentrations, quite plainly injected after the surrounding rock had crystallized, or else they are accompanied by some gangue materials like chlorite, quartz, garnet, epidote, etc. Such deposits were probably formed by highly-concentrated sulphide melts with water and other mineralizers. When the dominating influence is no longer the straight melt but is instead the concentrated and controlling gases, the mode of deposition grades into hydrothermal processes.

For the plainly magmatic ores, be they sulphide or oxides, Graton and McLaughlin[3] have suggested the term *orthotectic,* while for the slightly later processes in which the strictly magmatic influences were modified by mineralizers they would use the expression *pneumotectic.* Naturally, the boundaries are sometimes indistinct. The Sudbury ores would be considered pneumotectic.

In the orthotectic deposits the temperature may have been very high, but in the subsequent, though still magmatic, phases temperatures as low as 400° to 500° C. may have been reached. We are not well informed on this subject.

The magmatic sulphide ores have been discussed by W. H. Goodchild from a physico-chemical standpoint.[4]

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[1] *Econ. Geol.,* 22, 1927, pp. 44–61.

[2] *Idem,* 22, 1927, pp. 288–299.

[3] *Econ. Geol.,* 13, 1918, p. 85.

[4] *Mining Mag.* (London), January to June, 1918.

**Types of Deposits.**—Some of the magmatic sulphide deposits are simply basic rocks abnormal in containing much pyrrhotite, chalcopyrite, and pentlandite. Other occurrences are clearly related to contacts and bear evidence of later magmatic injection. There is still another class in which the magmatic origin is only dimly perceived on account of the metamorphic changes which the rocks have undergone. The basic igneous rocks are easily modified by pressure and more or less schistose amphibolites are developed, which, besides amphibole, contain garnet, quartz, epidote, and chlorite. Any primary sulphide segregations contained in such rocks will be correspondingly affected and a new type of deposit of metamorphic appearance will result; the sulphides themselves apparently undergo little change.

FIG. 308.—Thin section of olivine corroded by pyrrhotite and chalcopyrite, East Union, Maine. Magnified 15 diameters. (*After E. S. Bastin.*)

**Sulphides in Peridotites and Gabbros.**—E. S. Bastin[1] has described a rock from Knox County, Maine (Fig. 308), which shows convincingly the liquid-magmatic origin of sulphide ores. This rock, which occupies a small area surrounded by drift, consists of 60 per cent olivine, 21.53 per cent pyrrhotite, some andesine-labradorite feldspar, hornblende, and magnetite, 1.03 per cent chalcopyrite, and pyrite, biotite, and spinel. The analysis shows 0.94 per cent nickel sulphide, and the material is thus practically a low-grade ore. The constituents are intergrown, showing simultaneous crystallization except that the magnetite, enclosed in olivine, is the earliest mineral separated; the chalcopyrite is associated irregularly with the pyrrhotite; the sulphides corrode the silicates. There has been some later serpentinization but not enough to obscure the relations. The complete analysis is as shown on page 803.

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[1] *Jour. Geol.*, 16, 1908, pp. 124–138.

Vogt has described the numerous Norwegian occurrences in great detail. The ore-bearing intrusives are norites or allied rocks, often with biotite and brown hornblende, and are intruded in pre-Cambrian gneiss. In part the gabbros are pressed to amphibolites. The nickeliferous pyrrhotites occur largely along the contacts. They contain little copper and only 1 to 1.5 per cent nickel. In the amphibolitized deposits considerable migration has taken place. Garnet is formed along the streaks of pyrrhotite. The hornblende is in part transformed to bluish-green amphibole.

| | | | |
|---|---|---|---|
| $SiO_2$ | 28.04 | MnO | 0.24 |
| $Al_2O_3$ | 3.51 | $Fe_7S_8$ | 21.53 |
| FeO | 14.95 | NiS | 0.94 |
| $Fe_2O_3$ | | CoS | 0.03 |
| MgO | 21.97 | $CuFeS_2$ | 1.03 |
| CaO | 1.78 | $H_2O+$ | 2.54 |
| $TiO_2$ | 0.20 | $H_2O-$ | 1.48 |
| $P_2O_5$ | 0.04 | $CO_2$ | 1.01 |
| $K_2O$ | 0.08 | | |
| $Na_2O$ | 0.28 | | 99.65 |

A similar occurrence is that of Lancaster Gap, Pennsylvania, described by J. F. Kemp,[1] where the nickeliferous pyrrhotite lies along the contacts of a mass of amphibolite, contained in mica schist. Much nickel ore was mined here up to 1893.

Many copper deposits in amphibolite are really dynamo-metamorphosed forms of such magmatic deposits as have been described above. The examination of several such small deposits in Colorado[2] led to this conclusion. Deposits at Sedalia and Turret, in Chaffee County, are basic dikes in a pre-Cambrian sedimentary series, contact-metamorphosed by later granitic intrusion and still later altered to amphibolite. Chalcopyrite, sphalerite, and magnetite, are intergrown with bluish-green amphibole, garnet, spinel, and labradorite. The basic dikes are probably differentiated offshoots from a neighboring large mass of coarse diabase.

**Sudbury, Ontario.**[3]—The nickel deposits of Sudbury now yield the largest part of the world's production, and the once preeminent ores

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[1] *Trans.*, Am. Inst. Min. Eng., 21, 1894, pp. 622–631.

T. C. Phemister, *Jour. Geol.*, 32, 1924, pp. 498–510.

[2] W. Lindgren, Notes on copper deposits in Colorado, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 157–174.

[3] The literature is voluminous; only a few articles are cited here:

C. W. Dickson, *Trans.*, Am. Inst. Min. Eng., 34, 1904, pp. 3–67.

A. E. Barlow, Report on the nickel and copper deposits of the Sudbury mining district, *Fourteenth Ann. Rept.*, Canada Geol. Survey, pt. H, 1904.

W. Campbell and C. W. Knight, *Econ. Geol.*, 2, 1907, pp. 350–366.

A. P. Coleman, The nickel industry, Canada Dept. Mines, 1913.

E. Howe, *Econ. Geol.*, 9, 1914, pp. 505–522.

C. F. Tolman Jr., and A. F. Rogers, A study of the magmatic sulfid ores, of New Caledonia are of decreasing importance. Minor amounts of nickel are derived from deposits in Norway and elsewhere. No important nickel deposits have been found in the United States. The production from Sudbury in 1910 was 18,600 long tons of nickel, while in 1930 the nickel content of the matte reached 61,112 tons and was derived from several mines, which yielded 2,115,139 tons of ore. Besides nickel the ores contain an important percentage of copper, with a little gold, silver, palladium and platinum[1] (p. 779). It is not necessary to specify the uses of nickel; they depend on its properties of toughening, whitening, hardening, increasing the elasticity and preventing the oxidation of certain alloys. Nickel is extensively used for electroplating. Nickel steels are the most important of all alloy steels. The ores of the International Nickel Company are smelted and refined in Canada. The ores of the Falconbridge mine, opened in 1929, are smelted locally and the matte exported to Norway for refining.

The geology of the region is complicated. On a basement of Keewatin greenstone and the Sudbury quartzite of Archean age (Timiskaming) rests a syncline of Upper Huronian or Animikie rocks including conglomerate, tuff, and slates. This sycline is 36 miles long and 16 miles wide (Fig. 309). Between the basement and the Animikie there is intruded a thick sheet of igneous massive rocks which may be of Keweenawan age and is also referred to as the "nickel eruptive" on account of its unquestionable connection with the nickel deposits. This sheet is strongly differentiated, supposedly by gravitative settling of crystals in the magma. It grades from a norite or hypersthene gabbro in the lower parts to a micropegmatitic granite in the upper parts. Even in the lower section there is a certain amount of "acid extract" of micropegmatite between the other constituents.

The upper contact with the conglomerate is indistinct in many places. Lately there has been renewed discussion as to whether the differences in composition observed are really wholly caused by crystallization differentiation.[2]

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Stanford Univ., 1916, pp. 21–37.

A. P. Coleman, *Econ. Geol.*, 12, 1917, pp. 427–434; The Sudbury laccolitic sheet, *Jour. Geol.*, 15, 1907, pp. 759–782.

W. G. Miller and C. W. Knight, Nickel deposits of the world, *Rept.* Royal Ontario Nickel Comm., Toronto, 1917.

A. M. Bateman, *Econ. Geol.*, 12, 1917, pp. 391–426.

M. A. Dresser, *idem*, pp. 563–580.

A. Wandke and R. Hoffman, *idem*, 19, 1924, pp. 169–204.

*Eng. Min. Jour.*, Special Sudbury number, Nov. 10, 1930.

[1] In 1931, 44,700 ounces of platinum, 39,000 ounces of palladium, and 7,600 ounces of other platinum metals were produced.

[2] T. C. Phemister, *Jour. Geol.*, 33, 1925, pp. 819–824.

A. P. Coleman, E. S. Moore, and T. L. Walker, The Sudbury nickel intrusive, Univ. Toronto Geol. Studies, 28, 1929, 54 pp.

Granitic rocks are intruded in the basement, and there are even some granite dikes (of Killarney age?) in the norite at the Murray mine and possibly at the Creighton mine.

The deposits are found (1) as rudely tabular masses at the contact of the norite with the basement rocks (marginal deposits); they dip 30° to 60° toward the center of the syncline (Creighton, Gertrude, Murray mines). (2) As mineralized dikes or "offset deposits" in the basement rocks. These are steep, irregular or columnar and have been followed to depths of up to 3,000 feet (Copper Cliff, Frood, Levack, and Victoria mines). The Levack deposit is contained in gneissic granite but lies only 200 feet away from the norite contact to which it is parallel. It contains much pyrite as the earliest mineral. The Copper Cliff is a pipe-like shoot in a quartz diorite. The Frood is a large deposit of rich ore, and at present the greatest asset of the International Nickel Company.

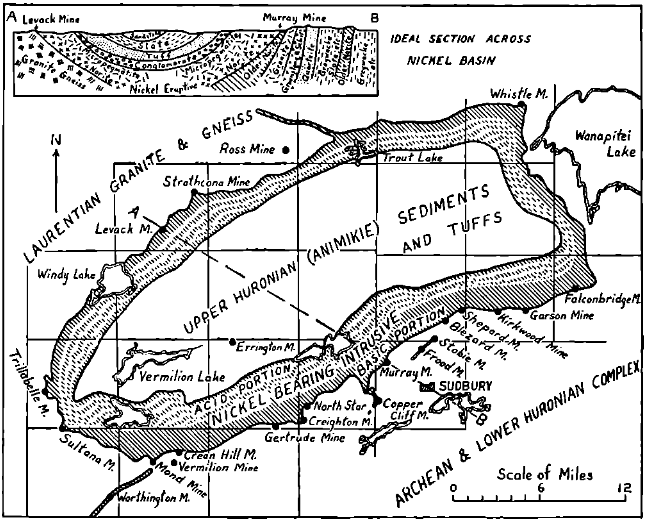

FIG. 309.—Geological map of Sudbury nickel region, Ontario. (*After A. P. Coleman.*)

The ore minerals are pyrrhotite, pentlandite, and chalcopyrite with occasional magnetite, pyrite, sphalerite, and sperrylite ($PtAs_2$). Among the probably secondary minerals are marcasite, millerite, and polydymite ($Ni_4S_5$?). Pyrite, arsenopyrite, sphalerite, and galena begin to appear in the offset deposits. The order of succession as established by Tolman and Rogers is silicates, magnetite, pyrrhotite, pentlandite, and chalcopyrite. Pentlandite often forms veinlets in pyrrhotite and can easily be distinguished from the latter in polished section by etching. The precious metals seem to follow the chalcopyrite and occur most abundantly in some of the "offset deposits," like Victoria and Vermilion.

The ores of the Creighton mine are the richest, containing in per cent about 3 nickel and 1.4 copper. Other deposits yield poorer ore with 2 nickel and 0.8 copper. The values of the precious metals aggregate $1 to $2 per ton.

The best-developed marginal deposit is at the Creighton mine where it has been followed to a depth of 3,500 feet (Fig. 310). It lies between granite and norite; but, as in other marginal deposits, the ore is a breccia or mass of subangular or rounded boulders of almost barren norite, cemented by the ore minerals, which often form a hard crust on the rock.

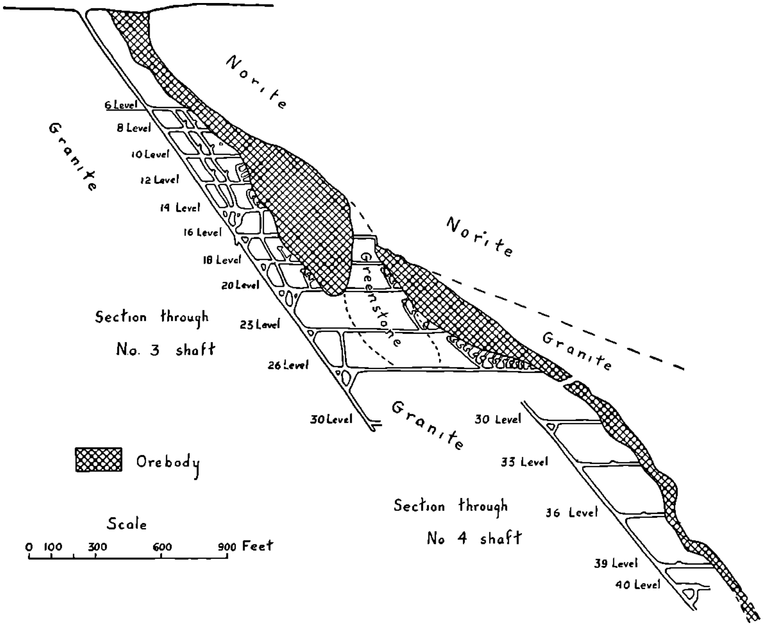

FIG. 310.—Vertical section through Creighton ore-body, Sudbury. (*After R. D. Parker, Eng. and Min. Jour.*)

The sulphides also enter norite and granite as veinlets. The massive ore contains abundant remnants of partly replaced rock minerals, and in the poorer ores the sulphides have corroded and replaced the rock minerals. However, the interpretation of the facts observed under the microscope varies considerably. No gangue minerals are ordinarily formed. The ore-shoots are from a few feet to 150 feet thick. Note similarity of Fig. 310 to sections of Norwegian deposits given by Vogt.

In the "offset deposits" the relations are similar although we find here more evidence of replacement with various gangue minerals, such as chlorite, amphibole, biotite, epidote, and quartz, indicating the action of hydrothermal solutions and lower temperatures.

The Frood mine lies to the south of the main norite contact in metamorphosed sediments and greenstones of the Sudbury series (Timiskaming). It has been known a long time, but rich ore has been disclosed in depth, carrying about 2 per cent of nickel and 10 to 20 per cent of copper. It has been opened to a depth of 3,000 feet and still deeper by drilling. The form is tabular or dike-like narrowing below the 1,200-foot level and continuing thence to 3,000 feet where it is about 150 feet wide (Fig. 311). The ore[1] contains in order of separation: magnetite, pyrrhotite, pentlandite; the remaining solid solution of iron-copper sulphides has unmixed to chalcopyrite, cubanite, and some pyrrhotite. The rock is a dark, medium- to fine-grained diorite with rounded or irregular grains of pyrrhotite and chalcopyrite, which seem to be later than the rock minerals; the latter consists of abundant lath-shaped, green amphibole, much biotite, basic feldspar, and quartz, probably also orthoclase. There are no "hydrothermal" minerals, like sericite, calcite, epidote, etc. The ore-body might well be an injected dike.

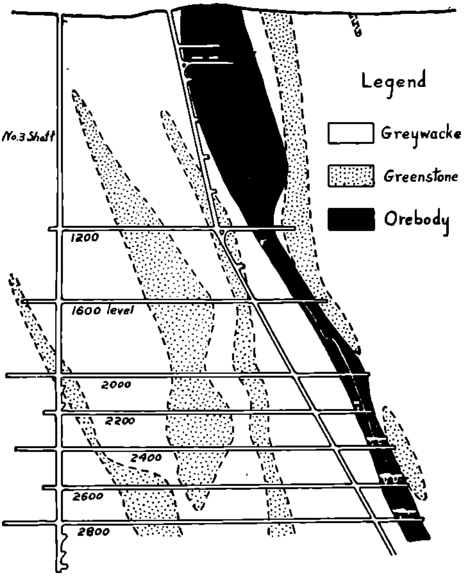

FIG. 311.—Vertical section through the Frood ore-body, Sudbury, Ontario. (*After H. J. Mutz, Eng. and Min. Jour.*)

The earlier view of a gravitative settling of the sulphides in the norite sheet has given way to the theory of an injection of sulphide magma more or less charged with mineralizers along certain brecciated or fractured zones.[2] In places the deposits show transitions to high-temperature veins and to the "injected pyritic deposits" (p. 810). In minor part they may have been formed by direct magmatic segregation from the

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[1] W. H. Newhouse, *Am. Mineralogist,* 16, 1931, pp. 334–336.

[2] See paper by W. H. Collins and E. D. Kindle, The life history of the Sudbury nickel irruptive, *Trans.,* Roy. Soc. Can., 28, 29, 1934, 1935.

nickel eruptive, but in greater part they have been formed at the end of the magmatic period by replacement of the silicates by a very liquid melt charged with sulphides and developed by differentiation in a magma reservoir in depth. Among the supporters of the liquid-magmatic theory of segregation in place are Coleman, Adams, Barlow, Kemp, Vogt, Goodchild; also Roberts and Longyear. Those who have supported the theory of sulphide injection are Knight, Tolman and Rogers, Howe, Bateman, and the staff of the International Nickel Company. Among those who advocate a hypothermal (hydrothermal) origin are Dickson, Knight, Wandke, and Hoffman. That the nickel ores are genetically connected with the norite admits of no doubt.

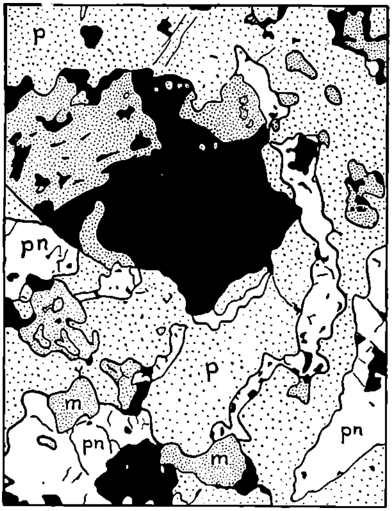

Fig. 312.—Polished section of ore from Creighton mine, Sudbury, Ontario. Black, silicates; *m*, magnetite; *p*, pyrrhotite; *pn*, pentlandite. Magnified 14 diameters. (*After Tolman and Rogers.*)

Much of the trouble experienced in explaining the genesis of the Sudbury deposits is evidently caused by lack of definition of terms. We speak of injection, replacement, corrosion, pneumatolytic stage, sulphide melts with or without mineralizers, and hydrothermal action without actually first defining the terms. In point of fact our knowledge of fundamentals is insufficient, and we try to explain our exact observations in terms of more or less hazily understood principles which can only be elucidated by chemical work at high temperature and pressure. First of all we do not understand how water cooperates with sulphide melts. That some water was present is clearly shown by the occurrence of biotite, amphibole, epidote, and chlorite, perhaps also sericite. On the other hand there was probably very little water compared to that in hydrothermal or even "pneumatolytic" solutions. It does not seem possible that the cases illustrated in Figs. 308 and 313 are simple *hydrothermal* replacements. They appear to be drusy openings in the consolidating magma filled with residual sulphide melt. That there has been some corrosion of the crystals does not invalidate this conclusion. In most of the Sudbury ores we find similar relations of silicates and sulphides (Fig. 312). In other cases again and especially in the offset deposits there is more evidence of ordinary replacement.

Probably there are deposits which represent liquid-magmatic processes of differentiation; certainly there are other deposits which are injected masses of the products of an earlier differentiation. Probably there are also diluted sulphide melts, mixed in some way with water and other mineralizers, which form deposits approaching but not identical with the hypothermal and pyrometasomatic ores.

**South Africa.**—A deposit at Insizwa, South Africa, similar to that of Sudbury, has been described by A. L. Du Toit and W. H. Goodchild.[1] Pyrrhotite, chalcopyrite, and pentlandite, with a little bornite and niccolite, occur in a thick sheet of gabbro or norite, at its contact with underlying sediments. The rock is intrusive in the sediments of the Karroo system, thus probably post-Triassic. The ores contain also a little palladium, osmiridium, gold, and silver. The sulphides separated from the magma at the end of the period of consolidation and corrode idiomorphic olivine (Fig. 313*A*), pyroxene, and feldspar.

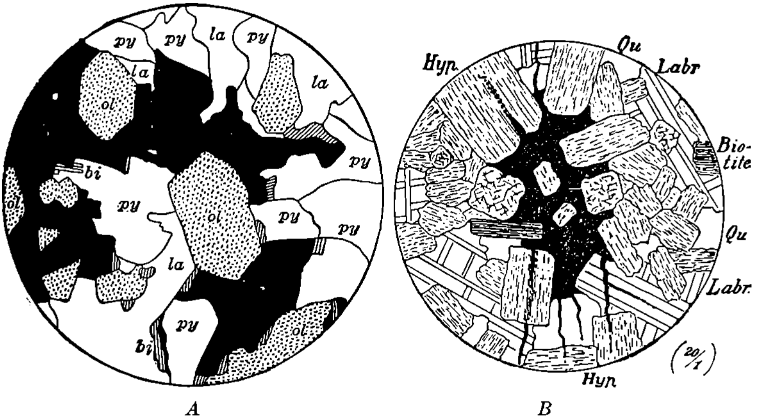

FIG. 313.—*A*. Thin section of olivine norite, Insizwa mine, Cape Colony. Black areas, pyrrhotite; *ol*, olivine; *py*, pyroxene; *bi*, biotite; *la*, labradorite. Magnified 18 diameters. (*After A. L. DuToit.*) *B*. Drawing showing pyrrhotite (black) in quartz-hypersthene-norite, Norway. Magnified 20 diameters. (*After J. H. L. Vogt.*)

In the Merensky platinum region, in the upper part of the norite sheet, in the Bushveld igneous complex, pyrrhotite, chalcopyrite, and pentlandite occur, and, according to P. A. Wagner,[2] are certainly of orthomagmatic origin (p. 782).

**Bornite Deposits.**—Bornite is occasionally recorded as a minor constituent of pegmatite dikes and sometimes occurs in the deep vein zone. A small but remarkable bornite deposit in an igneous rock, described by E. S. Bastin and J. M. Hill,[3] occurs at the Evergreen mine, Gilpin County, Colorado. Dikes of a monzonitic rock contain, intergrown with

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[1] *Fifteenth Ann. Rept.*, Geol. Comm., Cape of Good Hope (1910), 1911, pp. 110–142.
W. H. Goodchild, The economic geology of the Insizwa range, *Trans.* Inst. Min. Met. (London), 1916.

[2] P. A. Wagner, *Econ. Geol.*, 23, 1928, pp. 923–927.

[3] *Econ. Geol.*, 6, 1911, pp. 465–472.

the primary minerals, bornite and chalcopyrite, also garnet, calcite, and wollastonite. All these minerals are contemporaneous with the ordinary rock minerals. This seems to be a case of digestion of material from calcareous rocks; possibly the sulphides are also of foreign origin.

The remarkable and rich bornite deposits of Ookiep in Namaqualand have been regarded as magmatic by Stutzer, Kuntz, and other authors. However, here too, the idea of fractured zones directing or enriching the ores has been brought forward.[1]

Tolman and Rogers[2] have examined these ores and conclude that they are of typical magmatic origin. Magnetite, hematite, chalcopyrite and bornite replace the silicates in hypersthenite and diorite.

The Engels Mine, Plumas County, California, contains disseminated bornite and chalcocite in a mass of norite and slightly later quartz diorite. Some authors[3] have attributed the bornite to strictly magmatic processes while the chalcocite was believed deposited by concentrating thermal waters. Tolman and Rogers[4] regarded the bornite as deposited by the aid of mineralizers at a later magmatic stage.

Graton and McLaughlin[5] classified the deposit as of "pneumatolytic" (perhaps better, "late magmatic") origin with amphibole, albite, tourmaline, and magnetite as gangue minerals and bornite and chalcopyrite as ore minerals. This was followed by hydrothermal action producing chlorite, sericite, epidote and bornite, and finally, zeolites. The chalcocite is regarded as a product of descending meteoric waters.

Knopf and Anderson[6] found that the ore also occurs in the roof pendants of the various intrusions; thus it is post-intrusive and "pneumatolytic." Their conclusion that bornite and chalcopyrite are contemporaneous with the magnetite need not be accepted.

## INJECTED PYRITIC DEPOSITS

**General Features.**—Some deposits in which the ore consists mainly of solid pyritic minerals present features which can hardly be explained otherwise than by actual injection of molten sulphides, perhaps to be considered as residual solutions from adjoining intrusive bodies. A. Bergeat[7] first summarized these peculiar occurrences, among which, it must be confessed, are some of the most enigmatic of ore deposits.

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[1] A Schenk, *Zeitschr.* d. deutsch. geol. Gesell., Vortr., 53, 4H, 1902, p. 64.

[2] A study of the magmatic sulphide ores, Stanford Univ., 1910.

[3] H. W. Turner and A. F. Rogers, *Econ. Geol.*, 9, 1914, pp. 359–391.

[4] *Op. cit.*

[5] L. C. Graton and D. H. McLaughlin, *Econ. Geol.*, 12, 1917, pp. 1–38.

[6] *Econ. Geol.*, 25, 1930, pp. 14–35.
See also C. N. Fenner, *idem*, 25, 1930, pp. 420–425.

[7] A. W. Stelzner and A. Bergeat, Die Erzlagerstätten, 2, 1904, pp. 964–987.

That fluid sulphides may penetrate silicate rocks in veinlets and corrode the various primary minerals, like augite, has been shown in interesting experiments by O. Stutzer,[1] and by previous observations by von Cotta on the brickwork of old lead furnaces. In Stutzer's experiments the sulphide veinlets of pyrrhotite, sphalerite, and galena penetrated the rocks along minute cracks and along the cleavage planes of the minerals. In gabbros, the pyroxene grains were corroded in a manner similar to that noted in the ores of Sudbury. The sulphide melt would probably be under high pressure and would force its way into the adjoining rocks. Deposits of this kind are decidedly rare. The igneous rocks near whose contact injected deposits lie are of many kinds, not always of basic character.

A study of corroded furnace linings led Phemister[2] to conclude that practically anhydrous melts crystallizing as galena, chalcocite, and "matte" are capable of effecting replacement in silica bricks.

The minerals of the injected pyritic deposits include magnetite, pyrrhotite, pyrite, sphalerite, chalcopyrite, and rarely galena. The associated gangue minerals surely indicate high temperature and are present in scant quantity; they are quartz, orthoclase, plagioclase, amphibole, hypersthene, biotite, and spinel. The various minerals are practically contemporaneous. Some investigators would place the Sudbury deposits in this class.

**Norway.**—A number of remarkable pyritic deposits are found in Norway; among them are the well-known localities of Röräs, Vigsnäs, and Sulitjelma, all of which have been the subject of extended discussion. They occur in metamorphic schists, including clay slate, chloritic schist, amphibolite, and in part certainly in dynamo-metamorphosed gabbro intrusions. The ores consist of early pyrite, chalcopyrite, sphalerite and pyrrhotite. A little biotite and magnetite are present.[3] Flat ore lenses prevail, in some places strictly parallel to the schistosity, in other places, as at Röräs, distinctly cutting across it. In large part they are massive pyritic bodies, but the neighboring rock is usually impregnated with pyritic ore. One of the flat ore-bodies at Röräs extended along its dip for 1,900 meters and was 100 meters wide, averaging 8 meters in thickness.[4] At Sulitjelma the contact phenomena have been interpreted as injections. The ore brecciates the schist and enters into it on veins and seams. Feldspar, chlorite, and garnet of the schist are embayed but not sericitized or otherwise altered.

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[1] *Zeitschr. prakt. Geol.*, 16, 1908, pp. 119–122.

[2] T. C. Phemister, *Jour. Geol.*, 33, 1925, pp. 278–285.

[3] H. Ries and R. E. Somers, *Trans.*, Am. Inst. Min. Eng., 58, 1918, pp. 244–264.

[4] Stelzner and Bergeat, 1, 1904, p. 298.

Quartz, actinolite, garnet, epidote, and biotite accompany the ore minerals at some places. Th. Kjerulf and J. H. L. Vogt, among others, consider these deposits as igneous injections, the latter author placing them in genetic association with the gabbro intrusions.[1]

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[1] O. Falkenberg, Geologisch-petrographische Beschreibung einiger südnorwegischer Schwefelkiesvorkommen mit besonderer Berücksichtigung ihrer Genesis, *Zeitschr. prakt. Geol.*, 22, 1914, pp. 105–154.

R. Schreiter, Geologie . . . norwegischer sulfidischer Erzbezirke, *Freiberg Geol. Ges. Bericht*, 15, 1935, pp. 41–52.

# CHAPTER XXXII

## OXIDATION OF METALLIC ORES[1]

### GENERAL CONDITIONS

The upper part of a mineral deposit, within the zone of weathering, is usually more or less altered by surface waters containing free oxygen. The direct effects of this weathering cease in many deposits at the permanent water level, but in deposits of sulphides the indirect effects, due to the action of sulphates generated by the oxidation of primary sulphides, may persist to a considerable depth below the water level. Generally speaking, the zone above the water level is that of the oxy-salts, haloid salts, and native metals; underneath this in many deposits lies a zone of varying depth in which secondary sulphides appear and strong enrichment has taken place. Finally, beneath these zones of extensive changes and molecular rearrangements is found the original or "primary" ore.

The oxidation of mineral deposits is naturally a process analogous to rock-weathering. In deposits free from sulphides the changes are relatively simple, consisting of disintegration, solution, oxidation, and hydration. Siderite alters to limonite, carbonates of manganese to pyrolusite; calcite is dissolved; the rock minerals change to kaolin. The final products are likely to be residual quartz, limonite, pyrolusite, and kaolin. Where native copper is present malachite and cuprite are also found, if the leaching has not been carried too far.

