Mineral Deposits and Geology of Central Colorado

Red Rocks Park to Cripple Creek, Colorado
July 2–8, 1989

Field Trip Guidebook T129

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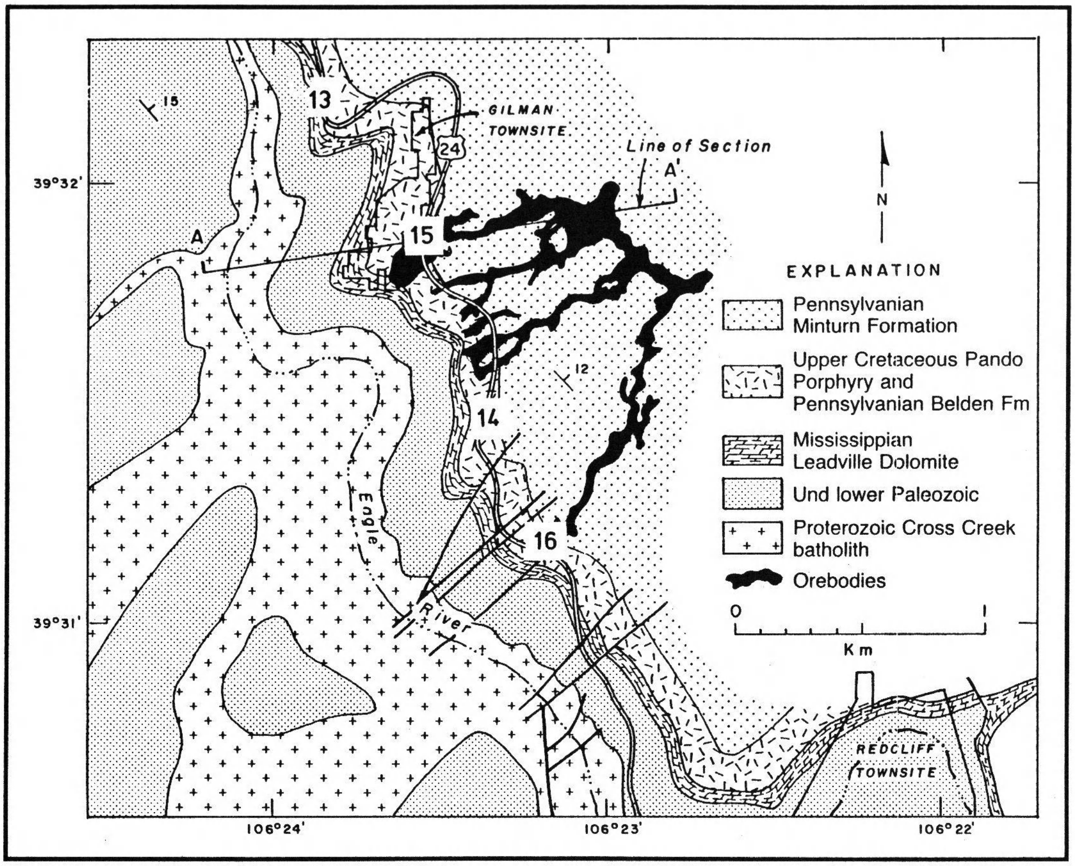
2000 Florida Ave., N.W., Washington, D.C. 20009

ISBN: 0-87590-645-1

Printed in the United States of America



COVER Map of the Gilman mine showing principal ore bodies in the Leadville Dolomite and locations of stops. See Figure 11, page 34 for details.

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IGC FIELD TRIP T129
MINERAL DEPOSITS AND GEOLOGY OF CENTRAL COLORADO

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This five day field trip is a general overview of the geology and tectonic history of central Colorado and of the geology of seven famous mining districts. Some of these mining districts had their most productive years in the latter decades of the 19th century, whereas others reached their peak production more recently. At present mining activity in some of them has ceased. In others it is at low levels compared to earlier years.

The regional geology and the road log are compiled from numerous sources. The principal ones are field trip guides by Tweto (1978), Reed and others (1988), Lipman (in press) and unpublished field guides of the Colorado Scientific Society and the Society of Economic Geologists. Figure 1 shows the principal geographic features along the route of the trip.

REGIONAL TECTONICS

Today central Colorado is part of a large topographic high named the Alvarado Ridge by Eaton (1986), who attributed its origin to lithospheric thinning accompanied by extensional strain and differential vertical jostling along the crest of the ridge during the Neogene. This restricted zone of extensional deformation is closely related to regional extension over a large region of the western United States (Fig. 2) known as the Basin and Range province. A well-defined graben system, the Rio Grande Rift, extends along the crest of the ridge through New Mexico and Colorado. This ridge, about 1,000 km long, accounts for the altitude of the region and is superposed on a long succession of older structures.

The next oldest regional structural features are the Laramide uplifts formed on the craton in front of the Cordilleran fold and thrust belt in Late Cretaceous and early Tertiary time (70-40 Ma) (Fig. 2). These uplifts trend north to northwest in Colorado and are part of a broad belt of uplifts whose trend becomes more westerly in Wyoming and Utah. In the past, geologists have disagreed on whether the deformation forming these uplifts was due to horizontal compressive or vertical forces. Seismic exploration and drilling (Gries, 1983) and COCORP reflection profiling (Smithson and others, 1979; Brewer and others, 1982) show that the major faults bounding the uplifts are thrust faults that carry rocks of the uplifts over adjacent basin fill. In a regional analysis of the timing of the development of Laramide sedimentary basins, Dickinson and others (1988) conclude that the intricate geometry of Laramide structural features in the region developed in a generally synchronous strain field imposed over a region of heterogeneous crust. One interpretation of the cause for the compressive stress and the pattern of the uplifts is that during the Late Cretaceous and early Tertiary time North America overrode buoyant oceanic lithosphere along a gently dipping subduction zone (Hamilton, in press). Because of drag against the overridden slab, the southwestern United States advanced slightly more slowly than the continental interior, resulting in a broad zone of crustal shortening. The regional pattern of the structures formed by compression indicates that the Colorado Plateau region has rotated a few degrees clockwise around an Euler pole near central New Mexico relative to the continental interior. Somewhat different plate tectonic scenarios have been given by Chapin and Cather (1981), Gries (1983), and Cross (1986).

The next oldest major deformation in the region occurred in late Paleozoic time when northwest-trending mountains and basins formed (Fig. 3). These mountains are called the Ancestral Rockies. The mainly clastic rocks, deposited in the basins, are locally

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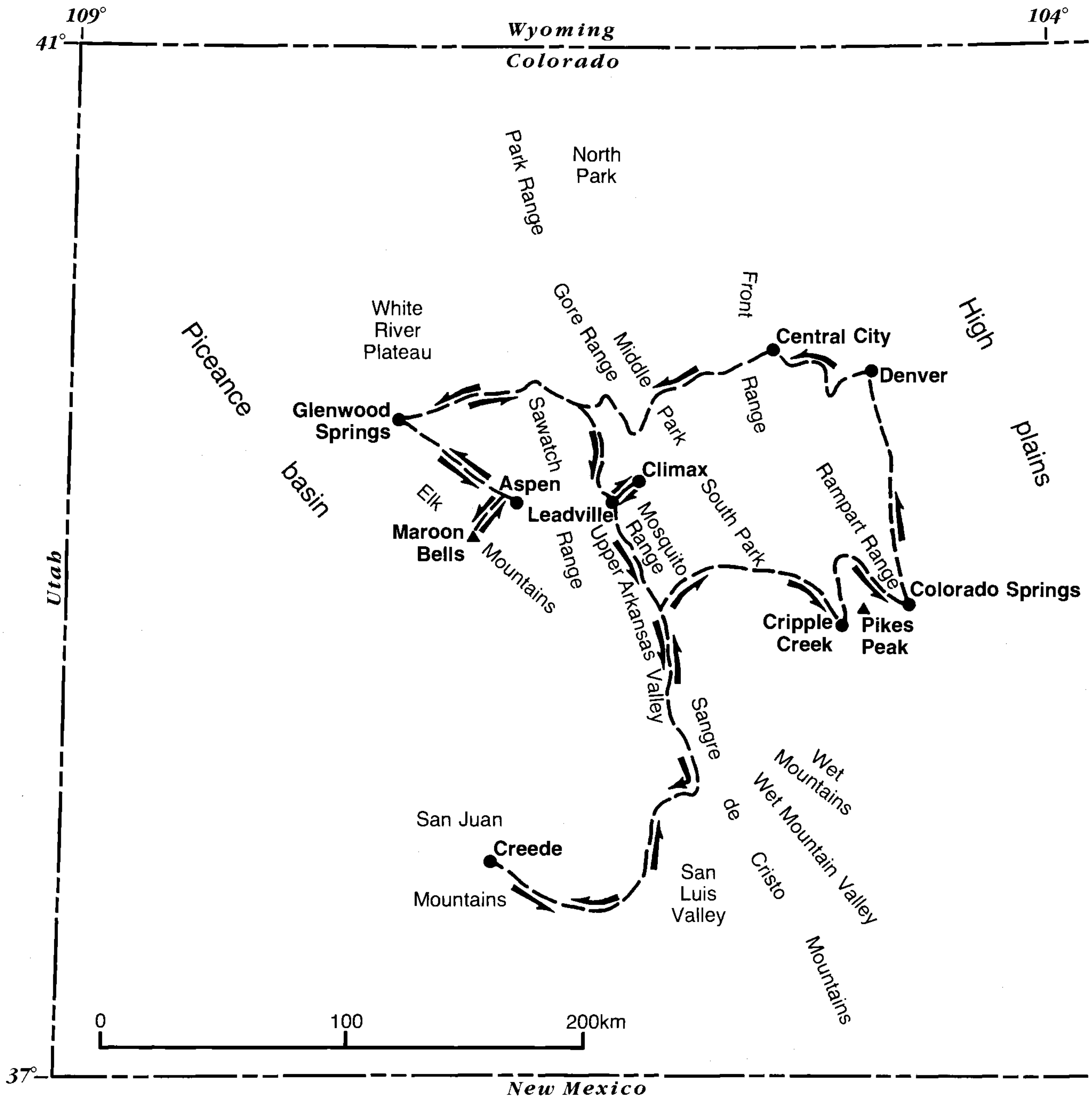


FIGURE 1 Geographic features in central Colorado. Route of trip shown by dashed line.

as much as 5 km thick and are well preserved in many areas. Many of these rocks were deposited in a semi-arid climate, and iron-bearing minerals, such as magnetite and hornblende, derived from basement rocks, were oxidized to produce strikingly red rocks that give the State of Colorado its name. Kluth (1986) related the formation of these tectonic features to the collision of the North America and the South America-Africa plates during the Ouachita-Marathon orogeny.

The oldest recorded event in the region was the formation of continental crust in Early Proterozoic time when a series of arcs were formed and welded to the Archean craton (Reed and others, 1987).

Much of the following discussion is condensed and paraphrased from Tweto (1975) and the references cited therein. The description of the Precambrian rocks is based on Tweto (1977a, 1987) and Reed and others (1987). In addition, the reader may wish to refer to the Geologic Map of Colorado (Tweto, 1979b) and the following 1^o X 2^o geologic maps: Denver (Bryant and others, 1981), Leadville (Tweto and others, 1979), Pueblo (Scott and others, 1978),

name Silver Plume specifically for the rock in the Silver Plume batholith, one of many plutons of that general age. Geochemical study of the Silver Plume batholith indicates that it was derived from a relatively dry, peraluminous oxidized lower crust and emplaced in the upper crust at depths of 9.7 to 11.2 km or shallower and at temperatures of 740-760°C at an elevated f_{O_2} (Anderson and Thomas, 1985).

The last major group of Proterozoic igneous rocks in central Colorado comprise the 1,000 Ma Pikes Peak batholith. The batholith consists principally of biotite and hornblende-biotite potassic granite, but includes subsidiary plutons of syenite, quartz monzonite, alkali diorite, and related rocks. Geochemical data suggest that the rocks in the batholith originated through a combined process of fractional crystallization of alkalic-basaltic liquid and assimulation of granodioritic and granitic roof rocks. Depth of emplacement may have been as little as 5 km (Barker and others, 1975; Barker and others, 1976).

The Proterozoic rocks are cut by an anastomosing swarm of northeast-trending zones of ductile deformation and cataclasis. These shear zones have long and complicated histories of movement, but movement along some of them began during the Early Proterozoic. The zones apparently provided the locus for the Laramide and younger plutons of the Colorado mineral belt (Tweto and Sims, 1963) and may have been part of a regional wrench fault system (Warner, 1978).

In the Front Range a set of northwest- and north-northeast-trending fault zones (fig. 4) is younger than the shear zones but formed in Precambrian time. Many of these fault zones were reactivated during Phanerozoic deformations (Tweto and Sims (1963).

PALEOZOIC, TRIASSIC, AND JURASSIC ROCKS

Paleozoic rocks ranging from Cambrian through Mississippian comprise a shelf sequence 300 m or less thick that was originally deposited across all of central Colorado. However, these strata were eroded from the major late Paleozoic uplifts and are preserved only in the intervening late Paleozoic basins. Older Paleozoic rocks are missing along the eastern flank of the Front Range north of the latitude of Colorado Springs, in North Park, Middle Park, and in the eastern part of South Park. They are preserved in the Mosquito and Sawatch Ranges, where they host the ore deposits at Leadville, Gilman, and Aspen.

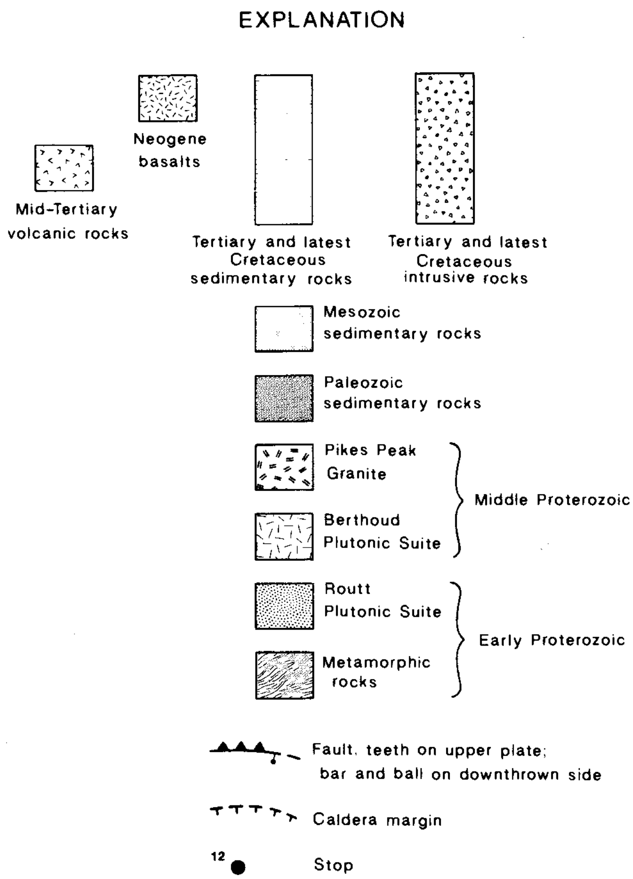
Major elongate uplifts, commonly referred to as the Ancestral Rocky Mountains, began to form in Early or Middle Pennsylvanian time (Fig. 3) Some of these uplifts developed into mountain ranges that attained altitudes of as much as 1,500 to 3,000 m above sea level (Mallory, 1972). As the uplifts rose, older Paleozoic strata were stripped, and Precambrian rocks were exposed over broad areas. Clastic sedimentary rocks of Pennsylvanian and Permian age shed from the uplifts accumulated in the flanking basins, locally attaining thicknesses of more than 5 km. The sediments become finer grained towards the interior of the basins, and locally thick sections of evaporite were deposited in sub-basins, which were at least in part controlled by intrabasin deformation (De Voto and others, 1986). The principal late Paleozoic uplifts in central Colorado were the Front Range highland, which occupied much of the area of the present Front Range, and the Uncompahgre highland that lay southwest and west of the present Sawatch Range and extended southeast to the present San Luis Valley (Fig. 1). Between the two lay the central Colorado trough (Fig. 3), the northwestern part of which comprises the Eagle basin.

The thick clastic sequences derived from the late Paleozoic uplifts are overlain by Permian, Triassic, and Jurassic fluvial, eolian, and shoreline deposits that record the gradual reduction and submergence of the Ancestral Rocky Mountain uplifts.

The general variation in thickness of these various sequences at several places along the route of the field trip is illustrated in Figure 5; the units are listed and briefly described in Table 1.

CRETACEOUS AND TERTIARY ROCKS

A blanket of intertonguing marine and continental rocks of Cretaceous age, locally nearly 3 km thick, was deposited across the entire Southern Rocky Mountain region, covering the eroded roots of the late Paleozoic uplifts and the intervening basins. The Cretaceous rocks were largely derived from sources to the west and record several cycles of southwestward marine transgression and northeastward regression. The beginning of the Laramide orogeny is recorded in structural basins throughout the region in the abrupt transition from fine-grained Upper Cretaceous sedimentary rocks of far western derivation to orogenic sediments of local derivation. In the Denver basin this change



is marked by the transition during the Maastrichtian from the marine Fox Hills Sandstone to nonmarine, coal-bearing strata of the Laramie Formation and sandstone and conglomerate of the Arapahoe Formation (Table 1). The Arapahoe is derived from the Front Range uplift, but at least locally near Colorado Springs the Laramie is unconformably overlain by the Denver Formation and the Dawson Arkose. In South Park on the other side of the Front Range uplift, the Laramie is unconformably overlain by volcanic detritus of the South Park Formation. In Middle and North Parks, Pierre Shale which is of Campanian and Maastrichtian age in this area is unconformably overlain by conglomerate of the Middle Park Formation of Late Cretaceous and Paleocene age. In western Colorado, south of Glenwood Springs, the Campanian part of the Mancos Shale intertongues with and is overlain by the nomarine Mesaverde Group, which contains coal beds. Upper Paleocene conglomerates of the Wasatch Formation disconformably overlie the Mesaverde.

