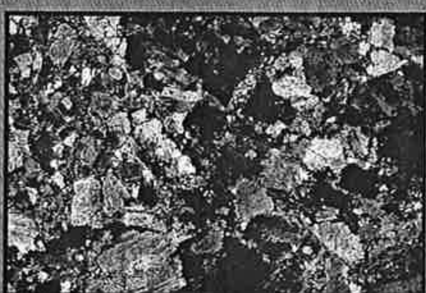
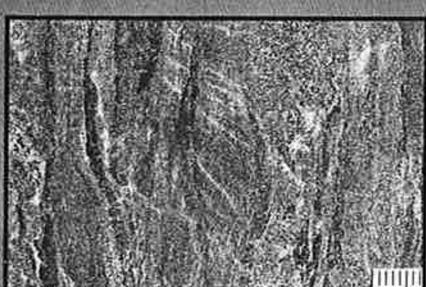
# ATLAS OF ALTERATION ASSEMBLAGES, STYLES AND ZONING IN OROGENIC LODE-GOLD DEPOSITS IN A VARIETY OF HOST ROCK AND METAMORPHIC SETTINGS

HELPING UNLOCK AUSTRALIA'S MINERAL SECRETS





**Pasi K. Eilu**
**Charter I. Mathison**
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THE UNIVERSITY OF WESTERN AUSTRALIA
*A Leading University*

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**(Centre for Strategic Mineral Deposits)**
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**COVER ILLUSTRATIONS:**

- Sulphidation front in banded iron-formation, Water Tank Hill mine, Western Australia. Greenschist facies
- Photomicrograph of proximal alteration and gold mineralisation in Granny Smith Granodiorite, Western Australia. Field of view 6.7 mm. Greenschist facies
- Cockade banding of quartz, ankerite and sulphides in a large vein, Wiluna deposit, Western Australia. Black elongated area is wallrock fragment. Sub-greenschist facies
- Proximal alteration in komatiite from Marvel Loch deposit, Southern Cross greenstone belt, Western Australia. Mid-amphibolite facies

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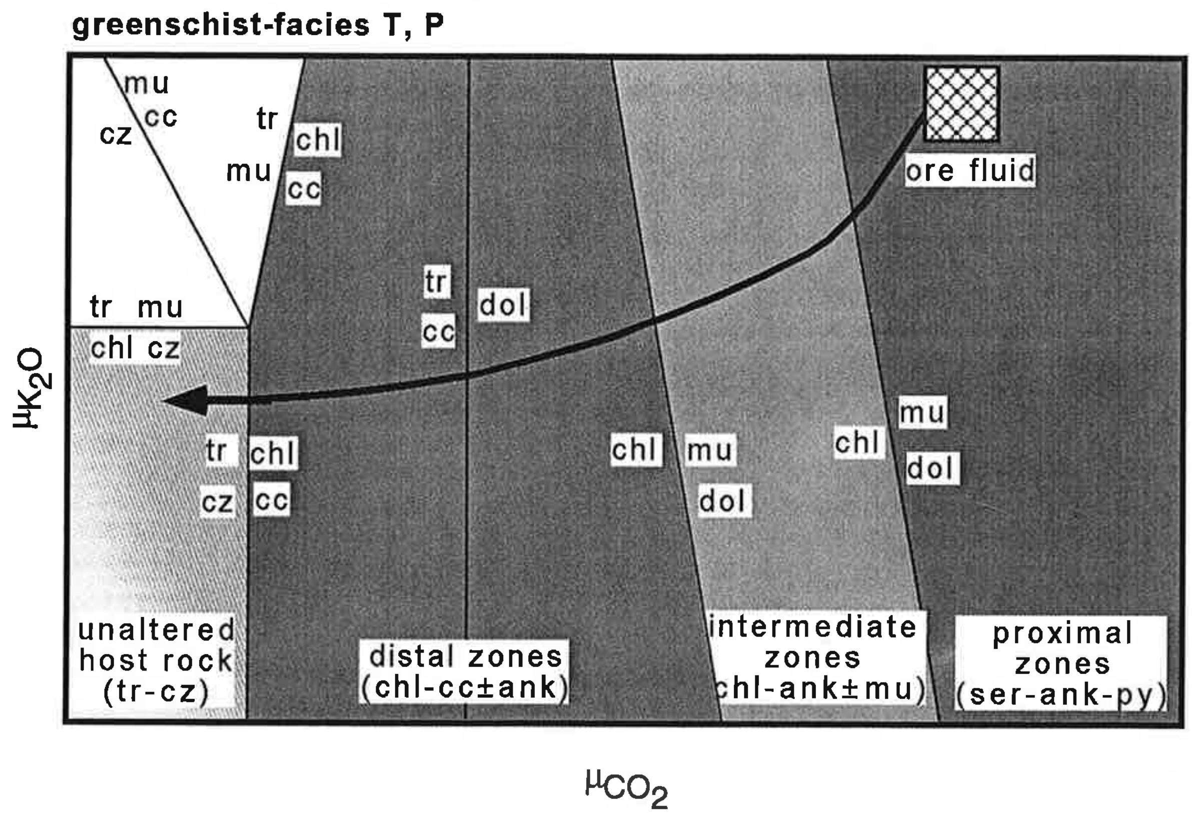
# PREFACE

OVERVIEW

Hydrothermal ore deposits form because large volumes of hydrothermal fluid have been focussed into relatively small volumes of the Earth's crust to generate economic concentrations of metallic resources where there are efficient metal depositional processes. As these fluids are normally out of equilibrium with the wallrocks, in terms of parameters such as P, T, pH, $fO_2$ and activities of components, wallrock alteration zones and primary geochemical and isotopic dispersion haloes commonly mark the passage of these fluids. The figure below shows, schematically, the generation of a zoned wallrock alteration halo through the interaction of ore fluid with a mafic wallrock under greenschist facies conditions. These haloes are normally much more extensive than the economic mineralisation itself, and hence have the potential to be used to target such mineralisation in exploration for hydrothermal ore deposits. In fact, wallrock alteration haloes have been used extensively in exploration for specific mineralisation styles. These include epithermal and hot-spring gold deposits, particularly of the acid-sulphate type (Berger & Bethke 1986, Henley et al. 1986, Pirajno 1992, Taylor 1996), volcanogenic massive sulphide deposits (e.g. Date et al. 1983, Hashiguchi et al. 1983, Elliot-Meadows & Appleyard 1991, Stanley & Madeisky 1993), and porphyry-style systems (e.g. Lowell & Gilbert 1970, Christie & Braithwaite 1986, Dilles & Einaudi 1992, Pirajno 1992, Sillitoe 1992, Kirkham & Sinclair 1996).




*Schematic figure showing the generation of a zoned wallrock alteration halo through the interaction of ore fluid with a mafic wallrock at greenschist facies PT conditions (designed by E.J. Mikucki 1998). Abbreviations: ank = ankerite, cc = calcite, chl = chlorite, cz = clinozoisite, dol = dolomite, mu = muscovite, py = pyrite, ser = sericite, tr = tremolite.*

ORGANISATION

In this atlas, the petrographic nature of wallrock alteration haloes around orogenic or "mesothermal" lode-gold deposits is described in a range of host-rock types over the complete range of metamorphic grades of the host rocks, with the emphasis on photographic presentation of alteration. This is preceded by a brief discussion of the deposit style and the constraints that the deposit parameters place on the nature of wallrock alteration.

CONTRIBUTIONS

Pasi Eilu, David Groves and Wendy Allardyce have compiled the text in this book and, unless otherwise stated, all photographs are by Charter Mathison based on specimens housed in the Centre for Strategic Mineral Deposits, Department of Geology and Geophysics, The University of Western Australia.

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# CHAPTER 1
# GEOLOGICAL CHARACTERISTICS OF OROGENIC LODE-GOLD DEPOSITS

## 1.1 TERMINOLOGY AND TECTONIC SETTING

In a recent review article, Groves et al. (1998) suggested that lode-gold deposits worldwide, which have been variously termed mesothermal, turbidite-hosted, slate-belt hosted, greenstone-hosted, Mother lode-type or gold-only deposits, are a coherent group of gold deposits with a common origin. They have suggested the term *orogenic gold deposit* for this deposit class, because all the deposits were formed during compressional to transpressional deformation processes at convergent plate margins in accretionary or collisional orogens. The deposits also occupy a unique depth range for hydrothermal ore deposits, with gold deposition ranging from 15-20 km depth to the near-surface environment, where antimony may also be present in mineable amounts and concentrations.

Figures 1 and 2 summarise the tectonic settings and depth ranges of these deposits in comparison with other gold (and Au-Cu and Au-Ag) deposits, and demonstrate that orogenic gold deposits may occur in the same orogens as Au-rich porphyry style, epithermal-style, or VHMS-style deposits, but normally over different depth ranges and in different geographic positions. The major consistent features of the orogenic gold deposits are outlined in the following sections. Unless otherwise stated, the summary of these features is taken from the recent reviews by Kerrich & Cassidy (1994), Robert & Poulsen (1997), Goldfarb et al. (1997) and Groves et al. (1998). These reviews have exhaustive lists of references on the deposit style.

## 1.2 GEOLOGY OF HOST TERRAINS

Perhaps the single most consistent characteristic of the deposits is their association with deformed metamorphic terrains of all ages. Observations from preserved Archaean greenstone belts and most recently-active Phanerozoic metamorphic belts throughout the world indicate a strong association of gold and greenschist-facies rocks. However, some significant deposits occur in higher metamorphic-grade terrains (e.g,, McCuaig et al., 1993; Bloem et al., 1994; Kerrich & Cassidy, 1994) or in lower metamorphic-grade domains within the metamorphic belts of a variety of geological ages (e.g. Hagemann et al., 1992; Gebre-Mariam et al., 1995). In the Archaean of Western Australia and southern Africa, a number of synmetamorphic deposits extend into granulite-facies domains (Böhmke & Varndell, 1986; Barnicoat et al., 1991; van Reenen et al., 1994). Pre-metamorphic protoliths for the auriferous Archaean greenstone belts are predominantly volcano-plutonic terrains of oceanic back-arc basalt and felsic to mafic arc rocks. Terrains dominated by clastic marine sedimentary rocks that were metamorphosed to metagreywacke, slate, phyllite, and mica schist host mostly younger ores, and these rocks are important in some Archaean terrains (e.g. Slave Province, Canada: Padgham, 1986).

## 1.3 DEPOSIT MINERALOGY

The orogenic gold deposits are typified by quartz-dominant vein systems with ≤3-5 percent sulphides (normally dominated by Fe-sulphides) and ≤5-15 percent carbonate minerals. Veins may also contain albite, white mica, chlorite, scheelite and tourmaline in greenschist-facies domains, or amphibole, diopside, biotite/phlogopite, tourmaline or even garnet in amphibolite-facies domains. Vein systems may be continuous over a vertical extent of 1-2 km with little change in mineralogy or gold grade: subtle mineral zoning does occur, however, in some deposits (e.g. Mt Charlotte, Golden Crown and Hill 50 in Western Australia: Mikucki & Heinrich, 1993; Uemoto, 1996; Kelly, 1998).

Gold/silver ratios range from 10 (most common) to 1 (less common), and gold grades vary at the deposit scale from 30 g/t (historical low-tonnage deposits) to 2-10 g/t (modern bulk-tonnage underground to open-pit resources). The siting of gold varies: in some deposits it is dominantly in veins, and in others it is dominantly in sulphidised wallrocks.

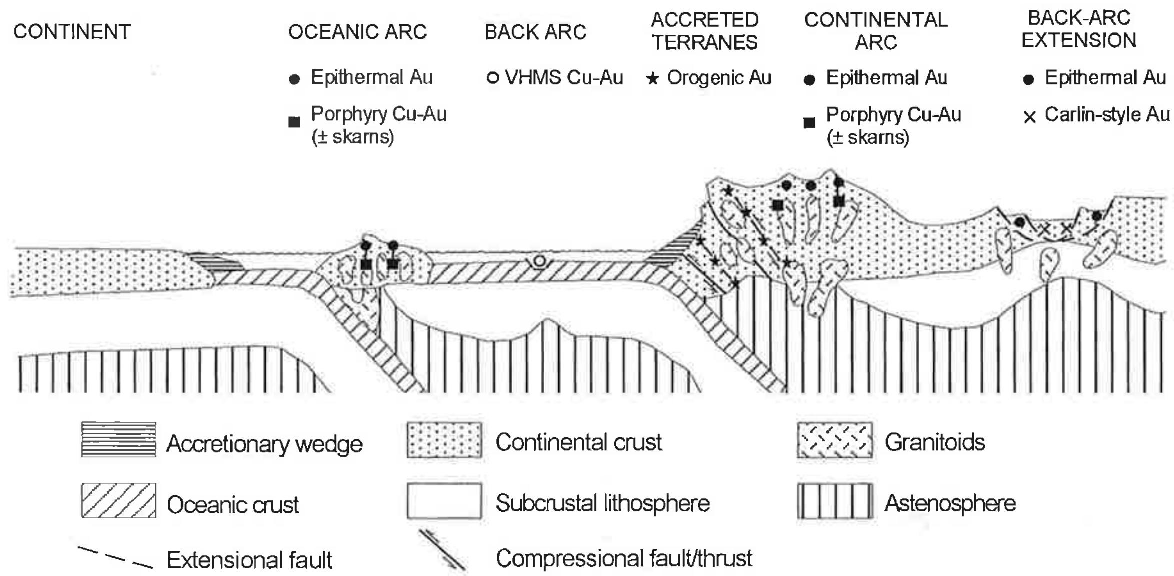



*Figure 1 Tectonic settings of gold-rich epigenetic mineral deposits. Epithermal veins, and gold-rich porphyry and skarn deposits, form in the shallow (<5 km) parts of both island and continental arcs in compressional through extensional regimes. The epithermal veins, as well as the sedimentary rock-hosted type Carlin ores, are also emplaced in shallow regions of back-arc crustal thinning and extension. In contrast, the so-called "mesothermal" gold ores (here termed orogenic gold) are emplaced during compressional to transpressional regimes, and throughout much of the upper crust, in deformed accretionary belts adjacent to continental magmatic arcs. Note that both the lateral and vertical scales of the arcs and accreted terranes have been exaggerated to allow the gold deposits to be shown in terms of both spatial position and relative depth of formation. From Groves et al. (1998).*

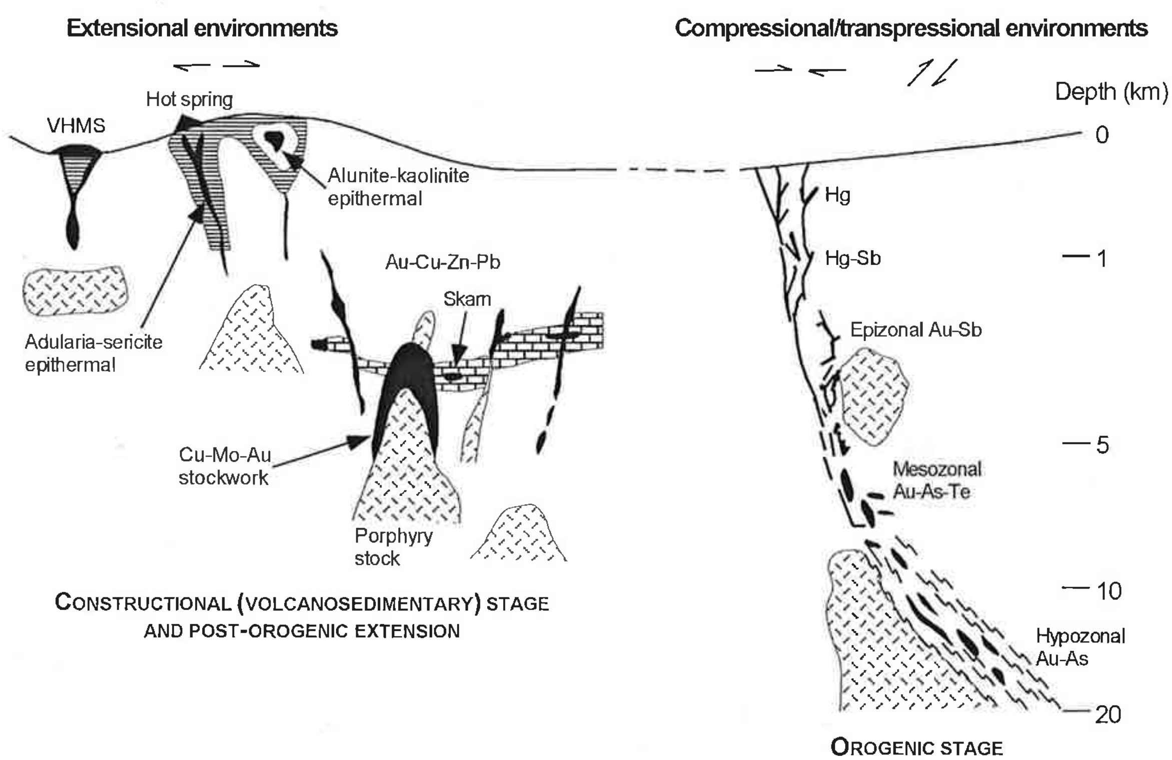



*Figure 2 Schematic representation of crustal environments for hydrothermal gold deposits in terms of depth of formation and structural setting within a convergent plate margin. This figure is, by necessity, stylised to show deposit styles within a depth framework. It is not implied that all deposit types or depths of formation will be represented in a single ore system. Adapted by Groves et al. (1998) from Groves (1993), Gebre-Mariam et al. (1995) and Poulsen (1996).*

Sulphide mineralogy may reflect the lithogeochemistry of the host rock. Arsenopyrite is more common in metasedimentary rocks, whereas pyrite or pyrrhotite (± arsenopyrite) are more typical in metamorphosed igneous rocks. Gold-bearing veins exhibit variable enrichments in As, B, Bi, Hg, Sb, Te, Se and W; Cu, Pb and Zn concentrations are generally close to background values.

## 1.4 HYDROTHERMAL ALTERATION

Wallrock alteration shows much stronger lateral variations around the ore zones than vertical variations in the plane of the orebodies. As described in detail in the sections below, the mineral assemblages and width of the alteration zones vary with wallrock type and crustal level of hydrothermal alteration. An alteration halo may extend laterally from gold mineralisation for a few centimetres up to 1-2 km. In general, there is a positive correlation between the size of the deposit and the lateral extent of alteration, and the alteration halo tends to be less extensive in the amphibolite-facies than in greenschist-facies host rocks.

Most alteration zones show evidence for the addition of significant amounts of $CO_2$, S, K, $H_2O$ and LILE (Large Ion Lithophile Elements). This is reflected by the replacement of most or all of the minerals present prior to alteration by calcite, dolomite, ankerite, pyrite, chlorite, sericite or fuchsite in greenschist facies, and by calcite, pyrrhotite, Ca amphibole, diopside, grossular, biotite or K feldspar at higher metamorphic grades. Sulphidation is, generally, most prominent in banded iron-formation (BIF), and carbonation is dominant in mafic and ultramafic host rocks.

Significant $SiO_2$ enrichment in many mineralised zones is evident from the commonly large volumes of quartz veins. However, silicification *sensu stricto*, that is addition of silica into the host rock and not just formation of quartz veins, has been rarely convincingly documented. Rather, $SiO_2$ released by alteration reactions is redeposited in host rock and quartz veins and, in some cases, partly removed from the mineralised site (Böhlke 1989). Well-documented examples of silica enrichment in the host rocks are mostly associated with the most proximal zones of alteration in lode-gold deposits from greenschist-facies metasedimentary rocks (Boyle, 1979; Read & Meinert, 1986; Peters et al., 1990; Kontak & Smith, 1993).

## 1.5 ORE FLUIDS

The wallrock alteration assemblages and fluid inclusion studies indicate that the ore fluid responsible for deposition of this deposit group was typically a low-salinity, near-neutral, $H_2O\text{-}CO_2 \pm CH_4$ fluid, which transported gold as a reduced sulphur complex. Fluids consistently have $CO_2$ concentrations of 5-25 mole percent and $\delta^{18}O$ values from 5-8 per mil in metavolcanic belts to 7-12 per mil in metasedimentary belts.

## 1.6 STRUCTURE

There is strong structural control of mineralisation at a variety of scales. Many would argue that structure is the single most important control on gold mineralisation. At the goldfield to deposit scale, most of the deposits are sited in second- or third-order structures, most commonly near regional-scale (typically transcrustal) deformation zones. An idea of the variety of structural styles that can develop in a deposit can be gauged from Figure 3, which is taken from Witt (1993). Although the hosting structures are commonly brittle-ductile, they can be of several types including:

(i) brittle faults to ductile shear zones with low-angle to high-angle reverse, strike-slip or oblique-slip movements;
(ii) fracture arrays, stockwork networks or breccia zones in competent rocks;
(iii) foliated zones with pressure solution cleavage; and
(iv) fold hinges and associated reverse faults in "locked-up" folds (Cox et al., 1995) in turbidite and/or BIF-bearing sequences.

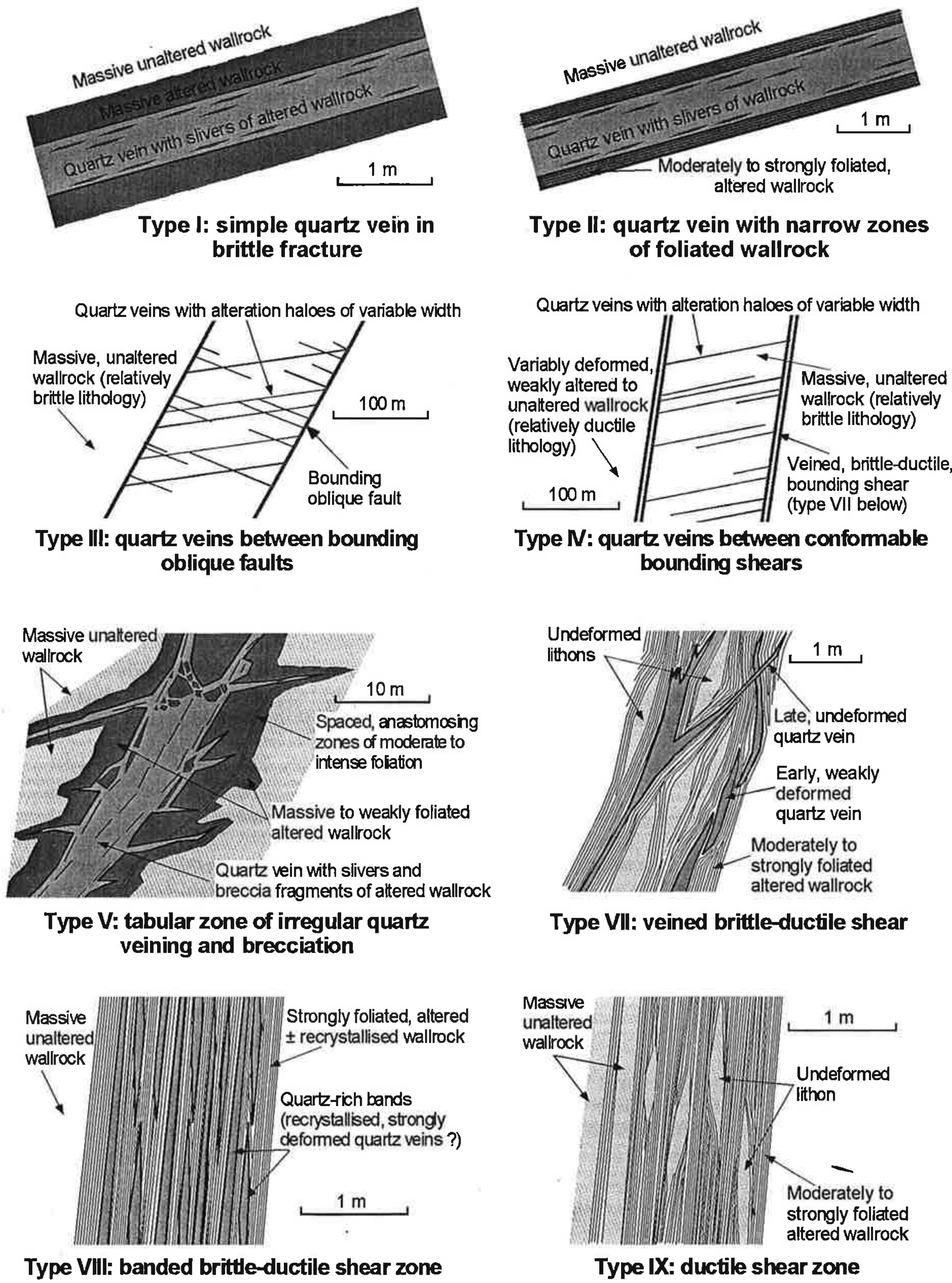



*Figure 3 Sketches of the types of structures hosting lode-gold mineralisation in the Menzies-Kambalda area. Note that the scale is variable and that the scale bar gives only an order of magnitude. Modified from Witt (1993). There may also be replacement-style deposits: for example, in BIF.*

## 1.7 TIMING OF GOLD MINERALISATION

There is increasing evidence that gold mineralisation was related to deep-seated thermal events, metamorphism and granitoid emplacement, that postdated the metamorphic peak (± granitoid emplacement peak in some terrains but not in others) at higher structural levels, i.e. in greenschist-facies domains. In contrast, deposits in amphibolite-facies domains tend to be syn-peak metamorphism. The deposits are therefore commonly late in the overall deformational cycle, although, in some provinces, they may be overprinted by much later deformational fabrics (e.g. Nurmi & Sorjonen-Ward, 1993). Robust geochronological data normally place these deposits about 20-70 m.y. after volcanism in the hosting volcanic belts (e.g. Kerrich & Cassidy, 1994; Yeats & McNaughton, 1997), although deposits in sedimentary belts may postdate sedimentation by a greater time period (e.g. Drew et al., 1996; Goldfarb et al., 1997).