In deposits which contain sulphides, but no pyrite, the changes are rather slow and inconspicuous. Galena changes slowly to anglesite and cerussite, sphalerite to calamine and smithsonite; galena and enargite often remain unoxidized close to the surface. The presence of pyrite, which easily gives off one atom of sulphur, changes and complicates the whole trend of the oxidizing processes. During oxidation the various metals behave very differently and thus many separations are effected.

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[1] For a fuller treatment of this subject see W. H. Emmons, The enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917.

P. Krusch, Primäre und Secundäre Erze, etc., *Compte Rendu*, XII$^e$ Congrès géol. internat., Ottawa, 1914, pp. 275–286; *Zeitschr. prakt. Geol.* 1907, pp. 130–139; *idem*, 1910, p. 165.

H. Schneiderhöhn, Die Oxydations und Zementationszone der sulphidischen Erzlagerstätten, *Fortschritte Mineral. Krist. Petr.*, 1924, pp. 67–160.

## DEPTH OF OXIDATION

Oxidation is a relatively slow process. Some of the more conspicuous cases of deep oxidation have required long geological time. Many copper deposits, where large bodies of oxidized ores are present, were probably attacked by oxidation in Tertiary time. In glaciated areas, such as Canada and northern Europe, oxidation has made little progress since the ice swept away the older accumulations of weathered products, and sulphide ores are usually found close to the surface. Comparatively little effect has been produced by an exposure of several thousand years.

In non-glaciated regions provided with a liberal rainfall the ground-water level is usually less than 100 feet below the surface and the oxidation is correspondingly shallow. On the other hand, in regions with deficient rainfall the ground-water may stand several hundred feet below the surface, and the oxygen has had an opportunity to decompose the ores to a similar depth. At the copper mines of Butte, Montana, the sulphides are oxidized to a depth of at most 400 feet; in the silver-gold veins of Tonopah, Nevada, 700 feet; at Tintic, Utah, in limestone, as much as 1,600 or 2,220 feet. At Bisbee, Arizona, also in limestone, oxidized copper ores are found at a depth of 1,400 feet. At the Durango lead mine, Mapimi, Mexico, the ground-water level is said to stand 2,300 feet below the surface.

In a general way the depth of the water table corresponds to the depth of the oxidized sulphides, but this is a rule with exceptions and qualifications. In most districts sulphide ores may be found in local masses almost or quite at the surface, and, on the other hand oxidation may penetrate to a depth of several hundred feet below the water level. In the Cripple Creek district, Colorado, for example, at the Golden Cycle mine, oxidized ores were found 200 feet below the water level. It is simply a question of the presence or absence of decided local movement of the oxygen-charged water downward along the fissures. The changes are, of course, greatest along the fissures, where oxidation is usually far in advance of the weathering of the general country rock. Changes in climate or elevation with corresponding changes in the water level must not be overlooked. In Arizona we find at many places—Clifton, Globe, and Ray, for instance—zones of secondary chalcocite which were formed below the water level, but which now lie high above the permanent water. On the other hand, elsewhere there is evidence of a depression of the block containing the deposits which may have the effect of raising the water level high above the position it formerly occupied. The facts observed in some districts can be explained only on the supposition of repeated and relatively rapid changes of water level.

The temperature also plays a part. We may expect a deeper oxidation in warm climates than in cold climates simply because heat accelerates reactions. Frequent alternation of moisture and dryness promotes oxidation. Porosity and fissuring of the rocks and ores are factors extremely favorable to oxidation.

In a region of dry climate where mountain ranges are separated by valleys filled with saline deposits, the winds carry the salt to the oxidizing outcrops and the development of chloride of silver, for instance, is facilitated.[1]

The essential factors entering into the problem of oxidation of ore deposits are, then, metal; ore; country rock; fissuring; permeability; climate, water level, and rainfall; topography; geological age and history of deposit.

## OUTCROPS

The outcrops of deposits in glaciated areas are likely to be inconspicuous, except where the principal gangue mineral is unusually hard, like quartz. In non-glaciated regions the outcrop form is determined by the difference in the rate of erosion of the deposit and the country rock.[2] A thick and hard quartz vein, or a mass of solid silicified rock, or garnet rocks in contact-metamorphic deposits will remain as little ridges or series of knobs above the general level of a softer country rock. The quartz veins of California, for instance, are ordinarily easily traceable on the surface. Where the quartz contains much pyrite, a honeycombed or cellular mass of limonite and quartz remains more or less conspicuously above the surrounding country rock. Such weathered outcrops the German miner calls "eiserner Hut," the Cornishman a "gossan," the Australian "ironstone." The Spanish-American gives these oxidized limonite ores the names "colorados," "pacos," "podridos," or "quemados," according to their reddish color, their soft or rotten character, or their burnt appearance.

Where the minerals of the deposit are softer than the country rock, a depression, or little gap, or saddle may mark its outcrop. At Butte, the outcrops of the rich copper veins, which contain little gangue, are inconspicuous, while the silver veins can be easily followed along the surface. Along a single vein there may be great variation in the outcrops. Barren parts tend to stand up prominently, while the ore-shoots, containing softer metallic minerals, may easily become effaced at the surface.

The typical gossan of pyritic bodies, under uniform conditions of high water level and slow erosion, probably remains without much change for long periods. When a gradual lowering of the water level

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[1] C. R. Keyes, Origin of certain bonanza silver ores of the arid region, *Trans. Am. Inst. Min. Eng.*, 42, 1911, pp. 500–517.

[2] W. H. Emmons, Outcrop of ore-bodies, *Min. and Sci. Press*, Dec. 4 and 11, 1909.

and a quickening of the erosion expose new parts of the deposit to the decomposing influence of oxygenated waters, the transfer of material downward becomes more active; and in a copper deposit, it may happen that the surface portion becomes entirely leached of metallic minerals and consists simply of cellular quartz and the more resistant parts of the country rock. Some such outcrops of pyritic copper ores contain scarcely a trace of iron or copper (Fig. 314).

## NOMENCLATURE

The terms primary and secondary as applied to ores are inconvenient and often misleading. They will, therefore, be avoided as far as possible.

A great number of ore deposits are formed by ascending waters. Such waters and such deposits are termed *hypogene*.[1] Most changes during direct and indirect oxidation are caused by descending surface waters. Such waters and the ores formed by them by the rearrangements of the hypogene ores are called *supergene*.

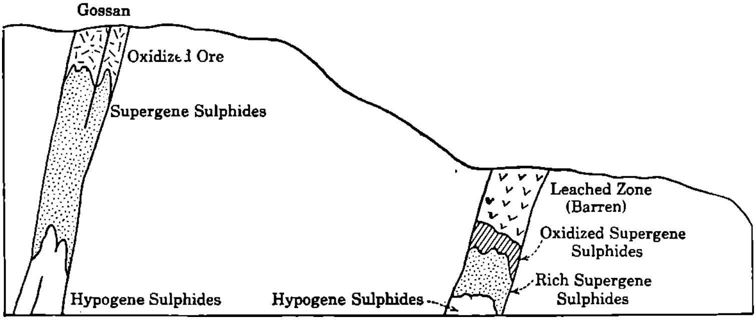

FIG. 314.—Diagram showing normal course of oxidation in pyritic veins and influence of rapid erosion on exposed secondary sulphide zone. In the deposit to the right the gossan has been eroded and the upper part of the secondary sulphide zone leached, causing a thinner but richer secondary zone.

In many cases, valueless but mineralized material has been worked over by descending surface waters and valuable ore-bodies have been formed from it. Ransome has proposed the term *protore*[2] to designate any primary material of too low tenor to constitute ore but which may be concentrated into ore. Thus, the low-grade pyritic dissemination underneath the chalcocite blanket at Ely, Nevada, is protore.

## PRINCIPLES OF OXIDATION

The powerful reagents of oxidation are oxygen, carbon dioxide, and sulphuric acid. The last combines with iron to form ferric and ferrous sulphate, the former being an especially active agent of oxida-

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[1] F. L. Ransome, *Bull.* 540, U. S. Geol. Survey, 1912, p. 152.

[2] W. H. Emmons, *Bull.* 625, *idem*, 1917, p. 68.

tion, while the latter is an important reducing agent. Sodium chloride and sulphuric acid yield hydrochloric acid, which easily combines with iron to make the strongly oxidizing ferric chloride. Under the influence of sulphuric acid the waters change from the calcium carbonate type characteristic of the normal surface conditions to the calcium sulphate type. The aluminous silicates are attacked by sulphuric acid and by carbon dioxide; sulphates, carbonates, and hydrous silicates result.

Insoluble minerals, like cassiterite, wolframite, and often also gold, remain without change in the outcrops, enrich them upon contraction of volume, or on their disintegration are concentrated into placers. Soluble salts, especially sulphates, are carried away. Newly formed compounds are precipitated, chiefly by reactions between carbonates and sulphates or between sulphates. Below a certain point, usually the water level, the free oxygen rapidly diminishes and sulphides are precipitated by reactions between sulphates and sulphides or by other processes.

Much of the dissolved material is naturally removed by the running water of the vicinity, but the greater part of it sinks in the deposit itself and there becomes re-deposited, thus contributing to the general process of enrichment by the descending waters. Some enriched deposits are the product of long-continued descending concentration from a great thickness of rocks now removed by erosion.

In ore deposits free from great amounts of resistant quartz gangue, oxidation obliterates structure. Heavy pyritic deposits appear at the surface as cellular masses of quartz and limonite; the sheeted gold-telluride veins of Cripple Creek, Colorado, which do not carry much quartz, appear as brown, clayey bands without visible structure. There is thorough rearrangement of metal association, and often also segregation of new minerals in large masses. Limestone country rock especially favors such changes. Lead and zinc closely associated in galena and sphalerite part company; the oxidized zinc ores wander farther away from the original deposit than does the cerussite. Copper and iron, so intimate in primary ores, separate more or less in the zone of oxidation, the former exhibiting a centripetal, the latter a centrifugal tendency,[1] and arrange themselves concentrically, just as happens in fragments of sulphide ore subjected to "kernel roasting."

Masses of nearly pure kaolin and alunite often form in this zone.

Some sulphates, like anglesite or basic ferric sulphate, are insoluble; others, like goslarite (Zn), mallardite (Mn), epsomite (Mg), ferrous sulphate, and aluminous sulphate, are most easily carried away. Gypsum, common as a product of exchange in reactions leading to the

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[1] W. Lindgren, L. C. Graton, and C. H. Gordon, The ore deposits of New Mexico, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 55; see also Pl. IV, *B*.

formation of insoluble carbonates in limestone, is also rather easily removed in solution. The native carbonates of zinc and copper are relatively insoluble and may remain for a long time in the gossan. Other minerals characteristic of the oxidized zone are native metals (copper, gold, silver, and mercury), chloride, bromide, and iodide of silver, phosphates, arsenates, antimoniates, molybdates, vanadates, rarely chromates; also hydroxides and oxychlorides; and a few hydrous silicates, like calamine and chrysocolla.

There are then, during oxidation, both dissipation and concentration of metals. The concentration may take place either in the deposit itself or may be effected by the precipitating influence of the country rock on the solutions; bodies of oxidized zinc ores often form in the limestone surrounding a deposit.

In the zone of supergene sulphides below the direct oxidation we meet the copper sulphides—mainly chalcocite and covellite, rarely chalcopyrite and bornite; also, argentite, and complex sulphantimonides and sulpharsenides, associated with native silver. Pyrite, galena and sphalerite are seldom found as products of this zone.

Generally speaking, solution prevails near the surface and precipitation and cementation[1] in the supergene sulphide zone. In the zone of direct oxidation enrichment may or may not take place. If there is a supergene sulphide zone this always involves enrichment. Even in this zone much material, particularly iron, has been carried away.

The character of the solutions changes gradually in depth. Oxygen is removed; the free acid decreases; reduction replaces oxidation; gases like $H_2S$ and $CO_2$ may be generated. The general character of gangue and wall rock is of great importance. If carbonates prevail, the minerals that form may differ from those that are developed in a quartzose gangue. The results show an infinity of variations and complexity.

## TEXTURES AND CRITERIA OF THE OXIDIZED ZONE

The action of meteoric waters, aided by sulphuric acid in pyritic deposits, opens spaces in the zones of oxidation resulting in vuggy, cellular, honeycombed structures, with clayey masses if silicates are present. Deposition proceeds in part by replacement without constancy of volume, in part by crustification in open cavities; mammillary, concretionary, and stalactitic forms of colloidal origin are common, alternating with crusts of crystallized minerals and pseudomorphs stable at the particular moment. Nodular textures are common coupled with rearrangement of the oxy-salts in shells, so that copper ores may surround limonite or zinc ores have a core of cerussite. Reticulating

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[1] P. Krusch, Die Eintheilung der Erze, etc., *Zeitschr. prakt. Geol.*, 1907, pp. 129–139. The word "cementation," used in German literature, is not very apt because the secondary sulphides do not always cement the primary ore.

fractures are filled with oxidized products. Concentric rings of the same products surround sulphides. Colloidal solutions and suspensions are about as common as electrolytes. Solutions change rapidly in composition so that calcite and even quartz crusts may alternate with limonite, kaolin and oxidized salts of copper, lead, and zinc. The volume is diminished and enrichment of relatively insoluble constituents follows. Oxidation tends to thorough change of composition and to obliteration of original texture and structure.

FIG. 315.—Polished section showing oxidation of enargite, Tintic, Utah. White, enargite; dark, copper arsenate; light gray, chalcocite. Magnified 360 diameters.

The presence of limonite with other hydrous oxy-salts is one of the safest criteria of oxidation, but the absence of sulphides is not necessary, for oxidation proceeds extremely capriciously and residual sulphides may be found at all levels in the zone of oxidation. To a limited extent supergene sulphides, like chalcocite, covellite and argentite, may be formed in the oxidized zone wherever there was a local shortage of oxygen (Tintic, Utah).

As oxidation usually has destroyed the sulphides at the surface and often has even carried away the resulting oxy-salts, it evidently becomes a question of great importance to determine from the aspect of the outcrops what sulphides may be expected below the oxidized zone. Oxidation of sulphide ores containing pyrite, chalcopyrite, galena, and sphalerite may result in a boxwork of leached cavities separated by thin walls of secondary jaspery silica. The color and structure of this cellular mass differ according to the original ore, and useful rules have been deduced by Boswell and Blanchard,[1] though they are not always conclusive or easy to summarize in brief.

Locke[2] and White[3] have done much work on the leached outcrops of disseminated chalcocite ores in arid regions. These outcrops are often quite barren at the surface. Such chalcocite ores consist mainly of disseminated pyrite with a coating of chalcocite. Some of the criteria are summarized as follows:

The oxidation of pyrite without copper sulphides in an inert gangue leaves little limonite or results in spreading thin blotches of this mineral. Surface capping with maroon dots of limonite in more concentrated form indicates copper sulphides intergrown with pyrite. Similar results were reached by Boswell and Blanchard. Locke emphasizes the importance of collapse of leached outcrops; this may result in an apparent compression or thinning out of the ore-body. Distinction is made between indigenous and transported limonite.

## TEXTURES OF THE SUPERGENE SULPHIDE ZONE

In the zone of supergene sulphides, important in copper and silver deposits, precipitation and deposition prevail and the textures become more compact, though in places loose, earthy ores are present. The so-called "sooty" chalcocite at Butte and many other places exemplifies this latter condition. The supergene sulphide ores are permeable and more or less porous. With increasing compactness replacement becomes more evident and proceeds volume for volume. The secondary sulphides replace the primary ore minerals in manifold succession and form. Reticulating veinlets are mostly formed by replacement; grains and crystals of pyrite are concentrically replaced by chalcocite (Fig. 319). Sphalerite is similarly replaced by argentite (Fig. 70). The supergene sulphides also fill veinlets or vugs or form thin coatings. Regular "graphic or 'pseudo-eutectic' intergrowths" of chalcocite and bornite similar to eutectic textures in metals are common. They are chiefly hypogene (possibly also supergene) and are believed to result from replacement processes. In Fig. 316 hypogene chalcocite replaces bornite. Thus chalcocite replaces bornite and covellite replaces galena; stromeyerite replaces chalcopyrite, tennantite, and galena; proustite replaces galena. Generally speaking, the supergene sulphides replace primary sulphides but rarely primary gangue minerals, although it is undoubtedly true that chalcocite may occasionally replace quartz.

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[1] B. F. Boswell and R. Blanchard, Notes on the oxidation products derived from chalcopyrite, *Econ. Geol.*, 20, 1925, pp. 613–641; Oxidation products derived from sphalerite and galena, *idem*, 22, 1927, pp. 419–453.

[2] Augustus Locke, Leached outcrops as a guide to copper ores, New York, 1926.

[3] C. H. White, Prospecting for disseminated copper by a study of leached croppings, *Eng. Min. Jour.-Press*, Mar. 22, 1924.

## SOLUTION

In the presence of water oxygen attacks the sulphides and carbon dioxide the silicates. Alkaline solutions would attack the quartz and the silicates but under the influence of free sulphuric acid generated by oxidation of pyrite they are generally absent. Distinction must be drawn between solution and decomposition; most of the changes in the oxidized zone involve decomposition. In general oxidation tends to transform sulphides, sulphosalts, arsenides, tellurides, etc., to oxygen salts and native metals, both of which may be further changed or carried away by surface waters. The silicates in the deposit are changed to a few stable minerals: kaolin, limonite, manganese dioxide and quartz. Carbonates of alkaline earths and alkalies are carried away; original quartz is rarely attacked but new silica from the decomposition of the silicates may be deposited as jasper, opal, and chalcedony.

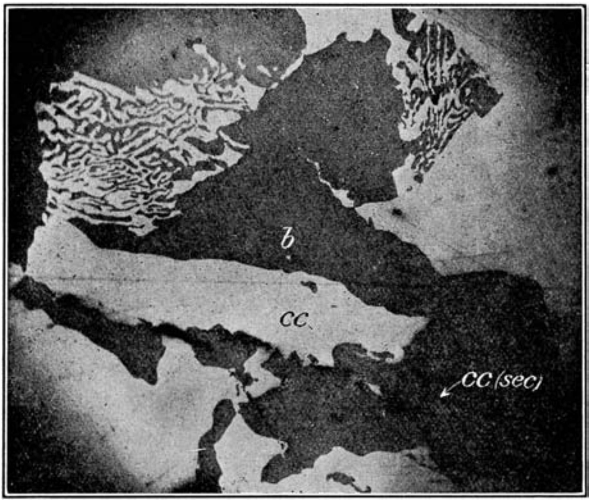

FIG. 316.—Intergrowth of bornite (*b*) and chalcocite (*cc*). Wall mine, Virgilina, North Carolina. Magnified 50 diameters. (*After Graton and Murdoch.*)

The simple sulphides are very slightly soluble in water at 18° C., the solubility decreasing as follows: Mn, Fe, Ni, Cd, Zn, Cu, Pb, As, Ag, Bi, Hg.[1]

In dilute sulphuric acid ($\frac{1}{10}$ normal), pyrrhotite, chalcopyrite, galena, sphalerite, and cadmium sulphide are dissolved or readily attacked. Pyrite, argentite, bornite, arsenopyrite, stibnite, pyrargyrite, and polybasite are slightly attacked,[2] while many others like cinnabar, molybdenite, realgar, orpiment, bismuthinite, covellite, and chalcocite are not attacked.

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[1] Oscar Weigel, *Zeitschr. physikal. Chemie*, 58, 1907, pp. 293–300.

[2] H. C. Cooke, *Jour. Geol.*, 21, 1913, pp. 1–28.

The decomposition or solution is often hastened if the dilute sulphuric acid contains ferric sulphate. Few sulphides resist this reagent.[1]

Some sulphides react with alkaline solutions at ordinary temperature.[2] Orpiment, realgar, stibnite, marcasite, and pyrrhotite are strongly attacked by a 1 per cent solution of $NaHCO_3$; many others are slightly attacked; arsenopyrite, enargite, chalcocite, bornite, light-colored sphalerite, and niccolite are practically resistant.

Experiments with pyrite have not given consistent results. A. N. Winchell[3] exposed powdered pyrite for 10 months to the action of distilled aerated water and obtained a very slow rate of oxidation, the solution containing $Fe_2(SO_4)_3$ and $H_2SO_4$. H. A. Buehler and V. H. Gottschalk[4] obtained a much more rapid attack, and in 3 months the filtrate yielded 2.5 to 3.7 per cent of the original weight of the iron in the sample. Sphalerite in the same time yielded only 0.2 per cent of its zinc, galena 0.005 per cent of its lead, covellite 2.7 per cent of its copper, and chalcopyrite 1 per cent of the same metal. On the other hand, enargite showed no solubility. When the various sulphides were mixed with pyrite the action was much more energetic. In the time specified sphalerite with pyrite yielded 4.2 per cent of its zinc, galena with pyrite 0.7 per cent of its lead, covellite with pyrite 2.7 per cent of its copper, covellite with marcasite 27.6 per cent, and enargite with pyrite 10 per cent of its copper. After an exposure of only 7 weeks, pyrite had oxidized to the amount of 0.1 to 0.28 per cent of its original weight. Walker and Parsons[5] found that a mixture of arsenides was rapidly attacked by aerated water. Carmichael[6] obtained similar results for "polydymite" from Sudbury which was extremely rapidly attacked.

In a second paper, Gottschalk and Buehler[7] show that while in a mixture of two sulphides there is a large increase in the solution of one, there is also a protective action exerted on the other; and further that there exists a difference of potential between the sulphides, which can be arranged in a series similar to the electrolytic series of metals. Acceleration of reaction is due to electric currents generated by contact of minerals of different potential; the currents flow from the mineral of the higher potential, and the mineral of lower potential will dissolve

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[1] For experimental data see G. S. Nishihara, The rate of reduction of acidity of descending waters by certain ore and gangue minerals, *Econ. Geol.*, 9, 1914, pp. 743–757.

[2] F. F. Grout, On the behavior of acid sulphate solutions of copper, silver and gold with alkaline extracts of metallic sulphides, *Econ. Geol.*, 8, 1913, p. 427.

[3] A. N. Winchell, *Econ. Geol.*, 2, 1907, pp. 290–294.

[4] *Idem*, 5, 1910, pp. 28–35.

[5] T. L. Walker and A. L. Parsons, Contributions to Canadian mineralogy, Univ. Toronto Studies, 20, 1925, pp. 41–48.

[6] Ferga Carmichael, *idem*, 22, 1926, pp. 29–37.

[7] *Econ. Geol.*, 7, 1912, pp. 15–34.

more rapidly. In mixtures with pyrite the amount decomposed is but a small portion of that obtained when iron disulphide is treated alone.

From this it follows that the order of oxidation in a mixture of minerals varies with conditions of mass, aggregate, and character of solution. No general rule can be formulated. It is known, however, that, for instance, sphalerite oxidizes before chalcopyrite and the latter before pyrite. In a pyrite-chalcocite ore the chalcocite is attacked first.

The relative solubility of the various carbonates and sulphates is important for the distribution of metals in the oxidized zone. The following data have been determined by Kohlrausch. The low solubilities of the carbonates are considerably increased in the presence of $CO_2$.

Near the surface in pyritic deposits ferric sulphates and even aluminum sulphate and free sulphuric acid may be abundant, but with increasing depth the ferrous sulphate predominates and the solutions tend to become neutral. In general, during oxidation there is a great dissipation of the sulphates of iron, zinc, and calcium.

SOLUBILITY OF SULPHATES AND CARBONATES AT 18° C., IN GRAMS OF ANHYDROUS SALT PER 100 GRAMS OF $H_2O$

| Salt | Grams | Salt | Grams |
|---|---|---|---|
| $BaSO_4$ | 0.00023 | $PbCO_3$ | 0.0001 |
| $PbSO_4$ | 0.0041 | $CaCO_3$ | 0.0013 |
| $CaSO_4$ | 0.20 | $Ag_2CO_3$ | 0.0017 |
| $Ag_2SO_4$ | 0.55 | $BaCO_3$ | 0.0023 |
| $K_2SO_4$ | 11.11 | $ZnCO_3$ | 0.0047 |
| $Na_2SO_4$ | 16.83 | $MgCO_3$ | 0.100 |
| $CuSO_4$ | 19.30 | $FeCO_3$ | 0.073[4] |
| $FeSO_4$ | 23.00 | $MnCO_3$ | .............. |
| $Al_2(SO_4)_3$ | 31.30[1] | $CuCO_3$ | .............. |
| $NiSO_4$ | 34.20[2] | $Na_2CO_3$ | 19.38 |
| $MgSO_4$ | 35.43 | $K_2CO_3$ | 108.00 |
| $ZnSO_4$ | 53.18 | | |
| $MnSO_4$ | 65.00[3] | | |

[1] With 18 mol. $H_2O$ at 0°; 89.1, with 18 mol. $H_2O$ at 100° C.

[2] At 15°.

[3] At 30°.

[4] Sat. with $CO_2$ at 7 atmospheres.

Hydrated sulphates and bicarbonates are more soluble than the normal salts.

## PRECIPITATION

Precipitation is effected by reactions between solutions, by hydrolysis, by coagulation, by gases, and by reactions between the solutions and solids. The last phase is very important. Many reactions take place by the action of solutions on sulphides or on gangue minerals or on country rock of sedimentary or igneous origin. The investigations of E. C. Sullivan[1] have shown that silicates may precipitate oxygen salts by chemical reactions. Thus cupric sulphate lost its copper when filtered through kaolin. This was formerly attributed to adsorption—that is, an accumulation of dissolved substance on the contact between liquid and solid—but Sullivan shows that a chemical change takes place. The natural silicates such as kaolinite, albite, orthoclase, amphibole, pyroxene, and mica precipitate the metals from salt solutions, while at the same time the bases of the silicates are dissolved in quantities nearly equivalent to the precipitated metals. The latter precipitates take the form of hydroxides or basic salts (Fig. 317), though silicates may also be formed to some extent. Thus, by a simple chemical exchange, the metal may be removed from a solution and fixed in the solid state and thus concentrated by contact with even the most insoluble of silicates.

FIG. 317.—Photomicrograph of thin section showing azurite crystals replacing kaolin, Morenci, Arizona. Magnified 15 diameters.

These experiments elucidate the deposition of brochantite and chrysocolla in granitic and porphyritic rocks, as well as the deposition of cuprite and azurite in shale.[2] A solution of silver sulphate yielded its metal completely to a powdered clay gouge, metallic silver being probably formed. The iron in ferric and ferrous sulphate is easily retained by kaolin as limonite.

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[1] The interactions between minerals and water solutions, *Bull.* 312, U. S. Geol. Survey, 1907; *Econ. Geol.*, 1, 1905, p. 67.

[2] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905, p. 191.

C. W. Correns[1] carried these investigations further and concluded by making a distinction between adsorption and chemical decomposition. High-grade copper ore, he states, cannot be produced by processes of adsorption.

The direct oxidation of galena, for instance, yields carbonate and sulphate of lead. By further reactions with ferric sulphate basic sulphates of iron and lead are formed, like plumbojarosite, and in this way the lead is further distributed. Many other difficultly soluble basic sulphates form during oxidation; alunite is one of the most common of these.

The reactions in the oxidizing zone are manifold and complex. Those minerals tend to form that are most resistant to the prevalent acid solutions. Kaolin, hydroxides, insoluble carbonates and sulphates, basic salts, and quartz remain. There is often a considerable precipitation of secondary chert mixed with sulphates.[2] Ferrous sulphate may form siderite ("boxwork siderite in limestone underlying oxidizing sulphide bodies").[3]

## SUPERGENE SULPHIDES

Secondary sulphides are formed by direct precipitation from solutions by means of hydrogen sulphide or other reducing solutions or gases; or from a metasomatic interchange between a solution and a solid, usually another sulphide. Dilute sulphuric acid generated by the decomposition of pyrite, for instance, attacks a few sulphides, with the evolution of hydrogen sulphide. This gas is produced in abundance by the attack on pyrrhotite and, to a less extent, when sphalerite is exposed to the acid. If copper is present in the solutions, a precipitate of cupric sulphide ($CuS$) will be formed, besides some cuprous sulphide ($Cu_2S$). Sulphides are formed mainly where the supply of oxygen from the surface becomes nearly exhausted.

Previous to the year 1900 the presence of secondary sulphides as indirect products of oxidation had been noted by some observers and had been definitely stated by L. de Launay.[4] In the year referred to, S. F. Emmons, C. R. Van Hise, and W. H. Weed in three notable papers[5] formulated the important law of the accumulation of sulphides as a concentration from the overlying oxidized zone, at or below the water

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[1] *Zeitschr. prakt. Geol.*, 1925, p. 19; see also P. Krusch, *idem*, 1921, p. 9; also in *Compte Rendu*, XII session, Internat. Geol. Congress.

[2] E. S. Simpson, *Proc.* Roy. Soc. Western Australia, 9(2), 1923, pp. 45–63.

[3] C. Trischka, O. N. Rove and D. M. Barringer, *Econ. Geol.*, 24, 1929, pp. 677–686.

[4] Les variations des filons métallifères en profondeur, *Revue générale des Sciences*, etc., No. 8, April 30, 1900.

[5] S. F. Emmons, The secondary enrichment of ore deposits; C. R. Van Hise, Some principles controlling the deposition of ores; W. H. Weed, Enrichment of gold and silver veins, *Trans.*, Am. Inst. Min. Eng., 30, 1901.

level. It was shown that in copper deposits chalcocite and covellite were precipitated by pyrite from sulphate solutions and that under similar conditions in silver deposits argentite, stephanite, polybasite, and pyrargyrite or proustite might form; it was also shown that sphalerite and galena were probably precipitated in a similar manner. The chemical reasons for these reactions were found in the so-called Schürmann's law,[1] which states that in the presence of the sulphides of certain of the metals the salts of other metals would be decomposed and the metals precipitated as sulphides. This was thought to indicate that the metals which were precipitated possessed a greater affinity for sulphur than the other metals.