LARAMIDE TECTONIC ELEMENTS

The principal Laramide features in central Colorado are the Sawatch and Front Range uplifts, the Denver basin, and the tectonic depression marked by North, Middle, and South Parks (Figs. 1, 4). The Sawatch and Mosquito Ranges are both part of the Sawatch uplift that rose in the Laramide but was sundered by faulting during development of the Arkansas graben (part of the Rio Grande Rift) along the crest of the Alvarado ridge in the Neogene. Igneous rocks apparently emplaced during early stages of the growth of the anticline have been dated at about 70 Ma (early Maastrichtian). Paleocene beds are involved in tilting on the east flank of the anticline, suggesting that growth of the structure may have extended over a long interval. The Uncompahgre highland was strongly rejuvenated during Laramide deformation at its eastern end where Precambrian basement rocks and the thick upper Paleozoic sequence of the central Colorado trough were thrust eastward. These structures are well exposed in the Neogene horst forming the Sangre de Cristo Mountains. Sediments in the Raton basin indicate that the Uncompahgre was a positive element by latest Cretaceous time. However, deformation of the west edge of the basin did not occur until Paleocene time and continued into the Eocene. The White River uplift has less structural relief and is largely the result of Eocene uplift.

Initial uplift of at least the southern part of the Front Range was apparently somewhat later than that of the Sawatch anticline. Conglomerates in the lower part of the Denver Formation along the eastern flank of the range indicate that a nearby uplift existed to the west in latest Cretaceous time. However, the Paleocene Green Mountain Conglomerate that disconformably overlies the Denver Formation east of stop 1 contains pebbles of Paleozoic rocks derived from west of the Front Range, showing that the uplift had not entirely blocked eastward drainage at that time. The sharp monoclinal upturn and steep reverse faulting along the east flank of the Front Range involved the Green Mountain Conglomerate and Dawson Formation and some of the deformation may be as young as middle Eocene. The Williams Range thrust and related folds and thrust faults on the west side of the Front Range uplift involve beds as young as early Eocene.

The Denver basin is in part a segment of a regional basin of Late Cretaceous age which has been modified by uplift of the

TABLE 1 Stratigraphic Sections

East Flank Of Front Range Morrison quadrangle (Scott, 1972)	Thickness (meters)
Green Mountain Conglomerate (Paleocene)	
Conglomerate, sandstone, and shale. Contains andesite pebbles	200
Denver Formation (Paleocene and Upper Cretaceous)	
Brown to olive claystone, siltstone, sandstone, and conglomerate. Commonly tuffaceous. Rich in andesite clasts.	290
Arapahoe Formation (Upper Cretaceous)	
White, gray , and yellow sandstone, siltstone, claystone, and conglomerate. Conglomerate clasts are Phanerozoic sedimentary rock and Precambrian igneous and metamorphic rocks.	120
Laramie Formation (Upper Cretaceous)	
Gray siltstone and claystone and yellow and white sandstone. Some beds of coal in lower part.	165
Fox Hills Sandstone (Upper Cretaceous)	
Olive to brown silty shale and sandstone and yellowish-orange sandstone.	55
Pierre Shale (Upper Cretaceous)	
Olive-green shale, some beds of olive to brown sandstone, and limestone concretions.	1890
Niobrara Formation (Upper Cretaceous)	
Smoky Hill Shale Member (125) at top; Fort Hays Limestone Member (10) at base; total.	135
Carlile Shale, Greenhorn Limestone, and Graneros Shale (Upper Cretaceous)	160
Dakota Group (Lower Cretaceous)	
Yellowish-gray sandstone and dark gray shale; yellow-brown sandstone and conglomerate at base. Divided into Lytle (lower) and South Platte (upper) Formations.	90
Morrison Formation (Upper Jurassic)	
Red and green siltstone and claystone. Some beds of brown sandstone and gray limestone.	90
Ralston Creek Formation (Jurassic)	
Purplish gray sandstone and siltstone, yellow silty sandstone	27
Lykins Formation (Triassic? and Permian)	
Maroon shale, sandy limestone, and maroon and green siltstone	40
Lyons Sandstone (Permian)	
Yellowish-gray conglomerate and sandstone	60
Fountain Formation (Permian and Pennsylvanian)	
Maroon arkosic sandstone, and conglomerate	500
Precambrian rocks	
Paleozoic Front Range Highland And West Margin of Laramide Front Range Uplift Blue River Valley near Dillon (Robinson, Warner, and Wahlstrom, 1974)	
Pierre Shale (Upper Cretaceous)	
Dark-gray to black shale	760±
Niobrara Formation (Upper Cretaceous)	
Smoky Hill Member (140±} at top; Fort Hays Limestone Member (5±) at base; total	145
Benton Shale (Upper and Lower Cretaceous)	
Dark-gray to black shale	60±
Dakota Sandstone (Lower Cretaceous)	
Gray sandstone; gray to black shale	55±
Morrison Formation (Upper Jurassic)	
Gray and greenish-gray claystone; local sandstone and limestone	180±

Entrada(?) Sandstone (Middle Jurassic)	
Gray crossbedded sandstone	0-45
Lykins Formation (Triassic? and Permian)	
Red and variegated siltstone and sandstone	60±
Maroon Formation (Permian and Pennsylvanian)	
Red, pink, and gray arkosic sandstone and conglomerate	30±
Precambrian rocks	

Northeast Flank of Eagle Basin
Minturn quadrangle (Tweto, 1977)

Dakota Sandstone (Upper Cretaceous)	
Medium-bedded to massive gray sandstone; shale at top	45-49
Morrison Formation (Upper Jurassic)	
Gray sandstone, green, gray, and purple shale, gray limestone	76
Entrada Sandstone (Middle Jurassic)	
Massive, crossbedded, buff to orange sandstone	18
Chinle Formation (Upper Triassic)	
Red and purple siltstone, mudstone, and sandstone	21
Maroon Formation (Lower Permian, Upper and Middle Pennsylvanian)	
Red sandstone, siltstone, grit, and conglomerate	520-1,280
Minturn Formation (Middle Pennsylvanian)	
Grit, conglomerate, sandstone, and shale, and some intercalated limestone and dolomite. Predominantly gray, but red in upper part and near base.	600-1,900
Belden Formation (Middle Pennsylvanian)	
Gray to black shale, limestone, and minor sandstone	0-60
Molas Formation (Lower Pennsylvanian)	
Green, yellow, or brown regolithic clay	
Leadville Limestone (or Dolomite) (Lower Mississippian)	
Gray limestone or dolomite	0-45
Chaffee Group (Lower Mississippian? and Lower Devonian)	
Gilman Sandstone	
Yellowish-gray sandstone, sandy and cherty dolomite, and breccia	0-15
Dyer Dolomite	
Thin bedded gray dolomite	0-25
Parting Formation	
White to tan sandstone and conglomerate; subordinate green shale	0-20
Harding Sandstone (Middle Ordovician)	
White, gray, and green sandstone and green shale	0-25?
Peerless Formation (Upper Cambrian)	
Brown, red, green, and buff sandy dolomite and dolomitic shale	0-21
Sawatch Quartzite (Cambrian)	
Medium-grained white quartzite	0-67
Precambrian rocks	

Southwest Flank of the Eagle Basin
Aspen region (Bryant and Freeman, 1977)

Mesaverde Formation (Upper Cretaceous)	
Light-gray, fine-grained sandstone	40
Mancos Shale (Upper Cretaceous)	
Upper shale member	1,450
Fort Hays Limestone Member	12
Lower shale member	122
Dakota Sandstone (Upper Cretaceous)	
Yellowish-gray, white, light-brown, and gray fine- to medium-grained sandstone with dark gray shale interbeds.	40-60
Burro Canyon Formation (Lower Cretaceous)	
Light-gray to white, fine- to medium-grained sandstone; chert and quartz pebble conglomerate, and light-bluish-gray claystone and siltstone.	40-60

Morrison Formation (Upper Jurassic)	
Grayish-red and greenish-gray claystone that grades downward into calcarous siltstone, sandstone, and yellowish-gray-weathering silty limestone and limestone.	90-160
Curtis Formation (Upper Jurassic)	
Oolitic limestone	3
Entrada Sandstone (Middle Jurassic)	
Yellowish-gray medium- to coarse-grained sandstone	0-50
Chinle Formation (Upper Triassic)	
Red siltstone, calcareous siltstone, and limestone; limestone-derived sandstone and conglomerate.	0-300
State Brudge Formation (Lower Triassic and Permian)	
Grayish-red to moderate-brown and orange siltstone, sandstone, and conglomerate.	0-800
Maroon Formation (Permian and Pennsylvanian)	
Grayish-red, pale-red, reddish-brown, and moderate-red siltstone, sandstone, and conglomerate and minor beds of gray limestone and evaporite.	3,500-4,600
Eagle Valley Formation (Middle Pennsylvanian)	
Greenish-gray, grayish-red, and reddish-brown calcarous siltstone, claystone, and sandstone; silty limestone and evaporite beds.	0-750
Gothic Formation of Langenheim (1952) (Middle Pennsylvanian)	
Light- to dark-gray, yellowish-gray, and brown sandstone, limestone, siltstone, and shale and minor white evaporite beds.	300-900
Belden Formation (Middle and Lower Pennsylvanian)	
Dark-gray to black limestone, dolomite, shale, and carbonaceous shale; minor beds of white evaporite.	150-450
Leadville Limestone (Lower Mississippian).	
Castle Butte Member. Blue-gray thick-bedded limestone	
Red Cliff Member. Gray to dark-gray, thin- to thick-bedded dolomite	30-60
Chaffee Group (Lower Mississippian and Upper Devonian)	
Gilman Sandstone	
Gray dolomitic sandstone, sandy dolomite, and sedimentary breccia.	1.2-4.6
Dyer Dolomite	
Light- to dark-gray thin- to thick-bedded dolomite and minor limestone.	15-36
Parting Formation	
White to yellowish-gray quartz sandstone, gray to yellowish-gray siltstone, gray, greenish-gray, and dusky-red shale and grayish-yellow dolomite.	15-30
Manitou Dolomite (Lower Ordovician)	
Gray, medium-bedded dolomite	50-85
Peerless Formation (Upper Cambrian)	
Grayish-orange to moderately orange-brown-weathering dolomitic sandstone and sandy dolomite and dusky-red to yellowish-gray and light-greenish-gray shale.	30-45
Sawatch Quartzite (Upper Cambrian)	
White quartzite; thin beds of quartz pebble conglomerate at base	45-75
Precambrian granite and gneiss	

Front Range to the west and the Las Animas arch to the southeast. It contains about 1 km of upper Cretaceous and lower Tertiary orogenic sedimentary rocks derived from the Front Range and uplifts to the west. North, Middle, and South Parks are parts of a structural low between the Sawatch anticline and the Front Range; North Park is a structural low between the main Front Range and its northwestern extension, the Park Range. These lows locally preserve more than 2,700 m of variously deformed Paleocene and Eocene orogenic sediments that rest unconformably on Mesozoic and older rocks.

THE COLORADO MINERAL BELT AND IGNEOUS ACTIVITY

The Colorado mineral belt is an irregular

and early Tertiary time (Lipman and others, 1972; Cross and Pilcher, 1978). The early Laramide (70-55 Ma) igneous activity was suppressed by thick cratonic lithosphere except along existing zones of weakness (Lipman, 1980). After a decline 55-40 Ma, renewed igneous activity occurred on a large scale blanketing much of the southern Rocky Mountains. This was part of a regional sweep south of igneous activity and west across the Cordillera perhaps related to foundering of the descending Farallon plate as its size and thickness and rate of convergence lessened (Lipman, 1981). Intermediate composition rocks tended to dominate during the earlier part of this activity, and compositions became more silicic with time. Magma became more silicic during the transition from a tectonic regime of subduction to one of rifting (Bookstrom, 1981). Some very large molybdenum deposits, such as the one at Climax, formed during the emplacement of composite stocks of these silicic magmas.

After about 26-20 Ma volcanism became predominantly basaltic or bimodal basalt-rhyolite, an association widespread in much of the western United States. This change reflects increasing interaction between the Pacific and American plates along the San Andreas transform fault (Lipman, 1981).

POST-LARAMIDE TECTONICS

Following the close of the Laramide orogeny in the middle or late Eocene the uplifts were largely reduced to a surface of low relief that beveled all of the Laramide structures. Paleo-valleys in this surface are filled by volcanic rocks as old as 36 Ma, so it can be no younger than earliest Oligocene. It is termed the late Eocene surface (Scott, 1975; Epis and Chapin, 1975). Floras and faunas in lake beds at Florissant in South Park indicate that the surface there was at an altitude of no more than 1,000 m in the Oligocene (MacGinitie, 1953). Large areas of this surface were covered by volcanic rocks in Oligocene time. The surface has been much disrupted by Miocene and younger faulting and regional uplift associated with the formation of the Alvarado Ridge (Eaton, 1986). Major Neogene structures include the Rio Grande rift and the Arkansas graben (a northern extension of the Rio Grande rift), and the faults along the western side of the Blue River Valley. Some of these structures show Holocene movement (Tweto, 1979a). Parts of the late Eocene surface have been uplifted as much as 3,000 m since the Oligocene, and much of the present topographic relief is the result of differential uplift and dissection of this surface. The importance of post-Laramide uplift in the shaping of the present mountain landscape is illustrated by the fact that of the 54 peaks in Colorado more than 4,267 m (14,000 ft) high, all but two lie either along the crest of the Alvarado ridge on the flanks of the Neogene Rio Grande rift-Arkansas graben or along the low-density composite batholith of the Colorado mineral belt, and that four of the five highest lie near the intersection of these two features.

ROAD LOG

First Day, Monday, July 3.

Road log begins at the intersection of Colfax Avenue and Indiana Street by the Holiday Inn. Go west on Colfax Ave. Basement rocks in the deepest part of the Denver basin are at 2.5 km below the surface only 4 km east of here. Fission-track dating of apatite on Mt. Evans, a 4,347-m peak east of the Continental Divide 40 km southwest of here indicates a total structural relief between basin and uplift of about 6.5 km (Bryant and Naeser, 1980).

Distance (kilometers)

2.2 Cross U.S. Highway 6. On Table Mountain at right potassic basalt (shoshonite) lava flows dated at about 64-65 Ma (Scott, 1972) overlie andesitic and arkosic sandstone and siltstone of the Denver Formation. Cretaceous-Tertiary boundary is about 40 m below the flow at the east end of the mesa.

3.2 Turn left on Rooney Road (Jefferson County Highway 53).

4.5 Cross Interstate Highway I-70.

4.8 On the left, white sandstone of the Cretaceous Fox Hills standing nearly vertical along the Golden fault zone. The Fox Hills, a delta front sandstone, interfingers with the Pierre Shale, a prodelta shale (Weimer and Tillman, 1980). These rocks were deposited near the shore of the Cretaceous sea 69 Ma just before the uplift of the Laramide Front Range. They are overlain by delta plain sandstone, siltstone,

here encountered the Dakota sandstones at four distinctly separate intervals, twice right-side up and twice upside-down (Fig. 6) (Smith, 1964). Apparently the uniform monocline exposed at the surface changes at depth to steeper and overturned strata cut by reverse faults. This is an excellent example of the vertical variations in structural style typical of Rocky Mountain uplifts. Similar variations in structural style occur along the strike of fault zones along the margin of the uplift.

To the north the Fountain Formation is cut off and the margin of the Proterozoic rocks offset by the Cherry Gulch fault, a moderate angle reverse fault having a displacement of only a few hundred meters.

Along the mountain front alluvium at several levels caps pediment surface. In the distance to the south the prominent surface is the Verdos Alluvium, which contains a 0.6 Ma volcanic ash.

12.8 Return to Colorado Highway 26 and turn left.

14.8 Cross under Interstate Highway 70; Dakota hogback to right; lower part of slope underlain by Jurassic Morrison Formation.

17.1 Turn left on Jefferson County Highway 93.

18.7 Turn left on U.S. highway 6. On right claypits in the Laramie Formation; Dakota Group to the left. The rest of the Cretaceous section is cut out along the Golden fault.

22.5 Enter Clear Creek Canyon and the Precambrian core of the Front Range uplift. Here the Fountain Formation is cut out almost entirely along the Golden fault.

Migmatitic biotite gneiss forms the lower part of Clear Creek Canyon. Hedge (1972) showed that the migmatites had no direct spatial relation to plutons of the Routt Plutonic Suite and that they were older than those plutons. Olsen (1982) found that some of the migmatites in Clear Creek Canyon formed by partial melting in a closed system, whereas others required some introduction of granitic or tonalitic material from external sources. She estimated that the migmatites formed at 4 kb and 650-700°C. Near the canyon some porphyroblastic orthopyryoxene has been found in biotite-hornblende gneiss, suggesting conditions transitional between upper amphibolite and granulite facies. About 10 km north of the canyon pelitic schist contains muscovite and undeformed porphyroblasts of andalusite. Swaze and Holden (1985) calculate a temperature gradient of 180°/kb and suggest that heat must have been added to the terrane from some external source, such as the Boulder Creek batholith to explain such low pressure and high temperature metamorphism.