# CHAPTER 2
## INTRODUCTION TO ATLAS

### 2.1 INTRODUCTION

As mentioned in Chapter 1, wallrock alteration haloes are typically well developed around orogenic lode-gold deposits and, due to their vertical and lateral extent, can be most useful in exploration and orebody definition. Such wallrock alteration has been described for different host rocks,,from deposits in variable metamorphic-grade domains in a variety of unpublished theses and some published papers (notably Mueller & Groves, 1991), and has recently been reviewed by Eilu et al. (1998). Emphasis has been placed on the mineralogy and petrology of proximal to distal alteration zones (e.g. Mueller & Groves, 1991), and the mass transfer and pathfinder element variations that might provide vectors to ore (e.g. Eilu et al., 1998). However, there is no systematic atlas of alteration assemblages to aid the field geologist or core logger to correctly recognise such alteration styles and their significance.

This atlas is designed to fill this gap by providing a pragmatic guide to wallrock alteration assemblages, styles and zones in a variety of host rocks at a variety of metamorphic grades. It is designed to be used as a field guide or as a guide during systematic logging of drill core intersecting alteration zones in environments where orogenic lode-gold deposits are known or expected.

### 2.2 ORGANISATION OF ATLAS

The alteration assemblages show greater variations with metamorphic grade than with host rock, and there are associated greater variations in easily determined characteristics, such as colour (e.g. bleaching) and reaction with dilute HCl. In concert with this, the atlas is ordered such that alteration zones are described in terms of ranges in metamorphic grade of the host rocks, with the alteration of different host rocks described at each metamorphic grade. A systematic description of alteration assemblages, from distal through intermediate to proximal zones, is given for each host rock within its metamorphic-grade category. Allowance is made for contrasting styles of gold mineralisation.

The associated photographs are selected to depict the visual appearance of distal to proximal alteration zones at the hand-specimen scale. Fields of view similar to those that would be visible under a hand lens are included. The scale-bar divisions are 1 mm. These photographs are mostly of diamond-drill core sawn and lapped to a flat surface, and photographed in sunlight, normally with a wet surface. Within each photographic plate, samples are arranged such that the most proximal alteration is on the right hand side. The hand-specimen photographs are supplemented with photomicrographs of thin sections where specific assemblages or variations can be demonstrated better at that scale. Although all photographs, except one, are from late Archaean deposits of the Yilgarn Block of Western Australia, the features depicted are similar to those in deposits in other orogenic belts and of other ages in similar host rocks and metamorphic grades.

### 2.3 DEPOSIT LOCATIONS

The locations of orogenic lode-gold deposits or deposit belts that are mentioned in this atlas are given on maps in Appendix A.

### 2.4 GRAPHICAL LOGS

In terms of assessment of alteration zones in drill cores and drill chips, it is important to develop a graphical log that depicts the alteration assemblages, which can be defined by using a hand lens, and their spatial variations. An example of a graphical log is given in Figure 4. The atlas is specifically designed to aid this process. This may be relatively straightforward where there are essentially homogeneous host rocks (Fig. 5), but may be more complicated where there is more than one rock type in the alteration zone.

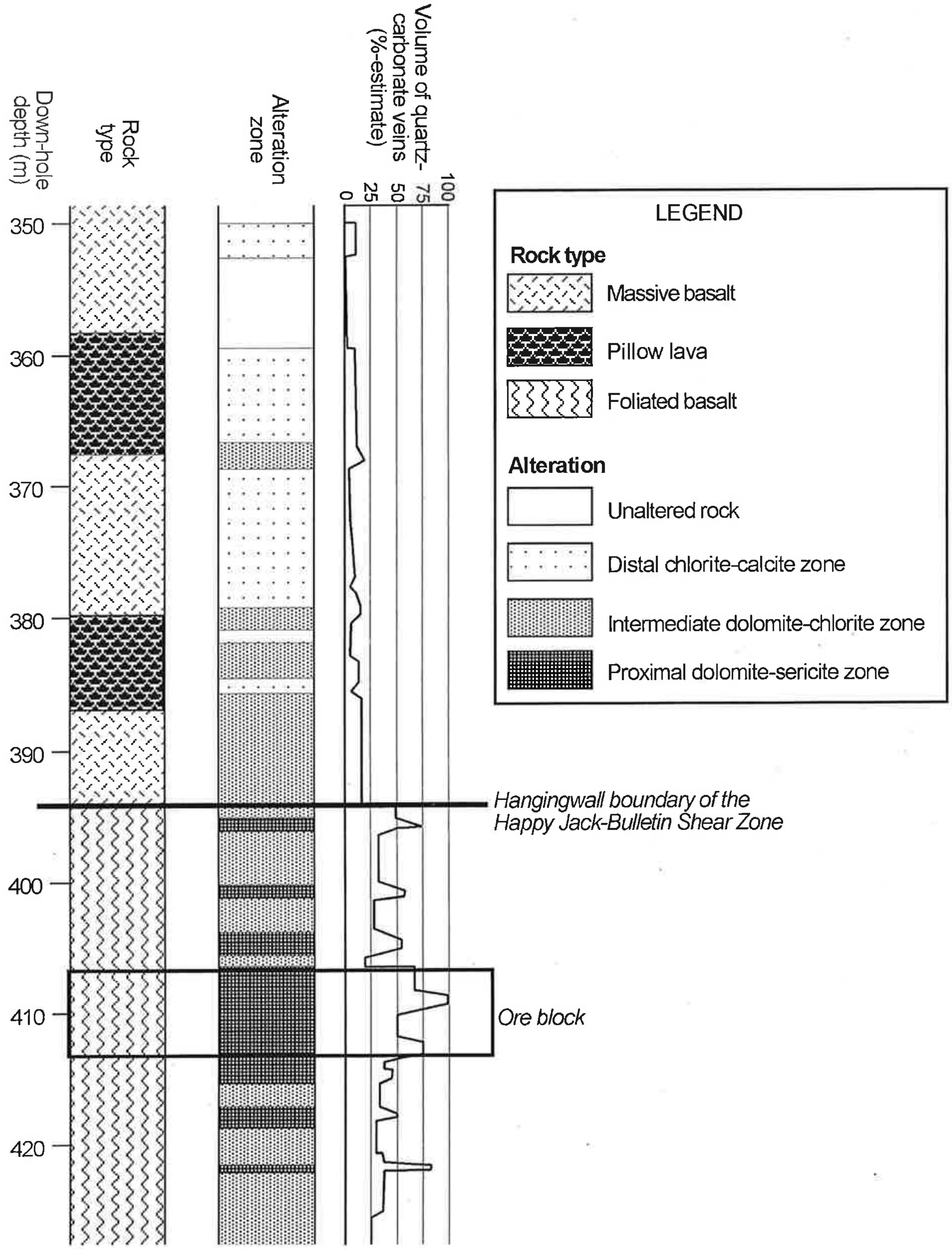



*Figure 4 An example of a graphic log. The log presents part of a diamond-drill core from the Bulletin mine, Wiluna (Eilu, 1996a). Several other parameters can be added into such a presentation: for example, sample locations, volume of veins, volume of sulphides, volume of bleached zones per metre (if these zones are frequent and narrow), degree of pre- or post-mineralisation alteration, gold grade, magnetic susceptibility, etc.*

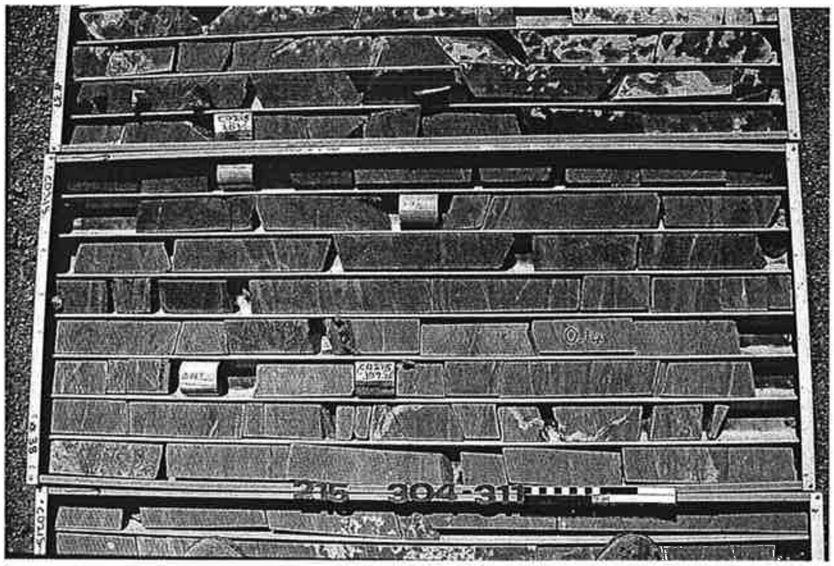


*Figure 5* *Bleaching, as can be displayed in drill core. Note the change from distal and intermediate, unbleached alteration to bleached proximal alteration in the intermediate volcanogenic sedimentary rock. The boundary of the main proximal zone is also marked by red tape. The yellowish colouring is due to initial weathering of ankerite in the intermediate and proximal alteration zones. Mid-greenschist facies conditions, Sunrise Dam mine, Laverton. The core tray is 1 m long. Photograph by Pasi Eilu.*

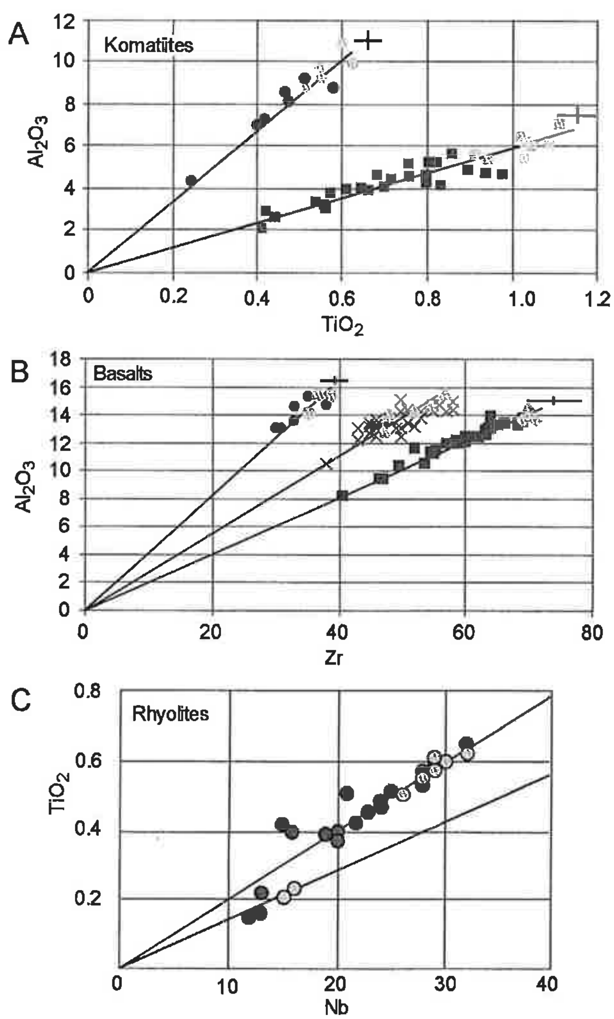



*Figure 6* *Immobile element plots for (a) metakomatiites from Cue and (b) metabasalts from Wiluna, both in Western Australia, and (c) metarhyolites from Noranda in Quebec, Canada. Each trend forms a comagmatic sequence. Blue and red symbols indicate unaltered and altered rocks, respectively. The elements plotted are immobile and conserved, so all samples within a sequence plot in the same trend (the diagonal line crossing the origin), although the data in each trend include samples from distal to proximal alteration zones. Data in (a) are from Eilu (1996a), (b) from Eilu (1996b), and (c) from Stanley & Madeisky (1995). Error bars in (a) and (b) indicate the total analytical error in the data.*

## 2.5 GEOCHEMICAL DISCRIMINATION OF HOST ROCKS

The rock types, even if intensely altered, can be readily and cheaply identified geochemically through the use of relatively immobile elements (e.g. Al, Cr, Nb, Ti, Y, Zr) which have distinctive ratios in the precursor rocks, as initially suggested by Hallberg (1985). Figure 6 shows the fields for typical Archaean volcanic and intrusive rocks in terms of some of these immobile elements. The ratios of these immobile elements are not only definitive in highly altered rock, they are normally preserved also in the saprolite zone of the weathering profile (e.g. Monti 1987).

## 2.6 ACID REACTIONS AND STAINING TECHNIQUES

Hand-specimen and hand-lens examination can be supplemented by reaction with dilute HCl and staining tests for carbonate minerals. The distinction between calcite and ankerite

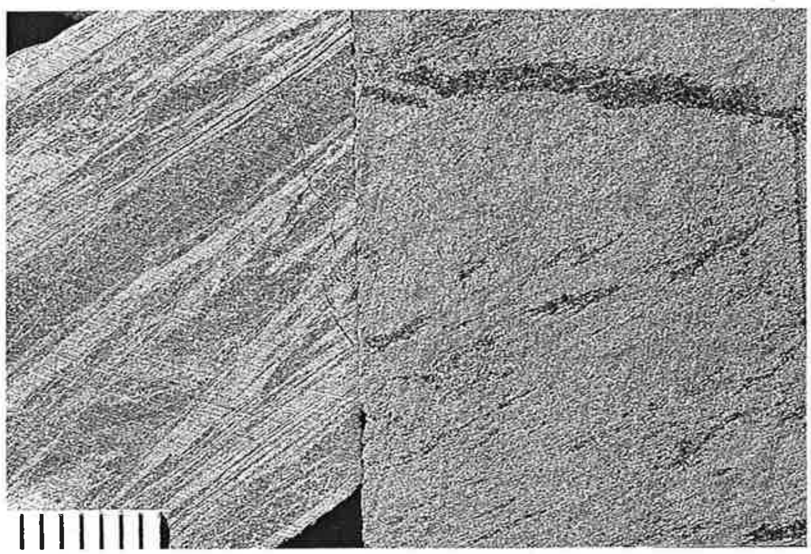


*Figure 7 Stained metabasalt samples from distal (at right) and proximal (at left) alteration zones. Distal sample is purple and proximal sample is blue, indicating the presence of calcite and ankerite or Fe-bearing dolomite, respectively. The distance between lines in the scale bar is 1 mm For this and subsequent photographs, except for photomicrographs, where the field of view is given.*

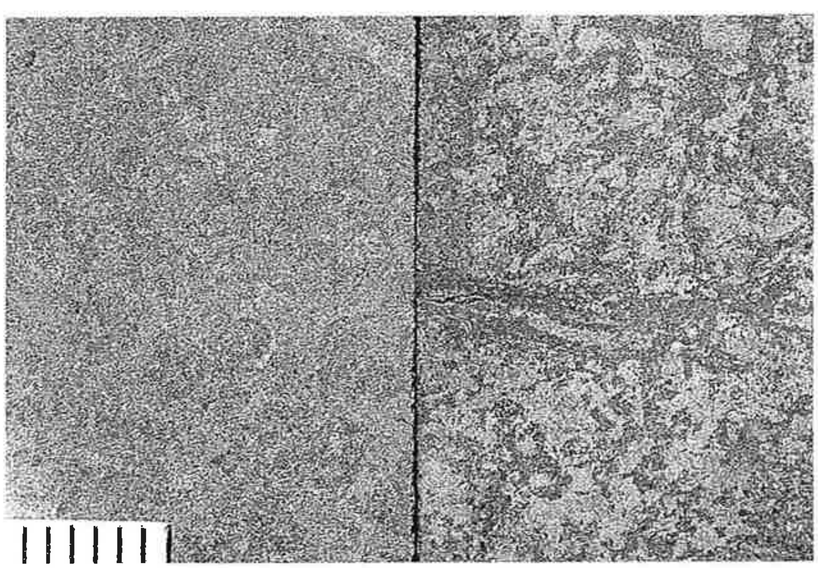


*Figure 8 Unstained (at left) and stained (at right) altered dolerite. Mineral assemblage albite-chlorite-ankerite-rutile (leucoxene). Note how clearly staining (blue colour) indicates the presence of Fe-bearing carbonate.*

or dolomite (or more rarely, siderite or magnesite) can be aided by the strong reaction of calcite with cold dilute HCl versus the weak reaction of ankerite and dolomite only with warm or hot dilute HCl, and the lack of reaction with siderite or magnesite. More definitive recognition of carbonate species, and particularly the textural relationships between the different generations of carbonate minerals, can be achieved with staining with Alizarin red-S + K-ferricyanide in dilute HCl (Hutchison, 1974). Importantly, this is a field technique, so a sophisticated laboratory is not needed. Examples of stained core illustrating different generations of carbonate minerals are shown in Figures 7 and 8, and the technique is summarised in Appendix B.

## 2.7 COMMON ALTERATION MINERALS

The common alteration minerals, their formulae, and diagnostic criteria for their recognition are given in Appendix C. It must be recognised that minerals in alteration zones in rocks at low metamorphic grade may be fine-grained and difficult to recognise. In such cases, a set of thin sections of standard alteration zones, defined at hand-specimen scale, may be a useful adjunct to alteration definition. Some photomicrographs are shown, together with photographs of hand specimens, where there is a need to illustrate specific textures and minerals in the alteration zones.

## 2.8 PATHFINDER ELEMENTS

As demonstrated in the review by Eilu et al. (1998), which also lists most of the critical papers on the subject, specific geochemical indices which define the degree of carbonation, sericitisation, etc., and specific pathfinder elements may show systematic variations from proximal to distal alteration zones in orogenic lode-gold deposits. In the atlas, a set of schematic diagrams exemplifies how the common alteration assemblages are accompanied by schematic variations in parameters such as carbonation indices and pathfinder elements. Integration of mineralogical and geochemical parameters, and definition of their spatial variations, is required to gain maximum benefit from alteration studies.

# CHAPTER 3
# ALTERATION IN ROCKS METAMORPHOSED TO SUB-GREENSCHIST TO MID-GREENSCHIST FACIES

## 3.1 GENERAL DESCRIPTION OF THE DEPOSITS

In the descriptions below, rocks that are at low strain, and in which original textures and structures can be recognised, are given precursor names: the prefix *meta-* is assumed but omitted. Where the rocks are pervasively deformed and metamorphosed, they are given metamorphic names.

### 3.1.1 Style of mineralisation

Rocks metamorphosed and altered at lower- to mid-greenschist facies form the most common host to orogenic lode-gold mineralisation. The dominant style of deformation in these deposits is brittle-ductile. At greenschist facies, in general, a brittle style of deformation dominates in the most competent rock types: for example, in granitoids and other felsic rocks and in BIF, and in all rocks at sub-greenschist facies (below the P-T range of lower-greenschist facies). Ductile deformation, on the other hand, tends to be more dominant in mafic and ultramafic rocks, and becomes more significant with metamorphic grades higher than mid-greenschist facies. Most of the structural subtypes developed in sub- to mid-greenschist deposits are Types I, II, IV and V, shown in Figure 3.

In rocks metamorphosed to sub- to mid-greenschist facies, as well as at higher metamorphic grades, the shape of the orebody is, in most cases, tabular, although this may be complicated if two or more mineralised structures occur close together (Fig. 9).

The most common structural types of veining are: breccia veins, folded and boudinaged veins, and laminated veins. Breccia veins, showing open-space filling (comb, cockade, crustiform and colloform textures), are most typical and, normally, the sole significant vein type in deposits in sub-greenschist facies domains and in competent host rocks. However, they may also be important in deposits in higher-grade metamorphic settings and in other rock types (brittle breccia: Figs 10 & 11; open-space filling: Figs 10, 12 & 13). The breccia veins may grade into a more widely spaced vein network, which is the dominant style in some deposits (Figs 14 & 15). Veins showing ductile deformation may contain intensely altered slivers of wallrock (e.g. Type I in Fig. 3) and, if there is gold in the veins, most of it is adjacent to these wallrock slivers and vein margins.