Schürmann's series was as follows: Mercury, silver, copper, bismuth, cadmium, lead, zinc, nickel, cobalt, iron, and manganese. The solution of a salt of any of these metals will be decomposed by the sulphide of any succeeding metal and the first metal will be precipitated as a sulphide. Thus from a solution of silver or copper salts the metal would be precipitated by a sulphide of lead, zinc, or iron. If secondary deposition of sulphides by reaction of pyritic ores on descending sulphate waters had taken place in an ore deposit containing silver, copper, lead, and zinc, these sulphides would theoretically be arranged in the following order: Argentite, chalcocite, galena, and sphalerite, the last at the lowest level. It was shown later by R. C. Wells[2] that the influencing factor was the relative solubility of the sulphides rather than the "affinity for sulphur."

The farther apart any two sulphides are in Schürmann's series the more nearly complete is the replacement. The full series is not represented by natural sulphides, and in ore deposits the reactions of copper and silver solutions are the most important. Supergene sulphides of bismuth and cobalt are not known and supergene cinnabar is rare. On the other hand chalcocite and covellite replace galena and sphalerite as well as pyrite, and they may also replace sulphosalts, such as enargite and tetrahedrite, and iron-copper sulphides like bornite and chalcopyrite; argentite replaces the sulphides of lead, zinc, and iron; galena replaces sphalerite though the reaction is of slight economic importance. The supergene sulphosalts of silver, such as polybasite, replace galena and other sulphides. Complex sulphosalts of lead, *e.g.*, jamesonite, may replace simple sulphides as well as galena.

In the German and English literature one may find statements that tetrahedrite, tennantite, and enargite are supergene products. This does not accord with American experience. Supergene sulphide deposition is accompanied by few characteristic gangue minerals; opal, chalcedony, kaolin, and alunite are occasionally present; quartz is rare.

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[1] E. Schürmann, *Liebig's Ann. der Chemie*, 249, 1888, pp. 326–350.

[2] *Econ. Geol.*, 5, 1910, p. 14.

The rôle of colloidal solutions may also involve the transportation of sulphides. It has long been known that sulphides may be transferred into colloidal solutions by certain dispersing agents. Experiments by John D. Clark and others[1] have shown the extent of this possibility.

Nearly all sulphides, arsenides and sulphosalts may become highly dispersed as colloids under the influence of hydrogen sulphide in solutions of mild alkalinity. These minerals are then in condition to migrate with the solutions. With the escape of $H_2S$ or by contact with calcareous and argillaceous material precipitation and replacement may occur.

Minerals may crystallize directly from colloidal solutions. Upward migrations may thus occur in connection with repeated invasions of hypogene solutions. Where $H_2S$ is locally developed this process may be of some importance in the supergene sulphide zone.

A. F. Rogers[2] has suggested that enrichment by secondary chalcocite has taken place by such ascending solutions, specially citing the case of the Butte deposits.

The temperature during direct oxidation of pyritic ores may in places rise considerably above 50° C. It is probable that even in the supergene sulphide zone fairly high temperatures of 30° or 40° C. may obtain at times.[3]

It has already been stated that in deep oxidized zones supergene sulphides may well form together with oxy-salts. Generally, however, the sulphides begin at the water level and extend for a varying distance below it thus forming a zone of secondary sulphides, which may be many hundreds of feet deep or which may occupy a thickness of only a few feet. Permeability of the primary ore has much to do with this but time and climatic conditions are also potent factors. The water level may have changed its position during geologic times and so we may find chalcocite zones "marooned" high above the water level and now in active process of oxidation. In the same way the products of direct oxidation of a former low water level may now be buried in the underground water.

In such cases physiography may come to the rescue and attempt to elucidate the changes which have taken place in the configuration of the ground.[4]

Such well-defined zones of supergene sulphides are common only in the case of copper and silver. In the case of silver deposits the products of direct oxidation and sulphide deposition are greatly mingled. No great zones of supergene lead or zinc sulphides are known.

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[1] C. F. Tolman, Jr. and John D. Clark, *Econ. Geol.*, 9, 1914, pp. 559–592
John D. Clark and P. L. Menaul, *idem*, 11, 1916, pp. 37–41.

[2] *Idem*, 8, 1913, pp. 781–794.

[3] W. H. Emmons, *idem*, 10, 1915, pp. 151–160.

[4] W. W. Atwood, The physiographic condition at Butte, Montana, *Econ. Geol.*, 11, 1916, pp. 697–740.

Certain elements like iron, zinc and arsenic, which may be common in the primary ore, may be completely eliminated in the oxidized ore and in the supergene sulphides.

## CRITERIA OF SUPERGENE SULPHIDE ENRICHMENT

The question whether or not secondary sulphides have been deposited in an ore-body by descending waters is most important. If the ore minerals are only a part of a shallow enriched layer and poorer ore is to be expected at lower levels, knowledge of this fact is greatly to be desired from the mine owner's standpoint.

The best geological evidence of enrichment consists in the progressive, uniform impoverishment of all similar sulphide deposits in a given district, coupled with the condition that the change in ore should be dependent upon post-mineral topographic development.[1] If the enriched zone is shallow such evidence may be conclusive. If it is deep there may be difficulties in arriving at a correct conclusion.

The occurrence of exceptionally rich ores just below the zone of oxidation is generally suggestive of enrichment. The microscopic structure of the ore is not always reliable because of the similarity of the latest hypogene replacements to those of the descending surface waters.

Generally speaking, it is believed that the presence of chalcocite and covellite in large amounts in the upper levels of a mine is a safe indication of supergene sulphide enrichment, while in silver deposits, the occurrence of rich silver sulphantimonides and argentite just below the zone of oxidation is also, in most cases, a reliable criterion. Replacement starting from the periphery or from cracks and cleavages often indicates supergene deposition.

## IRON

In iron deposits with siderite and iron silicates oxidizing conditions result in abundant limonite. Hematite and magnetite deposits are very slowly oxidized but ultimately form some limonite. Oxidation under tropical conditions generates hematite from ferrous silicates in rocks.[2]

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[1] F. L. Ransome, *idem*, 5, 1910, pp. 205–220.

[2] The oxidized iron ores have been discussed in Chap. XXI. A few references may be added here to indicate the more important foreign literature on the subject.

K. Hummel, Die Eisenerze der Lindener Mark, *Zeitschr. prakt. Geol.*, 1924, pp. 17–23. Residual limonite deposits in limestone.

H. L. F. Meyer-Harrassowitz, Die Verwitterungslagerstätten, *idem*, 1916, pp. 127–136; *idem*, 1921, pp. 65–72.

W. Witte, Die Eisen und Manganlagerstätten bei Oberrossbach, *Neues Jahrbuch*, Beil. Bd. 53-A, 1926, pp. 271–322. An interesting study of a limonite deposit in Devonian slates, caused by descending waters and dependent upon impermeable barriers. Enrichment by colloidal replacement of crushed slates by limonite.

It is known that magnetite may alter to hematite; pseudomorphs (martite) of hematite after magnetite are not uncommon. The nature of this alteration is as yet in some doubt. Some authors[1] noting the abundance of hematite near the surface in some magnetite deposits and the anastomosing veinlets of hematite in the magnetite itself believe that this slight oxidation of a resistant mineral is caused by descending surface waters. Undoubtedly the two minerals may form together independently of oxidation, or hematite may be introduced later than the magnetite during high-temperature processes (p. 701).

In sulphide deposits pyrite, pyrrhotite, and marcasite are the principal iron minerals.[2]

*Pyrite* is a persistent mineral forming in all deposits and at all temperatures even locally at the surface under reducing conditions. It may be reproduced in alkaline solutions, or, with marcasite, in slightly acid solutions.

The oxidation of pyrite is started by oxygen and hastened by the ferric sulphate developed. At the surface it proceeds as follows:

$$FeS_2 + 7O + H_2O = FeSO_4 + H_2SO_4$$
$$2FeS_2 + 15O + H_2O = Fe_2(SO_4)_3 + H_2SO_4$$

This last reaction involves several intermediate stages during which ferrous sulphate, sulphur dioxide, sulphur, or hydrogen sulphide may form. The well-known smell from old dumps containing pyrite indicates the development of sulphur dioxide, according to the equation $FeS_2 + 6O + H_2O = FeSO_4 + H_2SO_3$, and this sulphurous acid is further oxidized to sulphuric acid. Sulphur is often observed near the surface in the casts of dissolved pyrite crystals.

Ferrous sulphate easily changes to the ferric salt and to the hydroxide:

$$2FeSO_4 + H_2SO_4 + O = Fe_2(SO_4)_3 + H_2O.$$
$$6FeSO_4 + 3O + 3H_2O = 2Fe_2(SO_4)_3 + Fe_2(OH)_6.$$

Ferric sulphate hydrolyzes to hydroxide and free acid:

$$Fe_2(SO_4)_3 + 6H_2O = 2Fe(OH)_3 + 3H_2SO_4.$$

The ferric sulphate is a strong oxidizing agent, which attacks pyrite and almost all sulphides:

$$Fe_2(SO_4)_3 + FeS_2 = 3FeSO_4 + 2S.$$

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[1] W. Lindgren and C. P. Ross, The iron deposits of Daiquiri, Cuba, *Trans.*, Am. Inst. Min. Eng., 53, 1916, pp. 40–66.

[2] E. T. Allen, J. L. Crenshaw, and John Johnson, The mineral sulphides of iron, *Am. Jour. Sci.*, 4th ser., 33, 1912, pp. 169–236.

E. Posnjak and H. E. Merwin, The system $Fe_2O_3$-$SO_3$-$H_2O$, *Jour.* Am. Chem. Soc., 44 (2), 1922, pp. 1965–1994.

Sulphur may be oxidized to sulphuric acid. Ferric sulphate changes easily to various basic sulphates, like coquimbite, copiapite, or jarosite, often found in the lower part of the oxidized zone. Limonite is usually the final product. Melanterite ($FeSO_4.7H_2O$) often forms as crusts and stalactites from dripping mine waters.

Some of the numerous hydrous ferric sulphates are stable up to temperature well above 100° C. The most common is copiapite ($2Fe_2O_3.5SO_3.18H_2O$), stable up to 90° C. Jarosite, in which potassium enters, is also very common.

Ferrous sulphate and calcite yield limonite and soluble gypsum. In deposits in limestone this is a most important reaction.

In the deeper levels of the oxidizing zone much or all of the sulphate is likely to be transferred to the ferrous form.

*Marcasite* forms only in acid solutions, but may then crystallize together with pyrite. Above 450° C. it passes into pyrite. Low temperature and free acid favor its development. Marcasite is thus a relatively unstable mineral formed mainly near the surface. A few ore deposits of igneous affiliations like that at Goldfield, Nevada, contain marcasite, but it is usually a very late product forming in fissures and vugs. It is most prominent in the lead-zinc ores of the Mississippi valley type. Nevertheless it is not found among the supergene sulphides probably because descending sulphates are likely to attack it vigorously.

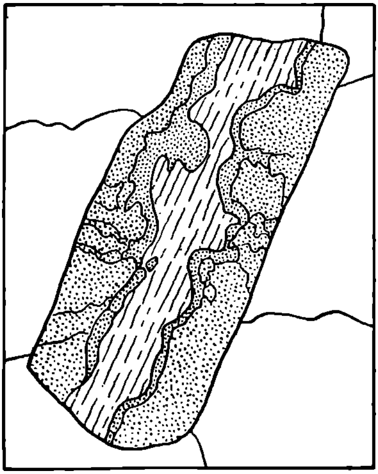

FIG. 318.—Polished section of crystal of pyrrhotite replaced marginally by colloform marcasite (stippled). In crystalline limestone (white). Magnification 350 diameters. (*After R. Kuhne.*)

It oxidizes much more easily than pyrite but the reactions are the same.

*Pyrrhotite*, regarded as a solid solution of sulphur in FeS, is in nature a high-temperature mineral not known to occur in ores of shallow or intermediate depths. Allen and Crenshaw made it, however, by synthesis in slightly acid solutions at temperatures as low as 80° C., and exceptionally it might form at even lower temperatures. It is readily attacked by dilute $H_2SO_4$ with evolution of $H_2S$, which in copper deposits may precipitate copper sulphide and prevent the development of deep chalcocite zones. It is also easily attacked by oxidation, the $H_2SO_4$ formed accelerating its destruction.

$$Fe_7S_8 + 31O + H_2O = 7FeSO_4 + H_2SO_4.$$

In the lower part of the oxidized zone, pyrrhotite, according to many recent observations,[1] is likely to be transformed into colloform marcasite Fig. 318. Such marcasite also replaces sphalerite.

Hydrous ferric silicates, like hisingerite and morencite, often form in the lower part of the oxidized zone and probably also below the water level.[2]

## COPPER

**Minerals.**—The most common primary sulphides of copper and iron include chalcopyrite ($CuFeS_2$) and bornite ($Cu_5FeS_4$). The pale yellow mineral cubanite ($CuFe_2S_3$)[3] is not uncommon and is generally a product of "unmixing" of pyrrhotite or chalcopyrite (p. 801). The soft valleriite[4] ($Cu_2Fe_4S_7$) may have a similar origin. Tetrahedrite ($Cu_8Sb_2S_7$) with its arsenical analogue tennantite, and enargite ($Cu_3AsS_4$), are important copper ores in some places. Bournonite ($CuPbSbS_3$) is sometimes an ore mineral, and there are a number of rare copper and lead-copper sulphbismuthides. Native copper is an important primary ore mineral in some districts while the arsenides are rare minerals. Chalcocite ($Cu_2S$) is in places of hypogene origin.

The supergene copper minerals are very numerous. Those of some economic importance include native copper, cuprite ($Cu_2O$), several indefinite minerals of colloidal origin containing $CuO$, $MnO_2$, $ZnO$, $SiO_2$, and $H_2O$ (copper pitch ores), also the oxychloride atacamite ($Cu_2Cl(OH)_3$) which is found where wind-blown sodium chloride is available.

There are further chalchanthite ($CuSO_4.5H_2O$) with its related minerals kröhnkite ($CuSO_4.Na_2SO_4 + 2H_2O$) and natrochalcite ($Na_2SO_4.Cu_4(OH)_2(SO_4)_2 + 2H_2O$), all important minerals in arid countries like Chile.[5] Phosphates are rare.

Malachite ($Cu_2(OH)_2CO_3$), azurite ($Cu_3(OH)_2(CO_3)_2$), and chrysocolla ($CuSiO_3.2H_2O$) are the most abundant of all the oxidized copper minerals. Basic sulphates, like brochantite ($Cu_4SO_4(OH)_6$), are locally abundant as are various arsenates most common among which is olivenite ($Cu_3As_2O_8.Cu(OH)_2$).

The secondary sulphides of prime importance are chalcocite ($Cu_2S$) and covellite ($CuS$). Bornite and chalcopyrite are rarely supergene.

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[1] G. Gilbert, Oxidation and enrichment at Ducktown, Tenn., *Trans.*, Am. Inst. Min. Met. Eng., 70, 1924, pp. 998–1023.

W. H. Newhouse, Paragenesis of marcasite, *Econ. Geol.*, 20, 1925, pp. 54–66.

W. Lindgren, *Bull.* 782, U. S. Geol. Survey, 1926, p. 140.

[2] D. F. Hewitt and W. T. Schaller, *Am. Jour. Sci.*, 5th ser., 10, 1925, pp. 29–38.

[3] B. L. Johnson, *Econ. Geol.*, 12, 1917 pp. 519–525.

[4] P. Ramdohr and O. Ödman, *Geol. För. För.*, 54, 1932, pp. 89–98.

[5] The numerous hydrous copper sulphates have been investigated by E. Posnjak and G. Tunnell (*Am. Jour. Sci.*, 5th ser., 18, 1929, pp. 1–34). They obtained crystallized brochantite at 100° C. and antlerite ($3CuO.SO_3.2H_2O$) from 50° to 200° C.

**Solution and Precipitation.**—Copper is one of the most easily transported metals and as it is also easily precipitated its supergene deposits have great importance. Copper migrates downward by stages through the oxidized zone and through the supergene sulphide zone so that a considerable concentration may eventually be reached.

Chalcopyrite is readily attacked by oxygen and by ferric sulphate. It is slightly attacked by dilute sulphuric acid.

$$CuFeS_2 + 8O = CuSO_4 + FeSO_4.$$

As the ferrous salt is easily transformed into limonite, pseudomorphs of that mineral after chalcopyrite are extremely common.

Bornite is more strongly attacked by dilute sulphuric acid than chalcopyrite,[1] but like all copper sulphides is easily decomposed by ferric sulphate.

$$Cu_5FeS_4 + 2H_2SO_4 + 18O = 5CuSO_4 + FeSO_4 + 2H_2O.$$

Chalcocite like covellite is very slightly attacked by dilute sulphuric acid but is decomposed by ferric solutions which transform it to sulphates, probably also to covellite:

$$Fe_2(SO_4)_3 + Cu_2S = CuSO_4 + 2FeSO_4 + CuS.$$

Enargite and tetrahedrite are likewise slowly decomposed by solutions of ferric sulphate. Normally the antimony remains as insoluble oxide while the arsenic is carried away unless fixed as arsenates of copper by carbonate solutions or limestone.

The universally resulting cupric sulphate[2] is more or less completely fixed as malachite, azurite, brochantite, chrysocolla and similar products by solutions containing carbonates or silica. Ultimately even these minerals will be leached, for they are slightly soluble in water containing carbon dioxide and easily soluble in dilute sulphuric acid. They may, however, be reduced to cuprite and the cuprite to native copper which again may go into solution with $H_2SO_4$.

The oxidation products of the chalcocite zone are described on page 842.

**Supergene Copper Sulphides.**—So far we have considered the products of direct oxidation, in the uppermost part of pyrite deposits; they form under the influence of acid solutions containing free oxygen and ferric salts; limonite and oxy-salts of copper result, mixed with residual quartz. At a certain depth, usually at the water level, if that remains in its

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[1] E. G. Zies, E. T. Allen, and H. E. Merwin, Some reactions involved in secondary copper sulphide enrichment, *Econ. Geol.*, 11, 1916, p. 476.

[2] Cuprous sulphate may form as an intermediate product in some reactions, but it is unstable and according to W. H. Emmons has not been discovered in any analysis of mine waters.

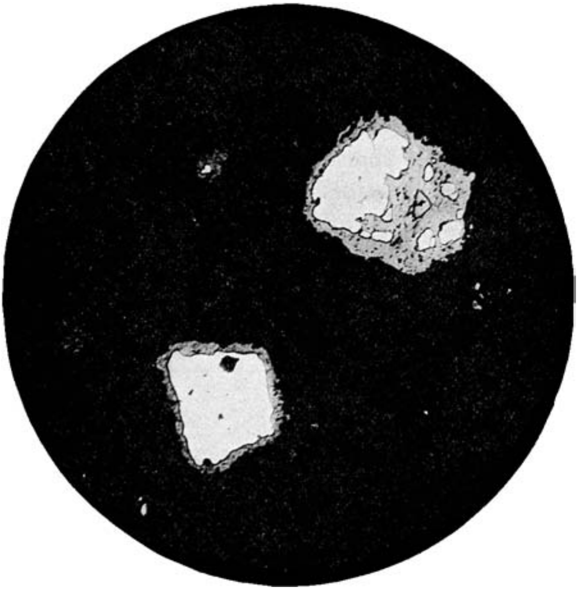

FIG. 319.—Chalcocite (gray) replacing pyrite (white), Ray, Arizona. Magnified 150 diameters. (*After L. C. Graton.*)

FIG. 320.—Supergene chalcocite (gray) replacing pyrite (white), Ray, Arizona. Magnified 120 diameters. (*After L. C. Graton.*)

original horizon, the material changes from a brown to gray or bluish color and copper sulphides begin to appear. At first they form pulverulent or sooty masses with little residual pyrite. Their quantity gradually decreases; we find grains of pyrite, chalcopyrite, bornite, or sphalerite covered by coatings of covellite or chalcocite, and the microscope gives evidence that the process is a replacement of primary sulphides by the two minerals mentioned (Fig. 319). Occasionally secondary chalcopyrite or bornite will appear but no secondary pyrite. The alteration proceeds from reticulating veinlets (Figs. 320 and 323) along grain boundaries, or in concentric shells. Below the upper part of the zone the chalcocite may be compact with dark gray metallic luster.

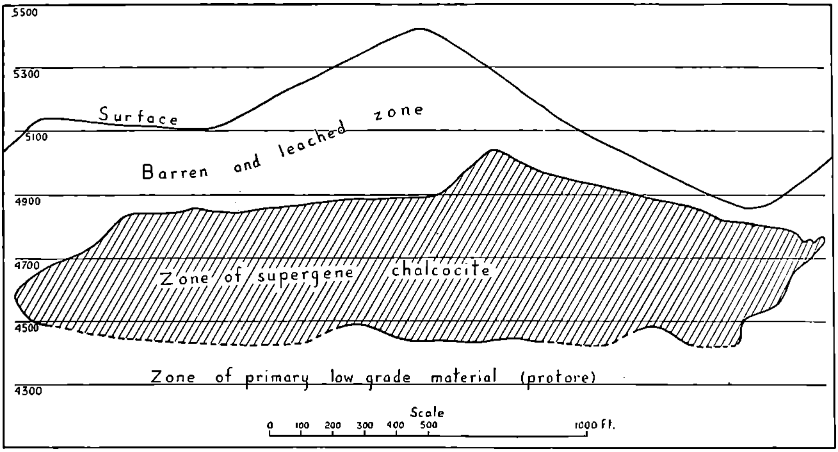

FIG. 321.—Longitudinal section of a chalcocite zone at Morenci, Arizona.

Veins of pyrite may be converted to sooty or massive chalcocite with only a few residual grains of the original mineral. In some deposits large and rich masses of chalcocite, more rarely covellite, may be formed in this manner.

Kaolin may form along with the secondary sulphides partly at the expense of the sericite; other minerals are chalcedonic and opaline silica, also alunite. Pyrite is unstable in this zone. Deposition of quartz is very unusual, though silicates are decomposed by the $H_2SO_4$ present and much silica is carried off by the acid solutions.[1]

The upper limit of the chalcocite zone is usually sharp. In depth, the secondary sulphides may cease equally suddenly (Fig. 322), but it is more common to find a gradual decrease. Chalcocite zones of wide extent are usually explored by churn drilling, and the plotting by graphic methods of the assays obtained gives a clear idea of the sharp changes or

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[1] T. S. Lovering, The leaching of iron protores, *Econ. Geol.*, 18, 1923, pp. 523–540.

gradual transitions due to the enrichment.[1] The depth of the chalcocite zone often reaches 1,000 feet and in exceptional cases considerably more. On the other hand, the enrichment may be confined to a thin layer with a thickness of a few feet only (Fig. 322).

The development of the chalcocite zones is dependent on climate and water level as well as on composition and texture of the primary rocks and ores. Permeable rocks like sericitized granite, porphyry, or schists are particularly favorable environments. In compact contact-metamorphosed shales and limestones, the secondary sulphides do not readily develop. In limestone the zone is usually irregular and shallow because the basic copper carbonates here form so easily.[2] Wherever the chalcocite zone is present a marked enrichment has taken place.

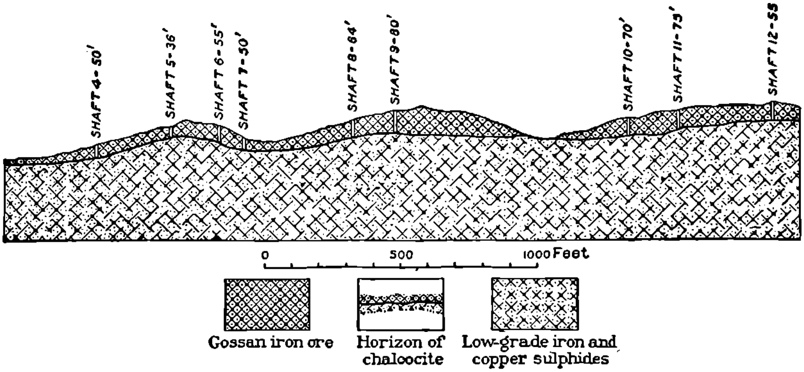

FIG. 322.—Chalcocite zone at Ducktown, Tennessee. (*After W. H. Emmons, U. S. Geol. Survey.*)

Chalcocite zones may develop in primary ores of economic value. They may also form by enrichment of low-grade material (protore, p. 816) whether this be contained as heavy pyrite in veins of low tenor in copper, as at Clifton, Arizona; or as disseminations of pyrite and chalcopyrite in larger mineralized areas as at Ely, Nevada, Miami, Arizona, and many other places. Such enrichments, usually of no greater thickness than 100 to 300 feet, but of considerable horizontal extent, are often referred to as "chalcocite blankets" and rarely contain more than 2 or 3 per cent of copper, except perhaps in their uppermost levels where progressive enrichment has been proceeding.

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[1] E. H. Perry and A. Locke, Interpretation of assay curves for drill holes, *Trans.*, Am. Inst. Min. Eng., 54, 1917, pp. 93–99. See also graphs in copper deposits of Ray and Miami, F. L. Ransome, *Prof. Paper* 115, U. S. Geol. Survey, 1919.

[2] A. C. Spencer (*Prof. Paper* 96, U. S. Geol. Survey, 1917, p. 82) states, however, that calcite does not precipitate copper carbonates from a solution of cupric and ferrous sulphate; also that secondary copper sulphides may form on pyrite and chalcopyrite in the presence of large amounts of calcite.

The secondary copper sulphides are not necessarily confined below the water level. They may be deposited at any place in the oxidized zone where there is a deficiency in oxygen and ferric sulphate, as well shown at Tintic, Utah, and other places.[1] Such supergene sulphides are, however, likely to be spotted and irregular in occurrence.

A lowering of the water level and resulting oxidation of the chalcocite causes a progressive enrichment and an enlargement of the zone, for wherever the cupric sulphate reaches the primary ore, fresh cupric or cuprous sulphide will form. Sphalerite, galena, and bornite are easily attacked, galena more easily than any other sulphide; next follows chalcopyrite, while pyrite is not readily replaced by secondary sulphides as long as the other minerals are present. The whole process involves removal of iron on a large scale. Zinc and arsenic are also carried away.

F. L. Ransome[2] presents analyses showing the chemical changes in the rocks during the mineralization of the original material to protore and during the chalcocitization of this protore.

| | 1 | 2 | 3 |
|---|---|---|---|
| $SiO_2$ | 66.84 | 66.92 | 70.63 |
| $Al_2O_3$ | 17.46 | 19.23 | 14.02 |
| $Fe_2O_3$ | 4.88 | 0.54 | 2.47 |
| FeO | 0.74 | 0.99 | 0.95 |
| $FeS_2$ | None | 1.40 | 0.52 |
| $Cu_2S$ | None | 0.79 | 2.56 |
| MgO | 1.14 | 0.97 | 0.70 |
| CaO | 0.37 | 0.27 | 0.13 |
| $Na_2O$ | 0.53 | 0.39 | 0.41 |
| $K_2O$ | 4.65 | 5.61 | 4.93 |
| $H_2O-$ | 0.17 | 0.24 | 0.14 |
| $H_2O+$ | 2.61 | 2.45 | 2.41 |
| $TiO_2$ | 0.67 | 0.58 | 0.61 |
| $CO_2$ | None | ...... | ...... |
| $P_2O_5$ | 0.03 | 0.08 | 0.13 |
| MnO | 0.02 | ...... | 0.01 |
| | 100.11 | 100.46 | 100.62 |

[1] Pinal schist (pre-Cambrian), Miami district, 1,500 feet west of Warrior mine. George Steiger, analyst.

[2] Metallized schist (protore). Inspiration mine. Chase Palmer, analyst.

[3] Chalcocite ore, Miami mine, 420-foot level. Chase Palmer, analyst.

The protore varies widely as to the amount of $FeS_2$ present. It is difficult to find protore entirely unaffected by chalcocitization. There

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[1] W. Lindgren, *Econ. Geol.*, 10, 1915, p. 236.

[2] *Prof. Paper* 115, U. S. Geol. Survey, 1919, p. 159.

is probably some chalcopyrite in 2. The changes in composition are surprisingly small; in a measure this is because the Pinal schist is rich in potash mica and sericite is the mineral predominating in the altered rocks.

**Theory of Supergene Copper Sulphides.**[1]—The recent literature on the subject of sulphide enrichment in copper deposits is voluminous, and it is not possible to follow it in detail in this place. The deposition is governed by Schürmann's reactions (p. 826) so that in general the simple sulphides of copper replace the simple sulphides of iron, lead, and zinc (Fig. 323). They also replace the copper-iron sulphides like chalcopyrite

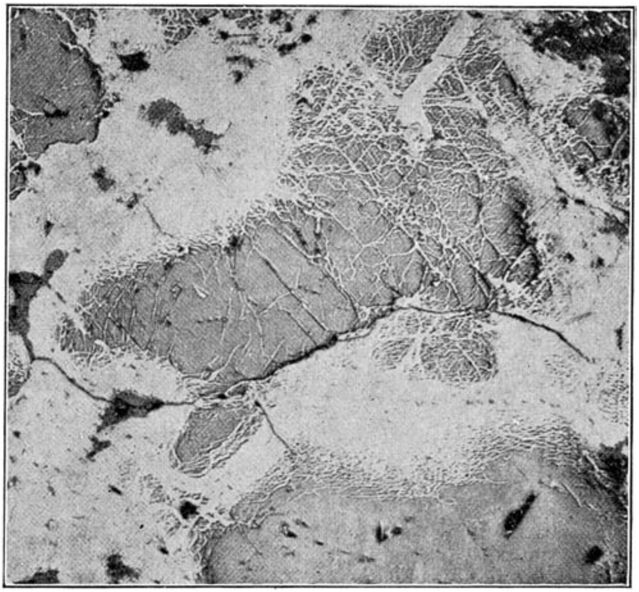

FIG. 323.—Polished section showing supergene chalcocite (light) replacing sphalerite (dark), Tsumeb mine. Magnified 80 diameters. (*After H. Schneiderhöhn.*)

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[1] H. N. Stokes, On the solution, transportation and deposition of copper, silver and gold, *Econ. Geol.*, 1, 1906, pp. 644–650. Also, *idem*, 2, 1907, pp. 14–23.