The map pattern of major rock units and the pattern of minor folds suggests that an early layer parallel isoclinal folding may make the apparent lithologic sequence have little stratigraphic significance. These folds are deformed by later, more open folds.

Quartz-plagioclase-microcline pegmatites, some of which contain biotite and magnetite, range from semiconcordant to discordant and were emplaced during and after the ductile deformation. The lack of tourmaline and muscovite in the pegmatites and the fact that some are deformed suggests that they belong to the Early Proterozoic Routt Plutonic Suite.

Younger, discordant dikes of hornblende lamprophyre cut the Precambrian rocks.

Along Clear Creek Canyon the foliation and layering and the trend of the rock units is subparallel to the overall trend of the canyon, but in detail the canyon winds back and forth across this relatively uniform trend. Dips are steep to moderate and mostly to the south. Zones of fractured, reddish weathering rock mark northwest-trending fault zones of Precambrian ancestry.

28.5 Quarry in fractured rocks along the Junction Ranch fault zone. During mining a cave containing typical calcium carbonate cave deposits was uncovered. The cave was formed by gravity separation of blocks along fractures; the calcium carbonate probably came from layers of calc silicate rock and marble in the metamorphic sequence.

32.2 This part of Clear Creek Canyon lies on a two mile wide transition zone between predomiantly metasedimentary biotite gneiss mapped to the north from predominantly metavolcanic felsic gneiss and amphibolite mapped to the south. The units are averages of the proportions of the interlayered rock types; contacts are transitional.

39.4 Gold placer workings in Clear Creek on left.

41.0 Turn right on Colorado Highway 119. From here we head northwest and cross a large northeast-trending synform containing superposed-northwest trending folds. The synform is cored by a unit rich in biotite-quartz-plagioclase-microcline gneiss and biotite-quartz-plagioclase gneiss containing layers of biotite gneiss, amphibolite, hornblende gneiss, and calc silicate gneiss.

50.2 Sillimanitic biotite gneiss forms a unit 1.6 km thick. Rocks here display a younger generation of northeast-trending structures formed under both ductile and brittle conditions along the Idaho Springs-Ralston shear zone, one of the northeast-trending Precambrian shear zones thought to furnish a regional control on the emplacement of Late Cretaceous and Tertiary intrusions and associated ore deposits of the Colorado Mineral Belt (Tweto and Sims, 1963).

52.1 Enter Blackhawk. Placer mining operations along Clear Creek were worked extensively during the Great Depression years of the 1930's, and several have been worked recently for gold on bedrock.

33.1 Turn left on Colorado Highway 279 up Gregory Guch to Central City. Gregory Gulch was the site of the first lode discovery in Colorado in 1859. Several mine openings, largely inaccessible, are visible from the road.

34.0 Central City.

CENTRAL CITY AND IDAHO SPRINGS DISTRICTS, FRONT RANGE, COLORADO

P.K. Sims
U.S. Geological Survey, Denver, Colorado

INTRODUCTION

History

The Central City and Idaho Springs districts represent the earliest mining activity in the state of Colorado. Placer gold was discovered in Clear Creek, near Idaho Springs, in January 1859, and shortly afterwards lode gold was discovered at a site midway between the present towns of Central City and Black Hawk. These early discoveries soon led to an influx of prospectors and miners, and eventually to the establishment of several villages and towns in the region, some of which remain today. Until the middle 1880's, the metal output of the districts exceeded the combined total of all other mining districts in Colorado. Production has declined sharply since 1914, and it has been negigible since the early 1930's. Gold, silver, copper, zinc, lead, and uranium were mined from the districts. Gold has accounted for about 80 percent of the value of the ore and silver for about 10 percent.

The early production exploited placer gold and oxidized lode gold ores. Rich silver veins were not discovered until 1877. At depths of 30-40 m or less, the oxidized gold ores gave way to primary sulfides, which could not be treated profitably by the simple amalgamation processes then used. Consequently, mining activity was drastically curtailed until the late 1860's, when milling methods were developed to satisfactorily treat the ores.

Mining and processsing

The ores in the districts were mined principally by shafts and open stopes, but as mining progressed to greater depths the handling of water and hoisting became a serious problem. As a result, around the turn of the century long tunnels were driven north from the Clear Creek drainage system to intersect the veins at depths of 300 to 500 m. The principal tunnel, the Argo (or Newhouse), was started at Idaho Springs and extended nearly to the town of Central City, a distance of nearly 8 km. Although it intersected several of the major veins in the Central City district at depth, it did not lead to a substantial increase in production.

Gold in the oxidized ores was extracted by using such simple devices as sluices, cradles, arrastras, and crude stamp mills. Attempts to process the primary gold-bearing sulfide ores by amalgamating the gold were largely unsuccessful, and it was not until metamorphic sequence.

32.2 This part of Clear Creek Canyon lies on a two mile wide transition zone between predomiantly metasedimentary biotite gneiss mapped to the north from predominantly metavolcanic felsic gneiss and amphibolite mapped to the south. The units are averages of the proportions of the interlayered rock types; contacts are transitional.

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1868 when the Hill smelter was opened in Black Hawk that these ores were successfully processed.

Two types of primary ores were mined in the districts: (1) "mill dirt," which required some form of concentration or amalgamation, and (2) "smelting" ore, which was higher grade and could be processed by smelting directly. The mills in operation since 1900 first employed amalgamation-concentration, and then, more recently, used gravity concentration and selective flotation (Bastin and Hill, 1917). In the early 1900's most of the ore was treated at the American Smelting and Refining Company plant at Leadville, Colo. Some zinc concentrates have been shipped to the company's plant at Amarillo, Texas.

Production

Ores have been produced from about 500 mines in the Central City district, but a substantial proportion came from relatively few large mines. Recorded production for the Central City district and mines in adjacent northern Gilpin County from 1859 to 1914 is about 70,000 metric tons of lead, 4,300 metric tons of copper, 8,700 tons of zinc, 28,350 kg of gold, and 155,630 kg of silver (Bastin and Hill, 1917, p. 174). Small amounts of uranium ore have been produced also. The production from the Idaho Springs district is about two-thirds that of the Central City district, with silver being more important than gold.

GEOLOGY

General geology

The country rocks in the area are mainly gneiss, migmatite, and intrusive rocks of Early Proterozoic age. The lithologic succession is dominantly an interlayered sequence of two principal rock types, biotite-quartz gneiss (commonly called Idaho Springs Formation) and a microcline-quartz gneiss. The protoliths of the gneisses were volcanic and sedimentary rocks of oceanic-arc affinity. The maximum exposed thickness of the succession in this area is less than 5 km. Older bodies of granodiorite and quartz diorite (ca. 1,700 Ma) and younger bodies of two-mica granite (ca. 1,400 Ma) intrude the layered rocks. Throughout the area, Proterozoic rocks are deformed into broad, mainly open anticlines and synclines whose axes trend north-northeast and are spaced one to two kilometers apart. In the southeast part of the area, a younger ductile deformation was superposed on the regional deformation pattern. This zone of deformation--folds, mylonites, and cataclasites--about 2 km wide, is part of a family of Precambrian shear zones that coincide approximately with the northeast-trending Colorado mineral belt. Tweto and Sims (1963) concluded that these shear zones were the dominant factor in localizing the Laramide and Tertiary intrusions (see below) and accompanying mineral deposits that constitute the mineral belt.

The Laramide and Tertiary igneous rocks are hypabyssal intrusions of calc-alkalic and alkalic affinities. They form generally small plutons and dikes.

The country rocks in the area are cut by abundant faults, some of which are Early Proterozoic structures that were reactivated during the Laramide orogeny. Most of the faults, however, are Laramide in age, having formed shortly before the mineralization.

Ore deposits

The ore deposits in the area are sulfide-quartz veins that contain precious metals, base metals, and sparse uranium. They were formed at about 60 Ma, and are related to the associated porphyritic igneous rocks. The deposits constitute the classic Laramide mining districts at Central City and Idaho Springs.

The veins are hydrothermal fillings in faults, and are similar in mineralogy, texture, and structure to the deposits classified by Lindgren (1933) as mesothermal. The principal ore minerals are pyrite, sphalerite, galena, chalcopyrite, and tennantite. Less abundant are enargite, telluride minerals, and molybdenite (Rice and others, 1985). The gangue minerals include quartz, barite, fluorite, and peripheral Ca-Mg carbonates.

Four stages of mineralization have been recognized (Sims and others, 1963): (1) uranium, (2) pyrite, (3) base metal, and (4) telluride. Most of the precious metals are associated with stages 2 and 3. Rice and others (1985) further recognized an early molybdenite-stage, which formed early in the paragenesis, and a later molybdenite stage with coeval fluorite, which postdates the base-metal stage mineralization.

The pyrite-base metal ores of the area have a well-defined concentric hypogene mineral zoning. A large central zone containing pyrite-quartz veins is surrounded by an area (intermediate zone) of pyrite-type veins that carry copper, lead, and zinc minerals (Fig. 7). Still farther out are areas (peripheral zone) containing predominantly galena-sphalerite-quartz-carbonate veins.

Rice and others (1985) have outlined an elliptical area of molybdenite- and fluorite-bearing veins that lies athwart the Dory Hill fault (Fig. 7) on the eastern margin of the Central City district. The molybdenite and fluorite zones are spatially distinct from the concentric pyrite-base metal zonation, although apparently of the same age (Rice and others, 1982).

The wall rocks of the fissure veins are altered to successive zones of sericitized and argillized rock (Tooker, 1963) which is restricted to vein margins.

Fluid inclusion data. Fluid inclusion studies by Rice and others (1985) indicate that primary inclusions associated with the precious metal-bearing pyrite and base metal veins are two-phase liquid vapor aqueous inclusions with homogenization temperatures in the range 220°C to 380°C. For early molybdenite veins, the fluid inclusions are of two types: CO_2-rich (common) and halite-bearing (rare), with respective homogenization temperatures of 340° to 420° C and 240° to 340° C. Those from late molybdenite veins are aqueous two phase inclusions with homogenization temperatures of 200° to 280° C. Salinities (as equivalent wt. percent NaCl) are between 32 to 42 for halite-bearing inclusions and 2 to 12 for all other types. No evidence of boiling was found by Rice and others (1985), but J. LeAnderson found evidence of boiling

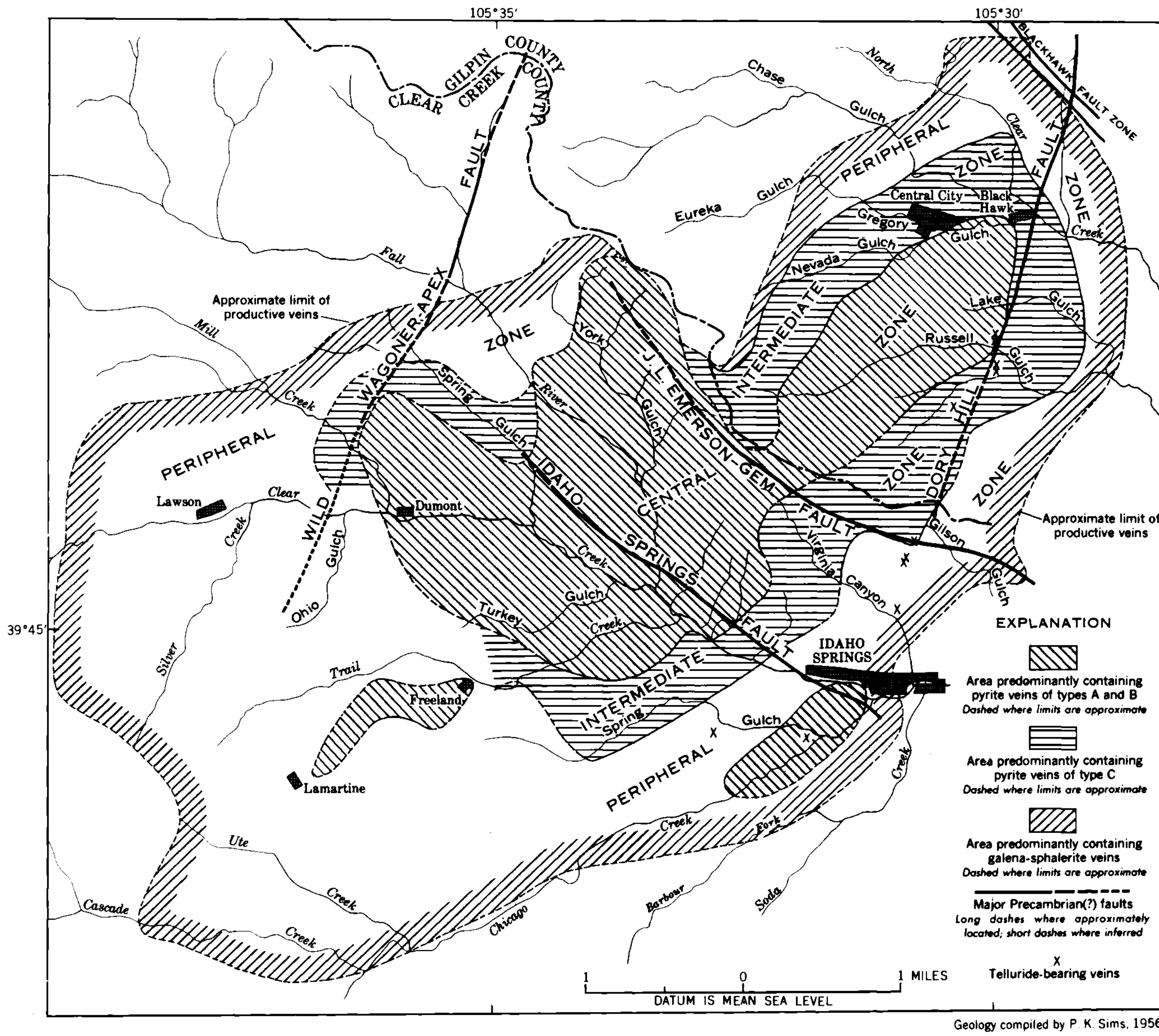

FIGURE 7 Map showing zonal arrangement of ores in the Idaho Springs-Central City area. From Sims and others (1963).

at the Phoenix mine at Idaho Springs (Alan Wallace, written commun., 1988). Lithostatic trapping pressure estimates of CO_2-rich inclusions are 0.46 to 0.75 kb, which suggest an emplacement depth of about 2 km for the early molybdenite mineralization (Rice and others, 1985).

Stable isotopes. The Precambrian country rocks have $\delta^{18}O$ and δD values respectively of +6.1 to 12.0 per mil and -100 to -59 per mil (Rice and others, 1985). The absence of negative $\delta^{18}O$ values indicates that the country rocks were not affected appreciably by large-scale hydrothermal interaction with meteoric ground waters. Thus, the mineralizing fluids were mainly confined to the discrete fractures (veins) and did not permeate the rocks.

For the veins, quartz $\delta^{18}O$ values range from +8.4 to +15.9 and fluorite fluid inclusion δD values range from -90 to -47. A distinct δD high is centered directly over the central zone (Fig. 7) in the Central City district, indicating that the ore-forming fluid here was nearly totally magmatic. Lower δD values in the early molybdenite veins reflect the effect of mixing of such a magmatic fluid with evolved meteoritic waters that occupied the fractures in the earliest stage of hydrothermal activity. The δD values of ore fluids in the intermediate and peripheral zones of the district are lower than in the central zone, indicating a significant (50-90 percent) amount of evolved meteoric water which mixed with the magmatic ore fluid.

Genetic model. Rice and others (1985) proposed that the molybdenum, uranium, precious and base metal, and telluride mineralization are genetically related and may represent the upper part of a Laramide (~60 Ma) alkaline porphyry molybdenum system. The inferred parent magma is a quartz bostonite or alkali rhyolite. Emplacement of the inferred parent magma was controlled by the intersection of the Dory Hill fault and the northeast-trending Idaho Springs-Ralston shear zone (Tweto and Sims, 1963). Early molybdenite mineralization was deposited at a temperature of about 390°C from CO_2-rich, moderately saline, mixed hydrothermal-meteoric fluids at a depth of about 2 km. Later pyrite-stage minerals were deposited from magmatic fluids, but the ore fluids of later stages contained varying amounts of meteoric waters due to recurrent movements along the fractures.

Alternatively, the telluride mineralization could represent a later mineralizing event that was superposed on the eastern part of the main ore mineralization, because K-Ar ages on sericite in this zone are as young as 52 Ma (Rice and others, 1982, Fig. 4 and Table 1) and the mineralogic and isotopic zonations are spatially distinct.

Figure 8 shows the route of the trip through the Central City district and the location of stops.