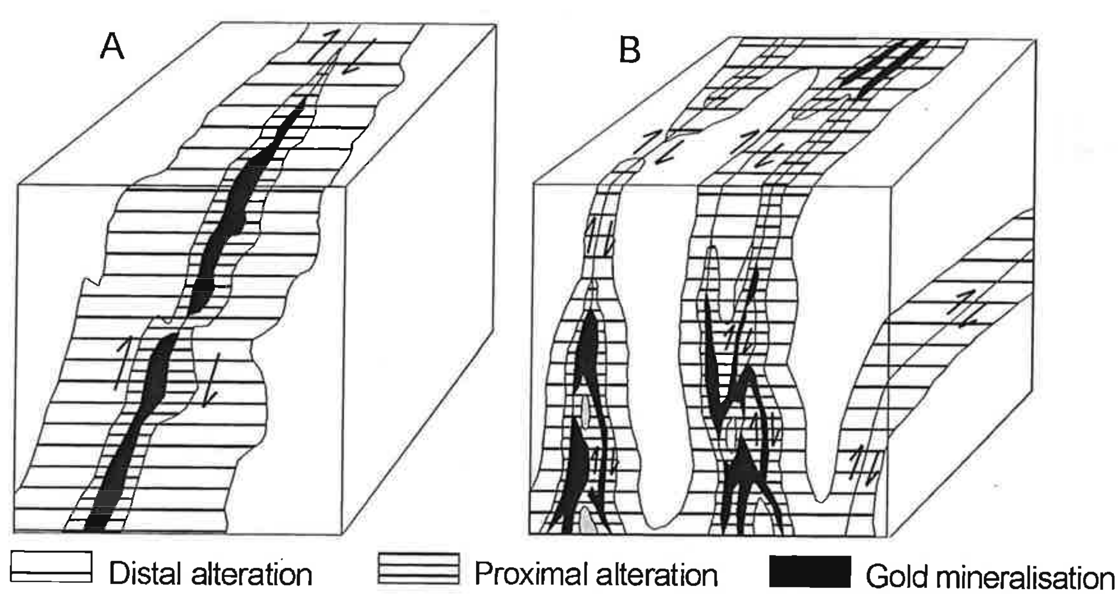



*Figure 9 Extent of alteration around a lode-gold deposit. A. A simple case with one quartz vein or shear zone. B. A complex case with several subparallel veins or shear zones adjacent to each other. Based on Colvine et al. (1988), Mikucki et al. (1990) and Ridley & Barnicoat (1990).*

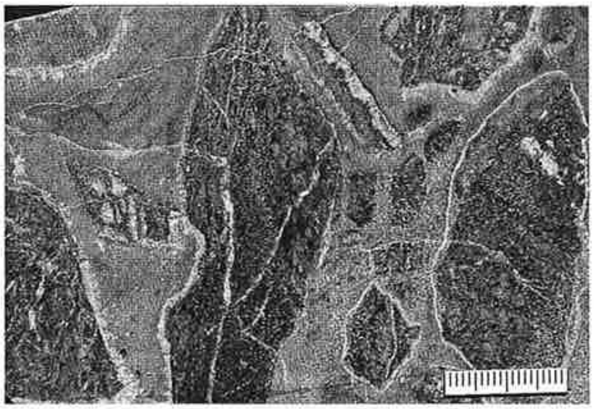


*Figure 10 Brittle deformation in a deposit in a lower-greenschist facies setting: brecciated granophyric unit in a dolerite sill, Golden Kilometre mine, 30 km N of Kalgoorlie. Breccia matrix consists of white carbonate adjacent to the fragments, and pale-grey, very fine-grained quartz and traces of pyrite. The texture of the breccia veins is typical for open space-filling. Most or all of the alteration in the fragments, e.g. intense bleaching and sulphidation, predates the major brecciation. Hence, no concentric zoning is present in the fragments. Scale bar divisions are 1 mm.*

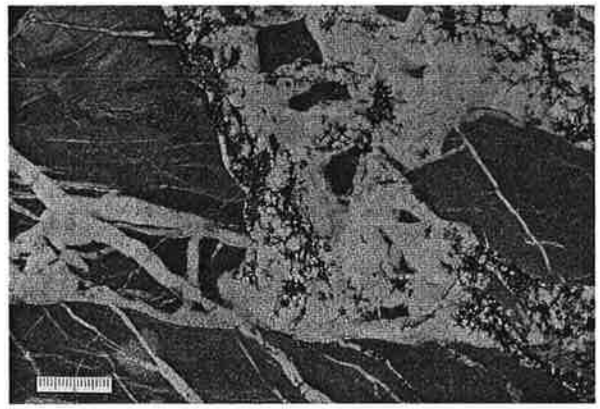


*Figure 11 Brittle deformation in a deposit in a sub-greenschist facies setting: brecciated basalt at the West Lode mine, Wiluna. Breccia veins comprise white carbonate and quartz, and dark-grey stibnite. The major brecciation postdates an event of weaker brittle deformation that is related to the numerous thin quartz-carbonate veins and pervasive bleaching of the host rock.*

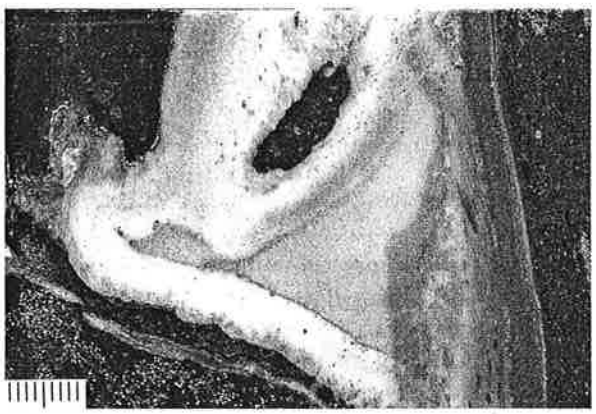


*Figure 12 Open space-filling in brittle shear zone at Sunrise Dam mine. Host rock is intermediate volcanogenic sedimentary rock and metamorphic grade is mid-greenschist facies. White vein margins are carbonate. Vein interior is very fine-grained, almost chalcedonic, quartz. Locally, host rock contains disseminated pyrite.*

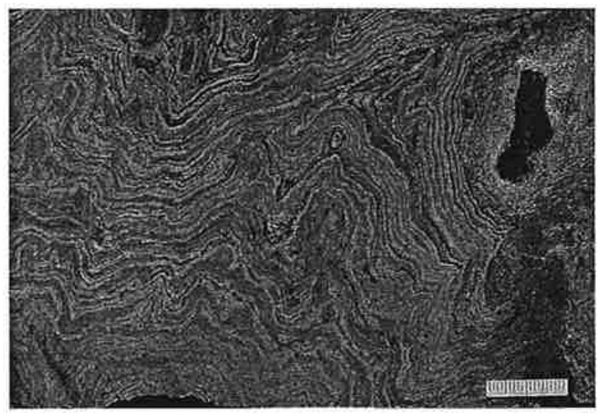
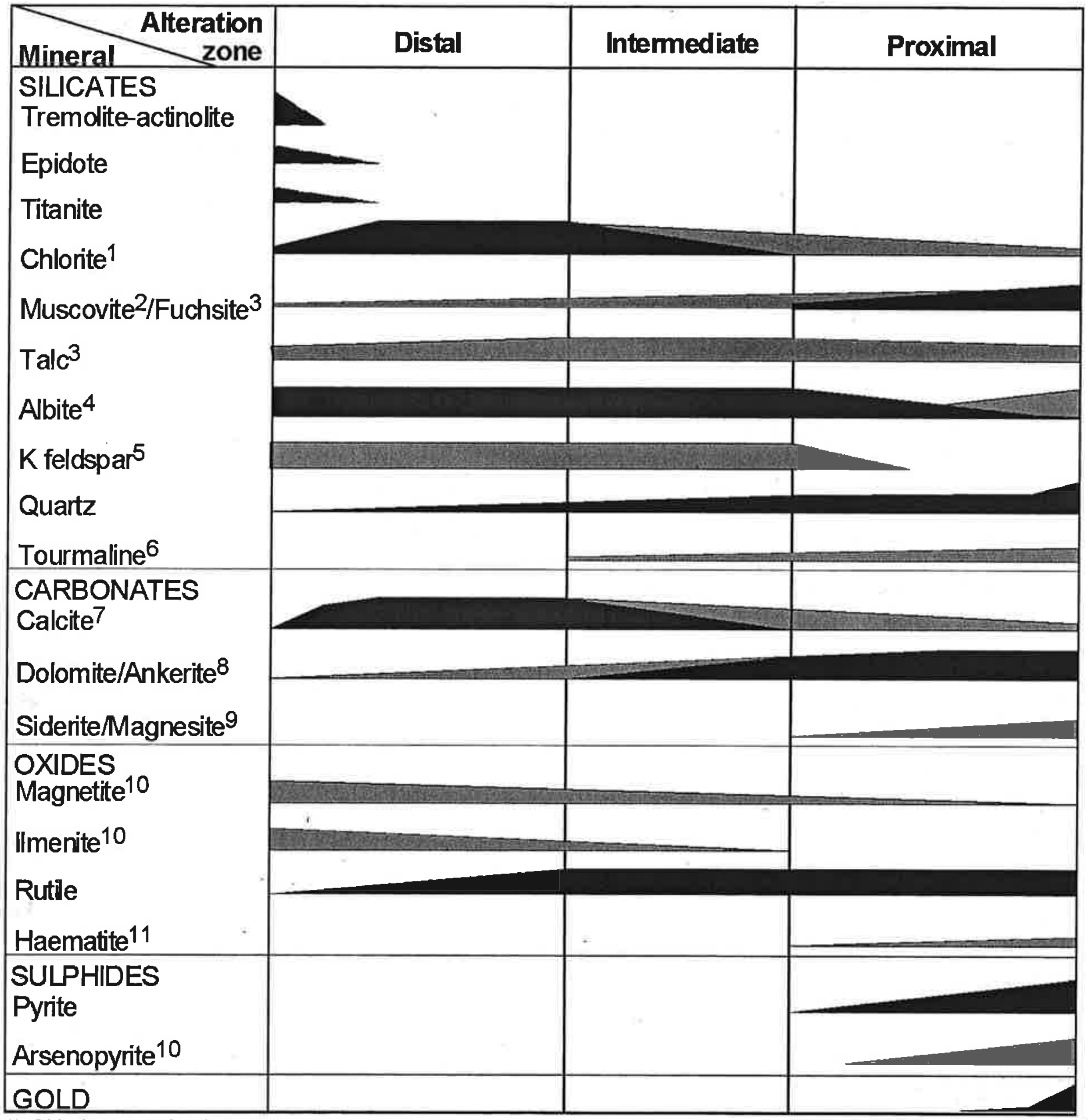


*Figure 13 Cockade banding (open-space filling) formed by quartz (grey), ankerite (pale-brown due to initial weathering) and sulphides (black) in a large vein in a deposit in a sub-greenschist facies setting at Wiluna. The black elongated area is a wallrock fragment in the vein.*

*Figure 14 Brittle deformation: mineralised vein sets enveloped by alteration at Mount Charlotte mine, Kalgoorlie. Field of view is 2 m. Metamorphic grade is lower- or mid-greenschist facies. Photograph by David Groves.*

*Figure 15 Brittle deformation in the Granny Smith gold mine, Laverton: network of conjugate shear veins with bleached haloes enveloping the veins. Host rock is granodiorite, altered under lower-greenschist facies conditions. The dark band on left is a blast hole. Width of the compass is 11 cm. Photograph by Juhani Ojala.*




1) Chlorite may also be part of the most proximal assemblage if muscovite content is low.
2) Alteration may produce some sericite in the intermediate zone. Retrograde sericitisation may be present in the distal and intermediate zones in intermediate and felsic rocks. Pre-alteration sericite may be present in clastic metasedimentary rocks.
3) Present only in ultramafic rocks.
4) Locally, significant albite formation occurs in the areas of most proximal alteration.
5) Detected in the alteration sequence where also present in unaltered rock.
6) The extent of tourmaline formation is unclear, as tourmaline is not present in all alteration envelopes.
7) In felsic rocks, calcite may be present also in the proximal zone.
8) In ultramafic rocks, dolomite or ankerite may be present also in the distal zone.
9) Siderite or magnesite are detected in a few cases, the former chiefly in BIF and the latter in ultramafic rocks.
10) May or may not be present.
11) Haematite may be present in the most oxidised systems.

*Figure 16 Schematic summary of the paragenetic alteration sequence around orogenic lode-gold deposits in sub- to mid-greenschist facies environments. Proximity to mineralisation increases to the right. Black represents the common case, and green indicates a less common occurrence. Zone widths shown here bear no resemblance to actual widths in the field. This sequence is based on data from Woad (1981), Fyon et al. (1983), Smith & Kesler (1985), Sandiford & Keays (1986), Kishida & Kerrich (1987), Colvine et al. (1988), Newberry & Brew (1988), Böhlke (1989), Mikucki et al. (1990), Thompson et al. (1990), Ames et al. (1991), Mueller & Groves (1991), de Ronde et al. (1992), Hagemann (1992), Couture & Pilote (1993), Mikucki & Ridley (1993), Witt (1993), Vielreicher (1994), LeAnderson et al. (1995), Ojala (1995), Eilu & Mikucki (1996), Mumin et al. (1996), Bierlein et al. (1997) and Robert & Poulsen (1997).*

The dominant veins in the mineralised zone are filled by quartz, variable amounts of carbonates (normally, ankerite or dolomite), albite and pyrite and traces (if any) of scheelite, tourmaline, apatite, pyrrhotite, arsenopyrite, gersdorffite, chalcopyrite, sphalerite, tetrahedrite, galena, molybdenite, gold, electrum, tellurides, selenides, stibnite and Bi-minerals (e.g. Spencer, 1906; Boyle, 1986; Colvine et al., 1988; de Ronde et al., 1992; Kontak & Smith, 1993; Phillips & Hughes, 1996). Quite commonly, the same ore minerals, or at least some of them, are sited in the mineralised wallrock. In many deposits, only quartz, carbonate, and pyrite ± arsenopyrite and gold are detected in the veins. There may be some variation due to variable metamorphic grade or host rock: for example, stibnite appears to be more characteristic in deposits in sub-greenschist facies domains (Hagemann, 1992; Gebre-Mariam et al., 1995; Goldfarb et al., 1997) and molybdenite, scheelite and tellurides in granitoids in slate belts (Phillips & Hughes, 1996). More commonly, however, there are regional variations in the accessory ore and gangue mineral assemblages: within a group of deposits, and despite variation in rock type, tourmaline or a set of tellurides may be characteristic, whereas a different minor- and trace-mineral assemblage may be present in an adjacent mining camp.

Examples of deposits at sub- to mid-greenschist facies are listed in Appendix D.

### 3.1.2 General description of alteration

In general, the alteration envelope is tabular and follows the shape of the structure hosting the mineralisation (Fig. 9A). This simple picture is, however, complicated if two or more structures with alteration envelopes related to mineralisation are adjacent. In such a case, the alteration haloes enveloping the structures may overlap, and form either a wider, continuous alteration envelope or a more complex interference pattern where the individual alteration haloes overlap and locally are separated by domains of unaltered rock (Fig. 9B). A continuous alteration envelope is more common in deposits where ductile deformation dominates, and in narrow shear zones, whereas a complex pattern is more typical of deposits with dominantly brittle deformation, and in areas of anastomosing shear zones in both brittle and ductile deformation environments.

The alteration sequence in sub- to mid-greenschist facies host rocks can be divided into three main zones. These can be named according to their diagnostic minerals as follows, from distal to proximal zones: calcite-chlorite, calcite-dolomite and sericite zones. Only in ultramafic rocks and BIF is the sequence somewhat different: there are calcite-dolomite, dolomite and sericite or talc zones in the former, whereas alteration in the latter displays weak carbonation and very weak sericitisation, if any. The alteration sequence is summarised in Figure 16.

The three-zone sequence is easy to detect in most, but not all, host rock types. In BIF, possibly only the proximal, sulphidised, zone can be easily defined. In intermediate and felsic rocks, it may be difficult to detect other than the proximal alteration zone or to distinguish between the alteration zones (e.g. Read & Meinert, 1986; Wigget et al., 1986; Peters et al., 1990), at least during early stages of exploration in an area when the local style of alteration is not yet well understood. In addition, the detection of alteration around a gold deposit may remain difficult through all stages of exploration in argillic sedimentary rocks, as visible alteration zonation may be weakly developed in these rock types (Goldfarb et al., 1986, 1997; Read & Meinert 1986; Mumin et al., 1996; Ramsay et al., 1996). The most obvious changes at the boundaries of the main alteration zones are summarised in Table 1.

The alteration halo around a deposit in a sub- to mid-greenschist facies domain varies from about 5 m up to 1-2 km in width, to ≥5 km along strike, and to ≥2.5 km down dip of the structure hosting mineralisation (Colvine et al., 1988; Newberry & Brew, 1988; Mikucki et al., 1990; McCuaig & Kerrich, 1994; Phillips & Hughes, 1996; Goldfarb et al., 1997; Robert & Poulsen, 1997). The extent of alteration depends on the size of the deposit and the size and duration of the hydrothermal system.

*Table 1 Changes that are easy to detect in drill core or outcrop at the boundaries of the major alteration zones in most greenschist-facies host rocks (Eilu & Mikucki, 1996). These are described in more detail in Section 3.2.*

| Alteration zone boundary | Change at the zone boundary |
|---|---|
| Unaltered rock → Chlorite-calcite zone | Calcite in, pervasive carbonation |
| Chlorite-calcite → Calcite-dolomite zone | Dolomite in |
| Calcite-dolomite → Sericite zone | Sericite in, bleaching |

In addition to the changes at the alteration zone boundaries, gradual mineralogical and textural trends exist across the alteration sequence. Calcite and chlorite characterise distal alteration and are gradually replaced by sericite and Fe-bearing dolomite or ankerite with increasing proximity to ore (Fig. 16). In addition, Ti-bearing magnetite, ilmenite and titanite are replaced by rutile (leucoxene), and the intensity of deformation, and volume of veins and sulphides, increase gradually from distal to proximal areas around all deposits. Other characteristic features of the alteration sequence are described in detail in Section 3.2.

It should be noted, however, that a sequence of different rock types within an alteration halo may produce a set of "unbleached and bleached zones" that does not strictly follow the trend in alteration intensity. For example, a set of felsic porphyries intercalated with ultramafic rocks in a shear zone may form such a sequence with "unbleached zones" being the ultramafic units and "bleached zones" being the felsic units, although no difference exists in the intensity of alteration between the rock types. If all of these rock units are intensely foliated and at least weakly altered, it may be difficult, without geochemical whole-rock analyses, to recognise the different rock types in outcrop or drill core, and an erroneous alteration sequence may be established.

Alteration zoning reflects chemical changes within the alteration envelope around an orogenic lode-gold deposit, as described in Section 1.4. These changes can be defined by geochemical alteration indices and pathfinder elements. In addition, geochemical vectors towards potential mineralisation can be defined by these parameters, as extensively discussed by Eilu & Mikucki (1996) and Eilu et al. (1998). Two examples of geochemical zoning around deposits in low-greenschist facies settings are shown in Figure 17.

## 3.2 ALTERATION IN DETAIL

### 3.2.1 Distal alteration zones

*Zone width*

The width of the distal alteration zones varies from 1 cm to 2 km (Smith & Kesler, 1985; Phillips, 1986; Clout et al., 1990; Skwarnecki, 1990; Gebre-Mariam, 1994; Vielreicher, 1994; Eilu & Mikucki, 1996).

*Colour and appearance at the meso- and microscopic scale*

There is rarely any difference in colour or texture compared to unaltered rock, although, in some cases, an originally very dark rock may be slightly bleached if carbonation is very intense. However, the volume of quartz-carbonate veins is somewhat greater than in unaltered rock.

A typical feature in mafic and some intermediate rocks is the development of leucoxene. This is visible in drill core as pale grey or pale brown specks or skeletal pseudomorphs after titaniferous magnetite or ilmenite, up to 5 mm in diameter, disseminated in an otherwise dark green or dark grey background (Figs 18, 19, 20 & 21; Section 3.3). Leucoxene is, in fact, titanite or rutile (pale grey and pale brown), titanite being common in the most distal parts of the alteration zone, but rapidly replaced by rutile towards the proximal parts of the zone (see Appendix C and compare samples in Fig. 18). Note that titanite is also relatively common in unaltered rocks: its presence alone does not necessarily indicate distal alteration.

*Mineral assemblage*

Distal alteration is, in most rock types, characterised by the mineral assemblage calcite+ chlorite+ quartz+ albite+ rutile (Fig. 16). In ultramafic rocks, the assemblage is calcite+ dolomite+ chlorite+ quartz ± talc, and in BIF, magnetite+ quartz+ chlorite ± hematite, Fe-amphibole, stilpnomelane, calcite and pyrite. Dolomite may also be present in sedimentary rocks and K-feldspar in sedimentary and felsic intrusive rocks, but this necessitates the occurrence of pre-alteration dolomite and K-feldspar, respectively.

*Response to HCl and carbonate staining*

A strong reaction with cold dilute HCl and the formation of a purple colour by carbonate staining (Hutchison, 1974), without indications of other carbonates (except in some ultramafic rocks), *in both veins and the host rock*, are diagnostic for the distal alteration zone (Fig. 7). If the rock investigated is fine-grained, it may be difficult to identify the stained calcite dissemination by the naked eye. Therefore, it is important always to view the stained samples using a hand lens or, preferably, a binocular microscope.

*Features to recognise in thin section*

In thin section, several replacement reactions are evident: olivine, pyroxene, and magmatic amphibole are completely replaced at the outer margin of the zone and, hence, are absent in the zone. Some tremolite-actinolite, serpentine, epidote, magnetite, ilmenite or titanite may be present, but these minerals are gradually replaced by chlorite+ calcite+ quartz+ rutile (+ dolomite ± talc in ultramafic rocks) within the zone. In BIF, all or most of the magnetite or hematite is preserved in the zone.

*Indications of zone boundaries*

In drill core or outcrop, the outer boundary of the distal alteration zone (or *distal boundary*) is easily detected by the change from rock with nil to weak reaction with dilute HCl to that with vigorous reaction, due to the pervasive appearance of calcite in altered rock. *This reaction must take place in the wallrock itself, not just in veins.* The boundary can

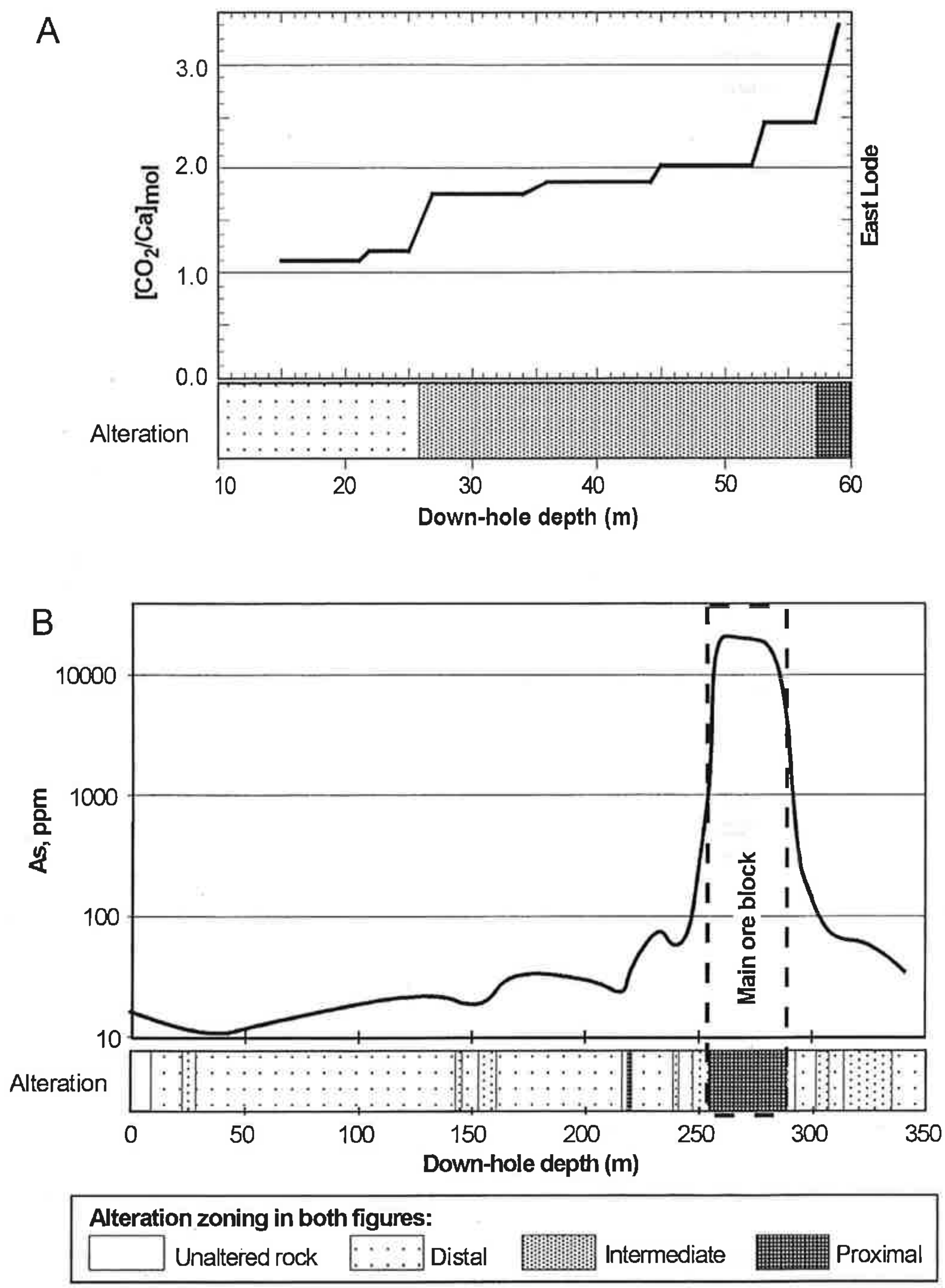



*Figure 17 A. Carbonation index trend from inner distal through intermediate and proximal alteration zones to the East Lode at Lake View, Golden Mile deposit, Kalgoorlie (after data from Phillips, 1986). B. Variation in arsenic concentration from unaltered rock, across a sequence of altered rocks, to the main ore block in the Bulletin mine (after Eilu & Mikucki, 1998).*

also be checked by carbonate staining: from unaltered rock to distal zone, there is a change from no staining, except possibly in some veins, to purple staining in both veins and in the wallrock.