E. Posnjak, E. T. Allen, and H. E. Merwin, The sulphide ores of copper, *idem*, 10, 1915, pp. 491–535.

E. G. Zies, E. T. Allen, and H. E. Merwin, Some reactions involved in secondary copper sulphide enrichment. Contribution No. 7. Secondary enrichment investigation, *idem*, 11, 1916, pp. 407–503.

C. F. Tolman, Jr., Observations on certain types of chalcocite, etc., *Trans.*, Am. Inst. Min. Eng., 54, 1917, pp. 402–442.

L. C. Graton and J. Murdoch, The sulphide ores of copper, *idem*, 45, 1914, pp. 126–181.

A. C. Spencer, Geology and ore deposits of Ely, Nevada, *Prof. Paper* 96, U. S. Geol. Survey, 1917, pp. 76–91. Excellent review of subject.

W. H. Emmons, The enrichment of ore deposits, *Bull.* 625, *idem*, 1917, pp. 154–249. Excellent and complete review. Bibliography on pp. 20–33.

and bornite and the sulphantimonides like tetrahedrite and the sulpharsenides like tennantite and enargite. There is not much evidence of deposition of copper sulphides by precipitation by hydrogen sulphide or by alkaline sulphides though no doubt such reactions may also take place, especially when pyrrhotite is one of the primary minerals.

Near the surface the mine waters are solutions of sulphuric acid and ferric sulphate; in depth their acidity decreases and ferrous sulphate increases; at greater depth the waters become neutral and finally alkaline.[1] Cupric sulphate is present all along but the secondary sulphides are evidently not readily precipitated in the presence of much ferric sulphate though they are normally found in the presence of ferrous sulphate of slight acidity. In 1906 Stokes had established quantitatively the reaction of chalcocite with pyrite. In the notable paper by E. G. Zies, E. T. Allen, and H. E. Merwin of the Geophysical Laboratory all of the various reactions were quantitatively determined at temperatures of 40° C. and 200° C.

The reaction with sphalerite is as follows:

$$ZnS + CuSO_4 = CuS + ZnSO_4.$$

The presence of sulphuric acid accelerates the reaction. When cupric sulphate and galena react at 35° C. cupric sulphide is first formed which is further attacked by cupric sulphate yielding cuprous sulphide. The attack on chalcopyrite at 40° C. and at 200° C. is expressed by the equation

$$CuFeS_2 + CuSO_4 = 2CuS + FeSO_4.$$

In this reaction the cupric sulphide again alters to cuprous sulphide on further attack by $CuSO_4$. The presence of sulphuric acid does not retard the reaction.

The action between bornite and cupric sulphate at the same temperature is expressed by the equations

$$5Cu_5FeS_4 + 11CuSO_4 + 8H_2O = 18Cu_2S + 5FeSO_4 + 8H_2SO_4.$$
$$Cu_5FeS_4 + CuSO_4 = 2Cu_2S + 2CuS + FeSO_4.$$

Bornite is attacked by $H_2SO_4$ resulting in CuS and $Cu_2S$, and $FeSO_4$, hydrogen sulphide developing at the same time. These products will react and form secondary chalcopyrite.

Pyrite alters to chalcocite and covellite according to Stokes' reaction:[2]

$$5FeS_2 + 14CuSO_4 + 12H_2O = 7Cu_2S + 5FeSO_4 + 12H_2SO_4.$$

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[1] G. S. Nishihara, The rate of reduction of acidity by descending waters, etc., *Econ. Geol.*, 9, 1914, pp. 743–757.

F. F. Grout, On the behavior of cold, acid solutions, etc., *idem*, 8, 1913, p. 429.

[2] Stokes verified this reaction at 180° and 100° C. with neutral solution. Some CuS was also formed, less at 100° than at 180°. Cuprous sulphate also forms as an intermediate product. Cupric and ferrous sulphate mix in all proportions without

The formation of covellite is expressed by the following reaction:

$$4FeS_2 + 7CuSO_4 + 4H_2O = 7CuS + 4FeSO_4 + 4H_2SO_4.$$

Sulphuric acid exerts a markedly retarding influence on these reactions.

According to Zies, Allen, and Merwin pyrrhotite alters to chalcopyrite and probably later to bornite when attacked by cupric sulphate but the reaction was not followed quantitatively. Most observers have assumed that covellite is earlier than chalcocite and Zies, Allen, and Merwin confirm this experimentally:

$$5CuS + 3CuSO_4 + 4H_2O = 4Cu_2S + 4H_2SO_4.$$

It is probable that this reaction is reversible for in many cases covellite is an alteration product of chalcocite.

Cupric sulphide is formed when cuprous sulphide is treated with dilute acid solution in the presence of oxygen. This probably explains the development of covellite in oxidizing chalcocite and its occasional crystallization together with products of oxidation like anglesite.

**The Crystallization and Relations of Chalcocite, Bornite, and Covellite.**—According to Posnjak, Allen, and Merwin[1] chalcocite crystallizes in the orthorhombic system, in striated prisms, up to a temperature of 91° C. By far the greatest number of crystals found in nature (they are not common) have this form. Above 91° C. chalcocite crystallizes in the isometric system usually in octahedral form. As isometric chalcocite passes through the inversion point a rearrangement of the molecules occurs and the crystal becomes an aggregate of orthorhombic grains. Chalcocite may hold covellite in solid solution; more or less is frequently present. When the amount exceeds 8 per cent there is no inversion point. Pure chalcocite is white in reflected light but when admixed with CuS it appears bluish. In some cases this "solid solution" is resolved into a mixture of covellite and chalcocite, when examined under high-powered objectives.

At first glance the inversion point would appear to be a suitable thermometric point for the distinction of supergene and hypogene chalcocite. It is admitted that all supergene chalcocite must have been formed below 91° C.; but it is not at all certain that all hypogene chalcocite must be isometric. The hypogene chalcocite is always developed at the latest stages of mineralization by ascending waters and, conceivably, well below 91° C.

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change, except at high temperatures (200° C.), when according to Stokes, cuprous sulphate and ferric sulphate form; the latter hydrolyzes to ferric hydrate and $H_2SO_4$, while cuprous sulphate deposits copper upon cooling. This reaction is not likely to take place at temperatures ordinarily existing in the oxidized zone.

[1] *Op. cit.*

The distinctly supergene chalcocite develops by peripheral replacement or along reticulating cracks (Figs. 319 and 320) and etching shows it to have always a granular texture, each grain being striated in one direction probably that of (001) (Fig. 324).

The merit of having recognized that chalcocite may be primary belongs to F. B. Laney[1] and to Reno Sales.[2] From a mineralogical standpoint the same has been proved by H. Schneiderhöhn,[3] at Tsumeb; by A. M. Bateman and D. H. McLaughlin,[4] at the Kennecott mines; by Bateman,[5] at Bristol, Connecticut; and by A. Locke, D. A. Hall and M. N. Short,[6] at Butte. Certainly the conditions at Butte are convincing. In smaller amounts hypogene chalcocite intergrown with bornite occurs in the deep levels of the United Verde mine. The question is now; of what nature is this hypogene chalcocite? It always occurs with bornite and is usually plainly later than the bornite. It shows a freedom of control by obvious channels; it includes corroded relics of older sulphides; it shows "graphic" or "eutectic" intergrowths; bornite

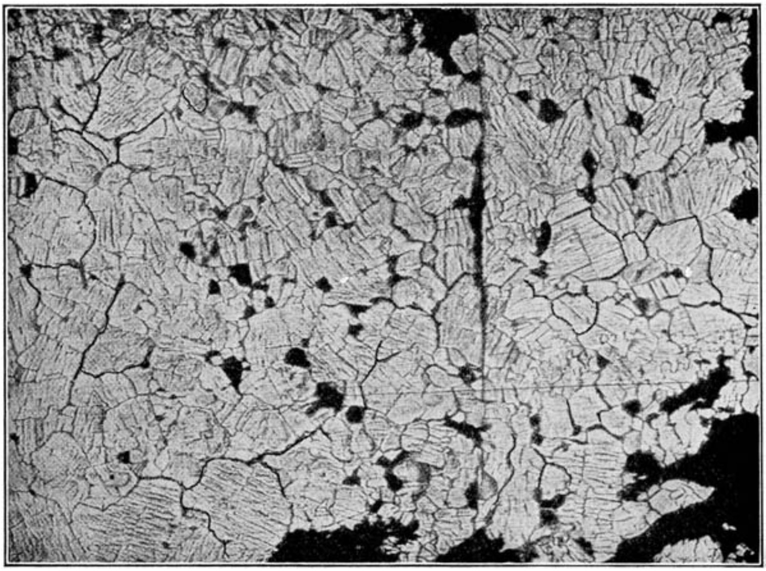

FIG. 324.—Photomicrograph of polished section showing etch pattern of rhombic (supergene) chalcocite, Tsumeb mine. Magnified 96 diameters. (*After H. Schneiderhöhn.*)

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[1] *Econ. Geol.*, 6, 1911, pp. 399–411.

[2] *Trans.*, Am. Inst. Min. Met. Eng., 46, 1914, pp. 3–106.

[3] Senckenbergiana, Bd. 2, Fasc. 1, Frankfurt a. M., 1920. *Metall und Erz*, 17 (13, 16, 19, 24), 1920; 18 (10, 11), 1921.

[4] *Econ. Geol.*, 15, 1920, p. 63.

[5] *Idem*, 19, 1923, pp. 122–166.

[6] *Trans.*, Am. Inst. Min. Met. Eng., 70, 1924, pp. 933–963.

is often minutely intermixed with it; obvious replacement relations with bornite are common.

Now Schneiderhöhn appears to have proved that chalcocite with octahedral cleavage occurs at Tsumeb, below the supergene zone (Fig. 325). This chalcocite is lamellar parallel to (111) and the individual lamellae show rearrangement into etch figures indicative of the orthorhombic system. This would show that it was formed above 91° C. but became orthorhombic after passing the inversion point.

At Kennecott, Merwin found several octahedral twins which contained over 8 per cent dissolved covellite. At Butte the distinctly orthorhombic crystals so far found are probably hypogene. Much of the chalcocite in the lower levels shows a lamellar structure which is interpreted by Locke, Hall and Short as a structure inherited from bornite. The idea that the lamellar structure is inherited from bornite is vigorously combated by Schneiderhöhn; and still the same author shows photographs of chalcopyrite in bornite, so controlled; this intergrowth is explained as a result of "unmixing."

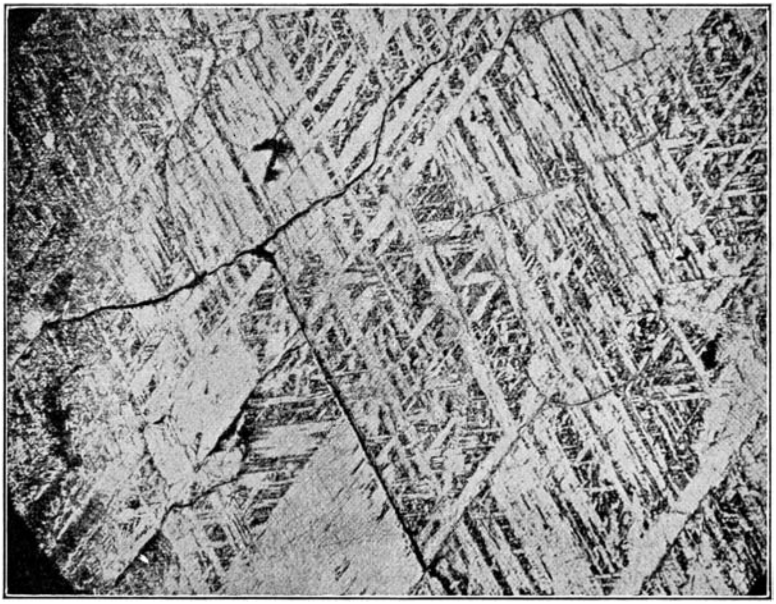

FIG. 325.—Photomicrograph of polished section showing triangular etch pattern of chalcocite, Tsumeb mine. Magnified 60 diameters. (*After H. Schneiderhöhn.*)

On the whole, the present writer believes that hypogene chalcocite exists; that it may be either orthorhombic or isometric; and that it is generally formed by replacement of bornite, or in rarer cases as separate free crystals. At Kennecott both isometric and orthorhombic chalcocite appear to exist, the latter as a product of supergene action. There is also another kind to which attention was first directed by Tolman (from Bingham); that is, the metacolloidal chalcocite; showing no etch figures but contraction cracks (Fig. 326). This is believed to be a product of the precipitation of colloidal copper sulphide by $H_2S$ in copper solution.

L. C. Graton,[1] J. C. Ray,[2] and others have also held that covellite may be of hypogene origin because it replaces, in crystal form, other sulphides like pyrite, enargite, and sphalerite. However, covellite formed by undoubtedly supergene solutions always shows an exceedingly strong tendency to develop blades and crystals. Covellite has not yet been proved a primary or hypogene mineral in sulphide deposits of hypogene origin, though there is really no valid reasons why it could not be formed in this manner.

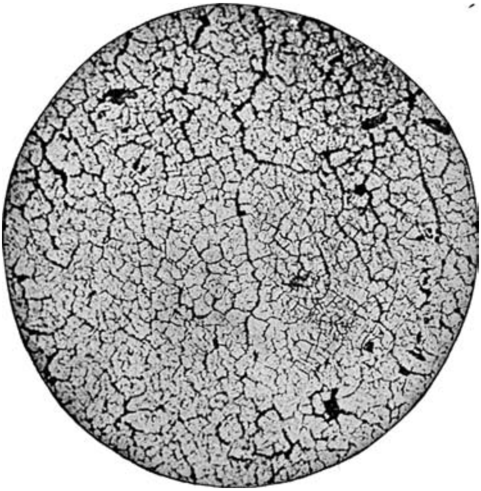

FIG. 326.—Polished section of "sooty" chalcocite etched to show "cracked porcelain" structure believed to indicate colloidal origin. Magnified 75 diameters. (*After Bateman and McLaughlin, "Economic Geology."*)

**Oxidation of Chalcocite Zones.**—Where erosion and water level have been stationary for a long time no changes take place in the chalcocite zone except by the gradual increase due to continued slow leaching of the oxidized zone. But when erosion is quickened or the water level subsides, the reactions of oxidation may invade the chalcocite zone. The gossan may be entirely removed and then oxidation will be working on enriched sulphide ore in which there will be comparatively little iron as pyrite. Other elements like arsenic may also have been removed during chalcocitization. The oxidation of such materials may result in almost complete leaching of copper and a residual outcrop consisting of quartz with some sericite in which even copper stains may be lacking. Such

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[1] *Trans.*, Am. Inst. Min. Eng., 45, 1914, p. 51.

[2] *Econ. Geol.*, 9, 1914, p. 473.

conditions exist at Clifton, Miami, and many other places in Arizona. The outcrops at Butte represent also a leached chalcocite zone and contain little copper.

The course of the oxidation depends largely on the amount of residual pyrite. Where much of this is present the chalcocite is decomposed to cupric sulphate while pyrite protected by the chalcocite remains a little longer until finally decomposed into iron sulphate. When there is little or no pyrite the chalcocite normally changes by oxidation to covellite and cuprite; the small amount of sulphuric acid and ferric sulphate available seems to be sufficient to dissolve these (Fig. 327). Again, the chalcocite may change to brochantite which seems to be particularly common in oxidizing supergene sulphide zones, or more rarely to malachite and chrysocolla.

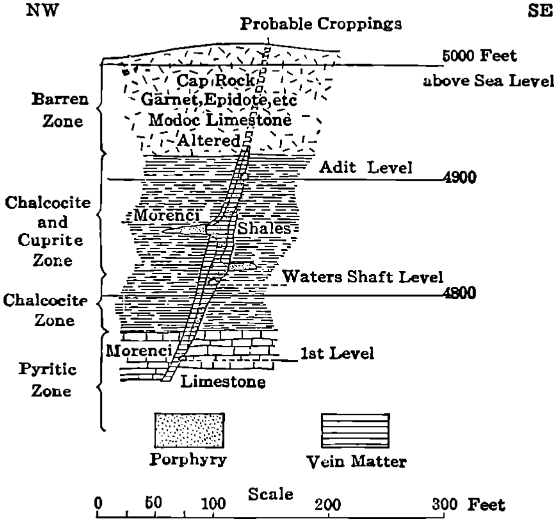

FIG. 327.—Secondary zones in copper veins in contact-metamorphic rocks, Clifton, Arizona.

$$2Cu_2S + O = 2CuS + Cu_2O.$$

$$2Cu_2S + 10O + 4H_2O = H_6Cu_4SO_{10}(\text{brochantite}) + H_2SO_4.$$

The cuprite is often reduced by ferrous sulphate to native copper, and this may again be dissolved by sulphuric acid. At times chalcocite is directly reduced to native metal which may preserve the structure of the black sulphide.[1] This might have been effected by ferric sulphate according to the following reaction, and probably in many other ways.

$$Cu_2S + 3Fe_2(SO_4)_3 + 4H_2O = 2Cu + 6FeSO_4 + 4H_2SO_4.$$

In the chalcocite blankets it is not common to find much native metal, but at Chino, New Mexico, much of the secondary sulphide appears to

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[1] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905, p. 101.

have been converted to copper.[1] At Chino there are several ore-bodies in quartz monzonite and quartz diorite within an area 2 miles in diameter. The areas also occupy parts of the intruded Carboniferous limestone and Cretaceous quartzites. The greatest depth of the chalcocite bodies is 800 feet.

## EXAMPLES OF OXIDATION OF COPPER DEPOSITS

**General Features.**—The study of the various modes of enrichment in copper deposits is a subject full of difficulties. We find the most diverse development even in a region of uniform general climate. Take, for instance, the Sonora-Arizona province, where the rainfall is small and the climate warm. At Los Pilares, Sonora, near Nacozari, a gossan of barren hematite 100 feet deep is underlain by an ill-defined zone with bornite and chalcocite, changing below the 500-foot level to primary chalcopyrite-pyrite ore. In other parts of Sonora, according to Finlayson,[2] are gossan and chrysocolla ores extending to a depth of 200 to 400 feet; below this is a shallow zone of secondary sulphides. Again, at Clifton, Arizona, there are in the contact-metamorphic deposits in limestone strong gossans, sometimes rich in copper, underneath which no secondary sulphides are found. Pyritic veins in porphyry at the same place have a barren siliceous outcrop without gossan and perhaps 150 feet thick, below which lies a rich chalcocite zone that in a few hundred feet or less changes to lean primary sulphides. Other veins nearby show chrysocolla from the surface down to a shallow chalcocite zone at 100 feet. At Ajo,[3] Arizona, a region of excessive aridity, disseminated ores and veins occur in a mass of monzonite. The ore-body occupies an area of 55 acres and has a maximum depth of 600 feet. Here there is apparently no chalcocite zone, but a great tonnage of carbonate and chrysocolla ore underlain by disseminated chalcopyrite and bornite, both said to be primary.

In the normal course of oxidation a gossan must form and the three zones should be distinct; if the gossan is not present, it has been eroded and the barren upper zone has then been formed by leaching of the zone of sulphide enrichment, the copper solution descending to further enrich the deposit in depth. In regions of deep erosion it is exceedingly rare to find a strong chalcocite enrichment in deposits exposed in the lower

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[1] L. C. Graton, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 316.

Sidney Paige, *Econ. Geol.*, 7, 1912, pp. 547–559; Silver City folio, *Geol. Atlas* 199, U. S. Geol. Survey, 1916.

T. A. Rickard, The Chino enterprise, *Eng. Min. Jour.-Press*, particularly, Dec. 8, 1923.

*Guidebook* 14, 16th Internat. Geol. Congress, 1932, p. 35.

[2] A. M. Finlayson, Economics of secondary enrichment, *Min. and Sci. Press*, July 16 and 23, 1910.

[3] J. B. Joralemon, *Trans.*, Am. Inst. Min. Eng., 49, 1915, pp. 593–609.

parts of the canyons. In glaciated or rapidly eroded regions almost all enrichments may be lacking.

**Rio Tinto.**—The pyritic deposits of Rio Tinto, southern Spain,[1] are situated in a country of sub-tropical climate, with an annual rainfall of about 25 inches and mature topography, where erosion makes slow headway. The primary deposits are thick lenses of pyrite containing less than 1 per cent of copper. There is a heavy gossan of massive hematite, 45 to 90 feet thick, containing no copper, over 50 per cent iron, and 10 to 15 per cent of siliceous and argillaceous matter. The depth of oxidation has everywhere been determined by the ground-water level. The lower limit of the gossan is sharp and the line is often marked by a thin earthy zone with notable quantities of gold and silver,[2] believed to represent an enrichment caused by leaching of the gossan by solutions containing chlorine and ferric sulphate. The top of the sulphide zone for a thickness of a few feet is composed of leached pyrite with a trace of copper, resembling the upper part of the chalcocite zone of Morenci, Arizona. Below this begins the zone of enriched sulphides, containing in the upper part 3 to 12 per cent copper, gradually becoming poorer downward and passing into lean pyritic ore assaying 1 per cent or less of copper. The depth at which the unaltered ore is reached ranges from 200 to 1,500 feet below the outcrop. The enriched pyrite contains mainly chalcocite. The bulk of the Rio Tinto copper production to-day is derived from enriched ore.

**Mount Morgan.**[3]—The great gold and copper deposit of Mount Morgan, in Queensland, which since 1886 has yielded about $65,000,000 in gold, shows the peculiar feature of great gold enrichment with almost entire absence of a zone of secondary copper ores. The region has a tropical climate and moderate rainfall; the topography is of the moderately mature type. The water level is probably deep. The irregular deposit is apparently a replacement in Carboniferous rocks, surrounded on both sides by intrusive granite.

At the outcrop there was an extremely rich zone with free gold in kaolin, limonite, and black manganese. Below this was found a zone of a cellular, almost pumiceous, siliceous mass, resulting from the removal of pyrite; this was poorer in gold, but the kaolin that was in places associated with it was rich in silver. The sharply defined lower limit of the oxidized ore was met at 180 to 300 feet below the surface, and the primary ore consisted at first of pyrite, then of pyrite with chalcopyrite, carrying

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[1] For literature see p. 622.

[2] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1899, p. 250.

[3] J. M. Maclaren, Gold, London, 1908, pp. 333–337.

W. F. Gaby, Petrography of the Mt. Morgan mine, *Trans.*, Am. Inst. Min. Eng., 55, 1917, pp. 263–283.

J. F. Newman and J. F. C. Brown, *Trans.*, Austral. Inst. Min. Eng., 15, pt. 2, 1910.

2 to 3 per cent copper and $1 to $8 in gold to the ton. It is difficult to account for the lack of a chalcocite zone. Unquestionably there has been concentration of gold on a large scale at the surface, probably caused by the presence of unusual amounts of chlorine. It is noteworthy that the gold has been precipitated mainly at the surface and could not be carried down into lower levels.

**Ely.**—An important deposit of secondary chalcocite is now being worked on a large scale at Ely, Nevada,[1] by the Nevada Consolidated Copper Company.

The ore production for 1929 was 5,200,000 tons, containing from 1.5 to 2 per cent of copper. The ore carries also, in ounces per ton, 0.018 in gold and 0.088 in silver.

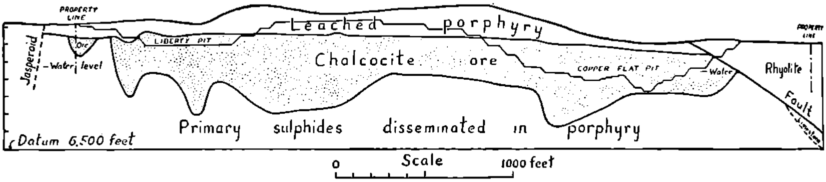

FIG. 328.—Longitudinal section of chalcocite deposit at Ely, Nevada. (*Annual Report of Nevada Consolidated Copper Co.*)

The geological relations are similar to those of the Arizona deposits. Intrusions of monzonite porphyry in Paleozoic limestone caused contact metamorphism of the limestone, silicification of both rocks, and some development of copper deposits, few of which are of economic importance. After intrusion, the porphyry became impregnated with disseminated pyrite with a little chalcopyrite, the silicates being replaced by sericite and pyrite. When the intrusive masses became exposed by erosion to the action of oxidizing waters, a downward migration of copper sulphate, either from the porphyry itself or from the overlying contact deposits, effected a chalcocitization over wide areas.

The leached zone is from 50 to 200 feet in depth and forms an iron-stained soft mass, in places containing oxidized copper ores; below this lies the chalcocite zone, consisting of white earthy porphyry with disseminated grains and flakes of chalcocite and a little pyrite. This zone has a maximum depth of about 500 feet, the copper minerals gradually diminishing downward to the pyritic valueless protore; the upper limit of the chalcocite is rather sharply defined. The general water level in the porphyry is said to be 385 feet below the surface (Fig. 328).

**Bingham.**—Relations similar to those at Ely exist at Bingham, Utah,[2] in a region of much sharper relief and medium aridity. A small mass of

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[1] A. C. Spencer, *Prof. Paper* 96, U. S. Geol. Survey, 1917.

[2] J. M. Boutwell, *Prof. Paper* 38, U. S. Geol. Survey, 1905.

J. J. Beeson, The disseminated copper ores, Bingham Canyon, Utah, *Trans.*, Am. Inst. Min. Eng., 54, 1917, pp. 356–401.

B. S. Butler *et al.*, Ore deposits of Utah, *Prof. Paper* 111, U. S. Geol. Survey, 1920.

monzonite is here intruded into Carboniferous (Pennsylvanian) quartzite and limestone, and has apparently caused the rich mineralization of the Bingham deposits. A large part of the monzonite, 1 mile in length and ½ mile in width has been subjected to hydrothermal metamorphism resulting in the development of disseminated pyrite, chalcopyrite, and bornite. This "protore," which contains less than 1 per cent copper, has been enriched by supergene solutions, depositing chalcocite and, in its upper part, also covellite, resulting in a low-grade ore containing (1918) 1.5 per cent copper and, in ounces per ton, 0.018 (37 cents) in gold and

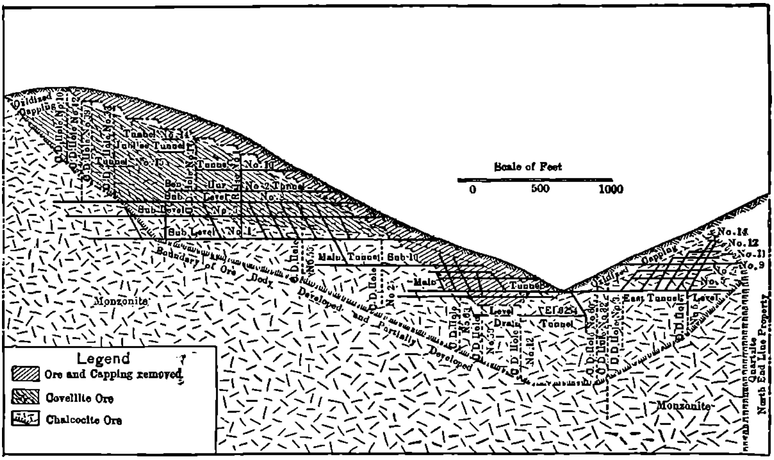

FIG. 329.—Longitudinal section through central portion of ore-body of the Utah Copper Company, Bingham Canyon, Utah.

0.25 in silver. Below a leached surface zone (with some oxidized ore) about 70 feet in depth, lies the chalcocite blanket. The total thickness of the enriched zone is not fully determined (Fig. 329).

Beeson has shown that some supergene chalcopyrite may be formed as an intermediate product between pyrite and chalcocite, and bornite as an intermediate mineral between chalcopyrite and chalcocite. The supergene sulphides replace the sulphides of the protore.

B. S. Butler has shown that the protore also contains hypogene biotite and orthoclase. This property of the Utah Copper Company is now the greatest copper-producing mine in the United States. It is worked as a gigantic open cut, and the ore is concentrated by flotation in plants near the Great Salt Lake. In 1929, 17,724,100 tons were mined by electrically operated steam shovels, a capacity of about 50,000 tons per day. About 148,000 tons of refined copper was obtained together with 116,087 ounces of gold and 1,050,075 ounces of silver. The ore reserves amount to 640,000,000 tons, believed to average a trifle over 1 per cent copper; 194,000,000 tons have been mined to 1926. The per ton operating cost was 87 cents; the per pound cost 6.65 cents. The copper content was 19.89 pounds per ton; the recovery was 17 pounds per ton. Since 1930 the production, owing to market conditions, has been curtailed.

**The Southwestern Chalcocite Deposits.**—In the arid country of southern Arizona and New Mexico we find an interesting group of secondary sulphide deposits similar to the last two examples given. They are sometimes called chalcocite blankets or disseminated chalcocite deposits, and excellent representatives of them are found at Clifton, Globe, Ray, Santa Rita, and in the Burro Mountains. In brief, the concentration has been proceeding in porphyry, granite, or schist containing disseminated pyrite with a little chalcopyrite. Enrichment through replacement of pyrite by chalcocite has in places occurred along fissures or fissured zones, or still more commonly in irregular areas of fractured and brecciated rocks. The result is a chalcocite ore containing 1 to 4 per cent copper and also some residual pyrite; this zone is from 100 feet or less up to several hundred feet in thickness. Above it lies a barren oxidized and leached zone reaching to the surface and from 50 to 1,000 feet in thickness; in places this zone contains some oxidized ore. Below the chalcocite, the primary pyritic dissemination extends to an unknown depth, the rock containing but a fraction of a per cent of copper. The upper limit of the chalcocite zone is sharply defined; the richest ore is found here, gradually decreasing in tenor as depth increases. The water level usually lies at or below the lower limit of the chalcocite zone, and the zone itself, or at any rate the top of it, is for the most part high above the present drainage level.