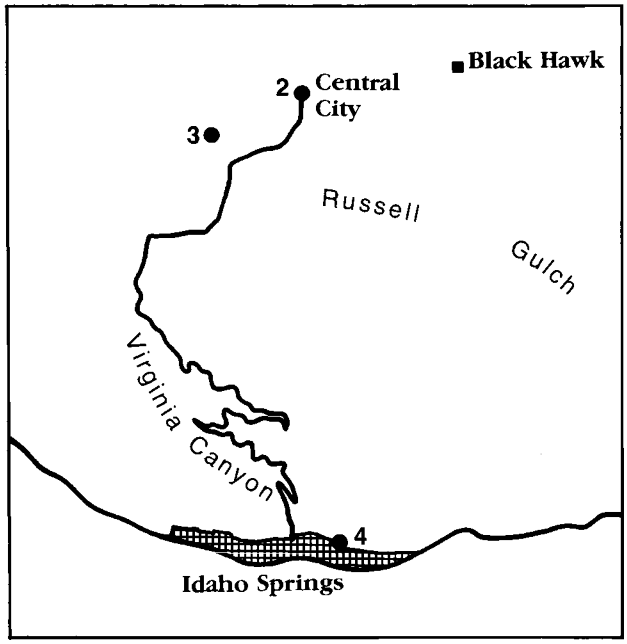

FIGURE 8 Map of Central City-Idaho Springs area showing route and field trip stops.

55.8 STOP 2. An overlook of the main mining area provides a perspective of the density of the veins and the nature of past mining. Quartz Hill, visible to the southwest, contains many of the more productive veins, including "The Patch," which is on the top of the hill. These veins are in the intermediate zone of the district zoning pattern, which consists of composite veins containing the richest gold and base metal assemblages.

Continue south on highway to mill site.

56.8 STOP 3. Walk up gentle road to "the Glory Hole," also called "the Patch." This is private property; permission must be obtained.

The Patch is an oval-shaped pipelike breccia that plunges steeply to the north. It is localized near the intersection of several veins, along the California-Gardner-Mammoth-Fisk structure, a persistent and highly productive vein zone that extends the length of the district. The mineralized breccia is described in Sims and others (1963). The patch is the only lode deposit in the district that has been mined successfully by bulk surface methods.

58.2 Entering village of Russell Gulch.

60.5 Virginia Canyon Road; keep left on county road 279, locally called "Prospectors Trail". This is the approximate boundary between the Central City and Idaho Springs districts. At the beginning of the descent into Clear Creek Canyon good views of Mt. Evans (4,347 m) and peaks on the crest of the Front Range to the west. The road winds down through the heavily mineralized country of the Idaho Springs district.

65.2 Virginia Canyon road; bear left.

68.2 Turn left at bottom of hill.

69.0 Left on Business I-70 in Idaho Springs.

69.7 STOP 4 at park opposite Argo mill. The Argo tunnel is the largest and most ambitious mining effort in the Central City-Idaho Springs district. Started in 1893 and completed in 1910, the tunnel extends 7.36 km northward under Quartz Hill. The tunnel was driven to intersect numerous veins in the district at depth greater than the deepest shafts, to dewater existing mines, and to provide easy transportation of ore to the mills in Idaho Springs. It was a bold concept, but the veins that were penetrated were generally subeconomic, and the operation was only marginally successful. The tunnel was operated intermittently until January 19, 1943 when drill holes penetrated water-filled stopes of the Kansas mine on Quartz Hill, triggering an enormous flood in which four men drowned. The tunnel has never been reopened. The pH of the water draining the tunnel is less than 1.0.

Continue west on Business I-70 through Idaho Springs and rejoin I-70 westbound at Exit 249. Outcrops here are fractured and iron-stained felsic gneiss in the northeast-trending Idaho Springs-Ralston shear zone.

For the next 10 km exposures are chiefly interlayered felsic gneiss, migmatitic biotite gneiss, and amphibolite, all cut by bodies of gneissic granitic rocks of the Routt Plutonic Suite and by dikes of Silver Plume Granite and related pegmatite. Outwash gravels form prominent terraces along Clear Creek in several places.

73.6-82.8 Pass through Lawson-Dumont-Fall River mining district, actually part of the Idaho Springs-Central City mineralized area (Fig. 7). Silver, lead, zinc, and gold were produced from narrow fissure veins occurring along minor faults (Hawley and Moore, 1967)

83.9 Exit 232. U.S. 40 turns up West Fork of Clear Creek toward Berthoud Pass and the Henderson molybdenum mine. Route continues on I-70 up main stem of Clear Creek.

Just beyond the interchange is the poorly defined terminal moraine of the late Pleistocene glacier in the Clear Creek valley. The terminal moraine in the West Fork is visible a few hundred meters to the west. The valley from here to Georgetown is chiefly in biotite gneiss cut by myriad dikes and irregular intrusive bodies of Silver Plume Granite. Landslide features are conspicuous in several places, particularly on the east side of the valley above the reservoir near Georgetown.

90.2 Georgetown. Entering Georgetown-Silver Plume mining district. High roadcuts on north side of I-70 along the steep grade above Georgetown are layered biotite gneiss and amphibolite cut by a network of dikes of Silver Plume Granite and pegmatite.

93.5 Exit 226 to Silver Plume. The Georgetown-Silver Plume district produced silver, lead, zinc, and minor gold from veins cutting Early Proterozoic schist and gneiss and Middle Proterozoic granite. The quantity of metals produced is not precisely known, but the value of production was about 15% of that from the Central City-Idaho Springs district. The principal discoveries were made in the 1860's, and the main period of production was in the 1870's and 1880's. After the demonetization of silver in 1893, zinc production became important. Little ore has been mined here since 1918 except during World War II.

The main productive veins trend northeast and west-northwest parallel major Precambrian fault zones. Dikes of leucocratic rhyolite-granite B porphyry 37 to 40 Ma are closely associated with the veins, which are interpreted as distal epithermal, magmatic hydrothermal veins related to a very large probably composite granite B batholith at depth. (For summary of district see A.A. Bookstrom in Reed and others, 1988).

At the west outskirts of Silver Plume is a quarry which is the type locality for the Silver Plume Granite, one of the Middle Proterozoic peraluminous granites that occur over a broad area on the North American craton (Anderson, 1983). Here the granite has been dated at 1409±40 Ma by Rb/Sr whole rock method (Hedge, 1969; corrected using presently accepted constants). This is the most widely known of the 1.4 Ga anorogenic granites in Colorado, for which Tweto (1987) has given the name Berthoud Plutonic Suite. Reported K-Ar ages of biotite and hornblende in Early Proterozoic metamorphic rocks from the Front Range are mostly younger than 1.4 Ga, indicating regional cooling after emplacement of rocks of the Berthoud plutonic suite.

95.1 Cross Browns Gulch, the site of a suburb of Silver Plume called Brownsville. Several distastrous debris flows composed of mine dump and colluvial material swept into the settlement. In 1912 a large flow wiped out the remaining structures in the village. Valley above here is largely in biotite gneiss cut by many intrusive bodies of Silver Plume, but exposures are poor at road level. Several conspicuous avalanche tracks scar the valley walls.

101.6 At Bakerville exit, brief glimpse of Torreys Peak (4,348 m) on the left. Torreys and its neighbor Grays Peak (4,349 m) are the highest peaks in the Front Range and are the only two fourteen thousand foot (4,267 m) peaks in Colorado whose summits lie on the Continental Divide.

About 3 km beyond Bakerville, enter a wide belt of fractured and sheared rocks along the north-northeast-trending Loveland Pass-Berthoud Pass fault zone. Outcrop is spotty in the lower valley, despite being in the interior of the Silver Plume batholith. Mount Bethel, the symmetrical peak north of the road is largely fractured Silver Plume. Snow fences on the ridge and diversion structures at the mouth of the prominent chute are attempts--not always successful--to prevent avalanches from blocking I-70.

109.2 Pass Loveland Ski Area and enter east portal of the 2.5 km-long Eisenhower Tunnel that carries I-70 under the Continental Divide. This is said to be the highest highway tunnel in the world. It passes through fractured and sheared biotite gneiss and Silver Plume Granite in the Loveland Pass-Berthoud Pass fault zone. The broken rock and the large diameter (necessary to provide adequate ventilation at an altitude of 3,400 m) led to delays and major cost-overruns during construction of the first bore.

Outcrops on the north at the west portal are near the western contact of the Silver Plume batholith. The valley of Straight Creek is controlled by a zone of faults that branches southwest off the Loveland Pass-Berthoud Pass fault zone. Rocks in the roadcuts are chiefly biotite gneiss; many landslides in till and along south-dipping foliation planes in gneiss plague maintenance of this section of I-70. Views of the Gore Range and Blue River valley ahead.

About 10 km west of the tunnel, near the bottom of the grade, road crosses the Williams Range thrust, passing from broken Precambrian rocks in the upper plate into Upper Cretaceous Pierre Shale in the footwall. This low-dipping thrust forms the west margin of the Laramide Front Range uplift. It is much longer and more continuous than the Golden fault, and map relations show a minimum of 8 km of westward displacement of the upper plate just south of here. About a 1.6 km beyond the fault, Dakota Sandstone is exposed in a large roadcut on the right.

123.1 Exit 205 to Dillon and Silverthorne. View of the Gore Range and Blue River valley on the right; Dillon Dam on the left.

The Blue River valley follows a belt of sedimentary rocks that is continuous along the west side of the Front Range uplift. The valley is a Neogene half-graben related to and en echelon with the north end of the Arkansas graben. There has been at least a kilometer of Neogene uplift of the east flank of the Gore Range along the Blue River fault on the west side of the half-graben. The Gore Range is composed of migmatite and granitic rocks of the Routt Plutonic Suite.

Yellowish gravel exposed across small valley on the right about a 1.5 km after crossing the Blue River is probably equivalent to the Miocene and Pliocene Dry Union Formation of the upper Rio Grande rift. To the north the downthrown side of the half-graben is almost concealed by glacial deposits, but a few exposures suggest the presence of numerous fault blocks of other Miocene sediments and Oligocene volcanic rocks.

127.7 Scenic view turnoff. STOP 5. Overview of Middle Park, west flank of Front Range, and east side of Tenmile Range. Precambrian rocks are thrust over Cretaceous shale along the slope across the lake above the highest breaks in the forest. To the south Mt. Guyot (3,983 m) and Bald Mountain (4,146 m) are formed by late Eocene intrusions in Mesozoic sedimentary rocks west of the Front Range uplift and separate Middle Park, which we are in, from South Park. Rocky slopes to south are the north end of the Tenmile Range, which is part of the same tectonic block as the Gore Range, which is behind us and out of sight. The west side of the valley is bounded by the Blue River fault, a major element of the rift system at this latitude. Patches of upper Tertiary Dry Union Formation are brought against Precambrian rocks along that fault.

The area is on the west margin of the Front Range highland, a range of the Ancestral Rockies that was uplifted in Pennsylvanian and Permian time. Upper Jurassic rocks lie on Precambrian basement to the southeast; near Dillon dam 30 m of redbeds of late Paleozoic or early Mesozoic age separate Jurassic rocks from the basement.

The low hills near the reservoir are late Pleistocene moraines.

The lake (2,751 m) is Dillon Reservoir, a part of the water system serving the Denver urban area. A 35 km long tunnel through the Front Range divide brings water from the reservoir across the continental divide into the South Platte River, which flows through Denver.

132.1 Exit 201 to Frisco. Enter the canyon of Tenmile Creek, which forms the physiographic boundary between the Gore Range to the west and the Tenmile Range to the east. The structural boundary between the ranges is formed by the Mosquito fault zone lying less than 1 km to the west.

In Tenmile Canyon the Precambrian rocks consist of layered hornblende and biotite gneiss and amphibolite cut by numerous light-colored anastomosing pegmatite dikes. Many of the gneisses are probably metamorphosed volcanic and volcaniclastic rocks similar to those preserved in less metamorphosed and deformed sequences near Salida and Gunnison. The rocks here locally contain small amounts of orthopyroxene coexisting with hornblende and calcic plagioclase suggesting metamorphic grade close to the amphibolite-granulite facies transition. Other rock types exposed in the canyon are migmatitic biotite gneiss and quartz-feldspar gneiss.

The bike path on the south side of Tenmile Creek follows the grade of the narrow-gauge Denver, South Park, and Pacific Railroad, which operated from 1884 to 1938. On the north side of the creek the narrow gauge line of the Denver and Rio Grande operated from 1882 to 1925. Competition for the business of the booming mining towns in the mineral belt in the 1880's was intense.

A scenic turnout at the head of the canyon lies nearly on the trace of the Mosquito fault; mylonitic rocks are exposed on both sides of the valley. The valley south of here is nearly parallel to the trace of the fault, which separates Precambrian rocks like those exposed in the canyon on the east from Pennsylvanian and Permian strata on the west. Exploration at the Climax molybdenum mine showed that the fault there dips 70^{o} W. and has a vertical displacement of 2,700 m and a left-lateral displacement of 450 ft.

138.7 Exit 195. Wheeler junction and Copper Mountain ski area.

143.5 To left Jacque Peak and east-dipping strata on skyline to west. Base of Maroon Formation of Pennsylvanian and Permian age marked by the Jacque Mountain Limestone Member of the Middle Pennsylvanian Minturn Formation well below skyline. Numerous Tertiary sills intrude the sequence. Exposures of red sandstone and conglomerate to right are in a downdropped block along the Gore fault, which we crossed in an area covered by surficial deposits.

146.9 Summit of Vail Pass; turnoff to Shrine Pass.

148.3 Turnoff to right. STOP 6 (optional). View of Gore fault and west side of Gore Range. Gore fault lies at about break in slope uphill to east. To north it is left of the peaks above timberline. Here the attitude of the fault is unknown, but to the north it dips northeast as little as 45^{o}. The Gore fault has had a long history of movement dating back to Precambrian time. In Pennsylvanian and Permian time it was active and helped define the west margin of the Front Range highland (Tweto and Lovering, 1977). It was active in Laramide time and the local overturning of the strata adjacent to the fault probably dates from that movement episode. North of the high peaks Miocene basalt is offest along the fault indicating Neogene displacement.

The highway descends into the valley of Gore Creek through Pennsylvanian redbeds of the Minturn Formation. At the time these rocks were deposited the Gore fault was not the margin of the basin, for sparse intercalated marine limestone maintain their identity right up to the fault. However, the thickness of the Minturn Formation decreases towards the fault; the margin of the Front highland at that time was to the east.

160.9 Cross Gore Creek. Enter the resort complex of Vail.

165.5 Mouth of Booth Creek on right. A debris flow from this valley damaged several houses in 1984. Urbanization has highlighted the potential hazzards of development in this steep mountain terrain. During the wet spring of 1984 debris flows, rockfalls, and landslides caused damage to structures at several locations. In the upper end of the valley avalanche tracks mark areas of potential danger to lives and property.

174.4-177.3 Large roadcuts in sandstone, siltstone, limestone, dolomite, and conglomerate of the Minturn Formation. Detailed sedimentological study reveals a number of cycles that represent shallow marine to alluvial fan deposition (Johnson and others, 1988).

177.9 Cross Eagle River. Exit 171, U.S. Highway 24 to Minturn and Leadville.

Debris from landslide on left blocked the eastbound lane of I-70 several times in recent years. Wet years in 1983, 1984, and 1985 increased activity. On curve to left just beyond is the toe of the Whiskey Creek slide, which extends more than 600 m vertically up the valley side. Any major movement on it would endanger the highway and railroad and might dam the Eagle River.

Two km beyond the interchange gray, massive, resistant, coarse-grained clastics in the Minturn Formation abruptly change facies northwestward into gypsiferous mudstone and siltstone of the Pennsylvanian Eagle Valley Evaporite.

184.4 Exit 167 to Avon. On the left, across the valley of the Eagle River, the west-facing light-colored steep-walled bluff is an outcrop of fairly pure gypsum of the Eagle Valley Evaporite.

West of Edwards are good views

south up the valley of West Lake Creek into the northern Sawatch Range. Peak on the right is New York Mountain, with dip slope of Cambrian Sawatch Quartzite on the ridge to the right. Jagged peaks are largely biotite gneiss and migmatite.

189.6 Turn right off I-70 into Wilmor Lake rest area.

STOP 7 (optional). The vertical monocline at the west end of the parking area forms the east flank of the Wolcott syncline. Vertical beds of sandstone and siltstone are in the transition zone between the uppermost part of the Eagle Valley Evaporite and the lower part of the Maroon Formation, the main body of which is visible to the west. Johnson and others (1988) interpret these rocks as representing several cycles of shallow marine or nonmarine to offshore marine deposition associated with changes in sea level but modified by the effects of basin-margin tectonics.

West across the Eagle River is a view of Bellyache Ridge. Dakota Sandstone forms the skyline. It is underlain by Morrison, Entrada, Chinle, State Bridge, and Maroon Formations (at river level).

Leave the rest area and continue west on I-70.

193-198.5 Valley cuts through Wolcott syncline which exposes Mesozoic rocks. Prominent sandstone ledge at the top of the red beds is the Jurassic Entrada Sandstone, which is overlain by nonresistant light-gray and greenish-gray Jurassic Morrison Formation, ledge-forming Cretaceous Dakota Sandstone, black Benton Shale and light gray calcarous siltstone and limestone of the Niobrara Formation. The contacts of the formations in the red beds below the Entrada are harder to see. The Triassic Chinle Formation is somewhat darker red and less resistent than the underlying formations. Its base is marked by a thin resistent pebbly sandstone. On the west flank of the syncline a light-gray sandstone bed within the red sequence marks the top of the Maroon Formation and is considerd to be a tongue of Permian Weber Sandstone from the northwest. The Permian and Triassic State Bridge Formation is between the Chinle and the Maroon Formations.

West of the Wolcott syncline to the mouth of Glenwood Canyon the route is in highly deformed evaporite, mudstone, siltstone, and dolomite of the Eagle Valley in a diapiric anticline.