The *proximal boundary* is marked by the disappearance of calcite in ultramafic rocks, appearance of abundant pyrite in BIF (sulphidation front), and appearance of dolomite or ankerite in other rock types. Generally, there is no change in colour on recently cut rock surfaces, and the zone boundary can only be delineated after determination of carbonate species by staining or XRD analysis. However, if the rock surface has been exposed to humid air for a few weeks or months, dolomite and ankerite develop a yellow surface layer due to initial weathering, as both are Fe-bearing in orogenic lode-gold systems, and so the boundary between the zones can be easily detected (Fig. 4). However, the yellow surface layer may be weak in medium- and coarse-grained felsic rocks, which have a pale colour throughout the alteration sequence, and this criterion cannot be readily applied to alteration zone mapping in such rocks. The easiest and cheapest way to check the change in carbonate species is by staining with Alizarin red-S + K-ferricyanide in dilute HCl (Hutchison, 1974): at the zone boundary, there is a change from a purple to purple + blue colour, reflecting the change from calcite to calcite + Fe-dolomite/ankerite, as shown in Figure 7.

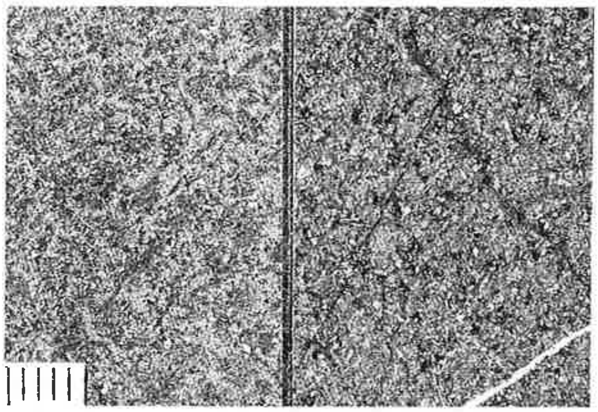


*Figure 18 Outer distal (left) and inner distal alteration (right) in a tholeiitic dolerite, Bulletin mine. Note the difference in the colour of the abundant green minerals and the disseminated Ti-mineral, leucoxene: white titanite and paler green amphibole in less altered rock, pale-brown rutile and darker green chlorite in the more intensely altered rock. In addition, greenish-yellow epidote is present in the less altered rock.*

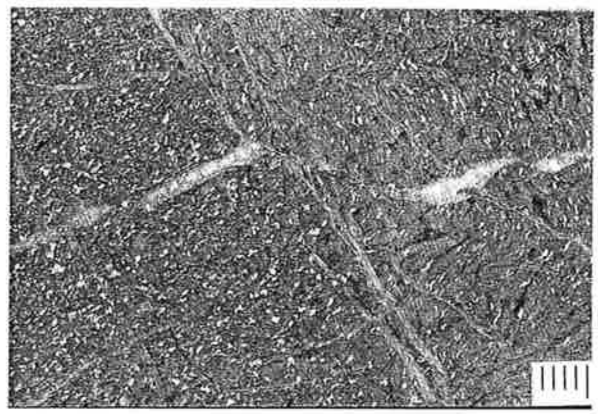


*Figure 19 Intermediate alteration in a tholeiitic dolerite, Bulletin mine. Note the pale-brown rutile (leucoxene) and, on the right, weak ductile deformation.*

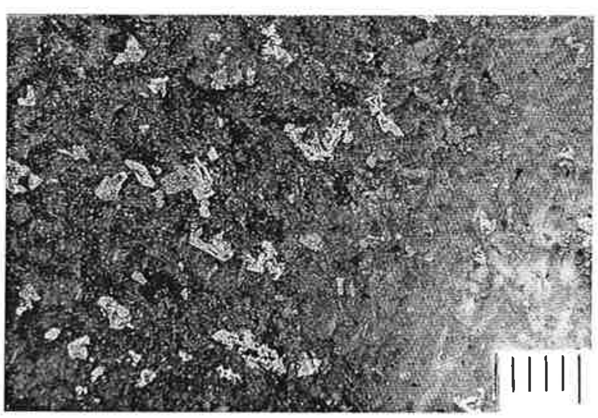


*Figure 20 Dark-green intermediate and bleached proximal alteration in the Golden Mile Dolerite in Mt Charlotte mine. Skeletal, pale-brown rutile (leucoxene), after the primary magmatic ilmenomagnetite, and yellow pyrite are present in both alteration zones.*

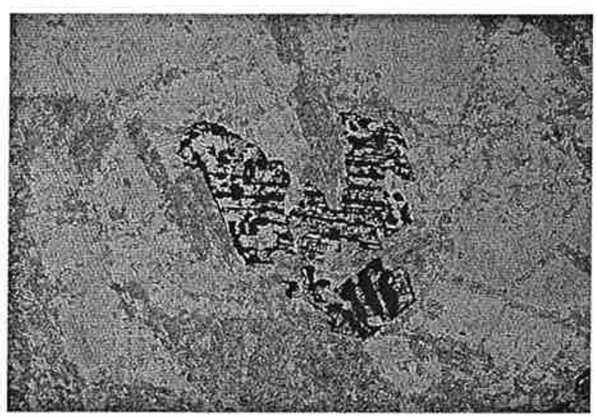


*Figure 21 Photomicrograph of leucoxene in Golden Mile Dolerite, Mt Charlotte mine. Intermediate alteration-zone mineral-assemblage: albite + ankerite + chlorite + quartz + rutile. Rutile with ankerite and quartz have replaced a primary magmatic ilmenomagnetite grain. Note the characteristic skeletal texture. Compare with rutile (leucoxene) in Figure 20, which depicts the same rock, alteration zone and similar skeletal rutile as this photomicrograph. Plane polarised light. Field of view 15 mm.*

### 3.2.2 Intermediate Alteration Zones

*Zone width*

The width of the intermediate alteration zones varies from 1 mm to 100 m (Smith & Kesler, 1985; Phillips, 1986; Skwarnecki, 1990; Eilu & Mikucki, 1996).

*Colour and appearance at the meso- and microscopic scale*

The intermediate alteration zone may be more deformed, foliated and/or brecciated, than the distal alteration zone, although most of the features present in the distal zone are generally still visible; for example, the leucoxene in mafic and some intermediate rocks is preserved. In the intermediate zone, the leucoxene is brownish in all cases, indicating that it is rutile (Figs 20 & 22).

If only a recently cut surface is available, the colour of rocks in the zone is generally the same as in the distal zone. Partial sericitisation, which is uncommon, may make an originally dark-coloured rock paler than its distal and unaltered counterparts (Fig. 23). After a few weeks or months of exposure to humid air, a distinct yellow or yellowish pale brown colour is developed in practically all rocks in the zone (Fig. 6).

*Mineral assemblage*

Ultramafic rocks are characterised by the mineral assemblage dolomite/ankerite+ biotite+ chlorite+ quartz+ rutile ± talc (Figs 16 & 24), and other rock types generally by calcite+ dolomite/ankerite+ chlorite+ quartz+ albite+ rutile (Figs 16, 19 & 22). In BIF, it is generally not possible to delineate an intermediate alteration zone at all (Woad, 1981; Thompson et al., 1990; Newton, 1991; Vielreicher, 1994).

*Response to HCl and carbonate staining*

A strong reaction with cold dilute HCl and the formation of purple (calcite) and blue (dolomite/ankerite) colours by carbonate staining (Hutchison, 1974) are diagnostic for the intermediate alteration zone (Fig. 6). Those ultramafic rocks that do not contain calcite are an exception, as there is no obvious reaction with dilute HCl and staining only produces a blue colour.

*Features to recognise in thin section*

Changes in mineral assemblages, which take place gradually within the intermediate alteration zone, include gradual replacement of calcite and chlorite by dolomite/ankerite + quartz (± talc in ultramafic rocks), and the complete replacement of magnetite and ilmenite by rutile (BIF excluded). In some intermediate and mafic rocks, chlorite is partially replaced by sericite.

An increase in the degree of deformation, if present, is also visible in thin section, either as an increase in the intensity of foliation or brecciation, or both. Increase in the degree of brecciation in some rocks (e.g. granitoids) may be detected only in thin section.

*Indications of zone boundaries*

The *distal boundary* of the zone is described in Subsection 3.2.1 (proximal boundary of the distal zone).

The *proximal boundary* of the intermediate alteration zone is, in most cases, marked by a striking change in colour, leading to complete bleaching (Figs 5, 14, 15, 20, 22, 25 & 26). This bleaching is related to the appearance of sericite (muscovite ± paragonite) and the simultaneous disappearance of chlorite. In many cases, the proximal boundary is also marked by an increase in pyrite content and a large increase in the intensity of deformation and volume of quartz-carbonate veins. In many ultramafic rocks, K-metasomatism results in the formation of fuchsite and the appearance of a bright-green colour at the zone boundary (Figs 27 & 28). In some ultramafic rocks, probably in the most magnesian compositions, there is no K-mica; chlorite is, instead, replaced by talc + carbonate + quartz, which also results in a bleached appearance. Likewise, sericitisation is not characteristic of BIF alteration, as the Al content of the rock is commonly too low to allow K-silicate formation.

### 3.2.3 Proximal Alteration Zones

*Zone width*

The width of the proximal alteration zones varies from 1 mm to 50 m (Phillips, 1986; Skwarnecki, 1990; Gebre-Mariam, 1994; Eilu & Mikucki 1996).

*Colour and appearance at the meso- and microscopic scale*

The single most diagnostic feature of the proximal alteration zone is bleaching (Figs 5, 14, 15, 20, 22, 25 & 26), as described above. BIF is an exception, as, in this case, the most striking feature is sulphidation. This sulphidation in orogenic lode-gold systems can be distinguished from sulphide-facies iron formation, where all sulphides are stratabound, by the fact that there is a sulphidation front that crosscuts the primary banding of the BIF (Figs 29 & 30) and is clearly related to mineralised quartz-carbonate veins (Phillips et al., 1984; Mueller & Groves, 1991; Bullis et al., 1994). The magnetite bands may be replaced by sulphides, whereas very fine-grained magnetite within the quartz-rich (chert) bands may be armoured from replacement even within the proximal zone. In addition to fuchsite in ultramafic rocks, there are rare cases where another green mica, roscoelite, has formed in the proximal alteration zone. It has been detected at the Golden Mile, for example, where the roscoelite-bearing unit is known as the Green Leader (Fig. 31).

Other characteristic features are the intense deformation (Figs 25, 26, 27, 28, 32 & 33) and veining in all deposits, formation of pyrite ± arsenopyrite, and formation of fuchsite in many ultramafic rocks (Figs 27 & 28). Brecciation may have been multi-stage, so that a breccia contains fragments of previously formed breccias. If hematite is present, the zone or part of it may be red, as is the case in certain areas at the Golden Mile (Fig. 22), although hematite need not be related to gold mineralisation at all, but be an indication of a later oxidising event in the area. If present in the alteration halo and related to the gold mineralisation, tourmaline, which is black to the naked eye and when viewed under a hand lens, and green in thin section, is also most abundant in the proximal zone.

*Ore minerals*

Pyrite and arsenopyrite are the most common sulphides, normally occurring as idiomorphic porphyroblasts (Figs 20 & 23). Instead of, or in addition to, arsenopyrite, gersdorffite may occur in rocks with a high Ni content, chiefly in ultramafic rocks (e.g. Böhlke, 1989; Eilu, 1996b). Abundant stibnite tends to characterise deposits in sub-greenschist facies environments (Fig. 11), although there are indications that high stibnite content may reflect a high Sb background in the host rocks, and not variation in the temperature of mineralisation (Goldfarb et al., 1997). Other ore minerals that may be present in the alteration zone are listed in Subsection 3.1.1.

The total concentration of sulphides increases gradually towards gold-bearing veins, and is highest in economic ore. The total volume of sulphides generally remains below 10%, commonly at 1-5%. Only in BIF are sulphides more abundant. Sulphides and gold may occur in both veins and adjacent wallrock. Arsenopyrite, in most cases, is less abundant and less extensive than pyrite: commonly, arsenopyrite is only present in detectable amounts within the gold mineralisation itself. This is not the case, however, for deposits in sedimentary host rocks, where the extent of arsenopyrite dissemination normally tends to be equal to that of pyrite.

*Mineral assemblage*

The proximal alteration zone is characterised by the mineral assemblage dolomite/ ankerite+ quartz+ muscovite+ pyrite+ rutile ± arsenopyrite and albite (Fig. 16). In some ultramafic rocks, talc occurs instead of muscovite or fuchsite, and magnesite and breunnerite (a Mg-Fe carbonate) may occur in addition to dolomite or ankerite (Skwarnecki, 1988; Böhlke, 1989). In BIF, the characteristic assemblage is quartz+ pyrite ± dolomite/ ankerite, siderite, chlorite, magnetite and arsenopyrite. In addition to Fe-Mg-bearing carbonates, calcite may be present in the proximal mineral assemblage in sub-greenschist facies deposits (Twemlow, 1984) and in felsic host rocks.

*Response to HCl and carbonate staining*

The diagnostic features are, generally, a lack of reaction with cold dilute HCl and the formation of only a blue colour by carbonate staining (Fig. 7). In some felsic rocks, where calcite is present in the zone, there is an obvious reaction with the acid and production of both purple and blue colour by staining. As magnesite and siderite do not react with dilute HCl, nor do they produce any special colour by the staining technique referred to here (Hutchison, 1974), not all carbonates are stained blue or purple.

*Features to recognise in thin section*

Gradual changes in mineral assemblage within the proximal alteration zone include the complete replacement of possible relict chlorite and K-feldspar by sericite, dolomite/ankerite and quartz (Mikucki et al., 1990; Eilu & Mikucki, 1996; Billay et al. 1997). Some of the

*Figure 22 Dark-green intermediate and bleached proximal alteration within the Golden Mile Dolerite in the Golden Mile deposit. Red hematite "pigment" visible in the bleached zone. Hematite probably formed after the gold mineralisation; the oxidising retrograde fluids have used only the preexisting fractures and, hence, mostly affected the bleached alteration zones. Note also the brown skeletal leucoxene in unbleached and bleached zones, and the relatively abundant pyrite in the bleached zone.*

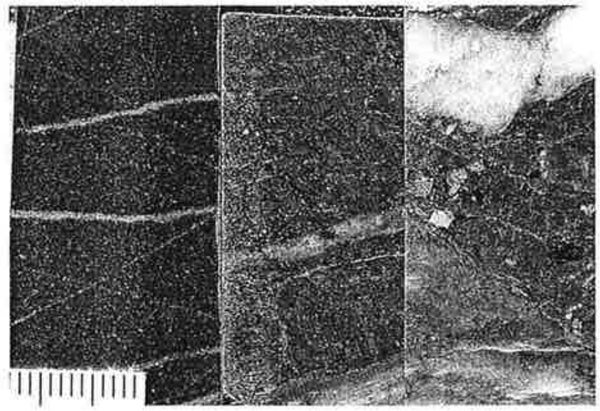


*Figure 23 Left to right: distal (calcite-chlorite), intermediate (dolomite-chlorite±sericite) and proximal (dolomite-sericite) alteration in fine-grained metasedimentary rock in the Twin Peaks deposit, Edjudina, Keith-Kilkenny shear zone. No bleaching in distal zone, partial bleaching in intermediate zone, and complete bleaching in proximal zone. Note the increase in volume of veins towards ore. Yellow surface is due to initial weathering of Fe-bearing dolomite in the intermediate and proximal samples. Arsenopyrite porphyroblasts are present in proximal sample.*

*Figure 24 Intermediate alteration in a mid-greenschist facies komatiite, Sunrise Dam mine. Mineral assemblage is talc-chlorite-ankerite. Veins comprise ankerite with minor amounts of quartz. The vertical lamina are saw marks on the sample surface. Distance between lines in the scale bar is 1 mm.*

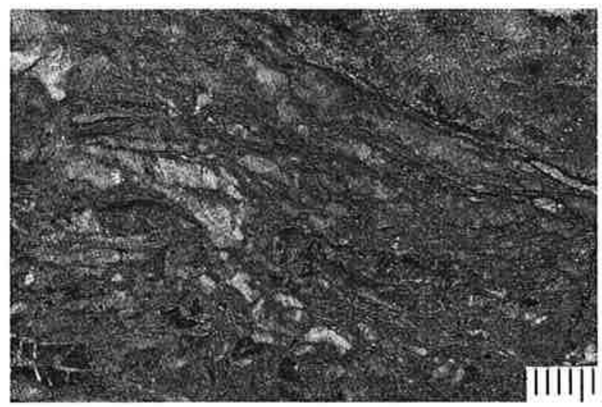


*Figure 25 Inner proximal alteration in ore (20 ppm Au), in a tholeiitic dolerite, Bulletin mine; same rock unit as in Figures 18 and 19. Intense shearing, multiple brecciation and complete bleaching. Mineral assemblage: dolomite + sericite + albite + quartz + rutile + pyrite + arsenopyrite. A relatively late quartz vein is evident on top right.*

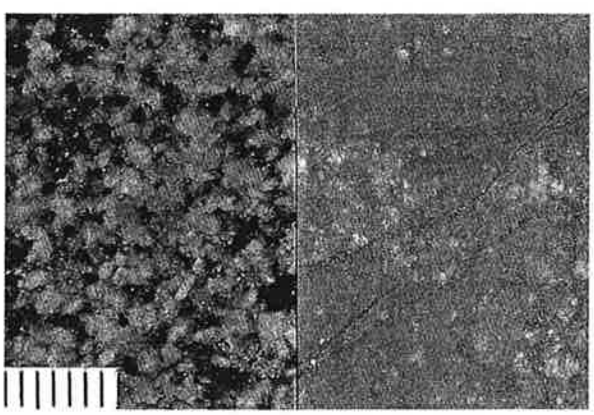


*Figure 26 Unaltered rock on left, proximal alteration (ore) on right, Granny Smith Granodiorite, Laverton. Note the total bleaching and abundant fracturing related to proximal alteration and mineralisation.*

*Figure 27 Proximal alteration in a mid-greenschist facies komatiite, Sunrise Dam mine. Intense green colour is due to abundant fuchsite (Cr-muscovite). This feature is characteristic for Cr-rich ultramafic rocks in proximal alteration zones of orogenic lode-gold deposits in sub- to mid-greenschist facies. Veins are filled by ankerite and quartz.*

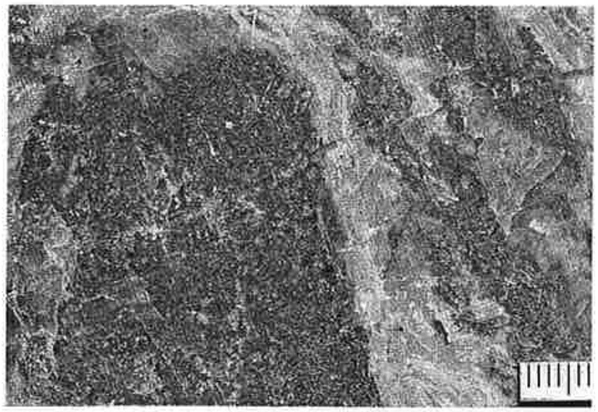


*Figure 28 Proximal alteration in sub-greenschist facies, brecciated basaltic komatiite, Phar Lap mine, Meekatharra. Abundant green fuchsite and disseminated, pale-brown rutile (leucoxene). Other minerals in the host rock are magnesite, dolomite, quartz and pyrite. The breccia veins are filled by ankerite, quartz and traces of pyrite. In addition, a post-brecciation, crosscutting quartz-calcite vein is evident across the centre of the figure.*

*Figure 29 Bleaching and a sulphidation front in interlayered BIF and tuffite in the Sunrise Dam mine. The alteration front marks the boundary between intermediate and proximal alteration zones. At the boundary, chlorite is replaced by ankerite-sericite, and magnetite is replaced by pyrite.*

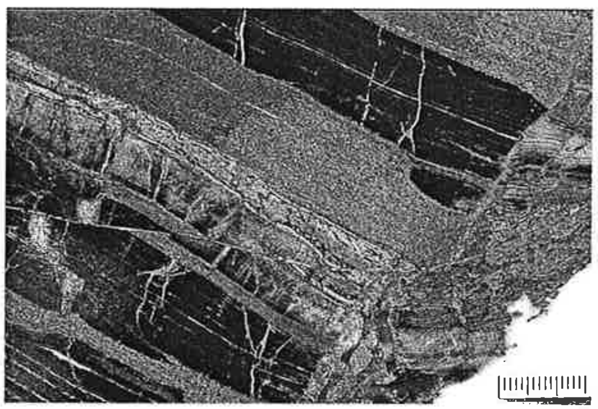


*Figure 30 Sulphidation front in the BIF-hosted gold deposit at Water Tank Hill mine, Mt Magnet. There has been selective replacement of magnetite by pyrite. Alteration is related to the discordant quartz veins visible in the figure.*

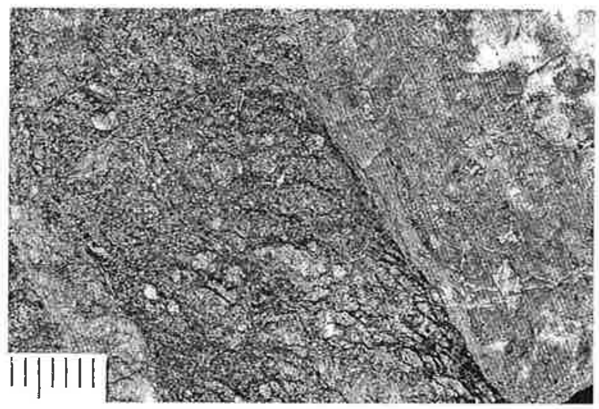


*Figure 31 Proximal alteration (extreme alteration intensity) in a roscoelite-bearing (green V-muscovite), breccia zone in the Golden Mile Dolerite, Golden Mile deposit.*

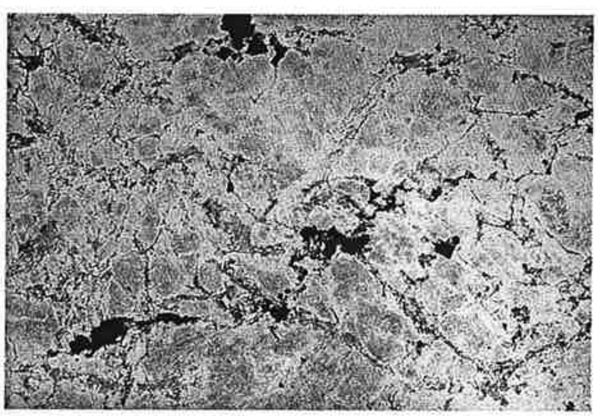


*Figure 32 Photomicrograph of proximal alteration and gold mineralisation in Granny Smith Granodiorite. Note the complete disappearance of mafic silicates and development of pervasive fracturing. Colourless quartz and albite, brownish-grey carbonate, and black rutile and pyrite. Plane polarised light. Field of view 6.7 mm.*

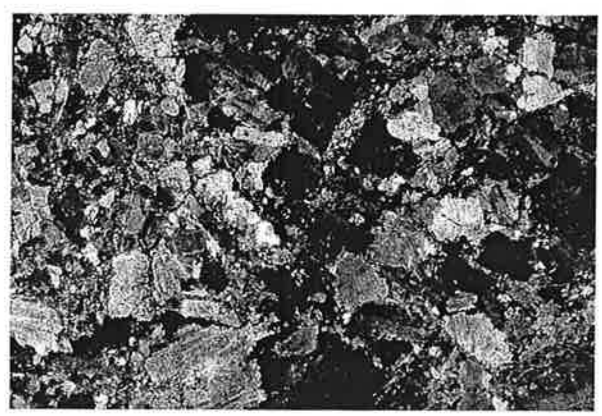


*Figure 33 Photomicrograph of proximal alteration and gold mineralisation in Granny Smith Granodiorite. Same sample as in Figure 32. Note the pervasiveness of brittle deformation. Crossed polarisers. Only carbonate grains show interference colours other than grey or white. Field of view 6.7 mm.*

dolomite may be replaced by siderite, and graphite may be replaced by carbonate(s). Most or all of the albite generated by regional metamorphism and distal alteration in the host rock is progressively replaced by sericite ± paragonite, but second-generation hydrothermal albite may precipitate in quartz-carbonate veins, and, in some cases, in the host rock in the most proximal areas. Pyrite replaces all of the pre-alteration pyrrhotite within the alteration zone. Tourmaline, if present, occurs mainly as a replacement of chlorite.