Evidently the secondary sulphides could not have been formed in their present places under present conditions, for their upper parts are now being actively oxidized. They give evidence of having been accumulated during a long period, probably beginning in the late Tertiary, when the climate was damp and the water level high, before erosion had cut to its present depth. The overlying lean porphyry was leached of its scant copper content, the copper descending as sulphate to become precipitated as chalcocite on the primary pyrite in depth.

These deposits are then old—marooned, as it were, high above their normal position and in an unstable condition. Probably they were once thicker and poorer than now and covered by a gossan. Erosion has carried away the surface gossan, and the scant rain waters have leached the upper part of the underlying chalcocite zone—now the barren zone—and driven the copper downward to replace the remaining pyrite at the level where the oxygen of the descending water became exhausted. Thus is explained the richness near the top, and it follows as a corollary that chalcocite may be deposited above the permanent water level, provided not much oxygen is present.

**Ray and Miami.**[1]—A few miles from Globe, in a region of moderate relief, there is an area of granite (Schultze granite) intrusive into the pre-Cambrian Pinal schist; in the latter, near the contact, several disseminated chalcocite deposits have been discovered.

At the Miami mine, the leached zone is about 200 feet deep and contains in places oxidized ores; a sharp line of demarcation separates it from the underlying chalcocite. The deposit forms a flattened mass which in depth gradually increases in extent. On the 270-foot level the chalcocite area occupies 1 acre; on the 370-foot level, 3 acres; on the 470-foot level, 16 acres. The average tenor of the ore is over 3 per cent of copper near the top of the chalcocite zone, but falls to 2.65 per cent on the 570-foot level. At greater depth the percentage of copper in the ore changes abruptly from 2 to 1 per cent or less. The mine produces a little water on the 450-foot level. In 1929, 5,000,000 tons of ore were mined.

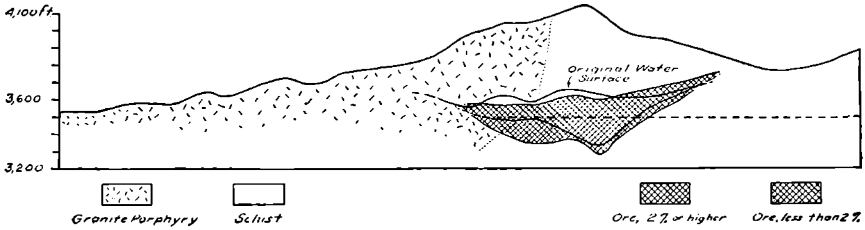

FIG. 330.—Section of chalcocite blanket at Inspiration, Arizona. (*After F. L. Ransome, U. S. Geol. Survey.*)

Somewhat similar conditions obtain at the neighboring Inspiration mine (Fig. 330). At one place the chalcocite zone reaches to a depth of 1,200 feet below the surface. At the present time (1933) both these mines produce ore averaging about 1 per cent of copper.

These deposits are thought to have been formed during the last part of the Tertiary period. Their oxidation is now in progress, with enrichment and concentration of the underlying chalcocite.

At Ray, Arizona, about 25 miles southwest of Globe, a similar but more extensive chalcocite blanket has been discovered and developed by churn drills. The Ray mines are situated in a basin at an elevation of about 2,200 feet. The deposits are in an area of pre-Cambrian schist, cut by dikes of granite porphyry and diabase. The upper leached zone, containing some oxidized copper ore, is from 50 to 150 feet thick. The chalcocite, disclosed by drilling and underground operations, extends over a large area, probably more than 100 acres; its thickness is from 20 to 300 feet and in a considerable part of the area averages 60 feet (Fig. 331). The chalcocite zone is richest at the top and gradually

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[1] F. L. Ransome, The copper deposits of Ray and Miami, *Prof. Paper* 115, U. S. Geol. Survey, 1919; also Ray folio, *Geol. Atlas* 217, *idem*, 1923.

becomes poorer in depth. The ore-body is said to contain about 100,000,000 tons, averaging above 1 per cent copper. The region had long been known as copper bearing, and futile operations on small masses of oxidized ore along diabase dikes had been undertaken. In 1929, 3,600,000 tons of ore were mined averaging about 20 pounds of copper per ton. Water begins to come in at the lower limits of the ore-body, which lies below the level of the creek.

**Chuquicamata.**—The great copper lode at Chuquicamata, in which the developed ore is estimated to be over 300,000,000 tons carrying an average of a little less than 2 per cent copper, is situated in northern Chile. Its production is now rivaling that of the Utah Copper Company. The climate is exceedingly arid. The deposit consists of a series of lodes in granitic rock, intrusive in Jurassic strata; from these lodes supergene mineralization has spread until it now occupies a large area, in which the ore is mined by open-cut methods. The mass of the ore so far developed

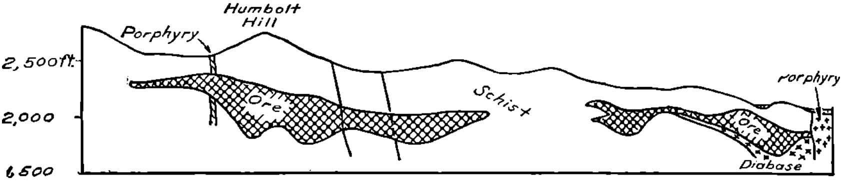

FIG. 331.—Section of chalcocite blanket at Ray, Arizona. (*After F. L. Ransome, U. S. Geol. Survey.*)

is oxidized and carries mainly brochantite, near the surface also atacamite, the latter formed by aid of windblown sodium chloride. The primary ore, as shown by borings and by adjacent smaller deposits, contains mainly quartz, pyrite, and enargite. The present great oxidized body is formed by the oxidation of a deep chalcocite-covellite zone of which the lower part only now remains. This is in part mixed with oxidized ore, and these ores are somewhat richer than those of the oxidized zone. During the chalcocitization the arsenic was removed and the oxidized zone now contains only traces of that element. During oxidation in the now nearly rainless climate very little copper has been carried downward. The brochantite, which contains 62.42 per cent copper (chalcocite 79.8 per cent), has evidently simply replaced the supergene sulphides. Further alteration of brochantite results in the poorer sodium-copper sulphates, kröhnkite, and natrochalcite. Close to the surface is an irregular and shallow zone of leaching, in which the copper is partly removed and basic iron sulphates, hematite, and gypsum have formed. Regarding Cananea, Mexico, see page 722.

**Relation of the Disseminated Chalcocite Ores to Intrusives.**—In a recent thoughtful paper W. H. Emmons[1] concludes that the protores of

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[1] Relations of the disseminated copper ores in porphyry to igneous intrusives, *Trans.*, Am. Inst. Min. Met Eng., 75, 1927, pp. 797–815.

these deposits are of the acrobatholithic type; in other words they occupy domes or cupolas of the intrusive mass—the high points of larger underlying masses. They occur either in the small areas of the (usually quartz monzonitic) intrusive or along or near the axes of not yet exposed ridges of the intrusive. The distribution of the disseminated ores on the cupolas is controlled by fracturing. The rising solutions that metallized the fractured rocks moved upward from the deeper parts of the batholith to the cupolas and particularly to the ridges on the cupolas. Descending cupric sulphate later enriched the upper part of this protore.

## ZINC

**Minerals.**—Sphalerite ($ZnS$) may be said to be the only important primary ore mineral of zinc. It is almost universally present in sulphide ores and is a persistent mineral ranging from deposits of magmatic origin to deposits formed practically at the surface where reducing conditions obtain. Wurtzite, the hexagonal modification of zinc sulphide, is questionable as to its hypogene origin. The absence of zinc sulpharsenides and sulphantimonides is remarkable and the small quantities of zinc reported in analyses of sulphosalts may well be caused by mechanically admixed zinc sulphide. Zinc spinel or gahnite ($ZnO.Al_2O_3$) occasionally occurring in high-temperature deposits is of no economic importance. The minerals zincite ($ZnO$), franklinite ($ZnO.Fe_2O_3$), willemite ($Zn_2SiO_4$), troostite ($Zn_2(Mn)_2SiO_4$), and several rare silicates are almost exclusively confined to the unique deposits at Franklin Furnace, New Jersey.

The oxidized ores of supergene origin comprise the most common smithsonite ($ZnCO_3$), the calamine ($ZnH_2SiO_5$) and the hydrozincite ($ZnCO_3.2Zn(OH)_2$). Willemite may also be formed during oxidation, as at Tres Hermanas, New Mexico, and Pioche, Nevada. There is also the rarer aurichalcite, a basic carbonate of copper and zinc, and several still more infrequent arsenates and vanadates. Goslarite ($ZnSO_4.7H_2O$) forms efflorescences but is of principal interest as the form in which zinc is usually transported in solution. Monheimite ($(ZnFe)\ CO_3$), the iron rich variety of smithsonite, is not uncommon.

**Solubility and Mineral Development.**—The sulphate and chloride of zinc are very easily soluble, whereas the carbonate and the silicates are difficultly soluble. Sphalerite is attacked by oxygenated water and the carbonate forms slowly. In sulphuric acid sphalerite is fairly easily soluble with development of $H_2S$ and hence the oxidation of the mineral proceeds most rapidly in pyritic deposits. Unless limestone or some other precipitant is available the zinc of the oxidized zone is rapidly dispersed as sulphate, and many examples are known of zinc-bearing sulphide deposits from the oxidized part of which the metal has wholly disappeared. Zinc is, in fact, the most mobile of the common metals in ore deposits. Like galena, sphalerite is also readily attacked by ferric sulphate,

$$ZnS + 4Fe_2(SO_4)_3 + 4H_2O = ZnSO_4 + 8FeSO_4 + 4H_2SO_4,$$

and the resulting free acid starts the decomposition again.

Smithsonite is supposed to form according to the reaction

$$ZnSO_4 + CaCO_3 = ZnCO_3 + CaSO_4,$$

but there is reason to believe that where the replacement of limestone is effected by equal volumes, the dilute solution of zinc carbonate is rather the reagent than the sulphate. It is not uncommon to find limestone replaced by smithsonite with perfect preservation of structure. The oxidized zinc ores are often inconspicuous earthy or admixed with clay and ferric hydroxide, and, therefore, easily escape attention. Zinc sulphate is often found in mine waters.

**Supergene Shoots of Zinc Ore.**—In calcareous rocks the descending zinc solutions are easily arrested and there it is common to find secondary zinc shoots below the primary ore (Fig. 332). In case of the extremely common combination of zinc and lead the latter metal remains in its original place as residual galena or cerussite while the smithsonite is found lower down or along convenient paths used by the downward moving waters. G. F. Loughlin[1] has described many cases of this kind from Tintic, Utah, and other places. Usually smithsonite forms first and hydrozincite, calamine, and aurichalcite are distinctly secondary after smithsonite.

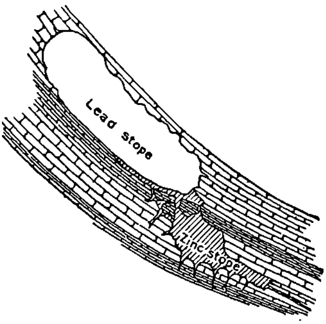

FIG. 332.—Diagram illustrating development of oxidized zinc ore in limestone below primary bodies of lead-zinc ore, May Day mine, Tintic, Utah. (*After G. F. Loughlin.*)

At the important zinc deposits of Goodsprings, Nevada, D. F. Hewett[2] found that hydrozincite is by far the most important ore. Originating from the oxidation of sphalerite, the mineral frequently replaces dolomite (Mississippian). Its formation is held to be favored by the arid climate, and it is shown that the mineral is much less common in more humid regions.

According to experiments by Y. T. Wang,[3] the following reaction appears to explain its formation:

$$3CaCO_3 + 4H_2O + 3ZnSO_4 = 3CaSO_4 + 2H_2CO_3 + ZnCO_3.2Zn(OH)_2.$$

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[1] *Econ. Geol.*, 9, 1914, p. 1.

[2] D. F. Hewett, Geology and ore deposits of Goodsprings, Nevada, *Prof. Paper* 162, U. S. Geol. Survey, 1931, 172 pp.

[3] Y. T. Wang, Formation of oxidized ores of zinc from the sulphide, *Trans.*, Am. Inst. Min. Met. Eng., 52, 1915, pp. 657–710.

Smaller amounts of calamine and smithsonite occur. Near the surface considerable masses of supergene chert occurred with which much plumbojarosite was intergrown.

**Supergene Zinc Sulphide.**—Zinc is not as a rule deposited as a secondary (supergene) sulphide and no case has been recorded where it replaces pyrite, as chalcocite so often does. Many cases have, however, been described showing that zinc sulphide may form below the oxidized zone. W. H. Weed describes such an occurrence at Neihart, Montana, and H. F. Bain regards a certain red variety of sphalerite at Joplin, Missouri (p.429) as of secondary origin. White amorphous zinc sulphide has been found precipitated by hydrogen sulphide from mine waters. Crystals of sphalerite have been observed in old workings opened after having been flooded for many years.[1]

Wurtzite, the hexagonal form of zinc sulphide, which, except by its optical qualities, is almost indistinguishable from sphalerite, has been discovered at several places in the United States, particularly at Joplin, at Butte, at the Hornsilver Mine, Utah, at Goldfield, Nevada, and at the Era district, Idaho.[2]

Allen and Crenshaw[3] have shown a remarkable analogy between pyrite and marcasite on one hand and sphalerite and wurtzite on the other. While pyrite and sphalerite may crystallize from alkaline or from acid solutions, the presence of free acid is essential for the formation of marcasite and wurtzite. It is believed that either wurtzite develops from sphalerite under the influence of acidic solutions or that both crystallize together from such solutions. Butler believed that the wurtzite which has formed abundantly in the lower levels of the oxidized zone at the Hornsilver mine[4] is of secondary origin. Occurrences of "Schalenblende" or concentric colloform intergrowths of sphalerite and wurtzite are known from Vieille Montagne, Belgium, and other places, but they are probably not of supergene origin. There is, as yet, much uncertainty as to the conditions under which wurtzite develops in nature.

## LEAD

**Minerals.**—Among the primary lead minerals galena (PbS) is by far the most common; bournonite ($PbCuSbS_3$), jamesonite ($Pb_2Sb_2S_5$), and a host of other lead sulphantimonides, such as boulangerite, geocronite, etc., are of little economic importance.

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[1] C. R. Keyes, *Trans.*, Am. Inst. Min. Eng., 31, 1901, p. 611.
W. P. Jenney, *idem*, 33, 1903, p. 470.

[2] For a review of these occurrences see J. B. Umpleby, *Prof. Paper* 97, U. S. Geol. Survey, 1917, pp. 87–89.

[3] *Am. Jour. Sci.*, 4th ser., 34, 1912, pp. 341–396; *idem*, 38, 1914, pp. 393–431.

[4] B. S. Butler, *Prof. Paper* 80, U. S. Geol. Survey, 1913, p. 154.

Far more numerous are the oxidized lead minerals. They comprise the red oxide, minium ($Pb_3O_4$), the yellow massicot (PbO), and the dark, colloform plattnerite ($PbO_2$). All these are comparatively rare, as are the chloride and the oxy-chloride. The really abundant oxidized lead minerals are anglesite ($PbSO_4$) and especially cerussite ($PbCO_3$). Of considerable importance are also a series of hydrous basic sulphates of lead with copper and iron, including the blue linarite ($(Pb,Cu)SO_4 \cdot (Pb,Cu)(OH)_2$), the yellow plumbojarosite ($PbO.3Fe_2O_3.4SO_3.6H_2O$), the microscopic foils of which cling to the finger like graphite, and several other yellow, earthy lead-copper sulphates of varying composition.

Lead chloro-phosphate, called pyromorphite ($Pb_5P_3O_{12}Cl$), and the corresponding arsenate (mimetite) and vanadate (vanadinite) are not uncommon and may be considered ore minerals. The same applies to wulfenite ($PbMoO_4$), crocoite ($PbCrO_4$), and stolzite ($PbWO_4$).

Silicates of lead with several other oxy-salts, like antimoniosilicate of manganese, are found sparingly at the two abnormal deposits of contact-metamorphic type, Franklin Furnace, New Jersey and Långban, Sweden.

**Reactions in the Oxidized Zone.**—Lead, in contrast to zinc, shows slight mobility in the oxidized zone. All the salts are difficultly soluble, particularly the carbonate. The sulphate is very slightly soluble. Most soluble is the chloride[1] by means of which some transportation may be effected. Galena is slightly attacked by dilute $H_2SO_4$ and especially by the same solvent together with ferric sulphate.[2] The first change in galena is usually to anglesite ($PbS + 4O = PbSO_4$). Residual nodules of galena are surrounded by dark concentric rings of anglesite, the color being caused by the remaining, finely disseminated lead sulphide. Anglesite is also seen well crystallized. Cerussite appears to form easily from anglesite and usually predominates ($PbSO_4 + H_2CO_3 = PbCO_3 + H_2SO_4$); it appears as beautiful crystal groups but is more commonly earthy, white or yellowish and of sandy texture (sand carbonate).

Once formed these two minerals are exceedingly stable. If there is free sulphuric acid or ferric sulphate or chlorides present, however, the lead may be rendered more mobile.[3] Soluble chloride and oxy-chlorides form, and a whole series of basic yellow lead-iron sulphates may be developed. Plumbojarosite, one of this series, has been used as an ore.[4] Considerable migration of lead in oxidized ore has been carried on by the aid of the two reagents just referred to.

"Steel galena" owes its fine-grained texture either to mechanical deformation of larger crystals or to a beginning transformation to

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[1] At 15° C., 0.909 gram in 100 grams $H_2O$; at 100° C., 3.340 grams.

[2] H. C. Cooke, *Jour. Geol.*, 21, 1913, p. 11.

[3] One liter of water dissolves only 4.4 milligrams $PbSO_4$ while the same amount of saturated NaCl solution dissolves 660 milligrams, slowly decomposing it to chloride.

[4] B. S. Butler, *Prof. Paper* 80, U. S. Geol. Survey, 1913, p. 109.

anglesite. It is supposed to be rich in silver but this is by no means always true. Oxidized lead ores are usually poor in silver, but here again there may be many exceptions noted.

Wulfenite is formed from the oxidation of galena and molybdenite; the latter is often present only in microscopic particles. The enrichment of lead in the oxidized zone is generally a consequence of the solution of associated minerals and the resulting reduction of volume of the ore.

**Supergene Sulphides.**—Well-defined zones of supergene lead sulphide have never been observed. According to its position in Schürmann's series lead should be deposited as sulphide ore on sphalerite and pyrite.

Examples are known of galena formed as well-defined crystals on iron spikes from old workings of a lead mine in Missouri. Thin films of galena are sometimes deposited on sphalerite; this has been observed by Irving and Bancroft at Lake City, Colorado; by Boutwell at Bingham, Utah; and by Ransome at Breckenridge, Colorado.

**Oxidation in the Coeur d'Alene District.**—In the Coeur d'Alene district,[1] northern Idaho, the precipitation is heavy, the topography is accentuated, and the water level stands near the surface of the ground. The veins, which are enclosed in quartzite country rock, contain galena, sphalerite, and siderite. Little pyrite is present. The lower limit of oxidation is very irregular. Complete oxidation is confined to the vicinity of the surface, but cerussite occurs in vugs and fractures several hundred feet below the surface, while the galena may in places occur in the outcrops. The minerals first attacked are pyrite and sphalerite, while the solid galena is very resistant. The chief products of oxidation are limonite, occurring as great masses at the outcrops, and cerussite, to which as the latest products pyromorphite, also plattnerite ($PbO_2$) are added. Owing to the prevalence of siderite, cerussite is the predominating oxidized lead mineral; anglesite is absent. The quantity of silver contained in the galena is small and there is no evidence of enrichment either in the oxidized zone or below it.

**Oxidation in the Mississippi Valley District.**—The effects of oxidation on the lead and zinc deposits of the Ozark region have been described by Bain,[2] Siebenthal and Smith,[3] and Buckley and Buehler.[4] The oxidation is mostly confined to the zone between the surface and the water level, which is rarely as much as 100 feet below the surface. The oxidized ore is, to a considerable degree, a product of the residual weathering of lime-

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[1] F. L. Ransome and F. C. Calkins, *Prof. Paper* 62, U. S. Geol. Survey, 1908, p. 132.

[2] H. Foster Bain, Lead and zinc deposits of the Ozark region, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 155–162.

[3] C. E. Siebenthal and W. S. Tangier Smith, Joplin folio, *Geol. Atlas* 148, *idem*, 1907.

[4] E. R. Buckley and H. A. Buehler, Geology of the Granby area, Missouri Bur. Geol. and Mines, 2d ser., 4, 1906.

stone and chert and thus consists of a confused mass of red residual clay, with layers and fragments of white chert, in which are found galena and the oxidized ores of lead and zinc. Galena is the only sulphide found in quantity above the water level.

Sphalerite alters either to calamine or to smithsonite, and nodular masses of each sometimes hold a kernel of the sulphide. During oxidation the solutions are probably acid only where considerable amounts of marcasite are present. A frequently occurring association is that of calamine surmounted by crystals of dolomite, which could not have been deposited from solutions containing free sulphuric acid.

In the upper Mississippi Valley (p. 436), according to C. R. Van Hise,[1] smithsonite and galena occur above the level of ground-water, which lies close to the surface. Encrusting the galena are some cerussite and less anglesite; with the smithsonite is some sphalerite. The smithsonite may extend 15 to 30 feet below water level, but at greater depth the oxidized products almost wholly disappear and sphalerite with much marcasite forms the principal ore-bodies. Above the water level, then, the iron sulphide has been dissolved, as well as much of the sphalerite, leaving a richer concentrate of galena.

## GOLD

Gold shows slight mobility in ore deposits; it is less easily transported than silver, and, compared to copper and zinc, it is almost stationary. The solubility of gold has been discussed briefly in the chapter on placers. Gold usually occurs native, in coarse or in fine distribution.

Excepting some alloys, like amalgam, maldonite ($Au_2Bi$), and palladium gold, the tellurides form the only definitely known combinations of gold with other elements. Among them are calaverite and the less common sylvanite and krennerite, all with the general formula of $(Au, Ag)Te_2$, and petzite ($(Au, Ag)_2Te$). The tellurides are apparently able to form under widely differing conditions, though they are generally absent from the magmatic and the pyrometasomatic deposits. They decompose easily above the water level, the tellurium is in part carried away as soluble compounds, in part fixed as tellurite ($TeO_2$) or tellurites of iron, like emmonsite and durdenite. The gold remains in minute brownish grains (mustard gold). In most cases there is little evidence of solution and transportation of this gold.

Certain deposits formed by hot ascending waters near the surface contain selenium, either alone or together with tellurium (Republic, Washington; Tonopah, Nevada; Radjang Lebong, Sumatra), and probably they carry a selenide of gold, though its existence has not been definitely proved. Little is known about the oxidation products of selenium.

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[1] *Trans.*, Am. Inst. Min. Eng., 30, 1900, pp. 102–109.

Much more commonly the gold occurs in native form, in gold-quartz veins with small amounts of sulphides. In such deposits there is little evidence of solution and transportation of the gold. Enrichment often takes place in them, but rather by reduction of volume of the ore than by solution of gold. The sulphides in these veins are usually rich in finely distributed gold which remains behind upon oxidation. The oxidation of a crystal of pyrite will generally result in a pseudomorph of limonite which contains flakes of native gold, indicating that within the crystal a certain amount of transportation of gold has taken place.

Very common also are the deposits in which the sulphides abound and which contain no visible free gold.[1] The Gilpin County veins, Colorado; Mount Morgan, Queensland; and the Haile Deposit, South Carolina, may serve as examples. It is in these deposits that most evidence is found of the solution and transportation of gold.

Gold has undoubtedly been transported in the form of chloride, though its migration in colloidal suspension derived by solution processes from native gold is probably much more common than has been suspected. Now, while chloride of gold is easily soluble in water, it is also most easily precipitated by reducing reagents, such as organic matter, ferrous sulphate, metals or sulphides, like pyrite. In the older literature ferric chloride and ferric sulphate were frequently given as solvents of gold, and the question had previously been discussed by Pearce, Don, and T. A. Rickard. H. N. Stokes[2] and J. R. Don[3] showed that ferric sulphate is ineffective, and the recent work of W. H. Emmons[4] and A. D. Brokaw[5] has shown that gold is insoluble in ferric chloride also and goes into solution only when nascent chlorine is present. The metal is precipitated easily and completely by ferrous sulphate.

In deposits containing much pyrite, oxidation results in the liberation of sulphuric acid. Sodium chloride is present in some degree in all mine waters and is abundant in some. Reaction between sulphuric acid and sodium chloride results in hydrochloric acid, and it should be recalled that Don actually found free HCl in a number of superficial mine waters from New Zealand.[6] If dioxide of manganese is present in the deposit, nascent chlorine will be generated according to the reaction:

$$MnO_2 + 4HCl = 2H_2O + MnCl_2 + 2Cl.$$

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[1] Very careful polishing of a section often makes such gold visible.

[2] *Econ. Geol.*, 1, 1906, p. 650.

[3] *Trans.*, Am. Inst. Min. Eng., 27, 1898, p. 599.

[4] The agency of manganese in the superficial alteration and secondary enrichment of gold deposits in the United States, *Trans.*, Am. Inst. Min. Eng., 40, 1910, pp. 767–837.

[5] *Jour. Geol.*, 18, 1910, p. 322; 21, 1913, pp. 251–267.

[6] J. R. Don, *op. cit.*

Ferric and cupric salts have similar power, but chlorine develops very slowly, if at all, in the cold. It should be expected, according to Emmons, that auriferous deposits which contain manganese would show the effect of solution and migration of gold more clearly than non-manganiferous ores. According to experiments by A. D. Brokaw, quoted by Emmons, gold is not dissolved in hydrochloric acid, ferric sulphate, or ferric chloride.[1] It is dissolved at 38° C. in a concentrated solution containing both ferric sulphate and hydrochloric acid; also at the same temperature in a concentrated solution of cupric chloride and hydrochloric acid; the dilute solutions are not effective. Brokaw's experiments verified the solubility of gold by nascent chlorine in the presence of manganese as outlined above.

Where much $MnO_2$ is present, ferrous sulphate is almost immediately transformed to ferric sulphate and the precipitation of the gold is delayed.[2]

The gold dissolved in the presence of $MnO_2$ and held in solution by the absence of $FeSO_4$ moves downward until the excess of acid is reduced, and simultaneously the iron and manganese compounds tend to hydrolyze and deposit oxides. At this stage $FeSO_4$ becomes increasingly prominent and effective as a precipitant. The transportation of gold is thus dependent upon the oxidation of ferrous sulphate by manganese dioxide. In the presence of $MnO_2$ gold may even be carried down and deposited below the water level.

The greater enrichment in gold will be found in the lower part of the oxidized zone. A slight enrichment is sometimes found in the upper part of chalcocite zones. In zones of deep oxidation as at Tintic,[3] Utah, supergene concentrations of gold may be found at all levels. Manganese is apparently always present here. The supergene gold is deep yellow and 900+ fine.

Many examples of actual redeposition of gold are mentioned in the literature, such as films of gold in fissures and on other minerals—for instance, on sphalerite at mines in Lake City, Colorado.[4]

Organic matter, hydrogen sulphide, carbon, sulphides, tellurides, and carbonates[5] precipitate gold from chloride solution. Palmer and Bastin[6] have shown that most sulphides easily precipitate gold; pyrite and galena, which do not precipitate silver, readily bring down the gold. With dilute solutions a protective colloid may inhibit precipitation.[7]

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[1] Don in 1897 had already stated that, in the absence of free chlorine, gold is insoluble both in $FeCl_3$ and $Fe_2(SO_4)_3$.

[2] W. H. Emmons, *op. cit.* (Experiment 20).

[3] W. Lindgren, *Econ. Geol.*, 10, 1915, p. 237.

[4] J. D. Irving, Oral communication.

[5] V. Lenher, *Econ. Geol.*, 13, 1918, pp. 161–184.

[6] *Idem*, 8, 1913, pp. 140–170.

[7] E. S. Bastin, *Jour.* Washington Acad. Sci., 1916, p. 64.

A peculiar feature in certain gold deposits where extensive kaolinization has taken place near the surface is the occurrence of white kaolin extraordinarily rich in gold so fine that it is scarcely visible when the material is washed in the pan.[1] This is undoubtedly an effect of oxidation, but the mode of this enrichment is not fully explained. Possibly the gold has been precipitated in this extremely finely divided form from colloidal solutions.

**Examples of Oxidation of Gold Deposits.**—In the Blue Mountains of Oregon,[2] a region of heavy precipitation but dry summers and rather high topographic relief, gold-quartz veins are contained in Paleozoic argillites and in intrusive diorite. The ores, which in places carry much free gold, are oxidized down to a depth of 100 to 300 feet. At the Sanger mine, on Eagle Creek, the uppermost 100 feet showed a narrow vein, caused by collapse of outcrops, yielding $25 per ton, while farther down the vein widened and its gold was reduced to $12 per ton.

In the formerly highly productive gold veins of the Cracker Creek and Granite districts, the sulphides and arsenopyrite are in fine distribution and much of the ore is rather hard. The water level is high, but on the steep hillsides the oxidized zone is in places 250 feet deep. The oxidation to this depth is only partial, but there is a surprisingly slight difference in tenor between the surface ore and the primary ore. In the latter the gold is contained mainly in the sulphides; free gold is present in the oxidized ore, but there is not enough to convert the material into free-milling ore. No great reduction of volume has taken place, and weathering has only slightly increased the tenor of gold, while the small silver content has been slightly leached.

The gold-telluride lodes of Cripple Creek, Colorado,[3] are mainly sheeted zones in which the seams are filled with quartz, fluorite, and calaverite [$Au(Ag)Te_2$]. These deposits oxidize to brownish clayey material in which the original vein structure is no longer apparent. As quartz is not abundant, the main product of the oxidation is kaolin, with some limonite. The fluorite is carried away, while the tellurides are very easily reduced to dark-brown powdery gold. The tellurium is partly carried away in solution but to some extent remains as colorless tellurite ($TeO_2$) and green ferric tellurites like durdenite and emmonsite.