At Eagle high wooded hill to south is Hardscrabble Mountain capped by State Bridge Formation in a north-dipping slab including beds up to the Mancos Shale and surrounded by an almost-closed diapiric contact with Eagle Valley Evaporite (Tweto, 1977b). To the southeast up tributary valley are the high peaks of the Sawatch Range.

Between Gypsum and the Colorado River the upper side of the valley is composed of Maroon formation above deformed Eagle Valley Evaporite.

231.1 On left, basalt flow on valley bottom came from a vent 200 m vertically up gulley to right. Dated by ^{14}C as 4000 Ka, this is the youngest known volcanic rock in Colorado.

233.6 Cross Colorado River.

252.1 Black shales exposed in railroad cut across valley to the left are in Lower Pennsylvanian Belden Formation.

254.5 Enter Glenwood Canyon. Outcrops of Mississippian Leadville Limestone on right. This canyon has been a transportation corridor since the railroad was constructed in 1886. Now a very expensive sretch of the Interstate highway system is under construction through the canyon. Descend gradually through the early Paleozoic shelf section.

Glenwood Canyon is cut through the White River uplift, which is a domal upwarp over 5,200 km^2 in area. It is capped by Paleozoic sedimentary rocks. Miocene alkali olivine basalts and interbedded sedimentary rocks and tuff cover the northeastern part of the uplift and form the highest area of the White River plateau, altitude 3,000-3,500 m (not visible from here). The uplift is bounded on the north by a series of lesser folds including those of the Axial Basin anticline. Bordering the uplift on the east is the Eagle River basin, a southeastward projecting tongue of the Sand Wash Basin, and on the west side is the Piceance basin. Dips on the east and north sides are relatively gentle, but on the south and southwest sides, along the Grand Hogback, the beds approach the vertical and in some places are overturned. Southwestward thrusting has been reported at depth along the Hogback, and large normal faults are common along the south flank near Glenwood Springs. Some uplift of this structure occurred during Paleocene, but the major movement began in early-middle Eocene. The ages and distribution of the basalts in the region indicates that some differential uplift of the area occurred later than 10 Ma, and that Glenwood Canyon was cut in late Miocene, Pliocene, and Pleistocene time (Larson and others, 1975).

256.0 Sawatch Quartzite forms cliffs at road level and is characterized by a banded appearance.

256.3 Belden Formation exposed near river level in a small graben; beyond the graben cliffs of Sawatch dominate the scenery along the road.

261.5- Basement rocks composed of Early Proterozoic granodiorite,
271.4 biotite gneiss, and pegmatite along road at bottom of canyon.

263.3 STOP 8 to view canyon and the lower Paleozoic section (Fig. 9).

The Sawatch Sandstone, Dresbachian or Upper Cambrian in age, is composed of about 100 m of quartz sandstone. From a distance the Sawatch can be recognized by the light to dark brown color banding in the lower cliffs of the canyon. Conformably overlying are 15 m to 25 m of sandy dolomite to dolomitic sandstone, which have been interpreted to represent the Peerless Formation (Ross and Tweto, 1980). Above the Peerless are 15 to 45 m of quartzite in an unnamed unit. All these rocks have been included in the Sawatch Sandstone by some workers, such as Campbell (1972).

Conformably overlying the Sawatch is the Tremplealeauan or Upper Cambrian Dotsero Formation. The Dotsero consists of thin bedded dolomite, dolomite or limestone flat-pebble conglomerate, and an

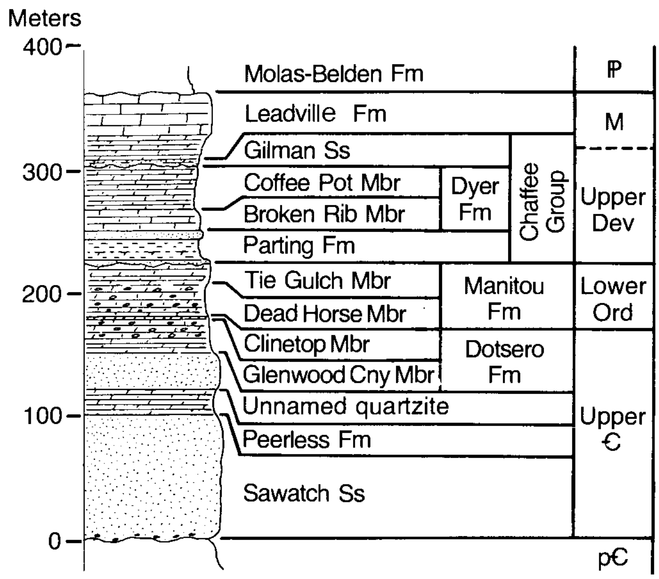

FIGURE 9 Statigraphic section in Glenwood Canyon. From Campbell (1972).

upper persistent lavender to white algal carbonate bed about 1.5 m thick. The lower carbonate beds are called the Glenwood Springs Member and the algal limestone and pebble conglomerate beds are called the Clinetop Member. The Dotsero generally forms a small grass or tree covered bench about two thirds of the way up the cliff in the canyon.

The Lower Ordovician (Canadian) Manitou Formation conformably overlies the Dotsero Formation. The Manitou has been divided into two members. The lower sequence consists of thin-bedded dolomite or limestone that contains flat-pebble conglomerate and is called the Dead Horse Member. The upper sequence consists of thin bedded dolomite and a few flat-pebble conglomerate beds and is called the Tie Gulch Member. The thickness of the Manitou ranges from 25 to 55 m, and it forms the brown upper one third of the cliff in Glenwood Canyon.

The unconformity between the Manitou and the overlying Chaffee Group is the major unconformity in the lower Paleozoic rocks throughout central Colorado. All of the Middle and Upper Ordovician, all of the Silurian, and the Lower and Middle Devonian are absent in the canyon. This unconformity is at the top of the cliff. The large tree and grass covered bench at the top of the prominent cliff in the canyon marks the position of the Upper Devonian Chaffee Group. The lower part of the slope consists of shales, quartz sandstone, and a few dolomite beds of the Parting Formation of the Chaffee Group. The Parting ranges from 20 to 30 m thick and from Frasnian to Famennian in age. The upper part of the bench consists of carbonate beds of the Dyer Formation of the Chaffee Group. The lower part of this carbonate sequence is the very fossiliferous, gray, massive limestone which has been designated the Broken Rib Member; the upper thin-bedded dolomite has been designated the Coffee Pot Member. The Dyer ranges from 50 to 55 m thick and in age from Famennian to Early Mississsippian(?) (Kinderhookian). The uppermost unit of the Chaffee Group is about 1.5 m of irregularly bedded sandy dolomite or dolomitic sandstone that is called the Gilman Sandstone of Late Devonian or Early Mississippian age.

Conformably overlying the Gilman Sandstone are the light colored carbonates of the Leadville Limestone (called Dolomite in the Leadville and Gilman districts). The lower one third to one half of the Leadville consists of thin-bedded dolomite of the Red Cliff Member, which are very similar to those of the Coffee Pot Member of the Dyer. The upper part of the Leadville consists of fossiliferous and commonly cliff-forming and oolitic limestone of the Castle Butte Member. The age of the Leadville ranges from Kinderhookian to Osagean and the thickness from about 15 to 60 m.

The irregularity in thickness of the Leadville is largely due to post-Leadville, Late Mississippian erosion, which produced karst topography and a soil zone on top of the Leadville called the Molas Formation. This soil zone is on top of the Devonian locally. The red and yellow streaks on the Leadville cliff are due to wash from the overlying basal Pennsylvanian Molas Formation. The Pennsylvannian Belden Formation unconformably overlies the Leadville in the canyon and characteristically forms colluvium mantled slopes.

264.2 Shoshone dam forming reservoir for hydroelectric plant 2.5 km downstream.

From 271.4 to mouth of canyon Lower Paleozoic rocks are cut by several west-trending faults. Hot springs issue from Leadville Limestone at the mouth of the canyon.

Glenwood Springs (1,751 m) is a transportation and tourist center at the confluence of the Colorado and Roaring Fork Rivers, was founded in 1882. The Glenwood Hot Springs pool was constructed in 1888 and the original bathhouse was completed in 1891. The main spring has a daily flow of 15,350,000 liters of water at a temperature varying from 51^{o} to 54^{o}C.

278.9 Exit 116, turn right twice and cross the Colorado River on Colorado Highway 82 and pass through downtown Glenwood Springs. From Glenwood Springs the route follows the Roaring Fork valley to Aspen. The Eagle Valley Formation, in part brought up in the Cattle Creek diapiric fold, forms much of the valley side to near Basalt. From there the valley is along or near the Castle Creek fault zone, traverses the eastern end of the Roaring Fork syncline, crosses the Castle Creek fault zone, and reaches the west margin of the Sawatch Range at Aspen.

280.8 On right, Pennsylvanian Eagle Valley Formation composed of predomiantly yellowish gray siltstone, sandstone, and gypsum overlain by red sandstone, siltstone, shale, and conglomerate of the Maroon Formation.

285.3 Large exposures of Pennsylvanian-Permian Maroon Formation on left.

287.1 Middle Pleistocene pre-Bull Lake outwash terrace at 1 o´clock may have been tilted away from the valley axis by Quaternary movement in the Cattle Creek diapir. A series of mild, shallow earthquakes that occurred near here in 1984 probably were related to salt diapirism.

291.7 Cottonwood Pass road and Cattle Creek on left. Across the river to the west a well encountered 633 m of interbedded evaporite and micaeous siltstone above 285 m of halite containing minor interbeds of anhydrite and siltstone to the bottom of the well (Mallory, 1966).

297.5 STOP 9 (optional) Scenic overlook. View of Mt. Sopris (3,947 m), the Crystal River valley, and the Grand Hogback. The four main terraces are of Pinedale (Wisconsin), Bull Lake (late Illinoian), and pre-Bull Lake (2) age. The lower of the two pre-Bull Lake terraces contains a lens of Lava Creek ash, about 600,000 years old. These terraces cannot be traced to terminal moraines near Aspen because of the steep sides of Snowmass canyon between Basalt and Aspen.

Mount Sopris is the northwesternmost outlier of the Oligocene calc-alkaline intrusions of the Elk Mountains. To the left of Mt. Sopris are rounded hills underlain by Mancos Shale in the west end of the Roaring Fork syncline. The shale is overlain by gravel which is probably of late Tertiary age like the basalts in the region. To the northwest is basalt forming a surface that appears to slope gently north. These basalts are correlated with a group of basalts 9 to 14 Ma by Larson and others (1975), but they have not been dated. To the west, ridge-forming sandstones of the Dakota and Mesaverde form the southern extension of the Grand Hogback at the east margin of the Piceance basin.

304.6 Crossroads at Catherine. To the left is a large plateau capped by 8.7 Ma alkali olivine basalt (Larson and others, 1975) associated with river gravels and overlying the Eagle Valley Formation. Basalts covered the floor of a broad valley eroded in the Eagle valley Formation, and the valley has been deepened only 150 m since that time.

307.2 At 11 o'clock is Basalt Mountain (3,311 m), a shield volcano formed by 20 or more flows of alkali olivine basalt about 9 Ma (Larson and other, 1975). The lavas rest on incompetent Mancos Shale. A landslide 26 km^2 in area extends from the edge of the basalt cliffs near the top of the mountain almost to the valley bottom.

312.8 Across the valley at 3 o'clock exposures of the upper part of the Maroon Formation, Entrada Sandstone, Morrison Formation, and Dakota Sandstone (the prominent ledge). State Bridge and Chinle Formations are missing. Thinning of the State Bridge by erosion prior to deposition of the Chinle and thinning of the Chinle by erosion prior to deposition of the Entrada probably due to waning stages of uplift of the ancestral Uncompahgre uplift, which lies to the southwest, can be demonstrated in this area.

314.1 We have crossed a fault bounding a graben that extends southeast along the Castle Creek fault zone and to the north under Basalt Mountain. Rocks of Jurassic and Cretaceous age occur beneath the valley floor to the southeast, and rocks of late Paleozoic and Triassic age crop out on the hills on either side. Interpretation of the structure as a simple graben is based on very poor exposures in the valley.

321.7 Cross Snowmass Creek. The Castle Creek fault zone probably extends along the valley here but produces little stratigraphic separation. The valley bottom is along the crest of an anticline southwest of the fault. At 11 o'clock hills formed by Maroon and State Bridge Formations.

325.0 Across the river in the lower outcrops, the beds of the Maroon Formation can be seen dipping eastward more steeply than the beds of the State Bridge Formation in the higher outcrops. The angular unconformity is about 30° here, and indicates movement along the Castle Creek fault zone in Permian time (Freeman, 1971). Ahead, white outcrops of Eagle Valley Formation in slices or as diapirs along the Castle Creek fault zone.

328.0 Intersection with road across Watson Divide, which separates the Snowmass Creek valley from the Roaring Fork valley and has a remanent of a gravel deposit containing boulders of basalt from the hill on the left. The deposit is correlated with a terrace of middle(?) Pleistocene age and shows that at that time the Roaring Fork did not flow through the narrow canyon from which we have just emerged. To the left across the river greenish-gray siltstones of the Jurassic Morrison Formation.

328.5 Road to Woody Creek on left. The lowest outcrops across the river are Cretaceous Burro Canyon Formation and Dakota Sandstone. Top of hill to left is capped by 1.5 Ma alkali basalt flows overlying river gravels (Larson and others, 1975). The river has incised about 370 m in Quaternary time here.

330.9 STOP 10 (optional). Looking northwestward back down the Roaring Fork valley the prominent hill to the west of the canyon is underlain by beds dipping southward 25° to 30° towards the Roaring Fork syncline. Resistant Dakota Sandstone is underlain in sequence by Burro Canyon, Morrison, and Curtis Formations, Entrada Sandstone, and Chinle and State Bridge Formations. These formations continue unbroken northeast across the canyon to the Castle Creek fault zone that passes through the crest of the hill northeast of the Roaring Fork. That hill is capped by the remains of a cinder cone resting on about 120 m of basalt flows that flowed out on the floor and side of the Roaring Fork valley 1.5 Ma.

The canyon east of the basalt capped hill exposes State Bridge Formation dipping about 20° eastward on the upper slopes. Maroon Formation beneath dips much more steeply and is vertical adjacent to the Castle Creek fault zone. Within the fault zone white gypsiferous beds of the Eagle Valley crop out in the lower slopes on both sides of the mouth of the canyon.

To the southeast are three major outwash terraces that we follow up the valley to their associated moraine deposits. Two miles upvalley, where the uppermost terrace is well preserved, the terraces are about 195, 145, and 50 m above the river. The terrace on which we are traveling heads at end moraines east and south of Aspen, which are breached by narrow gorges. The upper two terraces head at lateral moraines several kilometers downvalley. No ages of moraines or terraces are available, but it seems reasonable to correlate the lower terrace and moraine with the Pinedale, the middle one with the Bull Lake, and the upper one with the pre-Bull Lake glaciations of the Rocky Mountain region.

333.3 On right ledge-forming sandstone marker in the Mancos Shale reaches the valley bottom in the trough of the Roaring Fork syncline.

341.5 On right Buttermilk ski area is on dip slope of Mancos Shale on the southwest limb of the Roaring Fork syncline. To left is Red Butte composed of a slice of overturned rock along the Castle Creek fault zone. Prominent band of red soil formed by Chinle Formation. Upslope is State Bridge Formation and downslope is Entrada Sandstone, Morrison Formation, Burro Canyon Formation, and Dakota Sandstone. Cross faults of small displacement cut the formations. Ahead is West Aspen Mountain; Ordovician and Cambrian rock overlie Middle Proterozoic granite east of the Castle Creek fault zone.

343.0 View of Pyramid Peak (4,273 m) to right up canyon of Maroon Creek.

345.7 Downtown Aspen.

THE ASPEN MINING DISTRICT

Ralph J. Stegen[1], David W. Beaty[2], Bruce Bryant[3] and Tommy B. Thompson[4]

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INTRODUCTION

History

In 1879 prospectors from Leadville discovered outcrops of manto deposits which became the Spar mine on Aspen Mountain. Immediately following the discovery, many mining claims were located on both sides of the Roaring Fork Valley near the site of Aspen. After the first smelter was built in 1882 and the railroad reached town in 1887, Aspen grew quickly. In 1888, production from the district increased tenfold above the previous year. The peak in production was reached in 1892 when about 260,000 kg of silver was mined (Bryant, 1972). Aspen was one of the first mining districts to use electric powered hoists and pumps in the mines. The devaluation of silver in 1893 and the mining out of the large, very rich mantos forced a sharp decrease in the amount of silver mined. However, in 1894, some of the larger silver nuggets recorded were mined from the district; one weighed 770 kg and several others weighed between 225 and 450 kg each.

Mining continued into the 1920′s but the value of production declined. As the mines became deeper, great quantities of water (as much as 12,300 liters per minute) had to be pumped. After the middle 1920′s production from the Aspen mining district was minor compared to that from one mine in the Richmond Hill district 5 km to the south, which produced silver, lead, and zinc from 1928 to 1952. Since 1952 no mining has been done in either district.

Mining and Processing

Most of the mining in the Aspen district was by either square set timber or open stope mining methods. Locally, shrinkage-stoping was utilized in later years. Little is known about production rates from the mines.