The intense deformation is, naturally, easily detected in thin section (Figs 32 & 33). In the domains of intense ductile deformation, there may be bands 1-20 mm long and <0.05 mm wide formed by numerous, tiny rutile (leucoxene) grains. They may appear as yellow lamellae to the naked eye. These indicate the approximate locations of the original Ti-bearing oxide grains, as one band represents one primary Fe-Ti oxide grain.

*Indications of zone boundaries*

The *distal boundary* of the alteration zone is described in Subsection 3.2.3 (proximal boundary of the intermediate zone), except for BIF-hosted mineralisation. In BIF, the alteration boundary is marked by a distinct sulphidation front (Figs 29 & 30).

# CHAPTER 4
# ALTERATION IN ROCKS METAMORPHOSED TO UPPER-GREENSCHIST TO LOWER-AMPHIBOLITE FACIES

## 4.1 GENERAL DESCRIPTION OF THE DEPOSITS

### 4.1.1 Style of mineralisation

The style of deformation in orogenic lode-gold deposits in upper-greenschist to lower-amphibolite facies rocks is brittle-ductile. As the ductile style of deformation dominates, the most common structural subtypes present are Types II, IV, VII and VIII, shown in Figure 3.

Common structural types of veining are:
(i) folded and boudinaged veins,
(ii) laminated veins, and
(iii) brittle quartz vein arrays or stockworks.

The vein arrays and stockworks occur in the most competent rock types within a mineralisation environment. Laminated and folded veins commonly contain intensely altered slivers of wallrock (Figs 34 & 35). Most of the sulphides and gold in the deposits occur adjacent to wallrock slivers in veins (Figs 34 & 35) and in vein margins, both in the vein and wallrock (Figs 36-39).

The ore mineralogy in veins and in wallrock is similar to that in lower metamorphic grade deposits with one significant exception. Pyrrhotite, instead of pyrite, is commonly the dominant Fe-sulphide (Colvine et al., 1988; Kontak et al., 1990; Nurmi & Sorjonen-Ward, 1993; McCuaig & Kerrich, 1994; McMillan, 1996). However, pyrite may still be common in the most proximal domains in upper-greenschist facies rocks, even the dominant sulphide in some deposits. At lower-amphibolite facies, pyrite is much less common than in upper-greenschist facies.

Examples of deposits at upper-greenschist to lower-amphibolite facies are listed in Appendix E.

### 4.1.2 General description of alteration

Alteration haloes at upper-greenschist and lower-amphibolite facies have similar shapes as in lower metamorphic-grade rocks (Fig. 9). The deposits differ from those formed at lower metamorphic grades in the features listed below (Clark et al., 1989; Mikucki & Ridley, 1993; Bullis et al., 1994; McCuaig & Kerrich, 1994; Eilu et al., 1996; Robert & Poulsen, 1997; Williams, 1997; Witt et al., 1997).

- Biotite, instead of sericite, is the characteristic potassic mineral in the proximal alteration zones, although muscovite may occur in the areas of most proximal alteration in upper-greenschist facies rocks.
- Pyrrhotite, instead of pyrite, is generally the dominant sulphide throughout the alteration sequence.
- Calcite is present, with other carbonates, in the areas of proximal alteration in greenschist-facies rocks and, generally, as the sole carbonate in amphibolite-facies rocks.
- Talc is more common in ultramafic rocks.
- K-feldspar may be stable throughout the alteration sequence.
- In BIF, grunerite occurs instead of, or with, chlorite.

In lower-amphibolite facies rocks, there are a few features different to those in upper-greenschist facies rocks (Mikucki & Ridley, 1993; Bullis et al., 1994; McCuaig & Kerrich, 1994; Eilu & Mikucki 1996, Robert and Poulsen 1997, Witt et al. 1997), as listed below.

- Generally, only two alteration zones, distal and proximal, can be distinguished.
- Alteration haloes are laterally less extensive. In most cases, alteration is confined, apparently, to the mineralised shear zone, although it may extend several hundreds of metres, even kilometres along a mineralised structure. There are, however, notable exceptions to the restricted extent of alteration in amphibolite-facies rocks. Both Durocher & van Haaften (1982) and Smith (1996) described cases where the extent of alteration is similar to that around deposits in lower- to mid-greenschist facies, with a lateral extent of 100-1000 m, and along-strike extent up to 9 km.
- Ca-bearing amphiboles and calcic plagioclase may occur throughout the alteration sequence.

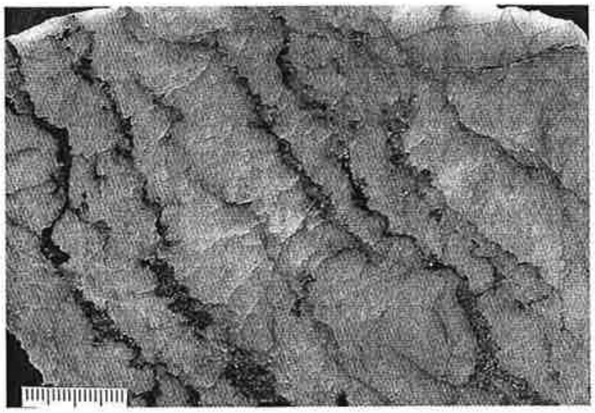


*Figure 34 Laminated quartz vein containing intensely sulphidised (pyrrhotite), stylolitic wallrock slivers in the Golden Crown mine, Cue. The host rocks are metamorphosed to upper-greenschist facies.*

*Figure 35 Intensely altered slivers of wallrock, green ultramafic rock and pale-brown mafic rock, in a quartz vein in the Bronzewing mine, Yandal greenstone belt. Note the difference in alteration due to the variation in the primary composition of the wallrock, despite the constant alteration intensity, and the concentration of sulphides (pyrite and pyrrhotite) in the vein-wallrock contacts. Host rocks were metamorphosed under upper-greenschist facies conditions.*

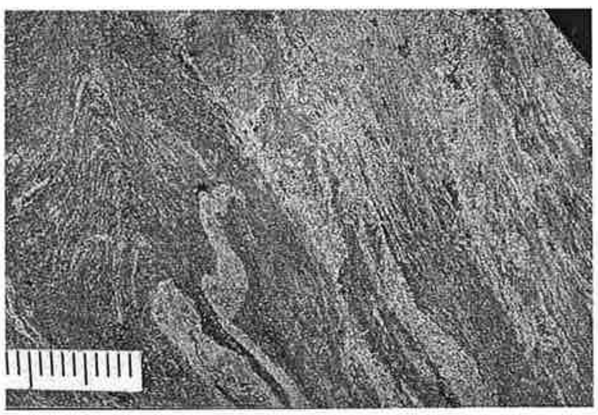


*Figure 36 Proximal alteration in tholeiitic basalt in the Bronzewing mine. Intense ductile deformation, biotitisation (brown) and local, retrograde, chloritisation (green) of biotite. Host rocks are metamorphosed under upper-greenschist facies conditions. Dry drill-core surface.*

*Figure 37 Same sample as in Figure 36, but the surface is wet and ground flat. Note how difficult it may be to distinguish between biotite and chlorite when the sample surface is not rough: chlorite is dark green and biotite dark grey to black. In contrast, it is much easier to distinguish pyrrhotite and pyrite from the background than in the Figure 36. Proximal alteration in tholeiitic basalt in the Bronzewing mine.*

*Figure 38 Proximal alteration and brittle deformation in tholeiitic basalt in the Hunt mine, Kambalda. Mineral assemblage in the host rock is biotite + albite + calcite + ankerite + quartz + pyrite + rutile. Biotite is more abundant in the darker areas. Host rocks are metamorphosed under upper-greenschist facies conditions.*

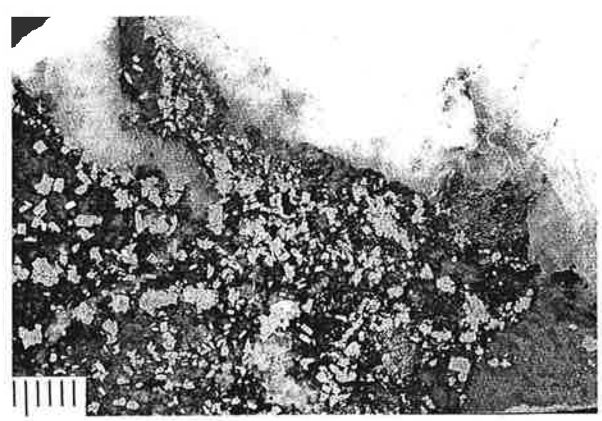
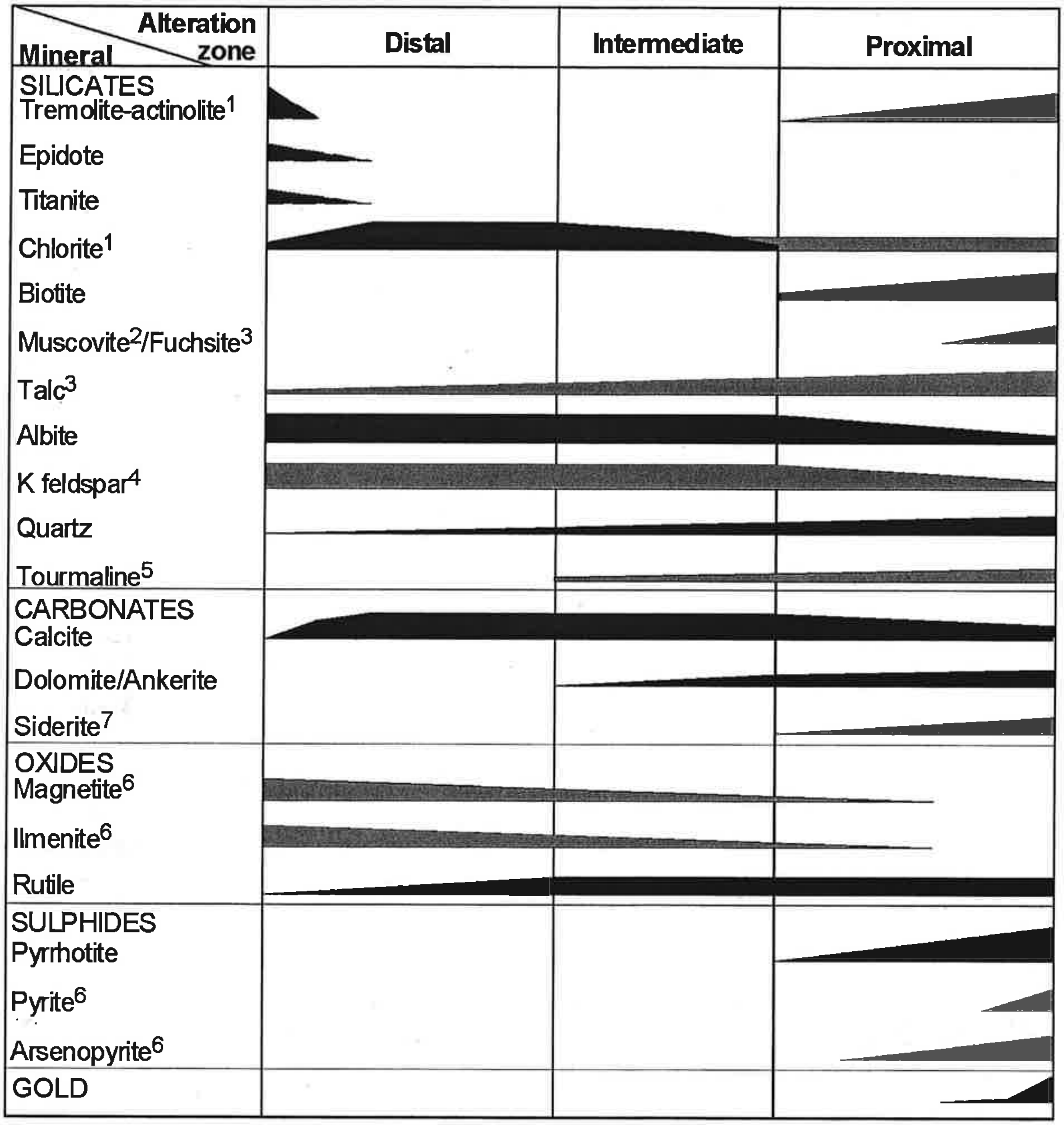


*Figure 39 Proximal alteration, with abundant arsenopyrite, in dolerite in the Harlequin mine, Norseman, where host rocks are at lower-amphibolite facies. The pale greenish colour is due to patchy retrograde sericitisation of plagioclase and biotite.*




1) Fe-rich amphibole and chlorite may be present in the proximal zone in BIF; chlorite also in other rock types.
2) Some sericite or fuchsite may be present in the most proximal domain of alteration.
3) Present only in ultramafic rocks.
4) Detected in the alteration sequence where also present in unaltered rock.
5) The extent of tourmaline formation is unclear, as tourmaline is not present in all alteration envelopes.
6) May or may not be present.
7) Siderite may be present in BIF.

*Figure 40 Schematic summary of the paragenetic alteration sequence around orogenic lode-gold deposits in upper-greenschist facies environments. Proximity to mineralisation increases to the right. Black represents the common case, and green indicates a less common occurrence. Zone widths shown here bear no resemblance to actual widths in the field. This sequence is based on data from Colvine et al. (1988), Böhlke (1989), Clark et al. (1989), Mikucki et al. (1990), Mueller & Groves (1991), Mikucki & Ridley (1993), Witt (1993), Bullis et al. (1994), Eilu & Mikucki (1996), Uemoto (1996), Robert (1996), Williams (1997) and Witt et al. (1997).*

- Calcic amphibole commonly forms selvages, 1-20 mm wide, to the mineralised quartz-calcite veins.
- Rutile gives way to ilmenite in distal, and to titanite in proximal, alteration zones as stable Ti-minerals; ilmenite and magnetite may be present throughout the alteration sequence.

The most diagnostic mineralogical feature of an alteration halo in upper-greenschist and lower-amphibolite facies rocks is the presence of biotite + pyrrhotite + calcite. This is reflected by a brown colour in altered rock. At upper-greenschist facies, the alteration sequence from the distal to proximal zone is: calcite-chlorite, calcite-dolomite/ankerite-chlorite and calcite-dolomite/ankerite-biotite zones. At lower-amphibolite facies, the sequence comprises a distal biotite and a proximal biotite-calcite zone. The alteration sequences are summarised in Figures 40 and 41, and the alteration zones are discussed in detail in Section 4.2.

Some studies have suggested that there is no distal alteration zone in lower-amphibolite facies BIF (Witt et al., 1994; McMillan, 1996). This is, most probably, an oversimplification due to difficulties in detecting anything other than the proximal sulphidised zone. For example, Smith (1996) described a sequence of four alteration zones around a BIF-hosted gold deposit at Mineral Hill (Montana, USA).

Vertical zoning may be present in some deposits, due to the change in proximal alteration from a mineral assemblage characterised by sericite and pyrite to one characterised by biotite and pyrrhotite (Robert & Brown, 1986; Clark et al., 1989; Mikucki et al., 1990; Robert, 1996; Uemoto, 1996). This change, which is gradual, probably reflects an increase in alteration temperature. Witt (1993) has pointed out that the evolution towards lower $CO_2$ content in the hydrothermal fluid could also result in biotite replacing sericite as the precipitating potassic mineral. This could explain the cases where biotite has partially replaced sericite, and no vertical zoning is obvious. However, if there also is a change from a pyrite- to a pyrrhotite-dominated alteration assemblage and a vertical zonation in a single rock type, a change in alteration temperature is the more likely alternative.

An example of subvertical zonation is demonstrated by the wallrocks of the BIF-hosted Lupin deposit in Yellowknife, Canada. The cordierite isograd, indicating a change from greenschist facies in the shallower areas to amphibolite facies in the deeper parts of the deposit, is roughly at a depth of 550 m below the present surface. The isograd is visible in the pelitic wallrocks, but there is no simultaneous change in the mineral assemblage in the BIF (Bullis et al., 1994). This difference is explained by the difference in composition between the rock units. Hence, the alteration assemblage in the BIF at Lupin, Fe-hornblende + quartz + pyrrhotite + arsenopyrite, most probably represents conditions covering a P-T range from upper-greenschist to lower-amphibolite facies.

Geochemical haloes also envelop deposits in upper-greenschist and lower-amphibolite facies settings, although there are less data published on the subject compared to deposits in lower-metamorphic grade settings. In upper-greenschist facies host rocks, geochemical anomalies probably have similar extents to lower- and mid-greenschist facies equivalents. For example, the alteration halo at Sons of Gwalia, in upper-greenschist facies host rocks, has a lateral extent of 500 m (Skwarnecki, 1990), equivalent to that at several deposits in lower metamorphic grade settings. At Bronzewing (a basalt-hosted deposit in the Yandal greenstone belt, Western Australia), $CO_2$, K, Rb, Sb, Te and W anomalies extend from 10 to >200 m beyond the Au anomaly, and K, Rb, Sb, Te and W anomalies extend, at least partially, beyond mappable alteration in the host rock (Eilu et al., 1996). There are indications that geochemical anomalies in lower-amphibolite facies settings may extend laterally further beyond the ore zone than the alteration halo, although most of the anomalies appear to be restricted to the shear zones hosting mineralisation (Mazzucchelli & James, 1980; Eilu & Mikucki, 1996).

## 4.2 ALTERATION IN DETAIL

### 4.2.1 Distal alteration zones

*Zone width*

The width of the distal alteration zones in upper-greenschist facies rocks is 1-80 m (Cassidy, 1992; Wang, 1993; Eilu, 1996b; Eilu et al., 1996; Uemoto, 1996) and in lower-amphibolite facies is 1-20 m (Bloem, 1994; Knight, 1994; Eilu & Mikucki, 1996).

*Colour and appearance at the meso- and microscopic scale*

Distal alteration in upper-greenschist facies rocks is similar to that in lower metamorphic grades. Excluding the easily detected leucoxene in mafic and intermediate rocks (similar to that shown in Figs 18, 20 & 21), and a minor increase in the volume of quartz-calcite veins, there is generally very little difference compared to unaltered rock. Due to decreasing grain size and replacement of primary hornblende by fine-grained chlorite or even biotite, there may be an obvious change in colour in intermediate and felsic plutonic rocks (Fig. 42).

In amphibolite facies rocks, the distal zones are clearly indicated by the appearance of a brown colour due to the presence of biotite (Figs 43 & 44). However, this may be difficult to detect in rocks with regional metamorphic or primary magmatic biotite; for example, in granitoids and sedimentary rocks. At both upper-greenschist and lower-amphibolite facies, the distal alteration zone displays more intense deformation than the unaltered rock (Fig. 42).

*Mineral assemblage*

In upper-greenschist facies rocks, the characteristic mineral assemblage is calcite + chlorite + albite + quartz + rutile (Fig. 40) and, in lower-amphibolite facies rocks, biotite + hornblende + plagioclase + quartz + ilmenite (Fig. 41). In the latter, regional metamorphic minerals, such as amphibole, talc, chlorite and magnetite, are partially preserved. Hence, the typical mineral assemblage in ultramafic rocks is biotite + tremolite ± talc, chlorite, anthophyllite, magnetite, ilmenite. In BIF, the distal alteration, if detected, may be characterised by an increase in the volume of ferro-hornblende and decrease in grunerite towards mineralisation (Smith, 1996).

*Response to HCl and carbonate staining*

A strong reaction with cold dilute HCl and the formation of a purple colour by carbonate staining (Hutchison, 1974), without indications of other carbonates, are diagnostic for the distal alteration zone in upper-greenschist facies. In lower-amphibolite facies rocks, there is no reaction with the acid nor with the chemicals used for staining carbonates, except in some ultramafic rocks where small volumes of calcite may be present.

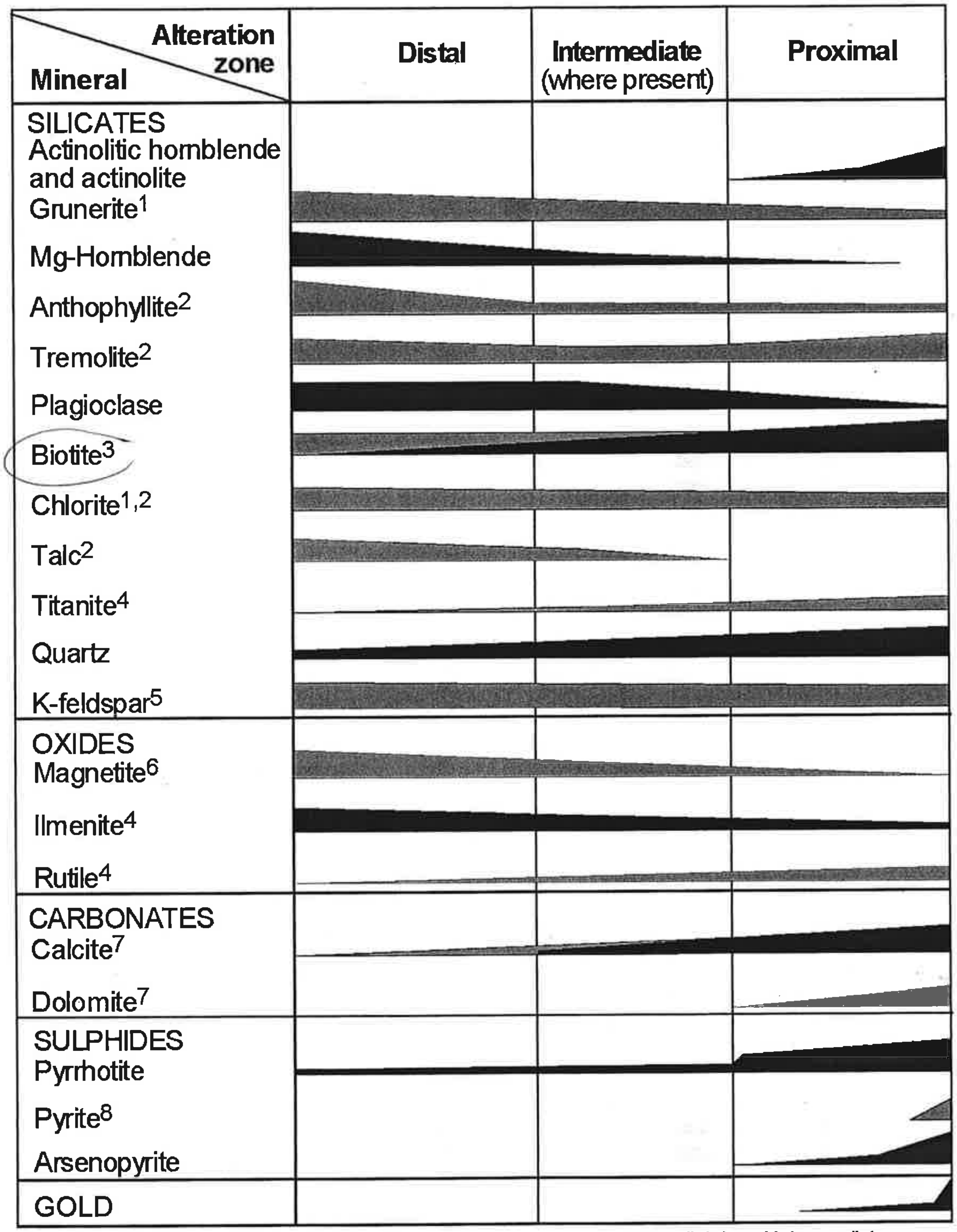
*Features to recognise in thin section*

A common feature detected in thin section is a gradual increase in the degree of deformation towards ore. In upper-greenschist facies rocks, the characteristic replacement reactions are the same as in distal zones in lower metamorphic grades: primary magmatic and regional metamorphic Fe-Mg silicates are replaced by chlorite + calcite, and titaniferous magnetite, ilmenite and titanite are replaced by rutile. In lower-amphibolite facies rocks, biotite gradually replaces the amphiboles of unaltered rock, whereas there is no significant change in Ti-bearing minerals nor an appearance of any carbonate species.