The oxidation extends to the water level, which is from 300 to 900 feet below the surface, and in places the ore is oxidized for some distance below the water level. There has been little or no enrichment of gold in the oxidized zone, but a decided leaching of the small amount of primary

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[1] W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, p. 171.
F. Guitermann, *Proc.*, Colorado Sci. Soc., 3, 1891, pp. 264–268.

[2] W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901.

[3] W. Lindgren and F. L. Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906.

silver originally contained as telluride or tetrahedrite. No secondary silver sulphides were detected, nor is there evidence of secondary deposition of tellurides.

At Creede, Colorado,[1] and in the Tomboy and the Camp Bird mines, Colorado, as well as in the Granite-Bimetallic mine, Montana, in the Tintic mines,[2] Utah, and in the Mount Morgan mine, Queensland (p. 845) examples are found of the deposition of supergene gold in the lower part of the oxidized zone.

## SILVER

**Minerals.**—Of the primary silver minerals argentite ($Ag_2S$) is easily the most important. Hessite ($Ag_2Te$) is not uncommon in gold-quartz veins but plays no great part as an ore mineral, and naumannite ($Ag_2PbSe$) is a rarity. Native silver may under some circumstances—as in the Lake Superior copper mines, and in the deposits of the Cobalt type—appear as a hypogene mineral. Electrum is of hypogene origin. Of more importance are tetrahedrite ($Cu_8Sb_2S_7$) and tennantite ($Cu_8As_2S_7$) for they usually contain silver sulphide in chemical combination. Finally we have a long series of sulphantimonides, sulpharsenides, and sulphobismuthides of silver or silver-lead. All of these are invariably of late though not necessarily of supergene origin.

Galena and silver are particularly frequently associated, and it is now known that this silver is contained in galena as minute inclusions of argentite,[3] easily made visible by etching a polished surface with HCl. The argentite is believed to have been separated by "unmixing," during slow cooling of a solid solution of the two sulphides. Occasionally tetrahedrite is also included in galena. It is probable that the argentite is also contained in other of the common sulphides and sulpho-salts. Galena and argentite form a eutectic at 77 per cent $Ag_2S$ at 630° C., but this is not developed where there is less than 2.70 per cent $Ag_2S$, and it has not been observed in nature. Silver sulphide may exist in solid solution in slowly cooled PbS but the limit of this lies below 0.2 per cent $Ag_2S$.

The distinctly supergene silver minerals comprise native silver, argentite, cerargyrite (AgCl), embolite (Ag(ClBr)), bromyrite (AgBr), and iodyrite (AgI), also argentojarosite. The first four are of common occurrence.

Stromeyerite ($(Ag,Cu)_2S$), closely allied to chalcocite, occurs supergene or hypogene.

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[1] W. H. Emmons and E. S. Larsen, Geology and ore deposits of the Creede district, Colorado, *Bull.* 718, U. S. Geol. Survey, 1923, 198 pp.; *idem, Bull.* 811, 1930, pp. 89–112.

[2] W. Lindgren, *Econ. Geol.*, 10, 1915, p. 237.

[3] A. E. Nissen and S. L. Hoyt, *Econ. Geol.*, 10, 1915, pp. 172–179.

Of the group of sulpho-salts, referred to above, the following are surely hypogene though they may also develop as a result of supergene enrichment. Stephanite and pearceite are frequently supergene.

| | | | |
|---|---|---|---|
| Pyrargyrite | $Ag_3SbS_3$ | Proustite | $Ag_3AsS_3$ |
| Miargyrite | $Ag\ SbS_2$ | | |
| Stephanite | $Ag_5SbS_4$ | | |
| Polybasite | $Ag_9SbS_6$ | Pearceite | $Ag_9AsS_6$ |

Dyscrasite is really an alloy of silver and antimony with varying composition; it has formed important ore at Broken Hill and Chañarcillo and is also known from Cobalt. It appears to be both hypogene and supergene.

**Solubility and Mineral Development.**—Argentite is oxidized to cerargyrite (horn silver), probably by way of the fairly soluble silver sulphate, as follows: $Ag_2S + 4O = Ag_2SO_4$; $Ag_2SO_4 + 2NaCl = 2AgCl + Na_2SO_4$. Pseudomorphs of cerargyrite after argentite are well known.

Argentite is very slightly attacked by dilute sulphuric acid, but is much more rapidly decomposed by ferric sulphate or by a mixture of the two solvents.[1] Pyrargyrite and polybasite are also decomposed by sulphuric acid, silver and a little antimony going in solution. The reaction is increased by ferric sulphate.

It is thus evident that the presence of oxidizing pyrite will greatly facilitate the movement of silver in ore deposits, and while the metal is not as mobile as copper or zinc it is transported much more easily than gold. The universal presence of chlorine in waters, particularly in those of arid climates, and the insolubility of the silver chloride account for the common occurrence of the latter mineral in the oxidized zones of ore deposits.

Though both cerargyrite and argentite are easily reduced to native silver they are very slightly soluble in solutions usually occurring in nature.

The solubility of the various silver salts is as follows, in grams of anhydrous salt per 100 grams of water: $Ag_2SO_4$ at 25° C., 0.83; at 100° C., 1.46. $Ag_2CO_3$ at 25° C., 0.003; in water saturated with $CO_2$, at 15° C., 0.08. AgCl at 13° C., 0.00014; at 43° C., 0.0004. AgBr at 25° C., 0.00001. AgI is still less soluble than the bromide. It will be noted that the solubility of AgCl increases rapidly with the temperature.[2]

Silver also enters at times into the composition of jarosite,[3] a basic ferric sulphate probably formed by decomposition of silver sulphides by ferric sulphate.

Another way to facilitate transportation of silver, which seems to have escaped general attention, is by colloidal solutions or suspensions.

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[1] H. C. Cooke, *Jour. Geol.*, 21, 1913, p. 13.

[2] A. G. Melcher, The solubility of silver chloride, etc., *Jour.* Am. Chem. Soc., 32, 1910, pp. 50–56.

[3] W. T. Schaller, *Jour.* Washington Acad. Sci., 13, 1923, p. 233.

Such solutions of AgCl, AgBr, AgI, and $Ag_2S$ have been prepared by methods which might easily be used by nature, and by adding protective colloids like silica they may be made very stable. The possibilities of such transportation are evident.[1]

**Precipitation.**—Native silver is easily attacked by ferric sulphate; the silver salts are easily reduced by ferrous sulphate.[2]

The reaction with ferric sulphate is reversible silver being dissolved on heating and reprecipitated on cooling. It may be written as follows:

$$2Ag + Fe_2(SO_4)_3 \rightleftarrows Ag_2SO_4 + 2FeSO_4.$$

Silver is further precipitated from its solutions by organic matter as well as by copper, cuprite, and various sulphides.[3] The precipitation by sulphides and arsenides has been investigated by C. Palmer and E. S. Bastin[4] and also by F. F. Grout[5] who found that alabandite, chalcocite, covellite, and enargite precipitate silver rapidly while pyrite, chalcopyrite, galena, and sphalerite are comparatively inactive. All simple arsenides are also very effective precipitants while the sulpharsenides do not react.

Silver is further precipitated by many silicates such as kaolin and orthoclase.[6] Whether important deposits are produced by such a reaction is doubtful; calcite, siderite and rhodochrosite do not precipitate silver from weak solutions.[7]

Silver sulphide is precipitated by hydrogen sulphide which may result from the attack of $H_2SO_4$ on pyrrhotite, sphalerite, or galena.

**Supergene Sulphide Enrichment.**—According to Schürmann's table (p. 826) silver sulphide should replace many other sulphides such as those of copper, lead, zinc, and iron, following the reaction $Ag_2SO_4 + ZnS = Ag_2S + ZnSO_4$. Such replacements are not uncommon (Fig. 68A), but the sulphides generally precipitate native silver (in places with some $Ag_2S$). Almost all of the common hypogene sulphides as well as tetrahedrite, tennantite, and enargite are easily replaced by the silver sulphantimonides,[8] or sulpharsenides, and to this action much of the supergene sulphide enrichment is due. These rich silver minerals as well as secondary argentite are also found in druses, veinlets, and crusts in which form they do not represent replacements but are perhaps precipi-

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[1] See E. S. Bastin, Experiments with colloidal gold and silver, *Jour.* Washington Acad. Sci., 5, 1915, p. 64.

[2] H. N. Stokes, *Econ. Geol.*, 1, 1906, p. 649.
H. C. Cooke, *Jour. Geol.*, 21, 1913, pp. 1–28.

[3] J. H. L. Vogt, Ueber die Bildung des gediegenen Silbers, *Zeitschr. prakt. Geol.*, 1899, p. 113.

[4] *Econ. Geol.*, 8, 1913, pp. 140–170; C. Palmer, *idem*, 12, 1917, pp. 207–218.

[5] *Idem*, 8, 1913, pp. 407–433.

[6] E. C. Sullivan, *Bull.* 312, U. S. Geol. Survey, 1907, pp. 37–64.

[7] L. G. Ravicz, *Econ. Geol.*, 10, 1915, pp. 368–389.

[8] F. N. Guild, *idem*, 12, 1917, pp. 297–352.

tated by hydrogen sulphide or by complex reactions between solutions. Enargite, tennantite, and tetrahedrite (including freibergite) are never supergene.

According to the investigations of F. F. Grout[1] and L. G. Ravicz,[2] the silver sulphantimonides and the less common sulpharsenides are formed where the descending solutions have become alkaline through the neutralizing reactions going on all the time during their downward travel.[3] Grout made a product of the approximate composition of stephanite by adding silver sulphate to a solution obtained by digesting stibnite in a 1 per cent solution of sodium carbonate. At higher temperatures of 250° and 300° C., de Senarmont and Doelter obtained proustite, pyrargyrite, and stephanite by treating alkaline sulpharsenite or sulphantimonite with alkaline carbonates.

It is a difficult task to separate sharply the effects of supergene alkaline waters from those of the ascending hypogene solutions. Both may yield argentite and sulpho-salts. In all silver deposits there is a progressive hypogene series beginning with arsenopyrite, pyrite, sphalerite, galena, and tetrahedrite, the later replacing the earlier; then follows a series of richer silver minerals, such as pyrargyrite and proustite, which may replace any and all of the earlier sulphides[4] besides filling druses and veinlets. These may be of either supergene or hypogene origin.

In a long series of investigations, E. S. Bastin[5] has examined the question of hypogene *versus* supergene deposition of silver sulphides. He arrives at the conclusion that these minerals as a rule are hypogene. In some cases they may be of supergene origin.

**Zones of Supergene Deposition.**—Zones of the regularity of those of copper deposits are rarely found in the case of silver ores. Still there is a more or less marked succession, dependent on the association of minerals. Silver enrichment apparently necessitates ferric sulphate as an active solvent; this may be furnished by pyrite or by arsenopyrite. In the gossan and down to the water level we find mainly cerargyrite, sometimes as large masses, though even here there may be local accumulations of native silver and argentite. At Lake Valley, New Mexico,[6] 100 tons of silver was obtained from cerargyrite in a comparatively small stope, called the Bridal Chamber; it was contained in Paleozoic limestone, and within 100 feet of

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[1] *Econ. Geol.*, 8, 1913, pp. 407–433.

[2] *Idem*, 10, 1915, pp. 378–484.

[3] G. S. Nishihara, The rate of reduction of acidity of descending waters, etc., *idem*, 9, 1914, pp. 743–757.

[4] F. N. Guild, *idem*, 12, 1917, pp. 297–352.

[5] E. S. Bastin and F. B. Laney, *Prof. Paper* 104, U. S. Geol. Survey, 1918.

E. S. Bastin, *Bull.* 735, *idem*, pp. 65–129 (San Juan, Colo.); pp. 41–63 (Comstock, Nev.); *Bull.* 750, *idem*, pp. 17–39 (Kingman, Ariz.); pp. 41–62 (Aspen, Colo.).

[6] Ellis Clark, *Trans.*, Am. Inst. Min. Eng., 24, 1895, p. 148.

the surface. Below the chloride zone we may find the chlorobromides and below this a horizon where the iodyrite is precipitated.[1] It would be expected to find the most difficultly soluble salt at the top instead of at the bottom, but according to W. H. Emmons[2] the explanation is that if in the solution containing the three halogens chlorine is in vast excess silver chloride though most soluble will be precipitated first. A. Knopf[3] suggests that the reason lies in the chemical relations of iodine and iron salts: no iodide ion can be precipitated until the ferric iron has been reduced to the ferrous state. The three halogens are probably of wind-blown origin, from the salt pans of the desert or from the sea.

Below the chloride zone we usually find native silver, and this may extend to considerable depths, in fact many hundred feet below the water level. The deposition of this native silver may, like that of argentite and sulpho-salts, be accompanied by calcite, barite, and other gangue minerals usually deposited by the ascending waters. Theoretically the zone of the argentite and the sulpho-salts should begin at the water level, but in silver deposits this transition line is far less sharply marked than in copper deposits. The bulk of the supergene argentite usually lies above the sulpho-salts.

The secondary silver sulphides may descend to considerable depths, in places 500 to 1,000 feet below the surface. Poorer ore of hypogene origin will be found below the enriched zone. In many cases the oxidized ore and the supergene sulphides appear inextricably mixed.

**Enrichment at Granite-Bimetallic Mine.**—The Granite-Bimetallic vein, in Montana, described by W. H. Emmons,[4] is contained in granite and has been opened to a depth of 2,600 feet. The early production amounted to $32,000,000 in gold and silver. The strong, steep fissure has a filling from 1 to 20 feet wide, of simple structure, and remarkable horizontal and vertical persistence. The primary ore consists of quartz and rhodochrosite with much pyrite, arsenopyrite, galena, and tetrahedrite, some galena and sphalerite, and specks of pyrargyrite. The ore is of comparatively low grade, containing 20 to 30 ounces of silver and about $2 in gold per ton (Fig. 333).

The uppermost, oxidized zone, from 50 to 300 feet deep, contains a poor iron-stained quartz, with little silver and a trace of gold. The material, which is so poor that many of the claims along the vein were abandoned during its early history, also carries some cerussite, malachite, etc., as well as remnants of unoxidized sulphides. Emmons considers

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[1] F. A. Moesta, Ueber das Vorkommen der Chlor-, Brom- und Jodverbindungen in der Natur, Marburg, 1870, 57 pp.

J. A. Burgess *Econ. Geol.*, 6, 1911, p. 13; *idem*, 12, 1917, p. 590.

[2] *Bull.* 625, U. S. Geol. Survey, 1917, p. 257.

[3] *Econ. Geol.*, 13, 1918, p. 622.

[4] *Prof. Paper* 78, U. S. Geol. Survey, 1913, p. 202.

that this leached zone is derived by incomplete oxidation of the lower zone of enriched sulphides.

Below this leached zone lies the zone of enriched oxidized ore; it is for the most part between the 100- and 400-foot levels. This ore is composed of quartz, stained by iron and manganese, but contains much cerargyrite and native silver as thin seams. There are also some argentite, galena, etc. This ore contains much silver and from \$5 to \$16 in gold per ton. Its value is generally less than that of the ore in the next lower zone.

The zone of enriched sulphides extends in the main from 300 to 800 feet below the surface. It is extremely rich in silver and yields from \$4 to \$8 in gold per ton. Quartz and rhodochrosite are the gangue minerals; there are much argentite, pyrargyrite, and proustite, as well as abundant remains of primary ore minerals.

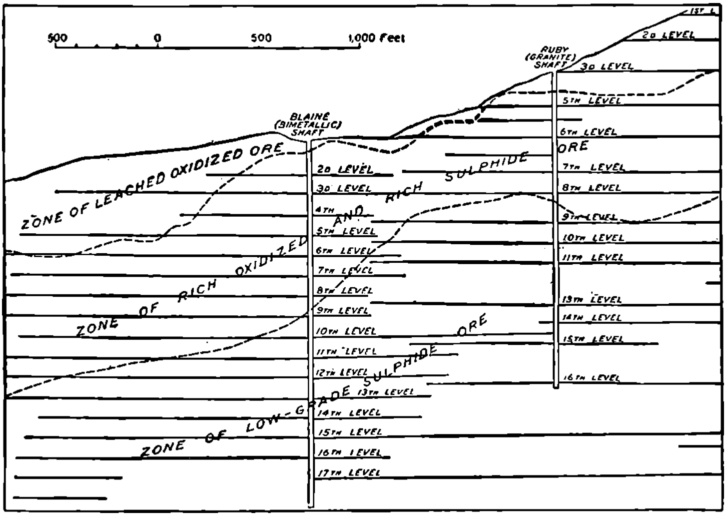

FIG. 333.—Longitudinal vertical projection of Granite-Bimetallic vein, Philipsburg, Montana, showing zone of enrichment. (*After Emmons and Calkins.*)

There are then here a distinct leached zone, a zone of moderate gold enrichment, and a zone of strong silver enrichment.

While it cannot be asserted that all of the primary ore was poor it is very unlikely that its value reached the high figures shown in the zone of secondary sulphides, which for a vertical distance of 300 to 400 feet averaged from 100 to 175 ounces of silver to the ton. Emmons concludes that the silver and gold from the upper parts of the vein, now eroded, have been carried down and that a successive enrichment has thus taken place. The moderate gold enrichment in the lower part of the oxidized zone is, according to Emmons, due to solution of gold by free chlorine (p. 857) and precipitation by ferrous sulphate.

**Enrichment at Georgetown.**—In the Georgetown district, Colorado,[1] argentiferous veins cut pre-Cambrian granites and schists. The climate is rigorous, the relief strong, the water level high. The zone of complete oxidation, in which the ores are rich in silver, extends at most 40 feet below the surface. Below this are friable black sulphides and galena rich in silver and with more gold than occurs at greater depth; the primary sulphides, which contained about 25 ounces of silver per ton, are here enriched and carry more than 200 ounces per ton. Below the zone where the soft secondary sulphides occur and irregularly overlapping the lower portion of this zone are rich ores containing polybasite, and ruby silver, better crystallized and more massive than the pulverulent sulphides, but also subsequent in origin to the primary galena-sphalerite ore. These richer ores diminish in quantity as depth increases, although gradually and irregularly. The best silver ore in most veins has been found in the uppermost 500 feet, although good ore extends locally down to 700 or 800 feet, and in one case 1,000 feet below the surface.

**Enrichment at Tonopah.**—At Tonopah, Nevada,[2] a series of rich silver-gold-quartz veins (Au:Ag = 9:1 by weight), containing hypogene gold, argentite, polybasite, pyrite, etc., with rhodonite, adularia, and carbonates, cut across Tertiary andesites. The climate is exceedingly dry, and the veins are situated in a group of hills rising from the desert. The deposits are oxidized down to a depth of about 700 feet; ground-water is lacking, but from 1,000 feet downward tepid and hot waters, containing mainly alkaline sulphates, are met. The oxidation is irregular and incomplete; the pyrite is changed to limonite, and much chloride of silver, with some bromide and iodide, has formed. In general, cerargyrite is found in the upper part of the oxidized zone; bromyrite in the middle, and iodyrite in the lower part. Other minerals of the oxidized zone are kaolin, hyalite, gypsum, limonite, hematite, jarosite, and wulfenite. On the whole the oxidized ore contains more silver than the primary ore.

Below the oxidized zone the fissures and cracks in the hard quartzose ore contain some secondary argentite, polybasite, pyrargyrite, and also chalcopyrite. Even the oxidized zone contains some veinlets filled with pyrargyrite. There is no well-defined zone of sulphide enrichment; secondary pyrite, sphalerite, and galena are absent. Spurr believes that "the secondary sulphides in the oxidized zone originated from descending surface waters, and probably part but not all of the sulphides in druses in the sulphide ore have a similar origin." Manganese is present but no evidence of transportation of gold has been found.

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[1] J. E. Spurr and G. H. Garrey, *Prof. Paper* 63, U. S. Geol. Survey, 1908, p. 144.

[2] J. E. Spurr, *Prof. Paper* 42, U. S. Geol. Survey, 1905.

J. A. Burgess, *Econ. Geol.*, 6, 1911, pp. 13–21.

E. S. Bastin and F. B. Laney, *Prof. Paper* 104, U. S. Geol. Survey, 1918.

Analyses show that the carbonates were removed from the primary ore and with them most of the lime and magnesia; iron, manganese, copper, lead, and zinc have been largely removed, likewise most of the selenium, arsenic, and antimony. The argentite has largely remained unaltered, while polybasite and the selenides have been decomposed. Bastin and Laney believe that most of the rich silver sulphide minerals are of hypogene origin, but list as supergene some argentite, polybasite, and pyrargyrite.

**Enrichment at Chañarcillo.**—The exceedingly rich silver veins of Chañarcillo, in the arid region of central Chile, have been described by F. A. Moesta[1] and by W. L. Whitehead.[2] The veins intersect Mesozoic limestone with associated volcanics. At a depth of from 500 to 1,000 feet, the primary ore appears. It consists of calcite and barite as the earliest minerals; a second stage is represented by pyrite, sphalerite, chalcopyrite, and galena; a third stage by arsenopyrite, cobalt arsenides, and quartz; and a fourth stage in which tetrahedrite, pearceite, proustite, polybasite, and pyrargyrite were deposited. The sulphide enrichment, in which processes of replacement predominate, occupies a vertical interval of 300 to 600 feet.

During the first stage of enrichment stephanite, pearceite, polybasite, stromeyerite, and argentite replace earlier sulphides. In a later stage silver and dyscrasite develop on a large scale mainly by replacing sulphosalts and sulphides but also by replacing calcite and by filling.

The zone of oxidation is from 150 to 600 feet deep and is marked by the rich development of iron oxides, cerargyrite, embolite, and iodyrite, the last being mostly found in depth. The halides replace silver and dyscrasite, but also sulphides and calcite; to a minor extent they are deposited in open space. The succession is closed by a development of argentite and native silver in local enrichment due to a reversal of oxidation reactions; these minerals again replace the halides.

**Silver Enrichment in Bolivia.**—The celebrated tin-silver veins of Potosi, Bolivia, carry quartz, cassiterite, pyrite, stannite, sphalerite, andorite, and pyrargyrite as primary minerals (p. 580). The bulk of the great wealth was extracted from the oxidized parts of the veins. From old accounts it seems that the principal ore minerals in this zone were cerargyrite, argentite and native silver, though ruby silver is often mentioned also. Recent examination[3] of the Potosi ores failed to show any evidence of notable sulphide enrichment below the oxidation. Pyrargyrite persists in small amounts to the deepest levels.

At the Porvenir mine,[4] Huanuni, a silver-bearing vein containing primary arsenopyrite, sphalerite, galena, and cylindrite is enriched in the

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[1] *Op. cit.*, p. 974.

[2] *Econ. Geol.*, 14, 1919, pp. 1–45.

[3] W. Lindgren and J. G. Creveling, *Econ. Geol.*, 23, 1928, pp. 233–262.

[4] W. Lindgren, Unpublished observations.

upper levels to a remarkable degree by silver-bearing jamesonite and franckeite, accompanied by a late generation of marcasite and pyrite. Jamesonite, in tiny needles and larger masses, replaces sphalerite and galena—very evidently this is a low-temperature process, and it has the appearance of having been effected by descending waters, though absolute proof of this was not obtained.

In conclusion, it seems that the present tendency is to minimize the importance of the silver sulpho-salts in supergene enrichment; argentite is of course admitted. Is it fully realized that this implies a zoning, which then must be hypogene? The supposedly hypogene sulpho-salts in many cases ceased to be deposited some distance below the level now occupied by the water table, which is curious to say the least. The distribution of ruby silver, if hypogene, should have no relation to the present topography. The literature gives evidence that no entirely reliable criterion has been found by which supergene sulpho-salts can be distinguished from the hypogene minerals.

## OTHER METALS

**Platinum and Palladium.**—The main source of platinum metals is in peridotites or in placers derived from them. Some sulphide deposits of the high-temperature types contain sperrylite ($PtAs_2$) and usually also more palladium than platinum; it is not known in what form the palladium appears. Platinum minerals are highly resistant to oxidation.

Sperrylite is not attacked by meteoric waters and this leads to concentration by reduction of volume. At Sudbury the shallow oxidized zone is richer in the platinum minerals than the primary ore.

The occurrence described on page 779, from southern Nevada, shows that oxidized ores with much plumbojarosite (basic plumboferric sulphate) contain finely divided gold as well as metallic platinum-palladium. It is probable that the metals have been precipitated from colloidal solutions or suspensions.

Palladium with some platinum is found in the blister copper from certain smelters, especially from some treating ore from disseminated chalcocite deposits. The Ely deposits, Nevada (p. 846), yield more palladium than others. Selenium, common in all blister coppers, is here also present in unusual amounts.

A. Eilers[1] showed that this copper contained in per cent: 0.0004 Pt, 0.00016 Pd, and 0.055 Se, but that there is no tellurium. This suggests a concentration of palladium and platinum, possibly as selenides, during the supergene chalcocite enrichment.

**Mercury.**[2]—Cinnabar, the principal ore mineral of mercury, is stable up to its sublimation point (680° C.). It is practically insoluble in water

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[1] *Trans.*, Am. Inst. Min. Eng., 47, 1913, p. 217.

[2] E. T. Allen and J. L. Crenshaw, The sulphides of zinc, cadmium and mercury,

(p. 821), but is soluble in alkaline sulphides forming the compounds $HgS.2Na_2S$ and $HgS.Na_2S$ (p. 473). From such solutions the mineral is easily formed as shown by Allen and Crenshaw. Cinnabar is not attacked by dilute solutions of sulphuric acid or by a mixture of ferric and ferrous sulphate, but it is dissolved by dilute hydrochloric acid and is readily attacked by nascent chlorine. The rare selenides tiemannite ($HgSe$) and onofrite ($Hg(SSe)$) are probably primary.

It is apparent that the enrichment by oxidation of cinnabar deposits is not easily accomplished. The secondary minerals are usually confined to a little native metal, which is often found in drops on cinnabar and probably is reduced by hydrocarbons common in many deposits; and occasionally to a little mercurous chloride or calomel. It is not impossible that some of the mercury may be of primary origin, for G. F. Becker in some of his experiments[1] obtained precipitates of mixed sulphide and native metal. The oxidized products of cinnabar consist, besides native metal and calomel ($HgCl$), of the red montroydite ($HgO$) and a number of oxychlorides, such as terlinguaite ($Hg_2ClO$) and eglestonite ($Hg_4Cl_2O$). They are generally found in the deposits occurring in arid, wind-swept regions, where sodium chloride abounds. Large masses of oxy-salts occurred at Terlingua,[2] Texas, but even here cinnabar is the principal mineral. According to Hill,[3] pyrolusite and pyrite are quite abundant in the deposits. All the conditions are thus present for the development of nascent chlorine so that the cinnabar can be transformed into $HgCl_2$, which again by reaction with the abundant calcite would produce oxychlorides.[4]

The mercurous sulphate is only slightly soluble in water (0.058 gram per liter), and the mercurous chloride $HgCl$ is still less soluble (0.002 gram per liter), while $HgCl_2$ is easily soluble.

It has been observed that when mercurial tetrahedrite occurs in a deposit, cinnabar may be found in the oxidized products. This is shown, for instance, in several mines near Sumpter, Oregon.[5] It is not known by what reactions this is accomplished.

K. Hummel[6] has described a case from Germany where cinnabar, with chalcocite, has developed as a supergene sulphide. The cinnabar is

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*Am. Jour. Sci.*, 4th ser., 34, 1912, pp. 367–383.

W. H. Emmons, Enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917, pp. 392–398.

F. F. Grout, *Econ. Geol.*, 8, 1913, p. 427.

T. M. Broderick, Some experiments bearing on the secondary enrichment of mercury deposits, *idem*, 11, 1916, pp. 645–651.

[1] *Am. Jour. Sci.*, 3d ser., 33, 1887, p. 199.

[2] W. F. Hillebrand and W. T. Schaller, *Bull.* 405, U. S. Geol. Survey, 1909.

[3] B. F. Hill, The Terlingua quicksilver deposits, *Bull.* 15, Univ. Texas, 1902.

[4] T. M. Broderick, *op. cit.*

[5] W. Lindgren, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, p. 708.

[6] *Zeitschr. prakt. Geol.*, 1925, pp. 154–167.

derived from tetrahedrite and the age of the oxidation is determined as pre-Variscian. The deposit itself is of Lower Carboniferous age.

**Supergene Sulphide of Quicksilver.**—From the position of quicksilver in the table of Schürmann's reactions it would appear that in the absence of oxygen the deposition of secondary sulphide should be easily effected. Broderick's experiments showed, however, that the sulphide was not precipitated from chloride solutions by marcasite, pyrite, or realgar. A precipitate was obtained with stibnite and chalcopyrite. Cinnabar replacing sulphides has not, however, been observed in nature. There exists a black amorphous modification of HgS, known as metacinnabarite, which has been found in the oxidized zone of many deposits. Allen and Crenshaw have ascertained that this is an unstable form and have obtained it artificially by precipitating dilute, acid solutions of mercuric salts by sodium thiosulphate, a reaction which might well take place in nature. Cinnabar was obtained only from alkaline solutions. At 100° C. the black sulphide changes into cinnabar.

**Cadmium.**—Cadmium is commonly contained in sphalerite, probably as an isomorphous mixture. The cadmium sulphate is as easily soluble as zinc sulphate, and it is present in some mine waters of Joplin, Missouri, and Butte, Montana (p. 62). The yellow cadmium sulphide, greenockite,[1] is often found as coating on sphalerite in agreement with Schürmann's rule, and it occurs in this way even in the zone of oxidation.

J. H. Müller[2] discovered germanium to the extent of 0.01 per cent $GeO_2$ in smithsonite from the Hudson mine, Kentucky; and also in the mine waters from the same place. It is widely distributed in silicates, and in many Bolivian minerals, also in coal.

**Nickel and Cobalt.**—Nickel contained as silicate in basic igneous rocks is concentrated during oxidation to hydrous nickel silicates like garnierite. Under similar conditions the cobalt separates as asbolite, a black earthy hydrous oxide with manganese. Both are colloidal precipitates.