The Aspen district produced two types of ore: 1) very high-grade silver-rich oxide and sulfide ore, which was shipped directly to smelters, and 2) lead and zinc-rich ore, which was processed by milling. The average grade of the high-grade silver ore is not precisely known, but it was as much as about 80% (Bastin, 1925; Henderson, 1926; Lakes, 1887; Spurr, 1898). The grade of the milling ore averaged 2 to 3 percent lead, about 100 grams per tonne (gpt) silver, and it contained some zinc (Knopf, 1926).

Production/Reserves

Cumulative production of all ore types at Aspen has been estimated at 2.3 million metric tons with an average grade of 1500 gpt silver, 5.9% Pb, 2.1% Zn, and 16.4% Ba (Stegen and others, in press). According to Rohlfing (1938) and Heyl (1964), low-grade lead-zinc ore remains between and below the mine workings.

GEOLOGY

General Geology

Aspen is located near the northwestern edge of the Colorado mineral belt on the west flank of the Sawatch uplift (Fig. 4). The district is located where the west margin of the Sawatch Range makes a sharp change in trend from N. 10° W. south of Aspen to N. 30° E. to the north. It is also located along the southwestern extension of the Homestake shear zone, one of the northeast-trending shear zones of Precambrian age. A 200-320-m-thick cratonal shelf sequence of Cambrian through Mississippian sedimentary rocks unconformably overlies Precambrian gneiss, schist, and quartz monzonite. The uppermost unit in this sequence is the Mississippian Leadville Limestone. The Leadville is in turn disconformably overlain by more than 5,000 m of Pennsylvanian through Cretaceous clastic sedimentary rocks. The sedimentary rocks dip westward off the margin of the Laramide Sawatch uplift.

Formation of the Sawatch uplift was accompanied by faulting along its western margin. The major structure near Aspen is the Castle Creek fault, which trends parallel to the Range south of Aspen, but to the north the fault diverges to a northwest trend parallel with the structural grain of the Eagle basin. A tight, south-plunging fold underlying part of the district on the south side of the Roaring Fork valley (Fig. 10) occurs where the Castle Creek fault diverges from the margin of the uplift. Numerous faults, which according to Spurr (1898) localized in part the Ag-Pb-Zn-Ba manto deposits, are also concentrated in this area.

In the southern part of the Aspen area, Bryant (1979) mapped four types of igneous intrusions, two of which have been dated by K-Ar methods at 69 to 74 Ma (Obradovich and others, 1969, corrected for currently

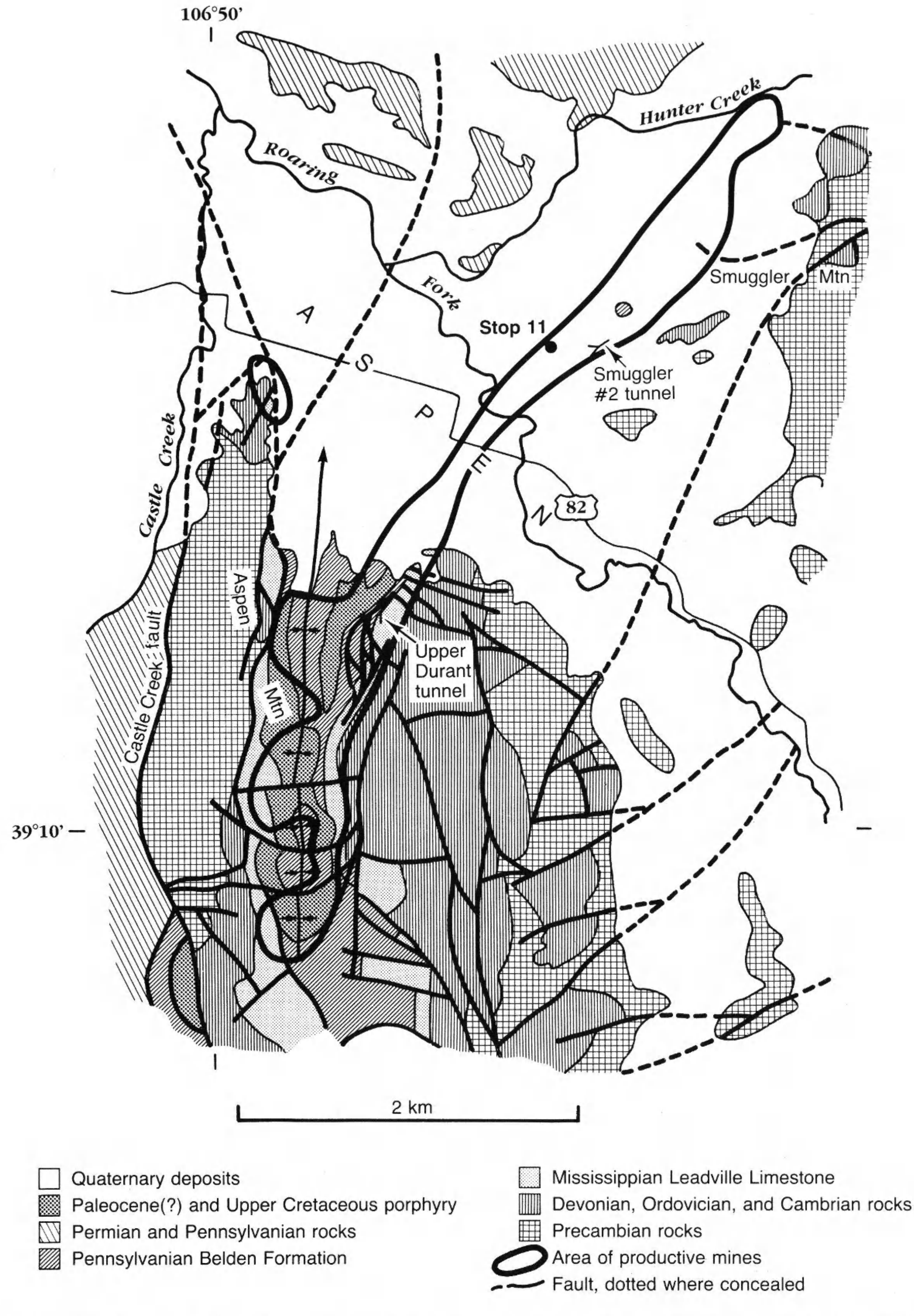

FIGURE 10 Geologic map of the Aspen mining district showing area of productive mines. After Bryant (1971a, 1972).

accepted decay constants). Within the Aspen mining district, by contrast, only one Laramide intrusion is exposed. Aplite porhryry forms a sill as much as 100 m thick near the base of the Pennsylvanian Belden Formation on Aspen Mountain and extends north to Smuggler Mountain where it is exposed only in mine workings, is much thinner, and pinches out in the more northerly workings (Spurr, 1898).

District host rocks. Most of the Aspen manto deposits are contained within the Leadville Limestone. The Leadville Limestone at Aspen consists of two members, a lower thinly laminated dolostone, the Red Cliff Member, and an upper massive limestone, Castle Butte Member. Both members have abrupt thickness variations which have been interpreted as the result of two separate dissolution events (Beaty and others, 1988). As a result of these dissolution processes, two laterally continuous breccia beds were formed. A lower bed (the Contact breccia) is at the contact between the Castle Butte and Red Cliff members and an upper bed (the Silver breccia) is at the top of the Castle Butte Limestone. The Contact breccia contained most of the orebodies. A younger breccia mass is present in the Smuggler mine, which cuts both the Contact breccia and the large baritic manto orebodies, and is mineralized by lead and zinc.

Ore deposits

The ore deposits of the Aspen district are of three principal types: Ag-Pb-Zn-Ba main manto deposits in the Contact breccia, Pb-Zn mantos in the younger breccia, and, in the Smuggler Mine, Pb-Zn manto ores in the Pennsylvanian Belden Formation. A supergene silver mineralization overprinted all ore types. Principal production was from large, laterally-continuous, stratabound mantos in the Contact breccia. Throughout the district, about 50 percent of the breccia was mineralized. The main mantos consist of galena, sphalerite, and chalcopyrite with tennantite, pearceite, acanthtite, and native silver in a barite-rich (30-70%) gangue. In the Smuggler mine the ore and gangue minerals were found in two main settings: as open-space filling between breccia fragments and as replacement of a basal bed of stratified dolomite sand. Virtually no replacement of the fragments occurred.

The young breccia-hosted mantos consist of varying proportions of galena, sphalerite, and pyrite that replaced the breccia matrix. Mineralized rock ranges from sparse, fine-grained disseminations to massive ore. The massive ore was mined along intersections of the brecciated strata with Laramide faults.

The mantos in the Belden Formation replaced carbonaceous shale and limestone. Principal ore minerals are galena and sphalerite. Seven separate manto orebodies have been mined, they are tabular-shaped, extend laterally 70 to 140 m, and are up to 1.5 m thick. The orebodies are found stratigraphically above the main mantos bodies in the Contact breccia, and we were unable to clearly connect the two by geological relations in the mine workings.

Hydrothermal alteration. Beneath the main manto orebodies, dolostone wallrock is typically unaltered; recrystallization occurs only within 0.3 m of ore. The orebodies within the young breccia and Belden Formation have no associated alteration. Jasperoid is localized along pre- and post-ore faults, but bears no obvious, direct relation to ore. Geologic and petrographic studies of jasperoid suggest a post-ore origin. Hydrothermal dolomite was reported by Spurr (1898) as forming an envelope around ore within the Castle Butte Limestone Member.

The aplite porphyry shows strong alteration of feldspar and biotite phenocrysts to sericite and minor kaolinite, extensive replacement of groundmass feldspar by quartz, and dissemination of pyrite (1-3%) throughout the rock. Alteration of aplite porphyry shows no zoning around the manto orebodies.

Fluid inclusion studies. Primary fluid inclusions were examined in barite, early sphalerite, tennantite, and pearceite from the main manto ores in the Smuggler mine. The late sphalerite at Aspen is too dark for study, and there is no quartz in the orebodies. Homogenization temperatures of fluid inclusions are 252-314^{o}C (average 252^{o}C) for barite, 237-268^{o}C (average 252^{o}C) for sphalerite, 250-261^{o}C (average 255^{o}C) for tennantite, and 254-269^{o}C (average 257^{o}C) for pearceite. Freezing measurements of fluid inclusions in barite indicate salinities of 1.5 to 7.3 (average 3.2) equivalent weight percent NaCl. No salinity data were obtained for sphalerite, tennantite, or pearceite.

Sulfur isotopes. The sulfur isotopic compositions of sulfide and sulfate minerals in the main manto ores in the Smuggler mine at Aspen have a wide range. The $\delta^{34}S$

values for five samples of barite range from +12.3 to +12.9. Twelve analyses of early sulfide minerals show relatively constant $\delta^{34}S$: pyrite (-11.4 to -11.6), chalcopyrite (-11.9 to -12.1), galena (-13.6 to -13.9). Sulfur isotopes measured from paragenetically later sulfides are significantly heavier: tennantite (+0.5), galena (-3.4 to -9.4), and pearceite (-3.9). These data indicate an early sedimentary source of sulfur that either evolved due to variations in SO_4 and H_2S to produce the heavier $\delta^{34}S$ compositions, or indicate a later influx of sulfur from an igneous source. Preliminary lead isotope data indicate the lead also had an igneous source.

Oxygen and hydrogen isotopes. Barite, the only oxygen-bearing gangue mineral in the mantos, has $\delta^{18}O$= +3.2 to +5.0. The calculated fluids (using T = 255°C) range in $\delta^{18}O$ from 0 to +2 (inferred basinal brines). Fluid inclusion waters in barite, galena, tennantite, and pearceite were analyzed for δD. For all samples, δD is within the range-106 to -124, and shows no systematic variation in δD with position in the paragenetic sequence. The hydrogen isotopic data indicate values intermediate between meteoric and magmatic water.

Altered aplite porphyry has a whole-rock value of $\delta^{18}O$ = +9.4 to +14.2 and δD = -80 to - 99, indicating a hydrothermal fluid of $\delta^{18}O$ =5.3 to 10.1 and δD = -42 to -62. This fluid is interpreted as magmatic water.

Genetic Model. The following mineralization model appears to best account for the data now avaiable: Laramide uplift of the Sawatch Range resulted in formation of the Castle Creek fault and associated faults along its west margin. A large batholithic mass with its northwestern margin below the Aspen area was concurently emplaced within central Colorado. A steep geothermal gradient developed, causing fluids containing sulphur and metals to migrate from evaporitic rocks deposited in the Central Colorado Trough towards the rising Sawatch Range. Igneous-derived fluids, sulfur and lead apparently were involved in at least part of the hydrothermal history. Fluid flow was localized by Laramide structures and then channeled through stratabound brecciated beds in the Leadville Limestone.

Second Day, Tuesday, July 4.

Stop 11. Overview of the Aspen mining district from below Smuggler mine. This area was once part of the dump from the Mollie Gibson mine. Near this site were the mills that processed ore, especially during the last 30 years of production from the district. Most of the city of Aspen is on a late Pleistocene outwash terace graded to a terminal moraine in the eastern outskirts of town.

In the distance to the west are Mesozoic rocks preserved in the Roaring Fork syncline. The Castle Creek fault is at the west base of the rugged ridge on the southwest side of the town, and it bends to a more westerly trend to parallel with the Roaring Fork valley northwest of Aspen.

To the southwest is the north-plunging Aspen Mountain syncline cored by dark gray and black limestone and shale of the Belden Formation, which is intruded by a sill of aplite porphyry (Fig. 10). Both sides of the syncline are faulted. The fault on the west side has the greatest stratigraphic separation; all the Paleozoic section below the Leadville is absent. On the east side the fault pattern is more complex, and displacement is distributed along several faults, most of which die out before they reach the valley bottom. To the north the Aspen Mountain syncline dies out, and Red Mountain is formed by red beds of the Pennsylvanian and Permian Maroon Formation only gently folded.

The Aspen mining district is underlain by an interconnected set of workings about 5 km long most of which lie on the east limb of the syncline. Workings extend about 300 m below the valley level and cover a vertical range of 850 m on Smuggler Mountain and 730 m on Aspen Mountain. Many of the mine dumps that used to dominate the landscape have been obscured by grading for ski trails and building developments. Most of the mine workings are inaccessible, but small parts of the Smuggler Mine, behind us, and the Durant mine, at the small building on the ski trail on Aspen Mountain, have been rehabilitated for purposes of exploration.

Aspen to Maroon Lake. The Maroon Creek valley is a glacial trough cut through redbeds of the Maroon Formation. Few outcrops occur at road level. Debris fans, talus, and colluvum covers the lower part of the valley sides. The view of the Maroon Bells from end of the road is one of the classic scenes of the Colorado Rockies. Road log begins at the junction of the Maroon Creek road and Colorado Highway 82.

1.8 On the right Cretaceous to Permian rocks exposed on the flank of the Roaring Fork syncline.

10.0 On left, sill of gray Oligocene porphyritic granodiorite in red beds of the Maroon Formation. This is the same sill visible north of and 400 m above Maroon Lake

15.6 STOP 12. Maroon Lake (3,143 m). Up valley across the lake the Maroon Bells (North Maroon Peak, 4,271 m; and Maroon Peak, 4,315 m) are composed of sandstones, siltstones, and some beds of pebble conglomerate of the Maroon Formation. The rocks on the lower slopes have been altered by heat from underlying Oligocene intrusive rock. A large rock glacier on the north face of North Maroon Peak was moving at the rate of about 60 cm a year in the middle 1960's (Bryant, 1971b). The gentle rocky deposit blocking the valley between Maroon Lake and the Maroon Bells is a landslide mostly composed of somewhat metamorphosed Maroon Formation, which is exposed on the south side of the valley. South and west of Maroon Lake, east-trending dikes of porphyritic granodiorite cut the Maroon Formation. Southeast of the lake, the front of an active looking rock glacier is the source of much talus. Just west of the active front, part of the rock glacier appears to be inactive. The Maroon Lake campground lies on a large debris fan heading in a steep gulley in red beds north of the lake. Sills of gray granodiorite are obvious in the red beds of that slope, and they are offset by a fault in the gulley. The debris fan forms the dam of Maroon Lake.

Return to Colorado Highway 82. Retrace our route to Glenwood Springs; drive east through Glenwood Canyon and the Eagle valley on Interstate Highway 70 to Exit 131.

193.6 Turn south on U.S. Highway 24. From here the route follows the gently dipping northeast flank of the Laramide Sawatch uplift. To the left the valley side is formed by sandstone, conglomerate, siltstone, and carbonate of the Pennsylvanian Minturn Formation and to the right Mississippian and older Paleozoic rocks form dip slopes overlying Precambrian basement rocks composing the core of the range. Morainal systems from major valleys in the Sawatch range reach the road in places south of Minturn.

196.5 Town of Minturn. Leadville Limestone, Gilman Sandstone, and Dyer Formation crop out in places along right side of road. South of Minturn late Pleistocene morainal complex from Cross Creek forms hills to right.

THE GILMAN MINING DISTRICT

David W. Beaty[1], C.W. Naeser[2],
C.G. Cunningham[3], and G.P. Landis[2]

INTRODUCTION

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The ore deposits at Gilman, Colorado cropped out on the side of Battle Mountain, and were discovered in 1879 by prospectors from the Leadville area. The town was

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founded in 1886, and was named after Henry M. Gilman, a mining man who represented Eastern capital. Battle Mountain was named for a confrontation in 1849 between the Ute and Arapahoe Indians. The New Jersey Zinc Company became interested in the district in 1912, and by 1915 had acquired the town and most of the larger mines, which were consolidated as the Eagle Mine.