*Indications of zone boundaries*

The *distal boundary* in upper-greenschist facies rocks is easily detected by the change from rock with nil reaction with dilute HCl, to rock with a strong reaction. The boundary can also be checked by carbonate staining (change from no staining to purple colouration; Fig. 7). In lower-amphibolite facies rocks, the boundary is clearly marked by the appearance of brown biotite. The boundary may also be indicated by the change from massive to foliated rock, but a change in the degree of brittle deformation may also be present, especially in felsic rocks (Fig. 42).

The *proximal boundary* is marked by the appearance of abundant pyrrhotite ± arsenopyrite in BIF (sulphidation front). In other rock types, the diagnostic features are the appearance of dolomite or ankerite in upper-greenschist facies rocks, and appearance of calcite in lower-amphibolite facies rocks. The presence of Fe-bearing carbonates can be detected by staining and, if the rock surface has been exposed to humid air for a few weeks or months, by the yellow surface layer due to initial weathering. In amphibolite-facies rocks, the boundary is easily detected by using dilute HCl and, if no intermediate alteration zone is present, there is also a distinct increase in the degree of deformation, veining and sulphidation at the zone boundary.




1) In BIF; grunerite may be present throughout the sequence or only in distal and intermediate zones.
2) In ultramafic rocks; antophyllite may also be present in proximal zones in BIF
3) Biotite throughout the sequence where biotite present in unaltered rock.
4) Ilmenite present with or without titanite and/or rutile.
5) Where present.
6) May or may not be present.
7) In ultramafic rocks, calcite may be present throughout the sequence and, if quartz is not present, also dolomite may occur in the sequence.
8) Pyrite is present in a few deposits.

*Figure 41 Schematic summary of the paragenetic alteration sequence around orogenic lode-gold deposits in lower-amphibolite facies environments. Proximity to mineralisation increases to the right. Black represents the common case, and green indicates a less common occurrence. Zone widths shown bear no resemblance to actual widths in the field. This sequence is based on data from Colvine et al. (1988), Ridley & Barnicoat (1990), Mueller & Groves (1991), McCall (1992), McCuaig et al. (1993), Mikucki & Ridley (1993), Witt (1993), Bloem (1994), Knight (1994), Witt et al. (1994), Crookes & Schaus (1995), Matthäi et al. (1995), McMillan (1996), Eilu & Mikucki (1996), Smith (1996), Robert & Poulsen (1997), and Witt et al. (1997).*

*Figure 42 Tonalite host rock in the Lawlers deposit, Leinster. Unaltered rock on left and distal alteration on right. Alteration has taken place in host rocks that are transitional between greenschist and amphibolite facies: magmatic amphibole has been replaced by biotite + actinolite + calcite.*

*Figure 43 Proximal alteration around a quartz-calcite-pyrrhotite vein in a Mg-tholeiitic basalt in the Kings Cross mine, Coolgardie. Mineral assemblage in the host rock is actinolitic hornblende + biotite + quartz + pyrrhotite. The most biotite-rich bands are nearly black. Green actinolite forms the selvage and margin of the vein. Host rocks are metamorphosed under lower-amphibolite facies conditions.*

*Figure 44 A narrow envelope of distal and proximal alteration around a quartz-calcite vein in a Mg-tholeiitic basalt in the Kings Cross mine. Regional metamorphic hornblende is dark green, biotite is brown to black, plagioclase and calcite are white, and pyrrhotite is yellow. There is a band of dark-green retrograde chlorite along the vein.*

*Figure 45 Distal or intermediate alteration in komatiite in the Agnew gold mine, Leinster. Pale-grey dolomite-quartz and dark-green talc-chlorite bands. Metamorphic conditions in the host rocks are transitional between greenschist and amphibolite facies.*

*Figure 46 Intense deformation and inner proximal alteration in a high-Mg basalt that has been metamorphosed under lower-amphibolite facies conditions. Central portion of the main shear zone in the Sons of Gwalia mine, Leonora. Abundant, sheared quartz veins. Mineral assemblage in the host rock is muscovite + quartz + dolomite/ankerite + albite + pyrite + rutile ± biotite.*

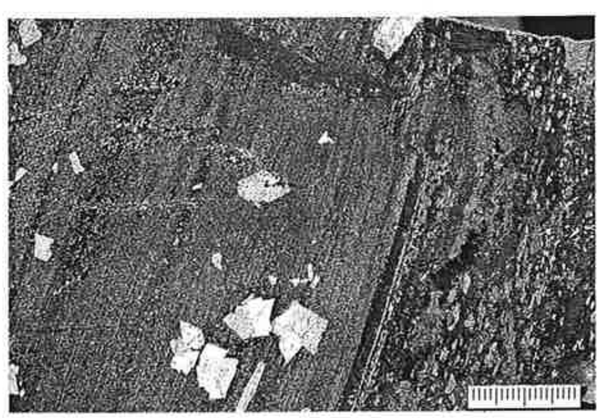


*Figure 47 Proximal alteration in BIF in the Santa Claus pit, Randalls deposit, 100 km ESE of Kalgoorlie. Note the large arsenopyrite porphyroblasts, pyrrhotite replacing magnetite along primary layering, yellow grunerite, and dark-green actinolite and chlorite. Host rocks at upper-greenschist facies.*

### 4.2.2 Intermediate alteration zones

*Zone width*

The width of the intermediate alteration zones in upper-greenschist rocks is 0.1-20 m, and in lower-amphibolite facies rocks is 0-20 m (Wang, 1993; Bloem, 1994; Knight, 1994; Eilu & Mikucki, 1996; Smith, 1996).

*Colour and appearance at the meso- and microscopic scale*

The degree of deformation, veining and, in some cases, sulphidation is higher in the intermediate zone than in the distal zone. In amphibolite-facies rocks, if the zone is present at all, it also contains more biotite and less amphiboles and, hence, has a more intense brown colour than the distal zone. Upper-greenschist facies rocks have a yellowish colouring if the rock surface has been exposed to humid air for a few weeks or months.

*Mineral assemblage*

The intermediate alteration zone is characterised by the mineral assemblage calcite + dolomite/ankerite + chlorite + quartz + rutile in upper-greenschist facies rocks (Fig. 40), and by biotite + calcite + hornblende + plagioclase + quartz + ilmenite ± titanite, pyrrhotite in lower-amphibolite facies rocks (Fig. 41). In ultramafic rocks at amphibolite facies, the assemblage may be biotite + tremolite ± calcite/dolomite, talc, chlorite, quartz, ilmenite, pyrrhotite (Fig. 44), and, in BIF, ferro-hornblende + quartz ± grunerite, pyrrhotite, chlorite.

*Response to HCl and carbonate staining*

A strong reaction with cold dilute HCl and the formation of purple (calcite) and blue (dolomite/ankerite) colours by carbonate staining (Hutchison, 1974) are typical of the intermediate alteration zone in upper-greenschist facies rocks. In lower-amphibolite facies rocks, there is also a strong reaction with cold dilute HCl, but staining indicates only calcite, except in some ultramafic rocks.

*Features to recognise in thin section*

Changes that take place within the zone in upper-greenschist facies rocks include gradual replacement of chlorite by dolomite/ankerite + quartz (+ talc in ultramafic rocks), and the complete replacement of magnetite and ilmenite by rutile (BIF excluded). In lower-amphibolite facies rocks, the most obvious replacement is that of hornblende by biotite; in addition, calcite + quartz partially replaces amphiboles and plagioclase. An increase in the degree of deformation is visible nearly everywhere in thin section at both metamorphic grades considered.

*Indications of zone boundaries*

The *distal boundary* of the zone is described in Subsection 4.2.1 (proximal boundary of the distal zone).

The *proximal boundary* of the intermediate alteration zone is marked by the start of significant sulphidation (pyrrhotite ± arsenopyrite, pyrite) and intense deformation in all rocks. In deposits in upper-greenschist facies rocks, there also is a striking change in colour: with biotitisation of chlorite, the rock becomes brown (Figs 36 & 38). In amphibolite-facies ultramafic rocks, talc and anthophyllite, if present in the rocks, disappear and, if quartz is not present in an ultramafic rock, dolomite may become part of the mineral assemblage at the zone boundary. In amphibolite-facies BIF, grunerite may disappear at the boundary.

### 4.2.3 Proximal alteration zones

*Zone width*

The width of the proximal alteration zones in upper-greenschist facies rocks varies from 2 mm to 50 m (Cassidy, 1992; Eilu et al., 1996; Wang, 1993; Uemoto, 1996), and at lower-amphibolite facies from 1 mm to 5 m (Bloem, 1994; Knight, 1994).

*Colour and appearance at the meso- and microscopic scale*

The intensity of deformation is commonly very high, as exemplified by Figures 36, 45 and 46, and, with a few exceptions, rocks are characterised by a brown colour (Figs 36 & 38). In the most proximal areas in upper-greenschist facies deposits, there may be sericitisation and bleaching (Figs 35 & 45), or the grain size of biotite may be very small, with the rock appearing black, not brown, to the naked eye. In BIF, Fe-amphibole may dominate and the rock may remain dark green (Fig. 47). Also, the degree of sulphidation may be high with coarse-grained sulphides (Figs 38, 39 & 47).

In amphibolite-facies deposits, quartz veins may have 1-20 mm wide selvages of tremolite-actinolite or actinolitic hornblende (Figs 43 & 49). This selvage may be monomineralic, or may contain one or two other minerals common in the proximal zone; normally calcite, biotite, K feldspar, pyrrhotite or arsenopyrite. In BIF, the selvage may be formed by grunerite ± anthophyllite (Bloem, 1994; Williams, 1997).

*Ore minerals*

Pyrrhotite and arsenopyrite are the most common ore minerals (Figs 38, 39, 47 & 49). Instead of, or in addition to, arsenopyrite, gersdorffite may occur in rocks with a high Ni content, chiefly in ultramafic rocks (Eilu, 1996b). The style and degree of sulphidation, and the other ore minerals which may be present, are essentially the same as in lower metamorphic grades, except that pyrite is generally absent, although it may abound in the most proximal domain of alteration (Fig. 38).

*Mineral assemblage*

The diagnostic assemblage in the proximal alteration zone is biotite-calcite-quartz-pyrrhotite (Figs 40 & 41). In upper-greenschist facies rocks, these minerals are accompanied by dolomite/ankerite, albite and rutile ± muscovite/fuchsite, chlorite, arsenopyrite and pyrite, and in lower-amphibolite facies rocks by hornblende/actinolite ± ilmenite, arsenopyrite, pyrite, titanite, magnetite. In addition, K-feldspar may occur in felsic rocks, talc and chlorite in ultramafic rocks, and grunerite and siderite in BIF in both upper-greenschist and lower-amphibolite facies settings.

*Response to HCl and carbonate staining*

One of the most typical features of the proximal alteration zones in upper-greenschist and lower-amphibolite facies settings is the strong reaction with cold dilute HCl and the formation of a purple colour during carbonate staining. In upper-greenschist facies rocks, and in some ultramafic rocks at lower-amphibolite facies, staining also produces a blue colour due to the presence of ankerite or ferroan dolomite.

*Features to recognise in thin section*

Gradual changes in mineral assemblages within the proximal alteration zone include the complete or near-complete replacement of chlorite and amphiboles, and partial replacement of albite or calcic plagioclase, by biotite + calcite + quartz, partial replacement of ilmenite and magnetite by pyrrhotite (+ titanite or rutile), increase in the degree of deformation and veining, and increase in the volume of sulphides. In greenschist-facies rocks in the most proximal domains of alteration, muscovite may partially or completely replace biotite and pyrite may be the dominant sulphide. The green amphibolite selvages around quartz-carbonate veins in lower-amphibolite facies rocks are easy to detect in thin section.

*Indications of zone boundaries*

The *distal boundary* of the alteration zone is described in Subsections 4.2.1 and 4.2.2 (proximal boundaries of the distal and intermediate zone) depending on the presence of the intermediate zone.

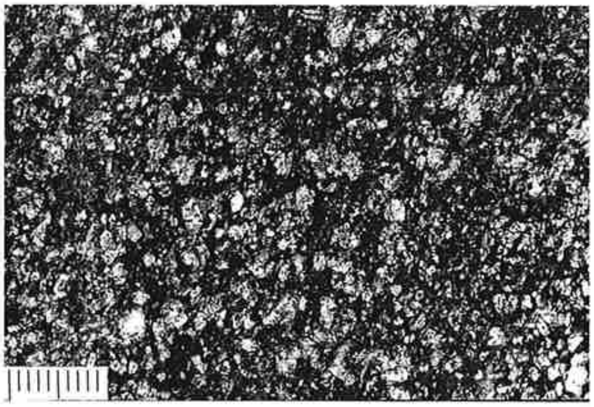


*Figure 48 Proximal alteration in the Lawlers tonalite. Due to intense deformation (i.e. presence of very fine-grained biotite and remnants of larger plagioclase grains) the rock appears black and white. Mineral assemblage: biotite + plagioclase + quartz + calcite + titanite + ilmenite + magnetite + pyrite.*

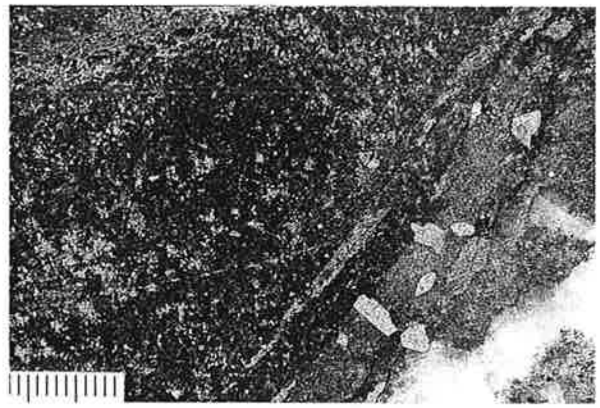


*Figure 49 Green actinolite selvage on a quartz vein, and dark brown proximal alteration (biotitisation), in dolerite in the Three Mile Hill mine, Coolgardie. Note arsenopyrite porphyroblasts within the selvage, and pyrrhotite in both the wallrock and selvage. Host rocks at lower-amphibolite facies.*

# CHAPTER 5
# ALTERATION IN ROCKS METAMORPHOSED TO MID-AMPHIBOLITE TO GRANULITE FACIES

## 5.1 GENERAL DESCRIPTION OF THE DEPOSITS

### 5.1.1 Style of mineralisation

The style of deformation in orogenic lode-gold deposits in mid-amphibolite to lower-granulite facies rocks is dominantly ductile. The most common structures hosting gold mineralisation are broad, ductile shear zones. Within the shear zones, deposits form lenticular bodies of altered, veined and mineralised rock, in most cases elongate parallel to the shear zone, foliation and lithological trend. Multiple, sheeted or laminated, foliation-parallel veins 1-20 cm wide are characteristic (Fig. 51). The most common structural subtypes present are Types II, VIII and IX, shown in Figure 3. Some deposits are characterised by brittle veins or vein stockworks. These occur in massive rock units, especially in gabbros or basalts, where the veins are typically a few centimetres to 1 m thick.

Ore minerals occur in veins and in the adjacent wallrock (Figs 52 & 53). Pyrrhotite ± arsenopyrite and loellingite are a major component in some veins, forming clots up to a few centimetres in diameter, but most veins have only minor sulphides. The ore mineralogy is similar to that of deposits in lower-amphibolite facies settings: pyrrhotite is the dominant Fe-sulphide and arsenopyrite is common, and these may be accompanied with a variable combination of trace amounts of other sulphides and As-, Bi-, Sb- and Te-minerals. The major differences compared to deposits in lower metamorphic-grade settings are that loellingite commonly occurs with arsenopyrite, and pyrite is generally absent (Potgieter & de Villiers, 1986; Fare, 1989; Barnicoat et al., 1991; Mueller & Groves, 1991; Cassidy, 1992; Lapointe & Chown, 1993; McCuaig et al., 1993; Neumayr et al., 1993; Bloem, 1994; Knight, 1994; Kontoniemi, 1998).

Most of the veins are quartz-dominated and, in addition to the ore minerals, may contain variable amounts of diopside, Ca-amphibole, garnet, plagioclase, biotite, scheelite, apatite or tourmaline. In some veins, diopside or Ca-amphibole may be of equal importance to quartz, especially in ultramafic host rocks (e.g. Figs 52, 54 & 55).

Examples of deposits in mid-amphibolite to lower-granulite facies settings are listed in Appendix F.

### 5.1.2 General description of alteration

Alteration enveloping orogenic lode-gold deposits in high metamorphic-grade rocks is characterised by a set of distinct features (Fare, 1989; Ridley & Barnicoat, 1990; Barnicoat et al., 1991; Mueller & Groves, 1991; Cassidy, 1992; Bloem, 1994; McCuaig & Kerrich, 1994; Mernagh & Witt, 1994; Witt et al., 1994).

- Alteration haloes are narrow, rarely extending beyond mineralised shear zones and, laterally, more than a few tens of metres from gold mineralisation. However, domains up to 100 m wide of altered rock exist in wide shear zones, and in areas where a number of subparallel mineralised structures occur in close proximity, so that their alteration haloes overlap.
- In most cases, only two alteration zones, a distal, brownish, biotite zone and a proximal, distinctly banded, diopside zone, can be detected. For many deposits, only the proximal zone has been described: the detection of the distal zone may be difficult, as the intensity of biotitisation may be low and biotite may also belong to the primary mineral assemblage in several host-rock types.
- The extent of the distal alteration zone enveloping a single mineralised zone is up to 10 m and that of proximal alteration up to 5 m.
- The most diagnostic feature is the repetitions of from one to five, narrow, mono- and bimineralic calc-silicate bands and Ca-silicate selvages along quartz veins in the domain of proximal alteration (Fig. 50, 51, 55, 56 & 57).
- Commonly, the individual bi- and monomineralic bands are only 1-2 cm wide, resulting in proximal alteration ≤20 cm in width around a single vein.
- Diopside is the most characteristic mineral produced by alteration.
- The degree of carbonation and sulphidation is low. Generally, only calcite is present and

the $CO_2$ content of even intensely altered rock is <5%, whereas it is 5-20% in lower metamorphic grade rocks (e.g. Kishida & Kerrich, 1987).

- In lower-granulite facies rocks, orthopyroxene, diopside, garnet and K-feldspar characterise alteration: orthopyroxene is absent at lower metamorphic grades.

When investigating a sequence of mafic (and ultramafic) rocks, prograde metamorphism of spilites under mid-amphibolite to lower-granulite facies conditions may result in misleading mineral assemblages. Mafic volcanic and sub-volcanic rocks, especially interpillow matrices and fracture fills, tend to alter to the mineral assemblage quartz + albite + actinolite/chlorite + epidote + calcite + pyrite in a synvolcanic submarine hydrothermal system (Reed, 1983). Under high-grade regional metamorphism, this assemblage may recrystallise to form diopside ± grossular garnet, calcite, hornblende, quartz, plagioclase, pyrite, pyrrhotite assemblages that superficially resemble proximal alteration assemblages related to lode-gold mineralisation. The diopside-bearing assemblages can, in most cases, be identified correctly by checking their context with respect to deformation and primary volcanic features of the rock sequence under investigation.

Retrograde minerals are common in the alteration haloes, just as they are common in all high metamorphic-grade terrains. Partial replacement of pyroxene, amphibole and feldspar by sericite, chlorite, prehnite and epidote, and the presence of low-temperature sulphides, tellurides, selenides and Bi- and Sb-minerals, is commonly detected in high-metamorphic grade deposits. However, several features argue against major retrograde gold mineralisation:

(i) there is no deposit-scale correlation between gold grade and intensity of retrograde alteration or retrograde structures;
(ii) retrogression is patchy;
(iii) only a minority of gold grains, if any, are hosted along the late fractures; and
(iv) gold also occurs as inclusions in, and in equilibrium with, hydrothermal amphibole, diopside and composite arsenopyrite-loellingite grains.

Geochemical studies show significant enrichment of $CO_2$, S, K, Au and several other trace elements, and depletion of Na and Sr within gold mineralisation and alteration haloes (Fare, 1989; Cassidy, 1992; Neumayr et al., 1993; Witt, 1993; Bloem, 1994; Knight, 1994). The relative chemical changes are essentially similar to those in lower metamorphic grades, except that no enrichment of Na has been detected. The lateral extent of geochemical anomalies seems to be restricted to a few tens of metres from a gold deposit, but, because very little has been published on the extent of the haloes, nothing more definite can be stated. It may well be that certain trace elements (e.g. As, Sb, W, Te) can define anomalies extending beyond mappable alteration in high metamorphic-grade rocks, just as is the case in greenschist-facies rocks (Eilu & Mikucki, 1996).

## 5.2 ALTERATION IN DETAIL

### 5.2.1 Distal alteration zones

*Zone width*

The width of the distal alteration zones at mid-amphibolite to lower-granulite facies is 2-30 m (Fare, 1989; Ridley & Barnicoat, 1990; Cassidy, 1992; McCuaig et al., 1993; Bloem, 1994; Witt et al., 1994).

*Colour and appearance at the meso- and microscopic scale*

Distal alteration zones are, in most cases, characterised by the presence of biotite, a brown colour, and a moderate degree, at least, of plastic deformation (Fig. 58). There is not necessarily much difference compared to unaltered rocks, which also may have a significant biotite content and display ductile deformation. Hence, it may be difficult to detect the distal zone at all, especially in felsic (Fig. 57) and pelitic rocks, and in BIF. Within the zone, there is commonly an increase in biotite content, with biotite replacing regional metamorphic or magmatic amphibole, an increase in the degree of deformation, and a minor increase in pyrrhotite content.

*Mineral assemblage*

The most common mineral assemblage in mid- to high-amphibolite facies rocks is biotite + plagioclase + hornblende + ilmenite with minor pyrrhotite and quartz (Fig. 50). In felsic rocks, there may also be K-feldspar. In BIF, the assemblage may be a combination of the regional metamorphic grunerite, ferro-actinolite, magnetite and quartz, with small amounts of hydrothermal grunerite, pyrrhotite and Fe-rich almandine garnet. In ultramafic rocks, regional metamorphic talc, tremolite, anthophyllite, cummingtonite, magnetite, olivine and/ or ilmenite may occur with hydrothermal phlogopite, chlorite (can also be retrograde) and pyrrhotite.

Lower-granulite facies mineral assemblages are characterised by the presence of orthopyroxene (hypersthene) and K-feldspar, with biotite, quartz and plagioclase.

*Response to HCl and carbonate staining*

Generally, there are no carbonates in the distal zone and, hence, no response to dilute HCl or carbonate staining chemicals.