The primary nickel ores consist of niccolite (NiAs), pentlandite ((FeNi)S), polydymite ($Ni_4S_5$), chloanthite ($NiAs_2$), and minerals allied to the last form. Millerite (NiS) is of little economic importance.

The nickel sulphides are easily attacked by oxygen, sulphuric acid, and ferric sulphate. Pyrrhotite with which they usually occur is also easily decomposed.

Nickel sulphate ($NiSO_4$) is easily soluble and does not hydrolyze like iron sulphate to form the trivalent oxide. Therefore, nickel and iron are separated during oxidation. Much of the iron remains in the gossan

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[1] E. T. Allen and J. L. Crenshaw, The sulphides of zinc, cadmium and mercury, *Am. Jour. Sci.*, 4th ser., 34, 1912, p. 341.

[2] *Indust. and Eng. Chemistry*, 16, 1924, p. 604. See also J. Papish, *Econ. Geol.*, 23, 1928, pp. 660–673.

while the nickel is carried away in solution.[1] Pyrrhotite ores are likely to lose nickel if exposed to oxidation in stopes and piles. Morenosite and bieberite are soluble sulphates.

If arsenic is present stable arsenates may be formed in small quantities, in the oxidized zone; like the apple-green annabergite ($Ni_3As_2O_8 + 8H_2O$) and the pink erythrite ($Co_3As_2O_8 + 8H_2O$). Basic sulphates of nickel are not known. There is some evidence that millerite and violarite (polydymite) may be locally precipitated as supergene sulphides.[2] Both nickel and cobalt are often found in acid mine waters.

The more common cobalt arsenides and sulphides include smalt ite ($CoAs_2$), cobaltite ($CoAsS$), and linnaeite ($Co_3S_4$). These minerals are easily attacked by reagents and, the sulphate being very soluble, the metal is readily dissipated during oxidation. The sulphate and the carbonate also occur as products of oxidation. The hydrous oxide, asbolite, seems to be a stable mineral.

**Chromium.**—The primary chromite of igneous deposits is very insoluble and often remains in detrital deposits. Chromium mica (mariposite and fuchsite) is known in sulphide deposits where they traverse serpentine or peridotite. Chromates such as the red crocoite ($PbCrO_4$) may be found in the oxidized zone of similar occurrences, and are probably derived from the decomposition of mariposite. Chromium sulphates of green or purple color have been found in quicksilver veins traversing serpentines.

**Manganese.**—The important primary manganese minerals in ore deposits are manganosiderite [$(MnFe)CO_3$] and manganiferous ankerite, more rarely rhodochrosite ($MnCO_3$) or rhodonite ($MnSiO_3$) and bementite ($H_6Mn_5(SiO_4)_4$), still more rarely alabandite ($MnS$). Hausmannite ($Mn_3O_4$) occurs with franklinite, tephroite and spessartite in some deposits of igneous metamorphism but may also be a supergene product.

Manganese is more soluble than iron and less easily precipitated. Calcite which readily precipitates iron has little effect on manganese solutions. These facts explain the separation of iron and manganese in oxidized deposits. Manganese is easily transported as sulphate or bicarbonate.

The universal products of oxidation are pyrolusite ($MnO_2$) or its hydrated derivatives like psilomelane and wad. Manganite ($Mn_2O_3.H_2O$) and braunite ($3Mn_2O_3.MnSiO_3$) in places accompany pyrolusite. Many of the hydrated forms of manganese dioxide contain in chemical combination the oxides of lead, copper, zinc, cobalt, barium, and potassium and are probably hardened colloidal precipitates.[3]

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[1] W. H. Emmons, *Bull.* 625, U. S. Geol. Survey, 1917, p. 460.

[2] W. Lindgren and W. M. Davy, *Econ. Geol.*, 19, 1924, pp. 309–319.

M. N. Short and E. V. Shannon, *Am. Mineralogist* 15, 1930, pp. 1–15. Authors find that violarite from Sudbury has composition $(NiFe)_3S_4$ and that it is hypogene.

[3] The supergene manganese minerals are sometimes difficult to diagnose, even in

**Tin.**—In tin deposits cassiterite is ordinarily the most abundant ore mineral and exhibits great resistance to solution and transportation. Frequently it remains after other constituents have been dissolved, the outcrops appearing enriched in tin. Some of the outcrops of the Freiberg veins are reported to have contained considerable amounts of cassiterite, probably, according to Stelzner, released from the sphalerite in which it was disseminated as minute crystals. Above the water level the Cornwall veins contained mainly tin, the accompanying copper having been leached. The occurrence of pebbles of wood tin—a colloform phase of cassiterite—is reported from placers in Saxony, Cornwall, several places in the Western states (as at Wood's Creek, Montana), and from Bolivia. Stelzner regarded the Bolivian pebbles as derived from stannite ($Cu_2FeSnS_4$) by processes of oxidation.

The solubility of cassiterite and enrichment of tin deposits have been repeatedly discussed. F. R. Koeberlin[1] asserts that supergene enrichment is important in Bolivian deposits. G. U. Greene[2] described instances of enrichment at Llallagua. He notes that the cassiterite in the Llallagua veins contains 5 per cent Fe and mentions the common occurrence at the same place of vivianite, a hydrous phosphate of iron. Gruner and Lin[3] found cassiterite practically insoluble in dilute sulphuric acid (5 to 8 parts Sn per million) and thought enrichment unlikely. Fink and Mantell[4] conducted a long series of experiments finding considerable solubility in ferric sulphate, in dilute sulphuric acid with sodium chloride, and in sodium sulphide (solubility up to 890 parts per million). In most of the reagents used, the solubility was up to 130 parts per million. In a later paper Singewald[5] discussed the whole question and concluded that the dissolving of cassiterite in supergene waters was improbable. Most of those who have studied the Bolivian deposits agree with him. All find stannite easily attacked by acid waters.

It is certain that hypogene cassiterite can be deposited in crystalline and in colloidal form. It is also extremely probable that stannite, on oxidation will yield colloidal $SnO_2$ solutions.

**Tungsten.**—Scheelite ($CaWO_4$) and wolframite ($(Fe,Mn)WO_4$) are common in many deposits, especially in those formed at high temperatures. These primary minerals are resistant and are found in eluvial

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polished sections; see G. A. Thiel, *Econ. Geol.* 19, 1924, pp. 107–145; W. V. Smitheringale, *idem,* 24, 1929, pp. 481–505; H. Schneiderhöhn, *Neues Jahrbuch,* Beil. Bd. 64, 1931, pp. 701–726; W. S. Savage, Solution, transportation and precipitation of manganese *idem,* May, 1936.

[1] *Eng. Min. Jour.-Press,* Apr. 17 and Nov. 20, 1926.

[2] *Idem,* Sept. 11, 1926.

[3] *Idem,* June 5, 1926.

[4] C. J. Fink and C. L. Mantell, *idem,* Feb. 4, 1928, pp. 201–206.

[5] J. T. Singewald, Jr., The problem of supergene cassiterite, *Econ. Geol.,* 24, 1929, pp. 343–364; *idem,* 25, 1930, pp. 211–218.

placers but rarely occur in alluvial deposits. A sulphide of tungsten ($WS_2$), looking like molybdenite, has been discovered in Utah.[1] Tungsten minerals are slightly attacked by dilute sulphuric acid.[2] The solutions are easily hydrolyzed to insoluble oxides, which also are precipitated by ferric salts, acids, etc. Tungstite[3] ($WO_3.H_2O$) usually results and forms a canary yellow coating.

**Vanadium.**—Few primary vanadium minerals are known. Igneous rocks, according to F. W. Clarke, contain an average of 0.018 per cent vanadium and many pyroxenes and micas carry notable amounts. Roscoelite is found in many gold-quartz veins. The metal is also widely transported in inorganic and organic cycles and thus found in sediments, clays, coal, etc. It is poorly represented in hydrothermal and pyrometasomatic deposits.

Vanadates, like vanadinite ($PbClPb_4(VO_4)_3$) or descloizite ($4(Pb,Zn)O.V_2O_5.H_2O$), are present in the oxidized zone of many lead-zinc deposits of igneous affiliations and may occasionally be of value for the vanadium contained, which probably is derived from the country rock. Important deposits are found near Tsumeb, Southwest Africa, and at Broken Hill, northern Rhodesia.

Regarding the complex vanadium oxy-salts in deposits in sandstone and the rare vanadium sulphide patronite, see pages 409–414. The former include vanadium sulphates, calcium vanadyl vanadates, uranium vanadates, and vanadium mica (roscoelite). These are undoubtedly deposited by meteoric waters, but the history of derivation and succession is imperfectly known. An extensive series of supergene vanadium minerals[4] is recorded from Mina Ragra, in Peru, where they are derived from patronite. The colors are red, green, yellow, orange, blue, and black.

**Uranium.**[5]—The principal uranium mineral is the isometric uraninite; this is probably originally $UO_2$ but always contains more or less $UO_3$; it is isomorphous with the dioxides of cerium and thorium; it contains many other rare-earth metals, also lead, and always more oxygen than the formula $UO_2$ demands. Since its formation, it is subject to continuous radioactive changes, the ultimate product of which is lead. Uraninite with many other rare-earth minerals, like samarskite, occurs in pegmatite dikes. Pitchblende is a colloform uraninite of undoubted gel origin found

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[1] R. C. Wells and B. S. Butler, Tungstenite, a new mineral, *Proc.*, Washington Acad. Sci., 7, 1917, pp. 596–599.

[2] R. W. Gannett, *Econ. Geol.*, 14, 1919, pp. 68–78.

[3] F. L. Hess, Tungsten minerals and deposits, *Bull.* 652, U. S. Geol. Survey, 1917, p. 34.

J. Morrow-Campbell, Tungsten deposits of Burma, *Econ. Geol.*, 15, 1920, pp. 511–534.

[4] W. Lindgren, Charles Palache and L. F. Hamilton, *Am. Jour. Sci.*, 5th ser., 3, 1922, pp. 195–203.

[5] G. Kirsch in Doelter, Handbuch der Mineralchemie, 4 (2), 1929, pp. 870–953.

in veins (Joachimsthal, Bohemia; Gilpin County, Colorado; Great Bear Lake, N. W. Territory, Canada); it is here one of the earlier minerals. Uraninite is easily decomposed, and oxidation gives rise to many uranyl vanadates, phosphates, silicates, and arsenates of yellow or green colors; also oxides like gummite. Many of the recently discovered oxidized uranium compounds come from the rich uraninite deposit at Kasolo, Belgian Congo. Regarding carnotite and allied minerals, see page 410. These secondary minerals, of which the phosphates (uranium micas) are most common, are widely spread, even in oxidized veins from which no uraninite has been reported. They also occur in sandstones (Bolivia) and peat (Madagascar) and in coal and bitumens. It is probable that uranium enters into the biochemical reactions of vegetable life.

**Molybdenum.**—From molybdenite, the common primary ore mineral, molybdite ($MoO_3$) or molybdic ocher[1] ($Fe_2O_3.3MoO_3.7\frac{1}{2}H_2O$) is formed as secondary yellow powder. The lustrous orange plates of wulfenite ($PbMoO_4$) are common in the oxidized zone of deposits containing galena and molybdenite, and it is in places, as at Mammoth, Arizona, of economic importance; rarely it is accompanied by powellite ($CaMoO_4$). W. H. Emmons states that molybdenite is not attacked by $H_2SO_4$ or HCl, and not even by ferric sulphate. The metal is not very mobile and is not commonly concentrated in the oxidized zone; neither does it appear in the supergene sulphide zone. The readily soluble blue ilsemannite, which by later investigations[2] appears to be molybdenum sulphate, is found in some mine waters to which it may give a deep blue color.

Most probably ilsemannite has been formed from molybdite or molybdic ocher by colloidal solution methods. Attention is also directed to the presence of molybdenum in vanadium deposits in sandstone, which probably were formed by meteoric waters and to the reported existence of a colloidal sulphide.[3] F. L. Hess doubts the derivation of ilsemannite from molybdenite but the probability is that it is at least indirectly formed from this mineral.[4]

**Bismuth.**—Among the primary bismuth minerals, bismuthinite ($Bi_2S_3$), the native metal and various lead-bismuth sulphides are the most important. The principal oxidized minerals of bismuth are bismite ($Bi_2O_3.3H_2O$), bismutite ($Bi_2O_3.CO_2.H_2O$), both of yellow color, and several arsenates, for instance, arsenobismite ($2Bi_2O_3.As_2O_5.2H_2O$). In general, the bismuth salts are difficultly soluble in water and the metal is in this respect like lead. They do not show great mobility and bismuth is not found abundantly in the supergene sulphide zone. Nevertheless it seems certain that bismuth has been concentrated in places in the zone of

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[1] W. T. Schaller, *Am. Jour. Sci.*, 4th ser., 22, 1907, p. 297.

[2] W. T. Schaller, *Jour.* Washington Acad. Sci., 7, July 19, 1917.

[3] F. Cornu, *Zeitschr. Chem. Indust. Kolloide*, 4, 1909, p. 190.

[4] F. L. Hess, *Bull.* 761, U. S. Geol. Survey, 1924.

direct oxidation as shown by masses of bismutite and hydrous bismuth arsenate found at Tintic[1] and elsewhere. Other observations point the same way: At the Great Cobar[2] copper mine (p. 688), in New South Wales, the smelting records show bismuth to have been abundant in the zone of oxidation while there is very little of it in the primary ore.

Another evidence of transportation of the metal is its occasional presence with lead in basic iron sulphates like jarosite.

**Arsenic.**—The primary arsenical minerals in ore deposits are arsenopyrite ($FeAsS$), löllingite ($FeAs_2$), smaltite ($CoAs_2$), chloanthite ($NiAs_2$), niccolite ($NiAs$), tennantite ($Cu_8As_2S_7$), enargite ($Cu_3AsS_4$), proustite ($Ag_3AsS_3$), and many rarer sulpharsenides of copper, silver, and lead.

Supergene minerals of arsenic include, arsenolite ($As_2O_3$), mimetite ($PbCl.Pb_4As_3O_{12}$), olivenite ($Cu_3As_2O_8.Cu(OH)_2$), and many other compounds of similar type, realgar ($AsS$), orpiment ($As_2S_3$), scorodite ($FeAsO_4.2H_2O$), pharmacosiderite (basic, hydrous, ferric arsenate), proustite, and pearceite ($Ag_9AsS_6$). The arsenate minerals tend to colloform development.

Arsenic acts as an acid-forming compound similar to phosphorus. The reactions in the zone of supergene waters are probably carried out by means of arsenates and arsenites of sodium and potassium. The sulphides, which are almost insoluble in dilute $H_2SO_4$, easily form soluble double salts with alkaline sulphides, *e.g.*, $Na_3AsS_3$. On the whole, arsenic is fairly mobile in the oxidized zone, much more so than antimony. Arsenopyrite oxidizes to ferrous sulphate and arsenic trioxide:

$$2FeAsS + 11O = 2FeSO_4 + As_2O_3.$$

Finally scorodite and pharmacosiderite are formed. Arsenopyrite is attacked by ferric sulphate and sulphuric acid. Smaltite oxidizes to arsenates:

$$3CoAs_2 + 14O = Co_3As_2O_8 + 2As_2O_3.$$

Enargite is very resistant to oxidation but yields arsenate, sulphate and sulphuric acid:

$$2Cu_3AsS_4 + 35O + 5H_2O = Cu_3As_2O_8 + 3CuSO_4 + 5H_2SO_4.$$

If there is not sufficient oxygen present, some chalcocite or covellite may form. T. L. Walker and A. L. Parsons (p. 822) found that a mixture of arsenides was rapidly oxidized by agitation in aerated water. Rammelsbergite was most easily attacked. $As_2O_3$ and an acid arsenate, $H_2Ni_2As_2O_8$, were found in the filtrate.

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[1] W. Lindgren, *Prof. Paper* 107, U. S. Geol. Survey, 1919.
A. H. Means, *Am. Jour. Sci.*, 4th ser., 41, 1916, p. 125.

[2] E. C. Andrews, Report on the Cobar copper and gold field, *Min. Res.*. 17. Geol. Survey N. S. W., 1913, p. 113.

Native arsenic in concentric shells clearly indicating colloform deposition is not uncommon. Fine concentric intergrowths of arsenic and antimony (allemontite) occur at the Engineer gold mine, Yukon Territory. Native arsenic[1] is probably a hypogene mineral.

Realgar and orpiment are in most cases supergene sulphides, but they are not found in the secondary zones of copper deposits. They are rather more characteristic of the oxidized zones, and often are derived from arsenopyrite. The chemistry of their deposition is uncertain. H. G. Ferguson[2] believes that realgar may be of hypogene origin.

Sulpharsenides of silver are characteristic of the supergene zone of many silver deposits.

In limestone the basic arsenates of copper are easily fixed; calcium, iron and zinc also enter in their composition.

In non-calcareous rocks the arsenates are scarce and, in fact, in many oxidized zones and supergene sulphide zones the arsenic of the primary enargite ore is almost wholly removed. This is the case at Chuquicamata, Chile, and at Butte, Montana.

**Antimony.**—The hypogene minerals of antimony in ore deposits are stibnite ($Sb_2S_3$) and a considerable number of sulphantimonides, in part of copper like tetrahedrite ($4Cu_2S.Sb_2S_3$), and famatinite ($3Cu_2S.Sb_2S_5$); in part of lead like jamesonite ($Pb_2Sb_2S_5$), and many others; in part of silver like pyrargyrite ($Ag_3SbS_3$).

The supergene minerals comprise, cervantite ($Sb_2O_4$), senarmontite ($Sb_2O_3$), and valentinite ($Sb_2O_3$); further, oxy-salts like bindheimite ($Pb_3Sb_2O_8.H_2O$) and stibiconite ($2SbO_2.H_2O$); finally many silver sulphantimonides of which stephanite ($Ag_5SbS_4$) and polybasite ($Ag_9SbS_6$) are the most important. Antimony is often associated with gold, silver, and lead. Pyrargyrite is mostly late hypogene but may be supergene. The same applies to the members of the jamesonite and boulangerite group (see p. 750).

Antimony is considerably less mobile than arsenic during oxidation and as the salts hydrolyze in water there is a strong tendency to form insoluble white or yellowish oxides. To these stibnite usually alters and once formed they are difficultly moved. When, as often happens, lead is present in antimony deposits insoluble compounds like bindheimite are likely to develop. Native antimony is rare and probably of hypogene origin.

Stibnite is very slightly attacked by dilute sulphuric acid and by ferric sulphate, but reacts readily with alkaline solutions and, like mercury and arsenic, forms soluble alkaline sulpho-salts.

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[1] Edward Sampson, *Pamphlet* 31, Idaho Bur. Mines and Geol., 1928. Occurs at Freiberg, Saxony; Washington Camp, Arizona; Pend Oreille, Idaho; and many other places.

[2] *Bull.* 723, U. S. Geol. Survey, 1924, pp. 100–104.

Though according to Schürmann's series antimony sulphide would be expected to replace various other sulphides no examples of this have been found. Stibnite is, however, often found as hairlike crystals in vugs where the mineral certainly is of very late origin.

Stibnite is said to be deposited at the present time in the West Gore mine, Nova Scotia;[1] also a "red sulphide" perhaps kermesite ($Sb_2S_2O$), both probably from alkaline water. Colloidal red sulphide (metastibnite) containing some lead occurs in the upper levels of the Socavon mine, Oruro, Bolivia,[2] as a product of supergene waters. The colloidal red sulphide and stibnite also are forming now in hot ascending waters at Steamboat Springs, Nevada (p. 66).

[1] W. Malcolm, *Mem.* 20-E, Canada Geol. Survey, 1912, p. 296.

[2] W. Lindgren and A. C. Abbott, *Econ. Geol.* 26, 1931, p. 468.

## CHAPTER XXXIII

# METALLOGENETIC EPOCHS[1]

## INTRODUCTION

Wherever universal geological processes are in operation—weathering, sedimentation, metamorphism, deposition by underground waters, and volcanism—there mineral deposits may be forming as they have done during the long ages of the earth's history. Over larger or smaller areas the conditions may at a given time be favorable for the deposition of useful minerals. Such areas are called minerogenetic or metallogenetic provinces. The Jurassic iron ores of England and France, the bauxite deposits of the southern Appalachian States, the salt deposits of Kansas, the gold-quartz veins of California occupy minerogenetic provinces. The provinces may widen into minerogenetic regions: Thus the Cordilleras of the Americas form such a region marked by gold and silver deposits of igneous genetic affiliations. The South African plateau contains large and unusual occurrences of chromite, platinum, and diamonds, likewise of igneous origin. In the same way the Paleozoic of the whole Mississippi Valley region is a metallogenetic region characterized by lead and zinc deposits.

The time intervals favorable for the deposition of certain useful substances are called minerogenetic or metallogenetic epochs. Usually speaking, they are short and transitory but they likewise may widen into periods and eras. Weathering and sedimentation may go on for periods, but the mineral deposits due to these processes commonly represent a comparatively brief time. The Clinton hematites, for instance, though widely spread, were formed during a limited period of littoral sedimentation and are covered and underlain by heavy Carboniferous and Ordovician limestones.

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[1] This conception was first developed by L. de Launay who has contributed much to the study of this subject (Gîtes métallifères, 1, 1913, pp. 241–288).

W. Lindgren, Metallogenetic epochs, *Econ. Geol.*, 4, 1909, pp. 409–420; Gold and silver deposits in North and South America, *Trans.*, Am. Inst. Min. Eng., 55, 1917, pp. 883–909.

J. E. Spurr has also called attention to this subject and has especially brought out the theory of "ore channels," certain lines or areas distinguished by similar metallization (Ore magmas, New York, 1922).

Of late years many other authors have taken up this general line of study. P. Niggli, Versuch einer natürlichen Klassifikation ,etc. *Abhandl. z. prakt. Geologie, etc.*, 1, 1925, pp. 40–68. Translated by H. C. Boydell under title "Ore deposits of magmatic origin," London, 1929, 93 pp.

This applies even more to deposits of igneous affiliations which often were formed during a very brief interval when the emanations of each igneous phase found opportunity for escape and transportation. Each igneous province usually includes many smaller elements in which successive intrusions and effusions were followed by short epochs of various metallizations.

Each mineral deposit has usually a well-marked paragenetic history due to the cooling or changing of solutions. In any older deposit subsequent epochs of mineralization may have left their imprints, or geologic processes like metamorphism and weathering may have modified them.

The minerogenetic development of the several continents is outlined in briefest form in the following paragraphs. It is a fascinating subject for it connects in logical order the science of general geology with that of mineral deposits.

**Main Epochs.**—In the nature of the case metallogenetic epochs of weathering, sedimentation and erosion are not confined to any particular geological period. On the other hand the deposits genetically connected with igneous rocks and metamorphism are most abundantly formed during the great periods of such disturbances which are more or less closely affiliated with folding and mountain building. In his studies on European deposits, de Launay divides them into those of: (1) pre-Cambrian age; (2) Hercynian age; and (3) Tertiary age.

It is necessary to realize, however, that the pre-Cambrian deposits, though generally formed at high temperature and great depth, include many ages far apart. The Hercynian deposits were formed during the great orogenetic disturbance, which falls between the end of the Paleozoic and the beginning of the Trias and which was accompanied by many intrusions. To this period belong such deposits as the tin veins of Cornwall and the silver-lead veins of Freiberg. "Hercynian" is a term that has been applied rather loosely and even metallogenetic epochs of distinctly Paleozoic or of Triassic age have been included in it. The "Hercynian" folds may be divided in a western, or Armorican, and an eastern, or Variscian arc. The Tertiary deposits are those which were formed during or following the great Alpine foldings and subsequent igneous activity. Folding is, however, a process not necessarily connected with the much more deep-seated sources of volcanism, and, therefore, we shall place less emphasis than de Launay has done on the connection between such tangential thrusts and mineral deposition.

## EUROPE[1]

**Pre-Cambrian Epochs.**—We find the pre-Cambrian deposits mainly in the great shield of Fenno-Scandia, in Sweden, Norway, and Finland.

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[1] G. Berg, Zonal distribution of ore deposits in Central Europe, *Econ. Geol.*, 22, 1927, pp. 113–132. (Footnote continued on following page.)

They carry mainly iron and copper, much more rarely lead and zinc, or tin. Gold, silver, arsenic, antimony, and mercury are very scarce. A few gold deposits with pyrite or arsenopyrite are worked. Deposits of contact-metamorphic or magmatic or hypothermal type prevail with structural and mineralogical features indicating depth and high temperature. Dynamic metamorphism frequently has been superimposed. Iron deposits of sedimentary origin in part similar to those of North America are present.

**Paleozoic Epochs.**—Sedimentary ores of hematite and iron silicates are rather widespread in Europe. We find them in the metamorphosed Cambro-Silurian of northern Norway, in Great Britain, France, Germany, Bohemia, and Spain. Among the deposits of igneous origin we note the ores of copper, nickel, titanium, and chromium of northern Norway, connected with gabbros intrusive in the Cambro-Silurian complex, and with the Caledonian folding. Some siderite deposits of Rhenish Prussia, in the Siegerland, appear to be of Devonian age.

**Hercynian Epochs.**—The Hercynian movements fell between the Carboniferous and the Trias and resulted in great mountain chains extending east and west across Europe. Intrusions of granitic rocks of that age are laid bare by erosion which has also truncated the deposits so that the veins which form the dominant type present the high-temperature or intermediate types. Tin, lead, zinc, copper, and silver prevail among the metals; antimony and arsenic begin to appear. Iron deposits are also found but do not attain the importance of those of Fenno-Scandia. In this age were formed the tin deposits of England,[1] France, Saxony,[2] and Spain all connected with late Carboniferous granitic intrusives.

The majority of the deposits of the Iberian Peninsula (including those of Rio Tinto and the Mesa Central), of the Central Plateau of France, and of central Germany belong to this age.

The Ural Mountains are also regarded as a Hercynian range truncated by erosion and at least many of its deposits of platinum, iron, copper, and gold are considered as induced by Hercynian intrusives. The deposits are generally of a deep-seated type.

**Permo-Triassic Epochs.**—Between the Hercynian ranges extended desert plains and saline lakes or bays. The salt deposits of central

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L. de Launay, *op. cit.;* La métallogénie de l'Italie, Internat. Geol. Congress, Mexico, 1906.

[1] The zonal arrangement of metals in the Cornish lodes, dependent upon a series of cooling solutions from central intrusives, has been well described by E. H. Davison, *Econ. Geol.*, 22, 1927, pp. 475–479, by H. Dewey, Special Repts. Min. Res., *Mem.* 21, Geol. Survey Great Britain 1921, and by H. B. Crenshaw, *Bull.* 204, Inst. Min. Met. (London), 1921.

See also R. H. Rastall, Metallogenetic zones, *Econ. Geol.*, 18, 1923, pp. 105–121.

[2] K. Dalmer, *Zeitschr. prakt. Geol.*, 1896, p. 148.

Germany, the sedimentary copper ores of the Mansfeld basin, the copper ores disseminated in sandstone in Germany and Russia belong to these epochs.

**Jurassic and Cretaceous Epochs.**—Broad transgressions of the seas with widespread formation of oölitic iron ores mark the Jurassic in England and on the continent. Glauconitic sands and chalk accumulated in the Cretaceous seas. The igneous forces were quiescent. Perhaps we do not err in attributing a Jurassic age to the concentrations of lead-zinc ores in the Paleozoic of Belgium and in the Trias of Silesia.

**Tertiary Epochs.**—Few mineral deposits of northern Europe date from the Tertiary but southern Europe was the scene of great activity mainly in connection with the igneous outbursts which culminated south of the great Alpine arches and overthrusts.

1. In the areas occupied by the Mediterranean seas, we find widespread sulphur deposits and phosphates, both in the sedimentary beds.
2. The great Alpine intrusions of granitic rocks originated replacement deposits of siderite and magnesite in the Triassic and Paleozoic limestones together with many minor metal deposits. Among the latter should probably be placed the widespread lead-zinc deposits of the Alpine Trias.
3. Farther south, in Italy, we note the contact-metamorphic specularite deposits in Elba in connection with intrusive granite. The magnetite deposits of the Banat, likewise of contact-metamorphic origin, also belong to this class.
4. The Eocene intrusion of greenstones (peridotites, gabbro, etc.) in the Alpine region and throughout the Mediterranean resulted in magmatic copper deposits, zeolitic copper deposits (Monte Catini), magmatic chromite deposits (Greece and Asia Minor), and contact-metamorphic emery deposits.
5. In Hungary and Transylvania, inside of the Carpathian arches, intrusions and effusions of andesite and dacite took place in the Miocene. Accompanying these was a metallization of gold and silver very similar to the Cordilleran (Pacific) types. These veins are of the type formed near the surface.
6. The last metallization, of cinnabar, accompanied the late Tertiary (and perhaps the present) igneous outbursts in Italy, Austria, and Serbia. It is throughout of the type formed near the surface.

Some effects of the Tertiary metallization is noted in connection with eruptives in Bohemia, Saxony, and the Harz as well as in the Central Plateau of France. Lead, silver, nickel, cobalt, antimony, and uranium are the principal metals supplied in these regions. Fluorite and barite are common. There has been much discussion regarding the age of the silver-lead veins of the Harz region. It is held by some that the metallization is of Hercynian age, while others consider it later, perhaps Tertiary.

It is believed that there was also a Tertiary metallization in Spain. Intrusions of diabase appear to be connected with deposits of iron (Bilbao) and mercury (Almaden).

## ASIA[1]

The mineral deposits of Asia have been described by de Launay, who in his work cited below, has collected a vast mass of data in the correlation of which there are still many blanks. According to Obrutschew, we find the pre-Cambrian deposits represented by gold-bearing veins in the great shield extending on both sides of Lake Baikal and in Korea. Again we meet them in the great Indian platform. Here are deposits of gold-bearing quartz (Mysore), of iron, of manganese, and of graphite; also copper.