The early production from the Gilman District consisted primarily of rich oxidized Ag-Pb-Zn ores and gold ore. By about 1910-1915 the miners had reached the unoxidized marmatite-bearing sulfide ore. In 1929, an underground flotation mill to concentrate the zinc was built. The mill closed in 1977, a victim of the deadly combination of falling zinc price and depletion of Zn-Pb reserves. Production of high-grade Cu-Ag-Au ore that could be shipped directly to the smelter continued until Spring, 1984. In 1984 the electricity was turned off, the mine was allowed to flood, and the town was abandoned.

Mining and processing

Most of the mining in the Eagle Mine employed square set timber and fill methods, although in some cases it was possible to mine open stopes. Timber consumption averaged about 15 board feet per tonne of ore. The mine operated at a production rate of 1,090 tonnes per day, and the entire facility employed about 500 persons. The town of Gilman (1963 population: 325) consisted of about 100 dwelling units, and was wholly owned by New Jersey Zinc Company. Most of the work force who did not live in Gilman lived in the nearby towns of Minturn and Red Cliff.

The Eagle Mine produced two ore types, Zn-Pb and Cu-Ag-Au. The Zn-Pb ore (1958 grade: 11.0% Zn, 1.75% Pb, 0.4% Cu, 78 grams per metric tonne (gpt) Ag) was milled, and the Cu-Ag-Au ore (1967-1977 grade: 2.5% Cu, 920 gpt Ag, 3.8 gpt Au) was shipped directly to the smelter. The mill produced a lead concentrate (4.0% Zn, 65% Pb, 3.0% Cu, 1190 gpt Ag), a zinc concentrate (49.5% Zn, 0.5% Pb, 0.8% Cu, 68 gpt Ag), and pyritic tailings (0.75% Zn, 0.26% Pb) which were piped 3.5 mi downstream to the disposal area. The lead concentrate was smelted at El Paso, Texas, and the zinc concentrate was sintered at Canon City, Colo., then smelted at Depue, Illinois.

Production/Reserves

Cumulative production of all ore types has been about 11.7 million metric tonnes with an average grade of 8.5% Zn, 1.5% Pb, 0.7% Cu, 228 gpt Ag, and 1.7 gpt Au (Beaty and others, in press). There are no known reserves.

GEOLOGY

General geology

At Gilman, a sequence of Cambrian through Mississippian sedimentary rocks (total stratigraphic thickness = 200 m) dip homoclinally eastward at 12-15°. These sedimentary rocks unconformably overlie Early Proterozoic basement (predominantly the Cross Creek batholith), and are in turn overlain by more than 3000 m of coarse clastics of the Pennsylvanian Minturn Formation. The sedimentary rocks in the Gilman area are intruded by one exposed igneous mass, the Pando Porphyry, which forms a sill 12-27 m thick in the Belden shale. The Pando Porphyry is a fine-grained quartz latite which crystallized at about 72 Ma (data of Pearson and others, 1962; recalculated by Beaty and others, 1987).

Ore deposits

The Eagle Mine developed a large interconnected sulfide deposit (Radabaugh and others, 1968; Lovering and others, 1978; Fig. 11) consisting of two principal ore types: Zn-Pb mantos and Cu-Ag-Au chimneys. The mantos consist of long, stratabound, subcylindrical Zn-rich replacement deposits within the Leadville Dolomite. Four main mantos, each 600 to 1200 m long, converge downdip and are connected by a NW-trending manto and the chimney deposits. Pyrite, marmatite (iron-rich sphalerite), and siderite are the most abundant minerals in the mantos. Galena and chalcopyrite are always present in small amounts. Typically, the mantos are surrounded by an irregular shell of siderite which is in turn surrounded by a shell of sanded dolomite. The up-dip ends of those mantos which have been exposed at the surface are oxidized for as much as 400 m downdip.

The chimney orebodies are funnel-shaped copper-, silver-, and gold-rich massive sulfide replacement bodies which cross-cut bedding and which extend from the upper Leadville through the Dyer Dolomite into the Parting Formation or Harding Sandstone. The chimneys are typically zoned, consisting of a pyritic core surrounded by irregular and incomplete shells of zinc-rich ore, siderite, and sanded dolomite. The chimney

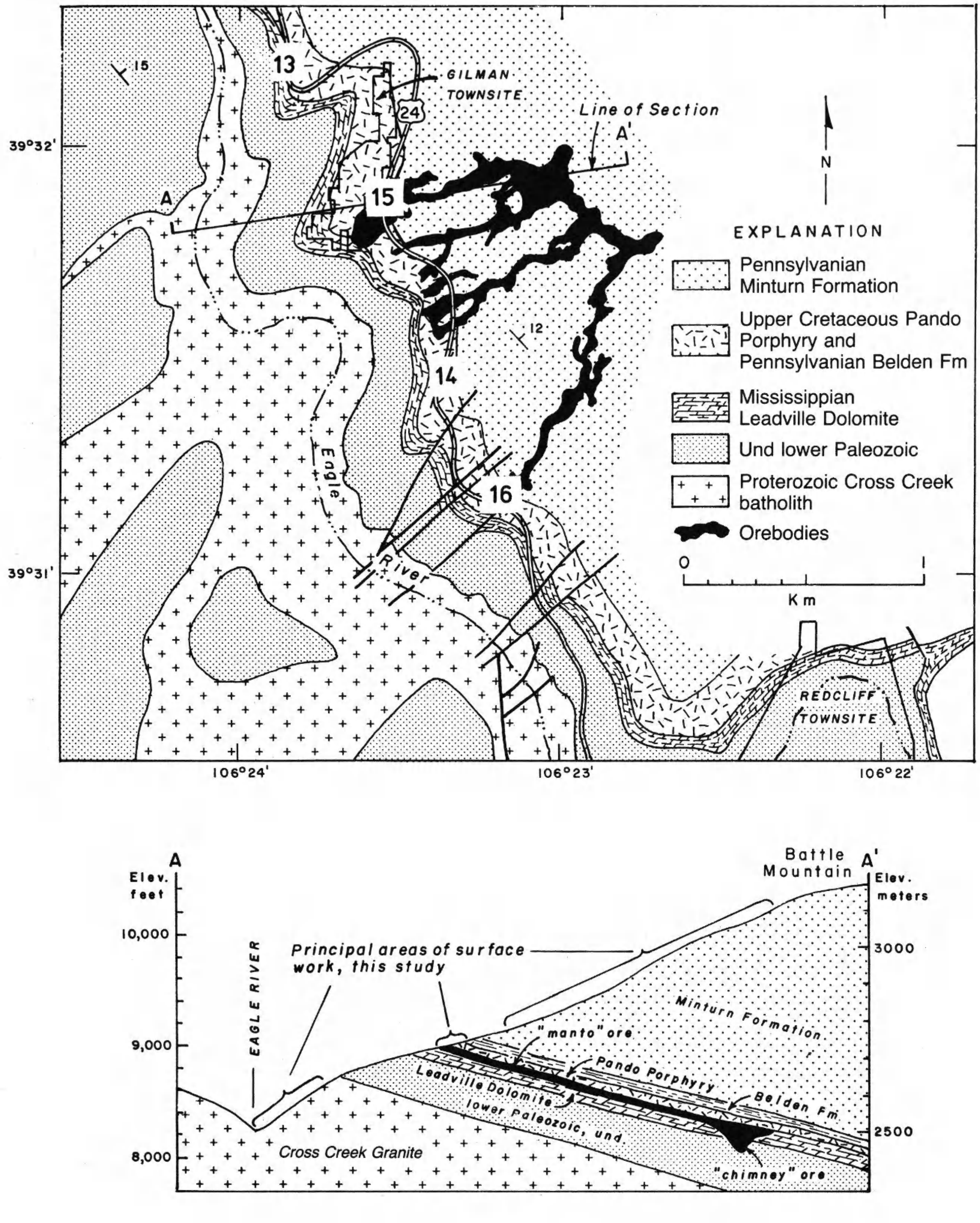

FIGURE 11 Map and cross section of the Gilman mine showing principal ore bodies in the Leadville Dolomite and locations of stops. Distribution of manto in cross section generalized. After Radabaugh and others (1968) and Lovering and others (1978).

ores contain the minerals chalcopyrite, tetrahedrite, freibergite (and other silver minerals), galena, and gold. There are eleven chimneys in all, ranging in width from 12 to 120 m at their tops and with a vertical extent of as much as 80 m.

District-scale metal zoning. The Leadville Dolomite down-dip and lateral to the main orebody was explored in 12 km of exploration workings and in numerous underground drill holes. A number of small orebodies were discovered, the size and abundance of which decrease to the northeast from the main orebody. These mineralization effects define concentric metal zoning patterns centered on the chimney deposits (from center outward: Cu-Ag-Au --> Zn --> Pb-Zn --> Mn). In addition, distinctive Ag-Pb-Zn orebodies are paragenetically early, and are distributed across the concentric geochemical zones.

Hydrothermal alteration. Above the manto/chimney orebodies, the Pando Porphyry sill is typically altered along its base, although it is essentially unaltered everywhere at its top. The altered rocks are mineralogically zoned away from ore: muscovite-rich --> kaolinite-rich --> chlorite-rich. The rocks of the Belden and Minturn Formations above the Pando Porphyry are unaltered and unveined throughout the Gilman area. Below the manto/chimney orebodies, the Precambrian basement was sampled in 11 deep core holes. The cores show local stockwork veining and alteration, but no post-Precambrian intrusives. The deep veins are mineralogically essentially identical to the manto/chimney complex. Zonal alteration about the veins (qtz-sericite-py-sphalerite to chlorite-clay-sericite-py to chlorite-rich) is very similar to that in the Pando Porphyry above the orebody.

Geochronology. The mineralization event was dated using the fission-track method. Late-stage hydrothermal apatite lining vugs in ore has an age of 34.5±2.6 Ma. Apatite from five samples of Proterozoic basement rock (three altered and two fresh) from the deep drill cores beneath the orebody yield average fission-track ages of 34.1±3.3 Ma (1 standard error of mean). Apatite grains from the Minturn Formation above the Eagle mine are also reset and have an average age of 36 Ma. On the basis of these data, the age of the orebody is inferred to be about 34 Ma. Zircon fission-track ages from altered Proterozoic granite below the chimneys do not show any significant resetting. This indicates that the principal source of the hot ore-forming solutions was not directly beneath the deposit.

Fluid inclusion studies. Primary fluid inclusions were examined in main-stage sphalerite, late-stage apatite, and in sphalerite/quartz-veins from beneath the chimneys. Unfortunately, most of the sphalerite at Gilman is too dark (marmatite) to use for conventional fluid inclusion studies, and the orebody is quartz-free. Homogenization temperatures of the fluid inclusions are 391-413°C for the manto sphalerite, 309°C for the late vug apatite, 180-259°C (average = 223°C) for deep vein sphalerite, and 174-306°C (average = 250°C) for deep vein quartz. Freezing measurements indicate fluid inclusion salinites of 2.7-7.7 wt. percent equiv. NaCl for the manto sphalerite, about 2.8 wt percent equiv. NaCl for the deep vein sphalerite, and 1.5-9.4 (average = 5.2) wt percent equiv. NaCl for the deep vein quartz; no salinity data were obtained from apatite.

Oxygen and hydrogen isotopes. The Pando Porphyry has a regional whole-rock $\delta^{18}O$ value of 11.0 to 11.5. Immediately above each of the manto orebodies, the porphyry is altered, and has whole rock $\delta^{18}O$ of 9.0 to 9.8. Clay from the altered porphyry has δD ranging from -95 near the orebodies to about -130 away from ore. $\delta^{18}O$ of quartz intergrown with marmatite (+14.4), in veins in the Leadville Dolomite (+14.1), and intergrown with carbonates surrounding the orebody (+16.1 to +17.1; data of Engel and others, 1958) are all comparatively high-^{18}O. Quartz veins (+14.0) and altered wallrock (+11.7) in the deep drill cores beneath Gilman are also high-^{18}O. These data indicate the ore fluid had a composition of $\delta^{18}O$ = +4 to +8 and δD= -55 to -75 permil, which is interpreted to be magmatic water. This high-^{18}O fluid was progressively diluted by meteoric water (δD= -120) peripheral to ore.

Sulfur isotopes. The sulfide minerals in the ores at Gilman have a very narrow range of $\delta^{34}S$ (pyrite= 0.6 to 2.7, sphalerite= 1.3 to 2.4, galena = -1.3 to -2.0, chalcopyrite = 0.8 to 1.9). The sulfide minerals in the rare, early Ag-Pb-Zn deposits ($\delta^{34}S$: amber sphalerite= 3.4, marmatite= 1.5, pyrite = 4.2), however, are somewhat heavier than those typical of the Gilman system. These data indicate a well-homogenized, igneous sulfur source for the overwhelming preponderance of the ore at Gilman, with a bulk system $\delta^{34}S$ of about 1.6. Sulfide mineral pairs at Gilman are difficult to establish, however, temperatures calculated from pyrite-sphalerite and pyrite-chalcopyrite pairs are consistent with those determined from fluid inclusion measurements. The pyritic cores of the chimneys apparently formed at about 400°C (py-sl-cpy). Temperatures of 297°C and 373°C (py-sl) are calculated for the manto ore, and 372°C (py-cpy) and 231°C (py-sl) for veins in the Sawatch Quartzite.

Genetic Model. The heating event associated with formation of the orebody can be dated by the fission-track method at about 34 Ma. The distribution of veins, small mantos, geochemical anomalies, and altered rocks around the orebody indicates that within the region of the manto deposits, the hydrothermal fluids flowed up-dip within the Leadville Dolomite. The base of the Pando sill, along with shales of the Belden and Molas Formations, acted as an aquitard, and the rocks overlying these mantos experienced no fluid flux. The only evidence that there is an orebody beneath these overlying rocks is the presence of annealed fission tracks in apatite. Within the Leadville Dolomite, mantos appear to have been localized by high-angle faults and fissures, by a preexisting paleo-cave system, and by zones of high permeability in the upper part of the Leadville Dolomite. The relative importance of these factors cannot be unambiguously established.

The sulfur isotope fractionation temperatures suggest that the highest temperature hydrothermal fluids entered from the chimney structures, and flowed updip into the mantos, cooling in the process. The conduit by which the hydrothermal fluids reached the site of ore deposition has apparently not been exposed by mine workings or by drill holes. Several obvious possibilities have been exposed, however, and can be eliminated from consideration. Even though a vein system is present beneath the chimney deposits, the paleothermal data indicate that this was not the main source of heat. The sampled portion of this vein system is thus either peripheral to the main source of the ore fluids, or perhaps part of the return flow of the hydrothermal system. The fluids also apparently could not have entered the region of the principal orebodies by simply flowing up-dip within the Leadville Dolomite, because the mineralization effects distinctly decrease down-dip from ore. The most logical hydrologic model involves updip fluid transport within the lower part of the sedimentary sequence (probably the Sawatch Quartzite) to the area of the chimneys, then up-section flow to the Leadville Dolomite (Fig. 12). The Leadville Dolomite is an important aquifer, and once the fluids entered it, they were transmitted dominantly up-dip.

Sulfur isotope data from throughout the main orebody indicate the sulfur came from an igneous source. Lead isotope data reported by Engel and Patterson (1957) indicate the lead also had an igneous source. Oxygen and hydrogen isotope data indicate that the orebody was deposited from a large mass of high-^{18}O, high-D water with isotopic properties indistinguishable from magmatic water. These relationships, combined with the thermal indications that at least portions of the deposit formed at 400°C, argue that the Gilman hydrothermal event is related to an unexposed igneous intrusion. This intrusion is inferred to be located a short distance northeast of the known sulfide deposit.

The genesis of the early Ag-Pb-Zn ores at Gilman can be only loosely constrained using the Gilman exposures alone. These deposits are tentatively correlated with those at Mt. Sherman, Colorado, which are of controversial age. These data indicate fundamental genetic similarities between Gilman and Leadville, and also one critical difference: Fluid transport was up-dip at Gilman, but up-section at Leadville.

204.4 STOP 13. Rock Creek outcrop. The road level at Stop 13 is at the contact between the Leadville Dolomite and the Belden Formation. The uppermost part of the Leadville, the Castle Butte Member, can be seen without climbing; it consists primarily of coarse grained dolostone which is of regional extent. This is the stratigraphic position of three of the large mantos at Gilman. Unaltered Pando Porphyry crops out above the highway.

207.6 STOP 14. Overview of Gilman mining district from the south. The roadcut here exposes sandstones of the Belden Formation. The Pando Porphyry and the Leadville Dolomite crop out a short distance downhill. Return north 2.6 km to STOP 15.