*Features to recognise in thin section*

In thin section, the gradual and partial replacement of regional metamorphic amphiboles by biotite + quartz and increase in the degree of deformation are the most common detectable features. Any gradual increase in the quartz and pyrrhotite contents, and decrease in amphibole, olivine and talc contents, can usually only be detected in thin section.

*Indications of zone boundaries*

The *distal boundary* is marked by the appearance of biotite ± minor pyrrhotite and a brown colour, and by an increase in the degree of deformation. The change from unaltered to altered rock tends to be gradual and, as discussed above, difficult to detect in many rock types, especially BIF and felsic and pelitic rocks.

The *proximal boundary* is clearly marked by the appearance of green diopside, mono- and bi-mineral banding, and a significant increase in the degree of deformation and veining. There may also be an increase in pyrrhotite content and appearance of arsenopyrite ± loellingite. In ultramafic rocks, cummingtonite and olivine disappear from the mineral assemblage (Fig. 50). Hydrothermal calcite, K-feldspar and titanite may appear at the alteration zone boundary. If calcite becomes part of the mineral assemblage, the boundary is easily detected by an increase in reaction to dilute HCl. In BIF, the boundary is clearly marked by a sulphidation front (Fig. 59).

Where there is also an intermediate alteration zone, mono- and bi-mineral banding is not detected, and the changes in mineral assemblage are more subtle at the alteration zone boundary than where there is an abrupt change from a distal to a proximal zone (Fig. 50).

### 5.2.2 Intermediate alteration zones

Intermediate alteration zones are not commonly described from around orogenic lode-gold deposits in mid-amphibolite to lower-granulite facies rocks. Hence, the following description cannot be very detailed nor comprehensive.

*Zone width*

The width of the intermediate alteration zones in mid-amphibolite to lower-granulite facies rocks is from being absent to, possibly, 5 m (Fare, 1989; Mueller, 1990; Cassidy, 1992; McCuaig et al., 1993; Bloem, 1994; Witt et al., 1994).

*Colour and appearance at the meso- and microscopic scale*

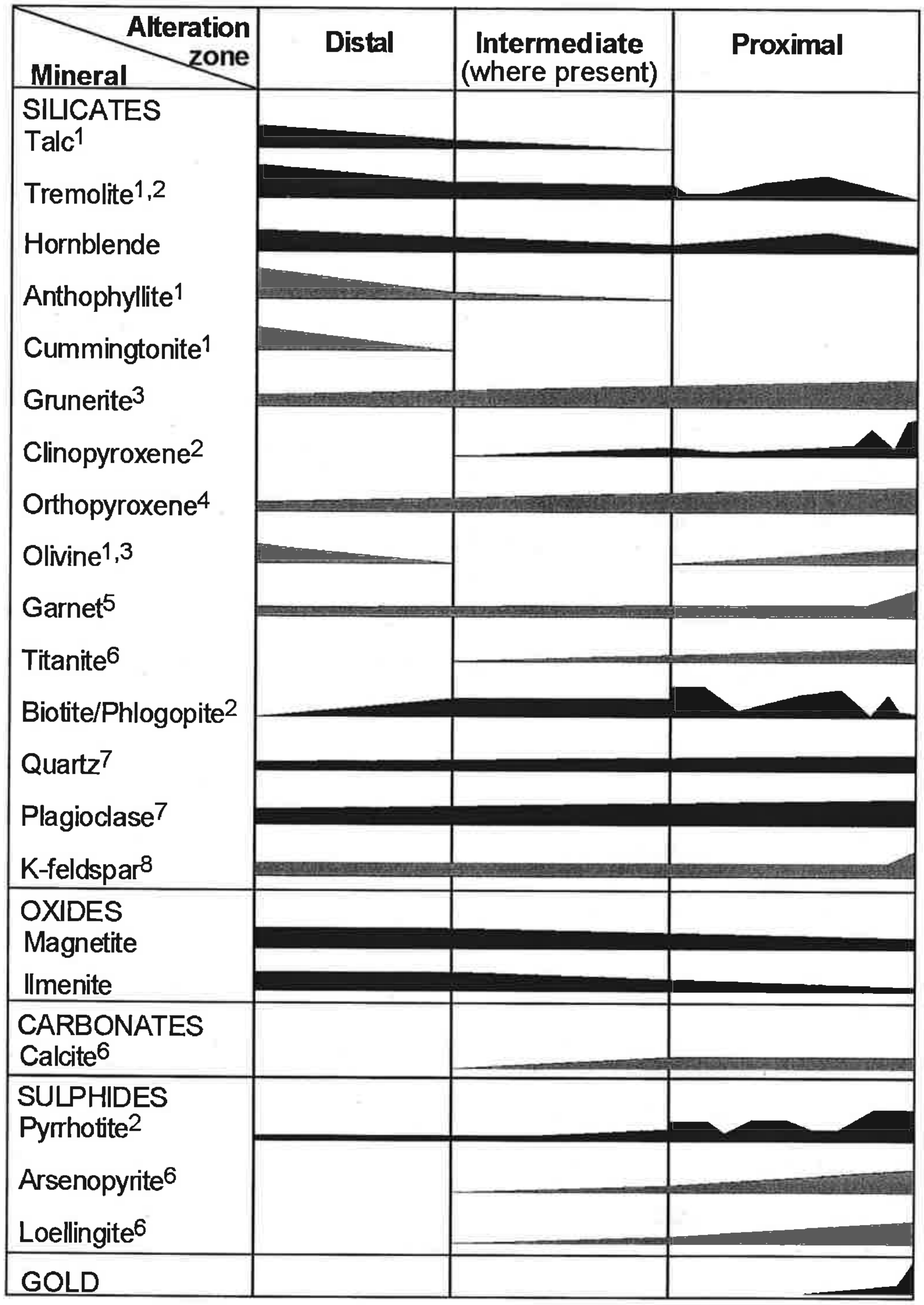
Most descriptions of intermediate alteration zones are from ultramafic rocks, where the zone is characterised by a brown colour due to biotite, by minor pyrrhotite ± arsenopyrite and loellingite dissemination, and variable amounts of calcite, diopside, tremolite, titanite, anthophyllite or talc. The intermediate alteration zone is more deformed and veined than the distal zone, but the differences tend to be small.

*Mineral assemblage*

A typical mineral assemblage in intermediate alteration zones in ultramafic rocks is talc + biotite + pyrrhotite with one or more of diopside, tremolite, hornblende, calcite, quartz and/ or ilmenite. In mafic rocks, the assemblage is plagioclase + hornblende + biotite + pyrrhotite + quartz ± calcite, ilmenite and titanite. In granulite-facies settings, hypersthene occurs in addition to the minerals mentioned above.

*Response to HCl and carbonate staining*

If calcite is present, there is a strong reaction with cool dilute HCl, and carbonate staining produces a purple colour on the surface of carbonate grains.




1) Only in ultramafic rocks, if present at all. Olivine in the distal zone is a regional metamorphic mineral.
2) May be part of thin, mono- and bi-mineralic subzones within the domain of proximal alteration. Note that the variation shown here in their volume in the proximal zone is only schematic, not exact.
3) Only in BIF; olivine (fayalite, if present) only in the proximal zone.
4) Only in the highest-temperature rocks.
5) In distal and intermediate zones, only regional metamorphic garnet is present.
6) Common, but not present in all cases.
7) Generally not present in ultramafic rocks.
8) Present throughout the sequence if in unaltered rock; otherwise, only in the proximal zone.

*Figure 50 Schematic summary of the paragenetic alteration sequence around orogenic lode-gold deposits in mid-amphibolite to lower-granulite facies environments. Proximity to mineralisation increases to the right. Black represents the common case, and green indicates a less common occurrence. Zone widths shown bear no resemblance to actual widths in the field. This sequence is based on data from Colvine et al. (1988), Fare (1989), Mueller (1990), Ridley & Barnicoat (1990), Mueller & Groves (1991), Lapointe & Chown (1993), Bloem (1994), van Reenen et al. (1994), Hay (1995), Neumayr et al. (1993) and Witt et al. (1997).*

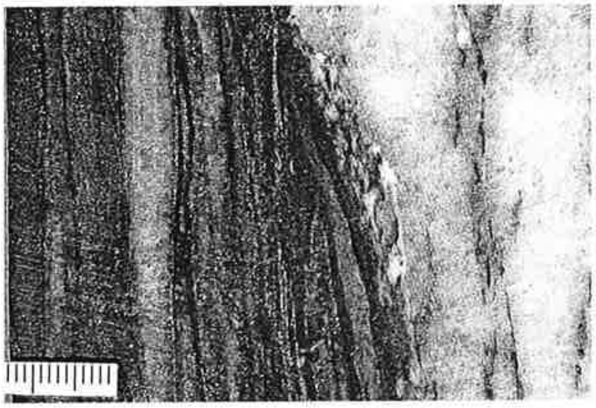


*Figure 51 Proximal alteration in komatiite from the Transvaal deposit, Southern Cross greenstone belt; metamorphosed to mid-amphibolite facies. Narrow banding is very characteristic: green bands and vein selvages comprise diopside, dark-green bands consist of hornblende with minor biotite, dark-brown bands comprise biotite with minor hornblende, and pale-grey bands are mainly quartz and calcite with minor diopside. All bands contain 1-2% pyrrhotite. White quartz veins contain host-rock slivers altered to green diopside ± white calcite.*

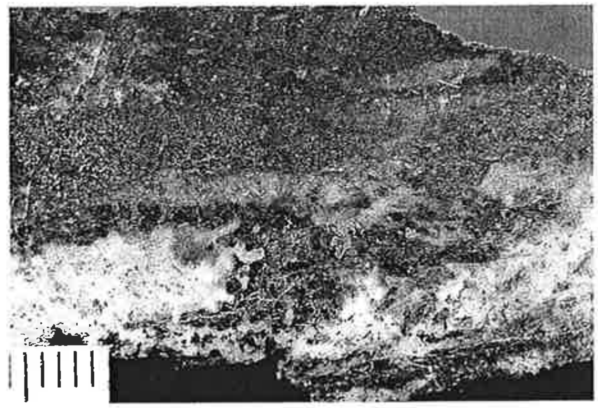


*Figure 52 Veins in ore from the lower granulite-facies sedimentary rocks at the Griffin's Find deposit, Southern Cross greenstone belt. Banding: red and brown garnet, green diopside and hornblende, and grey to green quartz-diopside-garnet vein. There is minor pyrrhotite in the main vein.*

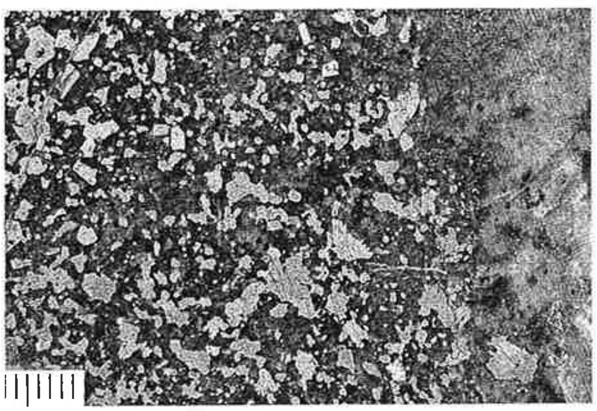


*Figure 53 Abundant pyrrhotite (bronze) and arsenopyrite-loellingite (white) dissemination near mineralised quartz vein (on right) in the lower-granulite facies metasedimentary host rock at Griffin's Find. Silicates are plagioclase, quartz, diopside and hypersthene. Yellow specks are reflections of light from the surfaces of sulphide grains beneath transparent quartz.*

*Figure 54 Vein comprising white quartz, yellow-green diopside and dark green to black tremolite in gold ore at Marvel Loch deposit.*

*Figure 55 Komatiite from the Hopes Hill deposit, Southern Cross greenstone belt; metamorphosed to mid-amphibolite facies. Grey-green zone at left is distal alteration, with tremolite + anthophyllite + talc + chlorite + ilmenite. Dark banded zone is proximal alteration with variable amounts of biotite, diopside, calcite and quartz, and minor titanite and pyrrhotite. Veins comprise quartz and diopside.*

*Figure 56 Strong banding in proximal alteration in basalt in the Polaris South deposit, Southern Cross greenstone belt; host rocks at mid- or upper-amphibolite facies. Dark-green bands are biotite-hornblende, pale-grey bands are plagioclase-calcite-quartz-diopside. There are also paler-green, boudinaged quartz-diopside veins. Bronze-yellow pyrrhotite occurs in both types of bands and in the veins.*

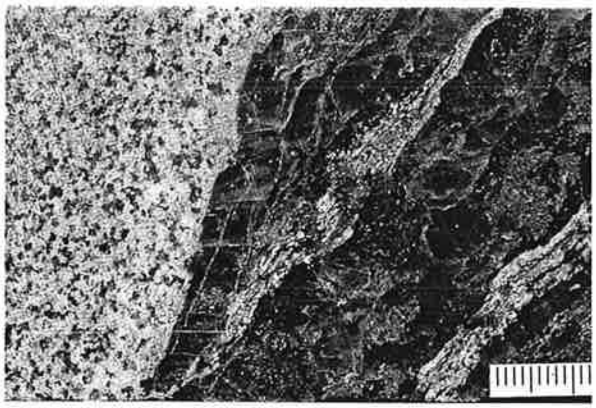


*Figure 57 Alteration in the granitoid-hosted Westonia deposit; host rocks metamorphosed at mid- to upper-amphibolite facies. On the left, there is weak distal alteration with little apparent change in the rock. In the middle and on the right, there is proximal alteration (grey, undulating slivers of wallrock) and quartz-diopside-hornblende veining. In the least altered area, black biotite decreases in abundance to the left and right, towards quartz veins: only those to the right are shown in the field of view.*

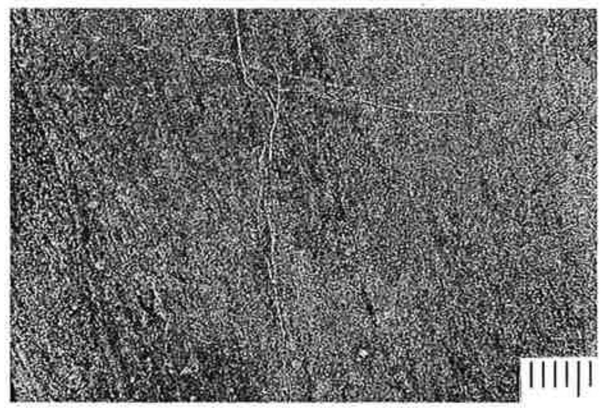


*Figure 58 Distal and proximal alteration in basalt from the Edward's Find deposit, Southern Cross greenstone belt, where host rocks were metamorphosed under mid-amphibolite facies conditions. Most of the sample represents distal alteration, with the mineral assemblage hornblende + biotite + plagioclase. Proximal alteration on left: bands dominated by green diopside and dark-brown biotite.*

*Figure 59 Sulphidation front in a BIF-hosted gold deposit in the Flin Flon-Snow Lake belt, Manitoba, Canada. Pyrrhotite replaces magnetite in the proximal alteration zone. Kinks in the sulphidation front indicate that alteration is clearly related to bedding-parallel veining and bedding planes, with veins and bedding planes forming the main fluid pathways. The BIF is metamorphosed to mid-amphibolite facies.*

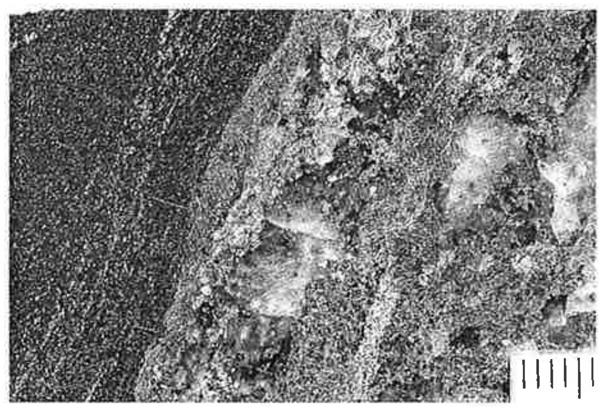


*Figure 60 Proximal alteration and a mineralised vein in basalt from the Chalice deposit, near Norseman, where host rocks have been metamorphosed under mid- or upper-amphibolite facies conditions. The proximal zone is characterised by the mineral assemblage hornblende + calcite + biotite + diopside + pyrrhotite. The vein comprises quartz, diopside and pyrrhotite. Note also a few mm-wide diopside selvage on the vein margin.*

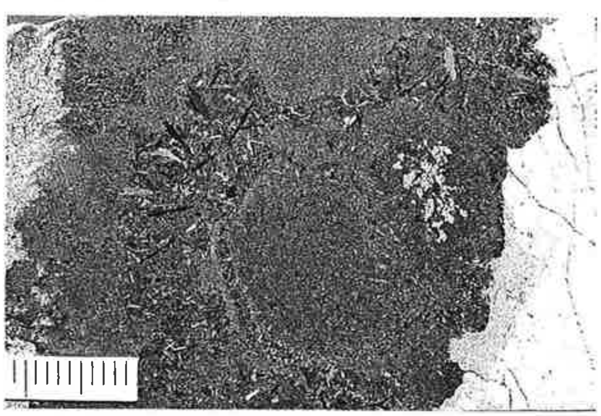


*Figure 61 Coarse-grained proximal alteration near BIF-hosted mineralised quartz vein (not shown), Nevoria mine, Southern Cross. Alteration assemblage is almandine garnet (red-brown) + pyrrhotite (bronze) + ferro-actinolite (dark green to black).*

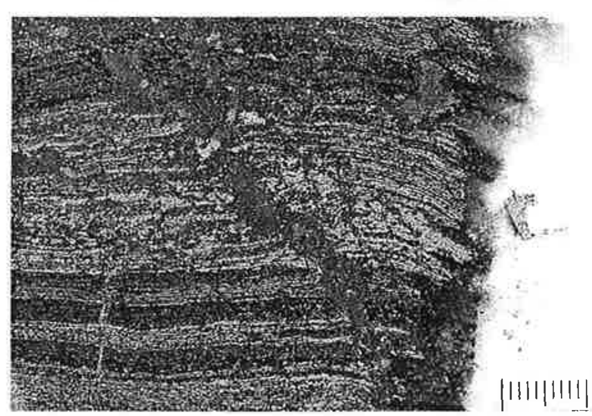


*Figure 62 Proximal, pervasive sulphidation of magnetite in BIF, Nevoria mine. Mineral assemblage is quartz + hedenbergite + ferro-actinolite + pyrrhotite with relict magnetite. Host rocks metamorphosed to mid-amphibolite facies.*

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# APPENDICES

# APPENDIX A LOCATION OF DEPOSITS

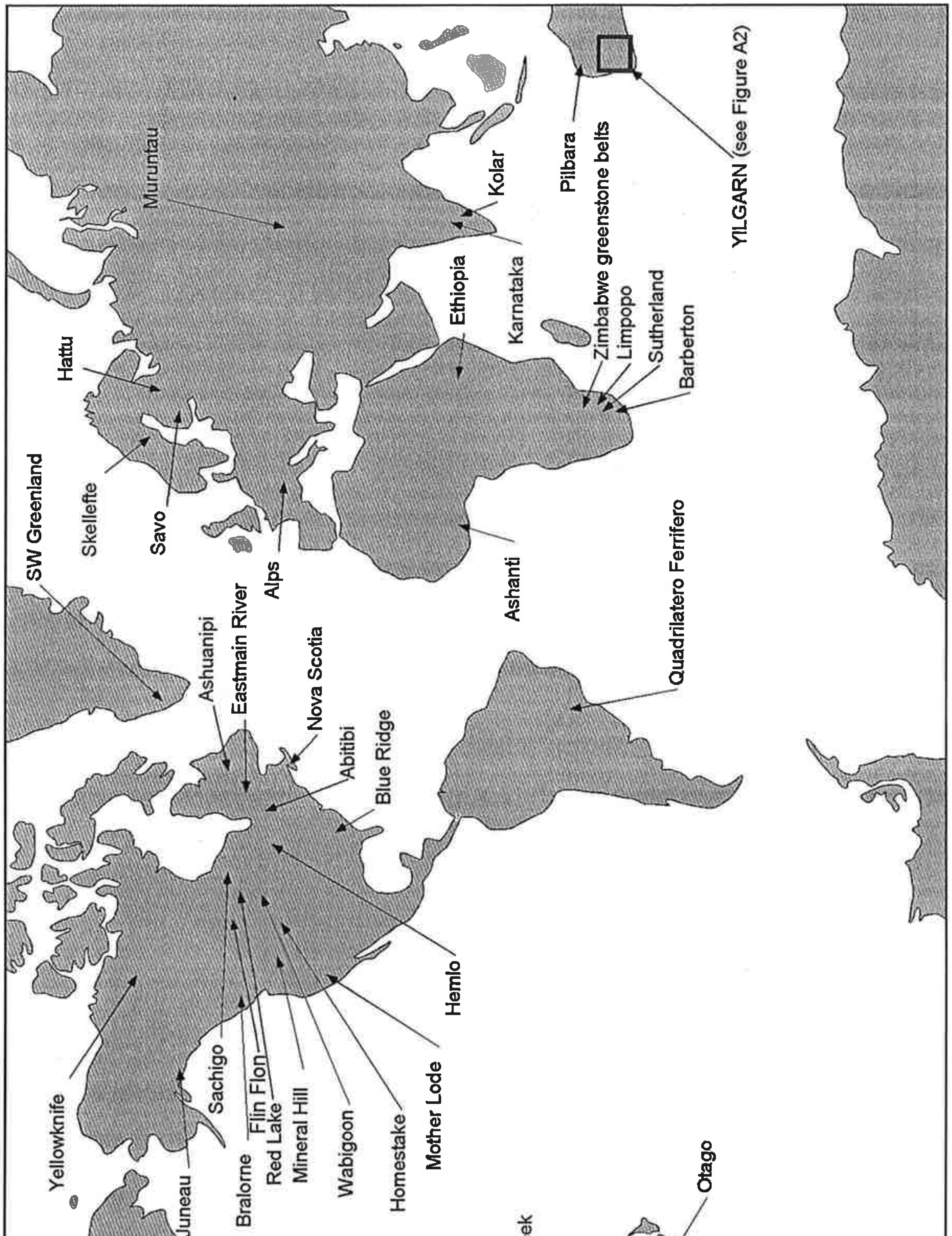


**Figure A1** Map of the world showing the location of orogenic lode-gold deposits and deposit belts mentioned in this Atlas

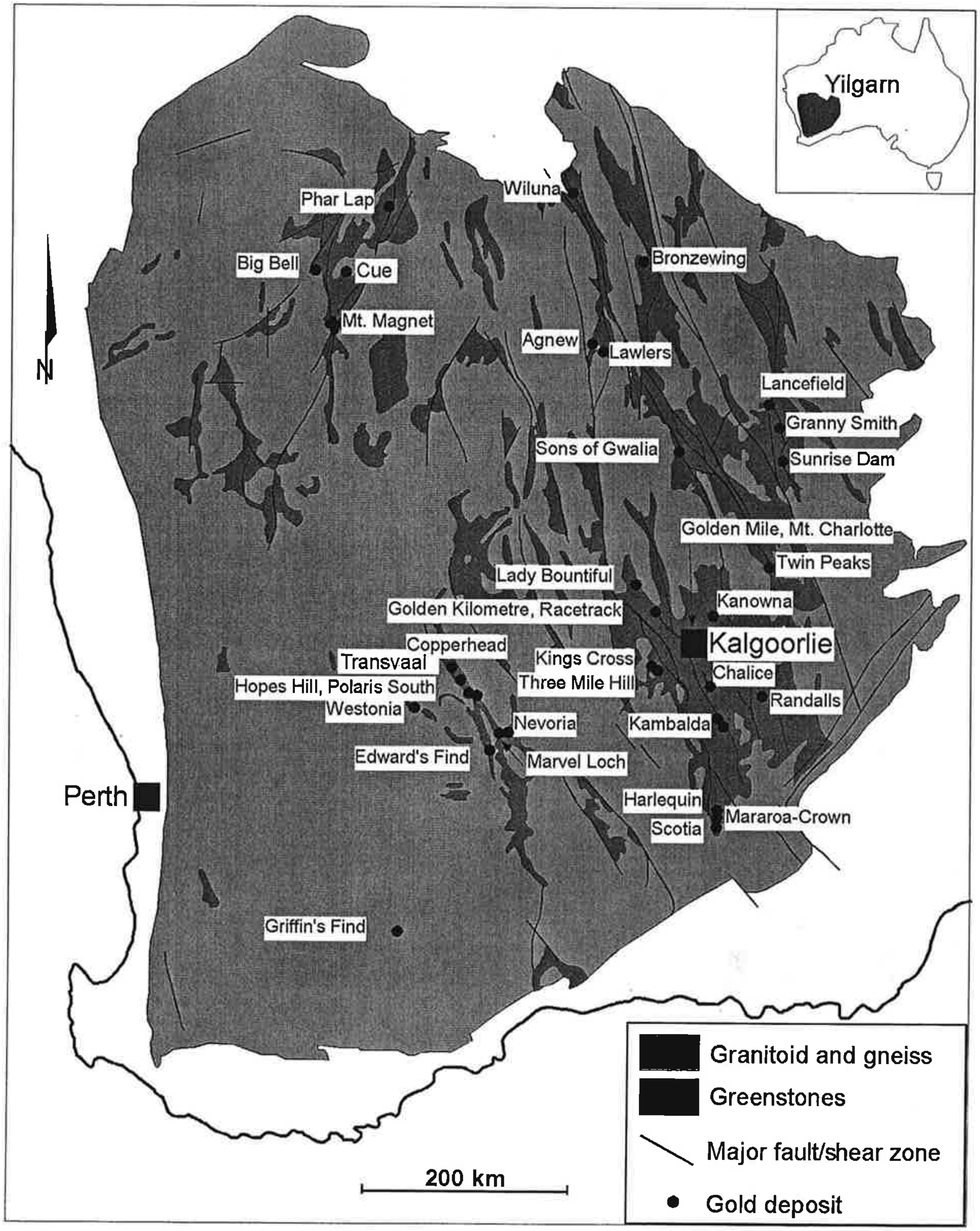


**Figure A2** Map of the Yilgarn Block, Western Australia, showing the location of orogenic lode-gold deposits and deposit belts mentioned in this Atlas

# APPENDIX B CARBONATE STAINING TECHNIQUES

The method presented here, based on a technique described by Hutchison (1974), is designed for staining pieces of diamond-drill core and RC and RAB drill samples, and other samples more bulky than thin sections.