The greater part of the gold production of Siberia is derived from the "Eozoic" (late pre-Cambrian) formations of sedimentary rocks, intruded by granodiorite, which extend south and west of Lake Baikal and elsewhere. Here belong the rich placers of the Lena gold fields and many other districts.

Deposits of Hercynian or late Paleozoic age are found in many ranges in central Asia, such as the Ural, the Altai, and the trans-Baikalian mountains. They are mostly well-defined veins carrying gold, lead, silver and other metals. There are also many contact-metamorphic iron deposits. To the same age are referred the tin and wolfram deposits of Malaya, Burma, Siam, and China, many of which are accompanied by much later placers. Copper deposits in Permian sandstone occur in northern Turkestan.

Wong recognizes in China two periods of metallization: (1) pre-Cambrian with gold, tungsten, and molybdenum; and (2) Hercynian with parallel belts of tin and zinc; zinc and antimony; antimony and mercury. These three zones are said to be genetically related to granite intrusions.

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[1] L. de Launay, Richesses minérales de l'Asie, Paris, 1911, 816 pp.

P. P. Goudkoff, New aspects of the geology of the principal ore-bearing provinces of Siberia, *Econ. Geol.*, 17, 1922, pp. 260–273.

W. A. Obrutschew, Die metallogenetischen Epochen und Gebiete von Siberien, Halle, 1926.

Manjiro Watanabe, The geological distribution of ore deposits in Japan, *Econ. Geol.*, 18, 1923, pp. 173–189.

W. Wong, Les provinces métallogéniques de Chine, *Bull.* 2, Geol. Survey China, 1920.

T. Kato, The last stage of magmatic differentiation as represented by Tertiary gold-silver veins, *Fifteenth Internat. Geol. Congress*, *Comptes Rendus*, 2, 1929, pp. 9–12.

The deposits post-dating the Himalayan and Alpine systems are apparently rare. A few of them occur in the Caucasus. The Tertiary mineralization accompanying the igneous outbursts bordering the Pacific is, however, well represented in eastern Asia. We find gold-bearing veins of the Cordilleran type in the island arcs, Sumatra, Celebes, Borneo, the Philippines, and Japan. In the last country there is also a more varied mineralization of the same date including veins of copper, antimony, and quicksilver.

In Japan, Watanabe describes the distribution of deposits in the Northeastern Islands as follows: (1) the inner zone of Tertiary sediments and effusives. Epithermal veins occur here in rhyolite and other effusive rocks. They carry gold and silver, or zinc-lead with fluorite, or copper. A peculiar feature is the "black ore," a fine-grained mixture of sphalerite, galena, and barite, with some copper. (2) The outer zone. This is a dissected plateau of Paleozoic and Mesozoic beds with intrusives; here occur replacement deposits with pyrite, pyrrhotite, and chalcopyrite (Hitachi).

In the Southern Islands there are (1) An inner zone of late Tertiary filled fissure veins with gold and silver; silver and copper; and copper. (2) A middle zone occupies a dissected plateau with granite intrusions of Paleozoic or Mesozoic age. Contact-metamorphic deposits characterize this area. (3) An outer zone of Paleozoic and Mesozoic sedimentary rocks, with few intrusives. Here we find bedded replacement deposits of post-Jurassic (?) age.

### AFRICA

In the main, Africa consists of a vast platform of ancient rocks. Arabia, India, and Madagascar have the same structure and were no doubt once connected with the African continent. Archean crystalline rocks are overlain by sediments, folded in places, of pre-Cambrian and Paleozoic age; there are granite intrusions of various ages and vast sheets of basic intrusive rocks (Bushveld laccolith); in the south these older rocks are covered by Karroo beds (Permo-Triassic). In the north, in Morocco, Algeria, and Tunis, there are Tertiary folds of the Alpine system with scant amounts of igneous rocks. Here we find a weak metallization with deposits of lead, zinc, antimony, and quicksilver; also siderite and hematite. An extensive deposition of sedimentary phosphates took place during the Eocene.

Tertiary and recent eruptives are absent from the main part of Africa except along a zone from Abyssinia to the Great Rift Valley where volcanoes rise above the old plateau.

The old pre-Cambrian land mass contains abundant gold deposits, usually of the high-temperature type as in Madagascar, in Rhodesia, and in Swaziland, and gold, probably of placer origin, is likewise found in wide distribution in the old conglomerates of pre-Cambrian age overlying the Swaziland crystalline rocks. Most famous among these conglomerates are those of the Witwatersrand, Transvaal, which furnish the larger part of the world's gold production. Somewhat later are perhaps the bed veins of Pilgrims Rest and other places in the Transvaal.

Liquid-magmatic platinum deposits are widespread in the basic rocks of the Bushveld complex, probably of pre-Cambrian age.

Other deposits of deep-seated origin in Rhodesia and Namaqualand carry chromite and copper ores of magmatic origin.

Tin deposits of pegmatitic or deep-seated vein type, with accompanying placers, are found in the Transvaal and in Nigeria.

The important copper deposits of the Belgian Congo are found in sandstone, shale, and limestone of pre-Cambrian or early Paleozoic age. Some regard them as formed by meteoric waters, while others explain their genesis by emanations of deeply buried intrusives. Similar large copper deposits in sedimentary beds have been opened in northern Rhodesia.

The latest mineralization in South Africa appears to be represented by the diamond-bearing peridotite pipes which are intrusive in the Karroo series. They are probably of Cretaceous age.

The metallization of South Africa is marked by widespread occurrence of siderophile elements, platinum, chromium, and diamond, and by high-temperature tin and gold deposits the latter accompanied by placer deposits of pre-Cambrian age.

## AUSTRALASIA[1]

On the old plateau of Australia no ore deposits (except gold placers) appear to have been formed since the beginning of Mesozoic time.

The pre-Cambrian era is marked by a widespread metallization of gold in Western Australia centering in Kalgoorlie and Pilbarra; this is apparently associated with intrusive granites. In New South Wales, we find the exceptional high-temperature lead-zinc deposit of Broken Hill. In South Australia are the iron deposits of Iron Knob and the copper deposits of Wallaroo, Moonta, and Burra Burra, all of liquid-magmatic or pegmatitic origin.

In Victoria and adjacent parts of New South Wales, also in Tasmania, a rich metallization of gold, in quartz veins, marked the period between

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[1] E. C. Andrews, Geographical distribution of ore deposits in Australia, *Econ. Geol.*, 18, 1923, pp. 1–25.

F. L. Stillwell, Correlation of ore deposits in Australia, Austral. Inst. Min. Eng., 1924, pp. 796–804.

A. G. Maitland, Ore deposits of Western Australia, Federal Handbook, Brit. Assoc. Adv. Sci., Australasian meeting, 1914.

E. F. Pittman, Mining fields of eastern Australia, *idem.*

the close of the Ordovician and the close of the Devonian;[1] it followed the intrusion of granitic rocks. The copper-gold ores of the Great Cobar, in western New South Wales, are probably also of early Paleozoic age, but the deposits are isolated among sediments far from outcropping igneous rocks. Gold-quartz veins of a similar age appear in the South Island of New Zealand. Tin, copper, lead, and platinum metals appear in Tasmania besides some gold-quartz veins.

South and west of the New England region, in the northern part of New South Wales, numerous intrusions accompanied by vein formation took place in the Devonian, while in New England and Queensland, deposits of tin, tungsten and bismuth were formed during and after Permo-Carboniferous granitic irruptions. To this age belong also, most probably, the gold, stibnite-scheelite veins of Hill Grove, and the gold-quartz veins of Gympie, Mt. Morgan, and Charters Towers.

Rich deposits of gold and silver of the Cordilleran type, also deposits of quicksilver, are found on the North island of New Zealand. They occur in the lavas of Tertiary age which form a part of the igneous girdle of the Pacific.

## SOUTH AMERICA[2]

In South America, as in the northern part of the continent, we must differentiate the eastern part, in which strong mountain building forces have rested since the close of the Paleozoic era, from the western margin, which is marked by the great Cordilleran ranges, in which, since early Mesozoic times, the scenes have been set for a tremendous display of volcanism and orogenic movements.

In the eastern part iron and gold are the principal metals. The former occurs in Brazil as sedimentary deposits of hematite of great extent and probably "Algonkian" age.

A little later than these and connected with intrusive pegmatites are the gold-bearing quartz veins of Brazil, mainly in Minas Geraes, in the Guianas, and in eastern Venezuela; all of these are of the high-temperature type. Many minor deposits occur in the pampean ranges of Argentina.

Along the Pacific Coast the pre-Cambrian is largely lacking and practically all of the deposits are of Tertiary age. A few of them, in

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[1] Stillwell says these deposits are of post-Silurian and pre-Permo-Carboniferous age.

[2] J. T. Singewald Jr. and B. L. Miller, Ore deposits of South America, 1919.

G. Steinmann, Gebirgsbildung und Massengesteine in der Kordillere Südamerikas, *Geol. Rundschau*, Bd. 1, 1910. Über gebundene Erzgänge in der Kordillere Südamerikas, Internationaler Kongress, Düsseldorf, 1910; Geologie von Peru, Heidelberg, 1929, 442 pp.

R. Beder, Los Yacimientos de la Republica Argentina, etc., *Revista*, Univ. de Cordoba, 14, Nos. 1 and 2, 1927.

Colombia, Peru, and Chile, are of the type formed near the surface and carry gold and silver. The great majority are of the mesothermal or hypothermal type and have suffered considerable erosion. They stand in genetic connection with smaller masses of intrusive granite or diorite, as well as with rhyolitic and andesitic porphyries. The principal metals are copper and silver. Great lead and zinc deposits are not common. Gold is widely spread but even the greatest of the gold districts, in Colombia, do not compare with those of California and Australia.

Tin, with silver, occurs in a belt extending through the eastern Cordillera of Bolivia. The tin veins as well as the tin-silver veins stand in genetic connection with intrusions of quartz monzonite, their porphyries and rhyolitic and dacitic flow rocks. Some of the South American geologists believe that the first mentioned group is of pre-Tertiary age. Tungsten deposits occur in Peru and Bolivia, mainly associated with cassiterite. Quicksilver deposits are found in Peru.

Among the many districts which have yielded great treasures of silver, we note particularly Cerro de Pasco, in Peru, and Potosi, in Bolivia; at both places supergene enrichment has played an important part. A widespread mineralization of copper often accompanied by tourmaline is found in the central and northern parts of Chile. Enargite characterizes many copper deposits in Argentina, Chile, and Peru. Few contact-metamorphic deposits have been described.

A later epoch of mineralization verging upon the Recent has yielded superficial deposits of borax and nitrates probably derived from volcanic exhalations. Gold placers of Quaternary age have been worked at many places.

## CENTRAL AMERICA

In Honduras, Nicaragua, San Salvador, and Costa Rica—a region in which volcanoes and effusive rocks abound—we find many representatives of gold-silver veins of Tertiary age which were formed near the surface.

## THE ANTILLES

A feeble mineralization of copper and gold is observed in the greater Antilles and in most cases it accompanied Cretaceous or Eocene intrusions of basic rocks. During the late Tertiary, superficial deposits of iron ores were formed by weathering of serpentine rocks.

## NORTH AMERICA

In describing the metallogenetic epochs of North America it will be convenient to separate the eastern and western parts of the continent for, with the exception of their earliest history, they have little in common. The western or Cordilleran part, including also almost the whole of Mexico and Alaska, contains few deposits older than the Cretaceous. The eastern part contains few that are more recent than that age and most of them are of pre-Cambrian age.

*In the eastern part we distinguish the following metallogenetic epochs:*

1. **The Pre-Cambrian Epochs.**—The pre-Cambrian era embraces a long time and many epochs of ore formation. Ages of deep weathering and sedimentation alternated with times of violent igneous action. The deposits of igneous affiliations are largely of the deep-seated, high-temperature types. Those connected with the latest pre-Cambrian (Keweenawan) igneous rocks are in part of mesothermal or even epithermal character. The metals contained are iron, copper, nickel, gold, and silver. Lead, zinc, antimony, and quicksilver are rare. Naturally the deposits are best exposed in the great ice-polished Canadian shield and adjacent parts of the United States, but many are also found along the Appalachian ranges, where the pre-Cambrian may be exposed.

Most ancient among the pre-Cambrian deposits are the "iron formations" contained in the sedimentary rocks associated with the Keewatin greenstones. At Vermilion, Minnesota, and many other places, hematite ores have developed by the oxidation of these iron formations. At Michipicoten, Ontario, siderite ores appear.

But little more recent are the many magmatic and contact-metamorphic deposits connected with granitic intrusions in the Grenville series of sedimentary rocks, in the Adirondacks and in Ontario. Among them we find titanic iron ores, magnetite, corundum, molybdenite, apatite, and phlogopite. The iron ores occur in large deposits. The great zinc and manganese deposits of Franklin Furnace, New Jersey, should perhaps be correlated with these concentrations.

Iron formations with large concentrations of hematite were repeatedly laid down during the Huronian sedimentations in the Lake Superior region, but epochs of intrusions of granite and porphyry intervened.[1] The majority of the gold-quartz veins of Ontario, Quebec, and Manitoba appear to have been formed in connection with widespread intrusions of granite, syenite, and porphyry accompanying or following the Temiskaming sedimentation. They are thus of pre-Huronian age. Though they are found in many districts, they have been most successfully worked in the Porcupine and Kirkland Lake regions of central Ontario.

The pre-Cambrian epochs close with the great Keweenawan intrusions and effusions of basic rocks, like gabbro, norite, diabase, and basalt. These magmas were rich in copper, silver, nickel, cobalt, and arsenic and many deposits of these metals were formed during this epoch. They are

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[1] W. G. Miller and C. W. Knight, Metallogenetic epochs in the pre-Cambrian of Ontario, *Twenty-fourth Ann. Rept.*, Ontario Bur. Mines, 1915, pp. 243–248.

distributed over a wide area in the Lake Superior region[1] and include the native copper deposits, the silver-cobalt veins of Cobalt, and the magmatic nickel-copper deposits of Sudbury.

2. **Paleozoic Sedimentary Epochs.**—During the long period of Paleozoic sedimentation in the Appalachian region, at least three epochs were characterized by deposition of iron ores. We find oölitic hematite of Ordovician age in Newfoundland, and similar oölites in the widespread Clinton formation in the Silurian, the latter reaching their maximum development in Alabama. Less important siderites were deposited as "black bands" during the Carboniferous. The phosphate beds of Tennessee also belong here.

3. **Paleozoic Intrusives.**—From Nova Scotia to Alabama granitic intrusions took place during the early Paleozoic and some of them even extended into the Carboniferous. A moderately strong metallization of gold-quartz veins followed these. We find these veins in Nova Scotia, at various places in New England, and particularly in the Southern states from Maryland to Alabama. Probably the majority of them are of Cambrian or Silurian age. Many deposits of pyrite and chalcopyrite in this region appear to date from the same period.

4. **Paleozoic Epochs of Saline Deposits.**—Epochs of arid climates with accompanying development of deposits of evaporation recurred at several times during the Paleozoic era. Thus there are deposits of gypsum and salt in the Silurian of New York, in the Carboniferous of Michigan, and in the Permian of Kansas and Texas.

5. **Epochs of Triassic Copper Deposits.**—The important period of the history of the igneo-genetic ores in the eastern part of the continent closed in the late Paleozoic. A feeble recurrence of ore formation took place during the early Mesozoic when the Triassic traps of New Jersey, Connecticut, and the Bay of Fundy were injected as sheets or overflowed as lava streams. Copper ores of various types and some contact-metamorphic iron and copper ores developed in places.

6. **Cretaceous and Later Periods of Lead and Zinc Concentration.**—Since the Triassic, volcanism has rested and in the eastern part of the continent metal deposits have formed only by the concentrating power of flowing surface waters, or of ground-water in decaying rocks, or of ascending waters of meteoric origin.

In the Mississippi basin and in the Appalachian valleys, the Paleozoic beds have been searched by saline and carbonate waters containing carbon dioxide and hydrogen sulphide. This solution and deposition have resulted in the many lead and zinc deposits which we find in the limestones of that region. It is believed by some geologists, however, that these lead-zinc deposits are of magmatic origin.

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[1] C. R. Van Hise and C. K. Leith, The geology of the Lake Superior region, *Mon.* 52, U. S. Geol. Survey, 1911, p. 591.

7. **Tertiary and Recent Periods of Rock Decay.**—For a long time, in fact since the end of the Paleozoic, large parts of this continent have been land areas and their rocks have been exposed to weathering and decay. Such periods have indeed not been absent during any extended time interval, and in the iron ores of the Lake Superior region as well as in certain manganese ores of Arkansas, we find indications of long pre-Cambrian and shorter Paleozoic epochs of weathering and oxidation.

The mild climate which prevailed in the southern Appalachian region during the Tertiary favored the development of deposits caused by such agencies. The ores contain limonite, manganese dioxide, phosphates, and bauxite.

*In the Western or Cordilleran part of the continent we distinguish the following metallogenetic epochs:*

1. **The Pre-Cambrian Epochs.**—The western part of the Cordilleran region is lacking in great exposures of the pre-Cambrian but wherever they are present we find deposits of gold, iron, and copper, rarely other metals. The correlation of these with the eastern epochs is often difficult but on the whole they are connected with late pre-Cambrian intrusions of granite and subsilicic rocks. As may be expected, the deposits are of the high-temperature type and consist mainly of veins and lodes. Among them may be mentioned the gold lodes of the Black Hills of South Dakota, the lenticular gold-quartz veins and copper deposits of Wyoming, many smaller deposits in New Mexico;[1] the gold, iron, and copper deposits of certain parts of southern California and Arizona (near Yuma and Parker), the copper deposits of the Jerome district in Arizona, and many others.

2. **The Early Mesozoic Epoch.**—Throughout the Paleozoic and the larger part of the Mesozoic the great interior province, now occupied by the Rocky Mountains, was the scene of almost uninterrupted sedimentation. The phosphate beds of Idaho date from this period. Not so along the Pacific Coast; for here we find in places evidence of intrusions and lava flows dating back to the early Paleozoic, and during the Trias and early Jurassic great effusions of subsilicic lavas took place. Copper deposits are found in these igneous rocks and many of these were doubtless formed shortly after this igneous activity. Among them we may probably count many minor deposits in California, the Kennecott deposit in the Copper River district, Alaska, and the ores of the Prince William's Sound in the same territory. In the Jurassic lavas of Vancouver Island, C. H. Clapp has found indication of a mineralization of the type formed near the surface.[2]

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[1] W. Lindgren, L. C. Graton, and C. H. Gordon, The ore deposits of New Mexico, *Prof. Paper* 68, 1910, pp. 48–51.

[2] *Econ. Geol.*, 10, 1915, pp. 70–88.

3. **The Late Mesozoic Epochs.**—The third and much more important ore-forming period followed and accompanied the great batholithic intrusions of the Pacific coast, to which an age from late Jurassic to early Cretaceous is generally assigned.

These intrusions of intermediate quartz-monzonitic or granodioritic character took place upon an immense scale and extended from Alaska to Baja California. The large batholiths of the Sierra Nevada and of British Columbia are the most striking features, but innumerable intrusions of smaller volume broke through the crust along both sides of the great batholiths and throughout Oregon, Idaho, California, southern Arizona, western Nevada, Washington, and Alaska. One of the largest of these is exposed in the great granitic area of central Idaho. But throughout this revolution and during the birth of mountain ranges on the coast the Cretaceous was being quietly deposited at sea level all over the eastern Cordilleran region from Canada to southern Mexico.

An epoch of intense metallization followed these intrusions, within the areas indicated. The great interior masses of the batholiths are usually free from deposits, as shown in the High Sierra, in the Clearwater region, and in British Columbia. But along their margins mineral deposits formed in abundance, as along the gold belt of California, and along the two contacts of the batholith of the Canadian and Alaskan coast regions. The great placer fields of Alaska derived their gold from deposits of this epoch. Gold, primarily, and copper, secondarily, are the characteristic metals. Along the Pacific coast, where there is little limestone in the intruded sediments, lead is practically absent, but in the interior, as in Nevada and Arizona, where the intrusions came into contact with Paleozoic limestone, this metal, with zinc, begins to appear. Silver is everywhere present, but scarcely ever important, except where lead appears. Arsenic and antimony are not abundant, mercury is nowhere present in commercial quantities.

The metal-bearing belts are well marked on both sides of the great granodiorite batholith of British Columbia. Schofield[1] distinguishes a belt along the western contact characterized by high-temperature copper deposits and another on the east side where gold, silver, and lead appear in mesothermal deposits. The northwest trending fissures carry mainly copper, the fissures with northeast trend carry chiefly lead, silver, and gold. Buddington[2] makes a more elaborate division into five belts. Lead

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[1] S. J. Schofield and George Hansen, Geology and ore deposits of the Salmon River district, *Mem.* 132, Canada Geol. Survey, 1922, pp. 65–66.

S. J. Schofield, The fissure systems of British Columbia, *Trans.*, Canadian Inst. Min. Eng., 28, 1926, pp. 428–433.

[2] A. F. Buddington, Coincident variations in types of mineralization, *Econ. Geol.*, 22, 1927, pp. 158–179.

A. F. Buddington and T. Chapin, Geology and mineral deposits of southeastern Alaska, *Bull.* 800, U. S. Geol. Survey, 1929, 356 pp.

deposits are said to occur on both sides of the batholith and their presence does not seem to be dependent upon limestone country rock. Gold-quartz deposits follow the western batholithic contact in Alaska.

4. **The Early Tertiary Epoch.**—By the end of the Cretaceous, the intrusions had gradually spread eastward, toward the region where the Cretaceous sea had deposited a plastic mantle of sediments, resting on Paleozoic rocks. About the beginning of the Eocene, the igneous activity had reached the eastern margin of the Cordilleran region.

The predominating magmas were again of intermediate character, and solidified as granular or porphyritic rocks, standing between the granites and the diorites; they contrast markedly with the potassic and acidic magmas of pre-Cambrian times. While it is not necessary to limit strictly this igneous activity to a certain time, there is little doubt that most of it took place in the Eocene. The eruptions mainly took the form of intrusions and largely that of laccoliths, undoubtedly because, unlike the conditions of the shattered rocks of the Pacific coast, they were covered by this heavy, tough and still yielding mass of Cretaceous sediments. We find an enormous number of these intrusions at various horizons between the Cambrian and the Cretaceous or as dikes or stocks that break through the underlying pre-Cambrian. They are not comparable in extent to the great batholiths of the coast. They extend from British Columbia, through Montana, Utah, Nevada, Colorado, New Mexico, eastern Arizona, and eastern Mexico. For reasons already indicated many, perhaps most, of these intrusives never reached the surface. Only in a few cases, as in Montana and in Colorado, near Denver, do the strata of Laramie or Eocene age contain volcanic detritus.

The fourth Cordilleran epoch of metallization followed these intrusions and really represents the unbroken continuation of the third. Contact-metamorphic deposits and veins were formed in abundance around their margins. The characteristic metals are silver and gold with much lead and zinc, especially where the intrusions cut the limestones. Copper and iron are also present at such limestone contacts. Arsenic and antimony are far more in evidence than during the second epoch, but mercury is still absent. Rossland, Coeur d'Alene, Butte, Bingham, Tintic, and Leadville are representative districts. The deposits are mesothermal or hypothermal.

5. **The Late Tertiary Epoch.**—Orogenic disturbances followed or accompanied the intrusions; the whole Cordilleran region was lifted high above sea level, warped, and faulted. These disturbances may have facilitated sub-aërial eruptions; at any rate it is certain that the middle and close of the Tertiary witnessed outflows of lavas upon a magnificent scale. This represents the last phase of the igneous activity which began along the Pacific Coast.

These eruptions spread over large areas of the western part of the continent; less pronounced in British Columbia and Alaska, they are abundantly represented in California, Washington, Idaho, Colorado, Utah, Nevada, New Mexico, Arizona and attained their greatest development in Mexico. Andesites and rhyolites are the predominating rocks. In some places the flows attained such thickness that during the later part of the volcanic epoch intrusions of magmas consolidated in them with granular structure.

During these eruptions, not strictly contemporaneous throughout, a fifth metallization took place, of which the characteristic metals are gold and silver. These epithermal deposits were often of great richness which is further accentuated by secondary processes; in fact most of the "bonanzas" belong to this class. Lead and zinc are not conspicuous except where the metallization took place in limestone. Copper is not abundant. Tellurium and antimony are common.

Representative districts are: Bodie in California: Tonopah, Goldfield, the Comstock, Tuscarora and Jarbidge in Nevada; the Black Mountains in Arizona; and De Lamar in Idaho; the San Juan and Cripple Creek in Colorado; Pachuca, El Oro, and Guanajuato in Mexico.

Large areas of volcanic rocks are barren. The metal deposits seem to have formed only near or at the foci of igneous activity, where connection could be established with underlying magmas. The most recent eruptions were mainly basalts, which, as a rule, were not affected by metallization.

Later studies by W. S. Burbank,[1] in the San Juan Region in the vicinity of Ouray and Rico, have shown that the deposits which were formerly held to be late Tertiary include an earlier mineralization formed after largely laccolithic intrusions of late Cretaceous or early Eocene age. These earlier deposits are often of the replacement type and tend to spread laterally on favorable sedimentary contacts. They contain lead-zinc-silver ores or pyritic gold ores, related to the fourth epoch.

A marked period of erosion separates them from the late Tertiary deposits, which are mainly gold-bearing veins, formed after the San Juan tuff was laid down.

6. **The Post-Pliocene Epoch.**—The youngest metallogenetic province is that following the Pacific Coast line. It is of very late age—post-Pliocene apparently—and is characterized by mercury accompanied by few other metals. It developed in the Coast Ranges of California, also to some extent in Nevada, Texas, and Mexico following volcanic eruptions and contemporaneous with it was a great development of hot springs. In part the deposition goes on at the present time.

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[1] Revision of structure and stratigraphy in the Ouray district of Colorado and its bearing on ore deposition, *Proc.* Colorado Sci. Soc., 12, 1930, pp. 151–232.

An active circulation of atmospheric water was undoubtedly established in the Coast Ranges long before the quicksilver deposits were formed.

7. **Cretaceous or Later Epochs of Copper Concentration in Sedimentary Rocks.**—In addition to these six classes, whose connections with igneous rocks are indubitable, the disseminated copper ores of the southwest should find mention. They occur in sandstones, shales, or conglomerates ranging from the Carboniferous to the Cretaceous, and, in most cases, chalcocite is the original ore; frequently small amounts of silver are present. New Mexico, Arizona, Utah, Colorado, and Texas offer numerous examples of this class. While their origin is not wholly clear, many observers believe that they represent concentration by ground-waters of small amounts of copper originally derived from the pre-Cambrian deposits and since distributed through late sedimentary beds. In similar deposits we find ores of uranium, vanadium, and radium.

Summing up we have then in the Cordilleran region:

| | Principal metals | Principal rocks associated |
|---|---|---|
| 1. Deposits of the pre-Cambrian period: | Gold and copper. | Granites.<br>Diorites, gabbro. |
| 2. Deposits of the early Mesozoic epoch: | Copper. | Basalt, diabase.<br>Gabbro. |
| 3. Deposits of the late Mesozoic epoch: | Gold. | Granodiorite.<br>Quartz-monzonite, gabbro. |
| 4. Deposits of the early Tertiary epoch: | Gold, silver, copper, lead, zinc. | Granodiorite.<br>Quartz-monzonite.<br>Monzonite, with corresponding porphyritic rocks. |
| 5. Deposits of the late Tertiary epoch: | Gold, silver. | Andesite, latite.<br>Rhyolite. |
| 6. Deposits of the post-Pliocene epoch: | Quicksilver. | Basalt, andesite. |
| 7. Cretaceous or later concentrations in sedimentary rocks. | Copper, vanadium, uranium, radium. | Sandstone, shale, conglomerate. |

In deposits that are clearly connected with igneous rocks metallization is certainly a function of varying pressure and temperature; these factors being dependent upon depths below the surface, proximity of igneous bodies, and other conditions; metallization is also dependent on the relative solubility of the components of the ore-forming solutions and on the nature of the rocks in which deposition takes place. Primarily, however, it is probably a consequence of magmatic differentiation in batholithic magmas, and in the deeper magma basins which supplied the effusive rocks.

It is well established that magmas of different types contain different associations of the rare metals. For instance, tin and tungsten characterize acidic rocks, while nickel and cobalt are found chiefly in magmas rich in ferromagnesian constituents. At the same time rocks of a given general composition may, in different localities, vary considerably in the quantity of rarer metals contained.

Ore deposits formed after the eruption of an igneous rock will contain the rare metals characteristic of it. It may be true that for each differentiated rock type there are corresponding types of deposits, varying with the conditions of deposition.

As periods of long continued differentiation may materially modify the composition of a magma, it is conceivable that this might find expression in a progressive change in the character of ore deposits successively formed during these periods. The quicksilver deposits of California, for instance, may be the ultimate result of such long-continued differentiation.

As we look back over the wide domain of mineral deposits, we perceive the strong tendency to concentration of common or rare elements, by magmatic differentiation, by solution, or by mechanical transportation; we perceive also cycles of transformations, based on the laws of stability of chemical compounds. Even when deeply buried, the deposits may suffer many changes. Near the surface they are constantly subject to transmutations involving both concentration and dispersion. A few stable compounds are formed while the rest of the elements are scattered by mechanical and chemical transportation. Some constituents are carried down into the earth by the underground circulation of water, perhaps to form new deposits in other rocks. Ultimately erosion sweeps away the wreckage of the old deposits into basins of sedimentation where the elements may be reconcentrated, be it by direct precipitation or by the aid of living matter. The sediments may again be lifted and corrugated, again destroyed by erosion, and new eras of concentration begin.

In one aspect the science of mineral deposits is frankly utilitarian, but from the viewpoint of pure knowledge it records the principles governing the cycles of concentration of the elements. It traces the processes by which the primeval gases and magmas have become differentiated into the manifold complexity of the earth's crust.

# INDEX OF LOCALITIES

**Boldface** type indicates principal references

# SUBJECT INDEX

Principal compounds and uses of elements in *italics*.