**Staining solutions**

I. 1.5% HCl
II. 2% potassium ferricyanide (PF) in 1.5% HCl (= 2 g PF in 100 ml of liquid)
III. 0.2% Alizarin red-S (ARS) in 1.5% HCl (= 0.2 g ARS in 100 ml of liquid)

**Recommended staining procedure**

1. Hold the whole sample, or only the surface to be stained, in 1.5% HCl (solution I) for 10-15 seconds to clean the surface(s) to be stained.
2. Hold the sample (surface) in a solution formed by a 2:3 mixture of potassium ferricyanide (solution II) and Alizarin red-S (solution III) for 30-45 seconds.
3. Rinse the sample with distilled water for ≤1 second.
4. Hold the sample (surface) in Alizarin red-S (solution III) for 10-15 seconds.
5. Rinse the sample with distilled water for ≤1 second.

Stages 3 and 5 remove excess staining solution from sample surfaces.

**Stain colours**

After the procedure described above, pure calcite ($CaCO_3$) has a pale pink stain, ferroan calcite (by far the most common type of calcite in lode-gold systems) a purple stain, and ferroan dolomite and ankerite a blue stain. As the dilute HCl is not able to dissolve magnesite or siderite, no stain will be developed in these carbonates.

**Handling**

The stain developed is water soluble and easily rubbed off. Hence, handle the stained surfaces carefully if preservation is required.

# APPENDIX C DIAGNOSTIC PROPERTIES OF MINERALS IN ALTERATION ASSEMBLAGES

MInerals are arranged according to composition and then broadly in order of increasing T-P of formation. The *most common* occurrence is given in terms of metamorphic grade of host rocks and position in alteration sequence; some minerals may also be present in host rocks at other metamorphic grades or in other alteration zones.

| Mineral | Formula (approximate) | MM Grade/Alteration[1] | Colour in hand specimen | Habit | Cleavage[2] | Other diagnostic properties |
|---|---|---|---|---|---|---|
| **CARBONATES** | | | | | | softer than feldspars |
| Calcite | $CaCO_3$ | G/D; A/P | white | granular | 3 | distinct reaction with cold dilute HCl; alizarin-red (AR) gives a pink stain[3]; no change in colour by oxidation |
| Dolomite | $Ca(CO_3)_2$ ± Fe | G/P | white to buff, darkens by oxidation | granular | 3 | no reaction with cold dilute HCl; no stain with AR, blue stain with K ferricyanide if Fe present[3]; darkens to yellow-brown by oxidation |
| Ankerite | $Ca(Mg,Fe)(CO_3)_2$ | G/P | white to buff | granular | 3 | no reaction with cold dilute HCl; staining as for dolomite; darkens to yellow-brown by oxidation |
| Siderite | $FeCO_3$ | G/P | pale-brown | granular | 3 | no reaction with cold dilute HCl; no stain with AR or K ferricyanide; typical for BIF; darkens to red-brown by oxidation |
| Magnesite | $MgCO_3$ | G/P | white to buff | granular | 3 | no reaction with cold dilute HCl; no stain with AR or K ferricyanide; typical for ultramafic rocks |

| Mineral | Formula (approximate) | MM Grade/Alteration[1] | Colour in hand specimen | Habit | Cleavage[2] | Other diagnostic properties |
|---|---|---|---|---|---|---|
| **PHYLLOSILICATES** | | | | | | flakes easily scratched and separated with a steel needle |
| Chlorite | $(Mg,Fe,Al)_6(Al,Si)_4O_{10}(OH)_8$ | G/D(-P) | dark-green | flaky or scaly | 1 | pale green streak |
| Talc | $Mg_3Si_4O_{10}(OH)_2$ | G/D-P | white to pale-grey | flaky or scaly | 1 | the softest of all minerals of the table; soapy feel; diagnostic for ultramafic rocks |
| Sericite / Muscovite | $(K,Na)Al_3Si_3O_{10}(OH)_2$ | G/P | white to pale-yellow | flaky or scaly | 1 | may give a silky appearance |
| Fuchsite | $(K,Na)(Al,Cr)_3Si_3O_{10}(OH)_2$ | G/P | pale bright-green | flaky or scaly | 1 | only occurs in rocks with a high Cr content; typical for ultramafic rocks |
| Roscoelite | $(K,Na)(Al,V)_3Si_3O_{10}(OH)_2$ | G/P | pale bright-green | flaky or scaly | 1 | detected in some mafic rocks |
| Biotite | $K(Fe,Mg)_3AlSi_3O_{10}(OH)_2$ | G/P; A/D-P | dark-brown to black | flaky or scaly | 1 | most typical in upper-greenschist and lower-amphibolite facies deposits |
| Phlogopite | $KMg_3AlSi_3O_{10}(OH)_2$ | G/P; A/D-P | pale-brown to medium-brown | flaky or scaly | 1 | occurs like biotite, but characteristic for ultramafic rocks |
| **FELDSPARS** | | | | | | harder than carbonates |
| Albite | $NaAlSi_3O_8$ | G/D-P | white | granular | 2 | commonly very thin twin lamellae on some cleavage surfaces |
| Ca plagioclase | $CaAl_2Si_2O_8$-$NaAlSi_3O_8$ | A/D-P | white | granular | 2 | commonly twin lamellae on some cleavage surfaces |
| K feldspar | $KAlSi_3O_8$ | A/P | white to pink | granular | 2 | no lamellae in cleavage surfaces |
| **AMPHIBOLES** | | | | | | harder than chlorite, slightly softer than diopside |
| Actinolite | $Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2$ | A/P | pale to dark-green | prismatic or fibrous | 2 | present only in unaltered rocks in greenschist facies |
| Tremolite | $Ca_2Mg_5Si_8O_{22}(OH)_2$ | A/P | white to dark-green | prismatic or fibrous | 2 | typical for ultramafic rocks; present only in unaltered rocks in greenschist facies |
| Hornblende | $NaCa_2(Mg,Fe,Al)_5(Al,Si)_8O_{22}(OH)_2$ | A/D-P | dark-green to black | prismatic | 2 | usually darker colour than other amphiboles |
| Grunerite | $Fe_7Si_8O_{22}(OH)_2$ | A/D-P | pale (yellow-) brown | prismatic or fibrous | 2 | typical for BIF |
| **OTHER SILICATES** | | | | | | |
| Epidote | $Ca_2(Al,Fe)_3Si_3O_{12}(OH)$ | G/D | yellow-green | granular | (1) | characteristic for unaltered rock; may be present in the most distal alteration zones |
| Diopside | $CaMgSi_2O_6$ | A/P | pale-green to grey-green | granular | (2) | grey compared to epidote |
| Grossularite | $Ca_3Al_2Si_3O_{12}$ | A/P | pale- to dark-brown | granular, euhedral | 0 | commonly porphyroblastic |
| Almandine | $Fe_3Al_2Si_3O_{12}$ | A/P | red to brown-red | granular, euhedral | 0 | commonly porphyroblastic |
| **OXIDES** | | | | | | |
| Magnetite | $Fe_3O_4 \pm Ti$ | G/D; A/D-P | black | granular, idiomorphic | 0 | strongly magnetic; octahedral crystals; black streak |
| Ilmenite | $FeTiO_3$ | A/D-P | black | granular, idiomorphic | 0 | nonmagnetic; black streak |
| Hematite | $Fe_2O_3$ | G/P | steel-grey | platy to fine granular | 0 | nonmagnetic; brown-red streak; red when fine-grained |
| Rutile (leucoxene) | $TiO_2$ | G/D-P | pale-brown to medium-brown | granular | 0 | most easy to detect in unbleached mafic rocks |
| Titanite | $CaTiSiO_5$ | G/D; A/P | grey to pale-brown | granular | 0 | white compared to the brown rutile |
| **SULPHIDES, SULPHARSENIDES, ARSENIDES** | | | | | | |
| Pyrite | $FeS_2$ | G/P | brassy-yellow | granular to cubic | 0 | harder than pyrrhotite |
| Pyrrhotite | FeS | A/P | brown-bronze | granular to massive | 0 | magnetic in most cases; dulls quickly |
| Arsenopyrite | FeAsS | G-A/P | silver-white | granular, prismatic | 0 | whiter than pyrite; small crystals needle-shaped |
| Loellingite | $FeAs_2$ | A/P | silver-white | granular | 0 | very similar to arsenopyrite |
| Stibnite | $Sb_2S_3$ | sub-G/P | silver-grey | granular, prismatic | 1 | similar to galena, but softer and, typically, elongate crystals |

***NOTES***

1 *Metamorphic Grade: G = greenschist, A = amphibolite, G-A = greenschist and amphibolite, sub-G = sub-greenschist facies; Alteration: P = proximal, D = distal, D-P = distal to proximal alteration*

2 *0 = no prominent cleavage, 1 = one, 2 = two, 3 = three distinct cleavages, brackets = cleavage not obvious*

3 *Refers to staining methods for distinguishing carbonates in Hutchison (1974)*

# APPENDIX D DEPOSITS IN SUB-GREENSCHIST TO MID-GREENSCHIST FACIES ROCKS

Selected orogenic lode-gold deposits in sub- to mid-greenschist facies rocks, listed in alphabetical order. Deposits are selected to represent all major host rock types and, if possible, include an example from several significant orogenic lode-gold terrains around the world.

**SUB-GREENSCHIST FACIES**

| Deposit | Main host rock | Reference |
|---|---|---|
| Commoner, Midlands, Zimbabwe | clastic sedimentary rocks | Twemlow (1984) |
| Phar Lap, Yilgarn, Australia | komatiite | Jackson (1990) |
| Racetrack, Yilgarn, Australia | basalt | Gebre-Mariam et al. (1993, 1995) |
| Ross, Abitibi, Canada | basalt | Troop (1986) |
| Wiluna, Yilgarn, Australia | basalt | Hagemann et al. (1992, 1994) |

**LOWER- TO MID-GREENSCHIST FACIES**

| Deposit | Main host rock | Reference |
|---|---|---|
| Bendigo-Ballarat (Victoria, Australia) | clastic sedimentary rocks | Phillips & Hughes (1996), Ramsay et al. (1996) |
| Björkdal (Skellefte Belt, Sweden) | granodiorite | Broman et al. (1994) |
| Bogosu and Prestea (Ashanti, Ghana) | clastic sedimentary rocks | Mumin et al. (1996) |
| Bralorne (British Columbia, Canada) | diorite, basalt | Leitch (1990), Ash et al. (1996) |
| Cam and Motor (Kadoma, Zimbabwe) | basalt | Foster and Piper (1993) |
| Fairview (Barberton, South Africa) | clastic sedimentary rocks | Wigget et al. (1986), de Ronde et al. (1992), Foster & Piper (1993) |
| Globe and Phoenix (Kwekwe, Zimbabwe) | granitoid, ultramafic rocks | Porter & Foster (1991), Foster & Piper (1993) |
| Golden Mile, Golden Kilometre, Mt Charlotte (Yilgarn, Australia) | dolerite | Phillips (1986), Clout et al. (1990), Gebre-Mariam et al. (1993), Mikucki & Heinrich (1993), Phillips et al. (1996) |
| Granny Smith (Yilgarn, Australia) | granitoid, clastic sedimentary rocks | Ojala et al. (1993) |
| Hollinger-McIntyre (Abitibi, Canada) | basalt, porphyry | Smith & Kesler (1985), Burrows et al. (1993) |
| Juneau[1] (Alaska, USA) | gabbro, clastic sedimentary rocks | Spencer (1906), Newberry & Brew (1988), Goldfarb et al. (1997) |
| Kanowna (Yilgarn, Australia) | clastic sedimentary rocks, felsic porphyry | Heithersay et al. (1994), Witt et al. (1994) |
| Kerr Addison (Abitibi, Canada) | komatiite, basalt | Kishida & Kerrich (1987) |
| Kirkland Lake (Abitibi, Canada) | syenite, clastic sedimentary rocks | Robert & Poulsen (1997) |
| Lady Bountiful (Yilgarn, Australia) | granitoid | Cassidy (1992) |
| Lancefield (Yilgarn, Australia) | sedimentary rocks, basalt | Hronsky (1993), Mikucki & Ridley (1993) |
| Macraes (Otago, New Zealand) | clastic sedimentary rocks | Christie (1996) |
| Mother Lode deposits, California, USA | granite, clastic sedimentary rocks, basalt | Böhlke (1989) |
| Mt Magnet deposits[2] (Yilgarn, Australia) | BIF | Thompson et al. (1990), Kelly (1998) |
| Mt Rosa (NW Alps, Italy) | clastic sedimentary rocks | Pettke & Diamond (1997) |
| Morro Velho and other deposits (Quadrilatero Ferrifero, Brazil) | BIF | Vieira et al. (1986), Ladeira (1991), Scarpelli (1991), Vieira (1991) |
| Sheba (Barberton, South Africa) | sedimentary rocks, komatiite | Wagener & Wiegand (1986), Foster & Piper (1993) |
| Sunrise Dam (Yilgarn, Australia) | clastic sedimentary rocks, BIF | Newton et al. (1998) |
| Twin Peaks (Yilgarn, Australia) | clastic sedimentary rocks | Eilu (1996c), Longworth (1996) |

***NOTES*** **1** *Most of the deposits in the Juneau area (Goldfarb et al., 1997)*
**2** *Parts of the deposits are possibly formed under upper-greenschist facies conditions*

# APPENDIX E DEPOSITS IN UPPER-GREENSCHIST TO LOWER-AMPHIBOLITE FACIES ROCKS

Selected orogenic lode-gold deposits at upper-greenschist and lower-amphibolite facies rocks, listed in alphabetical order. Deposits are selected to represent all major host rock types and, if possible, include an example from several significant orogenic lode-gold terrains around the world.

UPPER-GREENSCHIST FACIES

| Deposit | Main host rock | Reference |
|---|---|---|
| Alaska-Juneau[1] (Alaska, USA) | clastic sedimentary rocks, dolerite | Spencer (1906), Newberry & Brew (1988), Goldfarb et al. (1997) |
| Beaver Dam (Meguma, Nova Scotia, Canada) | clastic sedimentary rocks | Kontak et al. (1990), Kontak & Smith (1993) |
| Bronzewing (Yilgarn, Australia) | basalt | Dugdale (1996), Eilu et al. (1996), Phillips et al. (1998) |
| Cosmo Howley (Pine Creek, Australia) | clastic sedimentary rocks | Matthäi et al. (1995) |
| Golden Crown (Yilgarn, Australia) | dolerite | Uemoto (1996) |
| Homestake[2] (Black Hills, USA) | BIF | Caddey et al. (1991), Kerswill (1993) |
| Kambalda deposits (Yilgarn, Australia) | basalt, dolerite, komatiite | Neall & Phillips (1987), Clark et al. (1989), Lesher et al. (1991) |
| Lega Dembi (Ethiopia) | black shale | Billay et al. (1997) |
| Malartic (Abitibi, Canada) | komatiite | Sansfacon & Hubert (1990) |
| New Consort[2] (Barberton, South Africa) | pelites, komatiite, basalt | Maiden (1984), Voges (1986) |
| Randalls (Yilgarn, Australia) | BIF | Newton et al. (1997) |
| Sigma-Lamaque[1] (Abitibi, Canada) | intermediate volcanic and subvolcanic rocks | Robert & Brown (1986), Robert & Poulsen (1997) |
| Sons of Gwalia (Yilgarn, Australia) | basalt, komatiite | Skwarnecki (1990) |

**LOWER-AMPHIBOLITE FACIES**

| **Deposit** | **Main host rock** | **Reference** |
|---|---|---|
| Agnew[3] (Yilgarn, Australia) | komatiite | Broome et al. (1998), L. Rödsjö (pers. comm., 1998) |
| Copperhead (Yilgarn, Australia) | BIF | Bloem (1994) Southern Cross |
| Dickenson (Red Lake, Canada) | basalt | Colvine (1989) |
| Duport (Wabigoon, Canada) | komatiite, basalt | Smith (1986), Colvine (1989) |
| Gladstone[2] (Harare belt, Zimbabwe) | komatiite, basalt | Brown et al. (1988) |
| Harlequin (Yilgarn, Australia) | dolerite, basalt | McCuaig et al. (1993) |
| Hattu schist belt deposits[3] (Finland) | clastic sedimentary rocks, tonalites | Nurmi & Sorjonen-Ward (1993), Rasilainen (1996) |
| Hutti (Karnataka, India) | basalt | Raju & Sharma (1991) |
| Kings Cross (Yilgarn, Australia) | basalt | Knight et al. (1993), Knight (1994) |
| Lawlers[3] (Yilgarn, Australia) | granitoid | Cassidy (1992), Mikucki (1997) |
| Lily[3] (Barberton, South Africa) | pelites, BIF | Williams (1997) |
| Lowe (Blue Ridge, USA) | clastic sedimentary rocks | Stowell et al. (1996) |
| Lupin[3] (Yellowknife, Canada) | BIF | Bullis et al. (1994) |
| Madsen (Red Lake, Canada) | basalt | Andrews et al. (1986) |
| Mararoa-Crown (Yilgarn, Australia) | basalt, komatiite | McCuaig et al. (1993) |
| ZIA → Mineral Hill (Montana, USA) | BIF | Smith (1996) |
| Muruntau[3] (Kyzylkum, Uzbekistan) | clastic sedimentary rocks | Drew et al. (1996) |
| Musselwhite (Sachigo, Canada) | BIF | Hall & Rigg (1986), Colvine (1989) |
| Ouro Fino (Quadrilatero Ferrifero, Brazil) | clastic sedimentary rocks | Fonseca et al. (1991) |
| Three Mile Hill (Yilgarn, Australia) | dolerite | Knight et al. (1993), Knight (1994) |

***NOTES***
1. *Part of the deposit formed at mid-greenschist facies conditions and part probably formed at upper-greenschist facies conditions*
2. *Metamorphic grade interpreted from mineral assemblages described in the references*
3. *Transitional between greenschist and amphibolite facies*

# APPENDIX F DEPOSITS IN MID-AMPHIBOLITE TO GRANULITE FACIES ROCKS

Selected orogenic lode-gold deposits at mid-amphibolite to lower-granulite facies rocks, listed in alphabetical order. Deposits are selected, if possible, to represent all major host rock types and include an example from several significant orogenic lode-gold terrains around the world.

**MID- TO UPPER-AMPHIBOLITE FACIES**

| Deposit | Main host rock | Reference |
|---|---|---|
| Big Bell[1] (Yilgarn, Australia) | basalt | Phillips & de Nooy (1988), Wilkins (1993), Mueller et al. (1996) |
| Caue (Quadrilatero Ferrifero, Brazil) | BIF | Olivo et al. (1994) |
| Chalice (Yilgarn, Australia) | clastic sedimentary rocks | Thompson (1994) |
| Eastmain River (Canada) | basalt | Couture & Guha (1990) |
| Edward's Find (Yilgarn, Australia) | basalt | Bloem (1994) |
| Hemlo[1] (Ontario, Canada) | clastic sedimentary rocks | Pan & Fleet (1995), Fleet at el. (1997), Powell & Pattison (1997) |
| Hopes Hill (Yilgarn, Australia) | komatiite | Bloem (1994) |
| Fumani[2] (Sutherland, South Africa) | BIF | Potgieter & de Villiers (1986) |
| Kolar goldfield (Karnataka, India) | basalt | Hamilton & Hodgson (1986) |
| Marvel Loch (Yilgarn, Australia) | komatiite | Mueller (1990, 1995) |
| Mt York (Pilbara, Australia) | BIF | Neumayr et al. (1993), Neumayr et al. (1998) |
| Nevoria (Yilgarn, Australia) | BIF | Mueller (1990, 1997) |
| Osikonmäki (Savo, Finland) | tonalite | Kontoniemi (1998) |
| Polaris South (Yilgarn, Australia) | basalt | Schwebel et al. (1995) |
| Scotia (Yilgarn, Australia) | basalt | McCuaig et al. (1993) |
| Storö (SW Greenland) | basalt | Pedersen 1998 |
| Transvaal (Yilgarn, Australia) | sedimentary rocks, komatiite | Dalstra et al. (1997) |
| Westonia (Yilgarn, Australia) | granitoid | Cassidy (1992) |

**LOWER-GRANULITE FACIES**

| Deposit | Main host rock | Reference |
|---|---|---|
| Griffin's Find (Yilgarn, Australia) | clastic sedimentary rocks, basalt | Fare (1989), Fare & McNaughton (1990), Barnicoat et al. (1991) |
| Klein Letaba (southern Limpopo belt, South Africa) | komatiite | van Reenen et al. (1994), Gan & van Reenen (1997) |
| Lac Liliois (Ashuanipi, Canada) | BIF | Lapointe & Chown (1993) |
| Louis Moore (southern Limpopo belt, South Africa) | komatiite | van Reenen et al. (1994) |
| Osprey (southern Limpopo belt, South Africa) | basalt, BIF | van Reenen et al. (1994) |
| Renco[3] (northern Limpopo belt, Zimbabwe) | clastic sedimentary rocks, basalt | Böhmke & Varndell (1986), Hagemann & Brown (1996) |

***NOTES***
1. *Somewhat controversial with respect to origin and number of mineralising events*
2. *Metamorphic grade interpreted from mineral assemblages described by Potgieter & de Villiers (1986)*
3. *Still controversial regarding the mineralisation conditions; see discussion by Blenkinsop & Frei (1996, 1997) and Kisters et al. (1997)*