STOP 15. Roadcut behind the water tanks exposes the Minturn Formation 100-150 m directly above the No. 1 manto. No alteration or mineralization effects are visible,

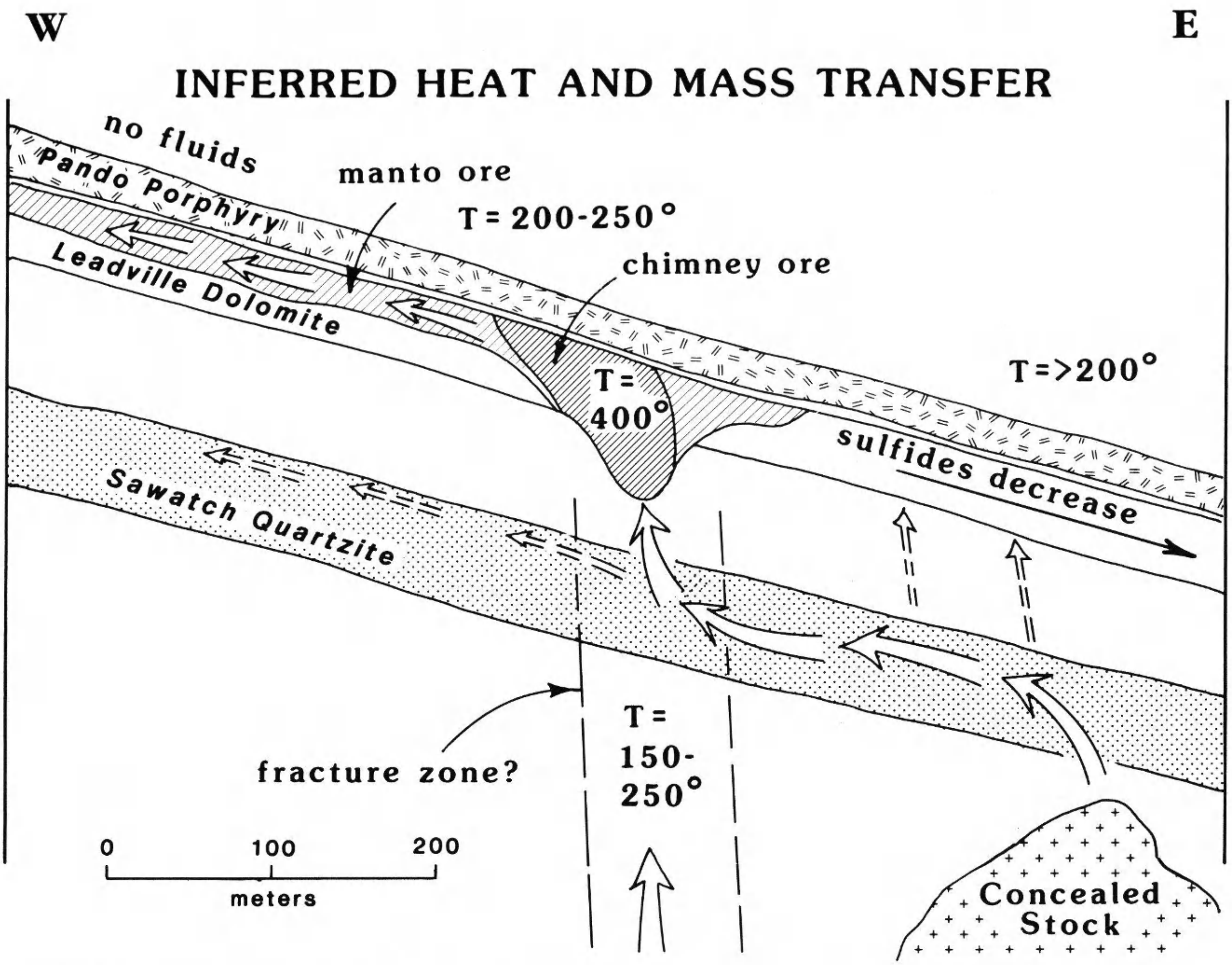

FIGURE 12 Inferred heat and mass transfer in the Gilman hydrothermal system.

despite the fact that this manto is 30 m thick. The total absence of hydrothermal effects indicates up-dip fluid transport within the Leadville. Return south on U.S. Highway 24.

208.2 STOP 16. Black Iron mine. Below the road is the outcrop of the No. 2 manto (the large dump is from the Black Iron mine). In the latter years of the Eagle mine, this adit was used for ventilation. The upper contact of the Leadville Dolomite is along the highway. Strongly altered Pando Porphyry is well-exposed along the drill road above the highway.

210.2 On left in roadcuts just before turnoff to Red Cliff and bridge unconformity between Cambrian Sawatch Quartzite and Early Proterozoic plutonic rock is exposed. About 2.5 km beyond on the right is the valley of Homestake Creek, which follows a northeast-trending shear zone of Proterozoic age.

220.9 Road to left to STOP 17 (optional). This large flat area is underlain by alluvial and lacustrine beds deposited behind a dam formed by the Homestake Creek glacier. During World War II, Camp Hale, a training facility for mountain troops, consisted of over 400 buildings and covered this flat area. Nearby the Leadville and Dyer Dolomites are extensively replaced by jasperpoid. These jasperoid masses have an unknown relation to Gilman-type mineralization processes, but it has been suggested that they represent the up-dip manifestation of those processes.

236.8 Tennessee Pass (3,172 m) on the Continental Divide. Tennessee Pass mining district consists of a 7 km belt of small stratabound gold deposits hosted by Leadville Dolomite (Beaty and others, 1987). Descend into the upper Arkansas graben, a northward extension of the Rio Grande rift. Large late Pleistocene morainal complex to right. Valley floor veneered by glacial outwash and Holocene alluvium above Neogene Dry Union Formation. As we approach Leadville the Mosquito Range is ahead. That range is composed of the eastern flank of the Laramide Sawatch uplift that has been separated from the modern Sawatch Range by the upper Arkansas graben.

City of Leadville at 3,094 m, the highest substantial town in the United States.

Third Day, Wednesday, July 5.

Head north of Colorado Highway 91 to Fremont Pass. Cross morainal deposits of the East Fork of the Arkansas glacier. Much faulted lower Paleozoic rocks intruded by upper Cretaceous-Paleocene porphyries form the valley side to the northwest. Where road crosses to the northwest side of the stream is a cliff formed by Sawatch Quartzite and a sill of hornblende latite porphyry. Just beyond and all the way to Fremont Pass, poorly exposed Minturn Formation intruded by sills of quartz latite porphyry.

15.4 On left, Chalk Mountain formed by stock of 28 Ma leucorhyolite.

18.7 Fremont Pass (3,450 m) on the Continental Divide. Enter the Climax mine.

THE PORPHYRY MOLYBDENITE DEPOSITS AT CLIMAX, COLORADO

S.R. Wallace
Private Consultant, Lakewood, Colorado
Geoffrey G. Snow,
Barranca Resources, Golden, Colorado

INTRODUCTION

Discovery

A gray metallic mineral, variously identified as galena, graphite, and silver, was discovered in 1879 by Charles J. Senter, amid the prominant iron-stained slopes of Bartlett Mountain, between the then-active mining camps of Leadville and Kokomo.

Market Development

With the correct identification of the mineral as molybdenite in 1895 and the onset of World War I in 1914, interest in the deposit grew, and shipments were made from 1915 to 1919 to meet the demand for new steel alloys used in the implements of battle. The end of hostilities presaged a collapse of the market, and there was no further production until 1924, when a small demand had developed for peace-time applications of superior iron and steel.

Mining History and Production

The Climax mine reopened in August of 1924 and was in essentially continuous production until mid-1982. The mine reopened in 1984 and operated intermittently until March 1987, when it shut down indefinitely.

From 1918 to March of 1987, the Climax mine produced 421.5 million tonnes of ore, yielding 858 million kilograms of molybdenum. Corrected for dilution in the mine and losses in the mill, these numbers equate to an average grade of 0.410 percent MoS_2 to a 0.2 percent MoS_2 cutoff[1].

Remaining geologic reserves of combined open pit and underground ore are estimated at 310 million tonnes with a grade of between 0.30 and 0.35% MoS_2 for a total of about 725 million tonnes of mined and mineable ore[2]. Considering that most of the Ceresco orebody, and a significant part of the high-grade portion at the top of the upper orebody, were removed by pre-mining erosion, the total geologic reserve may very well have exceeded one billion toness of plus 0.40% MoS_2. Climax, as far as is known, was the world's greatest deposit of molybdenite.

From 1948 to 1987 the recovery of tungsten as huebnerite and tin as cassiterite from by-product circuits in the

[1]Mining grade including dilution.

[2]Depending on price.

belt of small stratabound gold deposits hosted by Leadville Dolomite (Beaty and others, 1987). Descend into the upper Arkansas graben, a northward extension of the Rio Grande rift. Large late Pleistocene morainal complex to right. Valley floor veneered by glacial outwash and Holocene alluvium above Neogene Dry Union Formation. As we approach Leadville the Mosquito Range is ahead. That range is composed of the eastern flank of the Laramide Sawatch uplift that has been separated from the modern Sawatch Range by the upper Arkansas graben.

City of Leadville at 3,094 m, the highest substantial town in the United States.

Third Day, Wednesday, July 5.

Head north of Colorado Highway 91 to Fremont Pass. Cross morainal deposits of the East Fork of the Arkansas glacier. Much faulted lower Paleozoic rocks intruded by upper Cretaceous-Paleocene porphyries form the valley side to the northwest. Where road crosses to the northwest side of the stream is a cliff formed by Sawatch Quartzite and a sill of hornblende latite porphyry. Just beyond and all the way to Fremont Pass, poorly exposed Minturn Formation intruded by sills of quartz latite porphyry.

15.4 On left, Chalk Mountain formed by stock of 28 Ma leucorhyolite.

18.7 Fremont Pass (3,450 m) on the Continental Divide. Enter the Climax mine.

THE PORPHYRY MOLYBDENITE DEPOSITS AT CLIMAX, COLORADO

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INTRODUCTION

Discovery

A gray metallic mineral, variously identified as galena, graphite, and silver, was discovered in 1879 by Charles J. Senter, amid the prominant iron-stained slopes of Bartlett Mountain, between the then-active mining camps of Leadville and Kokomo.

Market Development

With the correct identification of the mineral as molybdenite in 1895 and the onset of World War I in 1914, interest in the deposit grew, and shipments were made from 1915 to 1919 to meet the demand for new steel alloys used in the implements of battle. The end of hostilities presaged a collapse of the market, and there was no further production until 1924, when a small demand had developed for peace-time applications of superior iron and steel.

Mining History and Production

The Climax mine reopened in August of 1924 and was in essentially continuous production until mid-1982. The mine reopened in 1984 and operated intermittently until March 1987, when it shut down indefinitely.

From 1918 to March of 1987, the Climax mine produced 421.5 million tonnes of ore, yielding 858 million kilograms of molybdenum. Corrected for dilution in the mine and losses in the mill, these numbers equate to an average grade of 0.410 percent MoS_2 to a 0.2 percent MoS_2 cutoff[1].

Remaining geologic reserves of combined open pit and underground ore are estimated at 310 million tonnes with a grade of between 0.30 and 0.35% MoS_2 for a total of about 725 million tonnes of mined and mineable ore[2]. Considering that most of the Ceresco orebody, and a significant part of the high-grade portion at the top of the upper orebody, were removed by pre-mining erosion, the total geologic reserve may very well have exceeded one billion toness of plus 0.40% MoS_2. Climax, as far as is known, was the world's greatest deposit of molybdenite.

From 1948 to 1987 the recovery of tungsten as huebnerite and tin as cassiterite from by-product circuits in the

[1]Mining grade including dilution.

[2]Depending on price.

mill has yielded 24.7 million kilograms of WO_3 and 1.5 million kilograms of Sn.

In the early years mining was done by shrinkage stoping. Dilution was essentially nil, extraction was from a high grade part of the upper ore body, and mill feed averaged 0.9% MoS_2. As demand increased so did production rates. The ore was sufficiently fractured that panel caving could be effectively used to mine large volumes at lower cost per unit. As production increased so did efficiency and cutoff grades dropped top 0.2% MoS_2. Since the early 1970's portions of the Ceresco and upper ore bodies have been mined by open pit.

GEOLOGY

General Statement

At Climax a composite stock, with associated dikes, of highly evolved rhyolitic composition was emplaced over a span of 9 m.y. from mid-Oligocene to earliest Miocene. Country rocks for this intrusive complex are 1,700 to 1,800 Ma metamorphic rocks, 1350 to 1480 Ma granitic rocks, and quartz monzonite porphyry dikes and sheets of Laramide age.

The textures of Climax rocks forming the composite stock range from aplitic to rhyolitic porphyries to porphyritic granite. Four separate productive phases of the stock were emplaced between 33 and 26 Ma. The ages of these intrusive events are shown in Table 2.

The Mosquito fault, a west-dipping normal fault, is a major structural feature and marks the western front of the Tenmile Range at Climax. It has displaced part of the Ceresco orebody between 2,700 and 3,000 m.

The orebodies

Molybdenite at Climax occurs in three distinct but overlapping stockwork bodies, each related spatially, temporally, and genetically to one or more of four productive phases of the Climax composite stock. The three orebodies are the Ceresco orebody, which is the uppermost and oldest orebody, the upper orebody, which in fact, is below the Ceresco orebody, and the lower orebody, which is the deepest and youngest of the three (Figs. 13 and 14[1]).

Each orebody is shaped like an inverted bowl or shell and is circular to ring-shaped in plan and arcuate in section. As outlined by the 0.2% MoS_2 grade contour, the ore shells are about 150 to 200 m thick with maximum outer diameters in plan of from 900 to 1,200 m. Each orebody is centered around and above its source intrusion. The Ceresco orebody is related to the southwest mass of the Climax stock, the upper orebody to the central mass of the Climax stock, and the lower orebody to a deeper phase of the central mass, the aplite porphyry.

Bodies of rock mineralized with low-grade tungsten stockworks are present within the upper part of the genetically related molybdenite zones and/or halos on the orebody hanging wall; tungsten-bearing veinlets are younger than their related molybdenite-bearing veinlets in places where the two zones overlap.

The upper orebody is itself a dual orebody related to two separate intrusive phases of the central mass of the Climax stock. These two phases are very nearly spatially coincident as are the mineralized zones which they produced (Fig. 15). The zonation of MoS_2 in plan and on section is mimicked by that of genetically related WO_3 above and peripheral to the molybdenite. These patterns of distribution clearly indicate the multiple origin of the upper orebody.

The late phase of the central mass (but predating the lower aplite porphyry phase) is the presumed source of the second major pulse of mineralization which formed the inner strands of molybdenum and tungsten.

Mineralized zones of the "late barren stage". This hydrothermal event is about 1.7 m.y. younger than the lower orebody and is related to the porphyritic granite phase of the Climax stock. Near the top of the porphyritic granite phase the rock contains small sparse flakes of disseminated molybdenite. Above the disseminated zone, this phase of the stock and the rock above its upper contact contains a small, weakly developed stockwork of quartz-molybdenite veinlets. Within this zone base-metal sulfides and coarse-grained fluorite and rhodochrosite are locally abundant. Mineralized zones of this stage do not make ore.

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[1]Recent work has demonstrated that the late rhyolite dikes are cut off by the porphyritic granite on the 929 level (Fig. 2 of White and others, 1981). The present authors have retained the older illustrations since they are uniform with respect to scale, map symbiols, and nomenclature and show features not presented elsewhere.

TABLE 2 Ages of Climax events
[Data from White and others (1981) and Bookstrom (in press)]

Material and/or event dated	Age (Ma)	Method	Mineral
Rhyolite porphyry from the southwest mass of the Climax stock (Ceresco orebody).	33.2±2.1	Fission track	Zircon
One of two phases of upper orebody (one of two phases of central mass of Climax stock).	29.8±0.4	K/Ar	Sericite
Aplitic porphry phase of the central mass of the Climax stock (lower orebody).	26.1±1.2	Fission track	Zircon
Late rhyolite porphyry dikes.	25.1±1.2	Fission track	Zircon
Mineralized zone of the "late barren stage" (porphyritic granite phase of the Climax stock).	24.4	K/Ar	Sericite

Hydrothermal Alteration

General Statement. Hydrothermal alteration at Climax has been intense in many parts of the mine, particularly on the Phillipson level and above. Where alteration zones related to upper orebody events are over-printed by those of lower orebody events, texture-destructive alteration is common, and several generations of geologists have disagreed on the identity of original rock types. Points of contention have been:

(1) Laramide quartz monzonite porphyry vs. Proterozoic schist, and

(2) Mid-Tertiary rhyolite porphyries vs. Proterozoic granite.

Beyond this was the problem of separating the different phases of the ore-related intrusions.

A generalized zonal arrangement of hydrothermal alteration products from the top of the causative intrusion upward and outward are:

Footwall silica zone,
K-feldspathized zone,
Phyllic zone,
Argillic zone, and
Propylitic zone.

The most common minerals are quartz, potassium feldspar, sericite, and fluorite.

Silicified zone. The apices of the causative intrusions and the rocks above these cupolas have been replaced by fine-grained hydrothermal quartz. Silica flooding is virtually complete over large areas on the footwall of the upper and lower orebodies, especially the upper orebody where the north footwall drift cut almost 300 m of light gray to white "quartzite" containing a few small and scattered remnants of pre-existing, but generally unidentifiable, rock types. Geologic projection indicates Precambrian rocks cut by Laramide porphyries.

Feldspathized Zone. The orebodies are best developed in the K-feldspathized zones above and peripheral to the footwall silicified zones. Conversion of Precambrian rocks to pinkish-tan feldspar is essentially complete over distances of a few hundred meters on some parts of the Phillipson Level, and above.

Phyllic Zone. Outward from the feldspathic zone is a halo of fine-grained quartz-sericite rock containing disseminated pyrite. These replacement products are clearly related to quartz-pyrite (sericite) veinlets; a single 0.5 cm- veinlet can produce a distinct halo of 8 cm on either side of the fracture, in which the color is changed and the rock fabric is obliterated. Tungsten and tin are best developed in the phyllic zone, as is topaz. Fluorite is common.

Peripheral Zones. Beyond the phyllic zone is an argillic zone and still farther

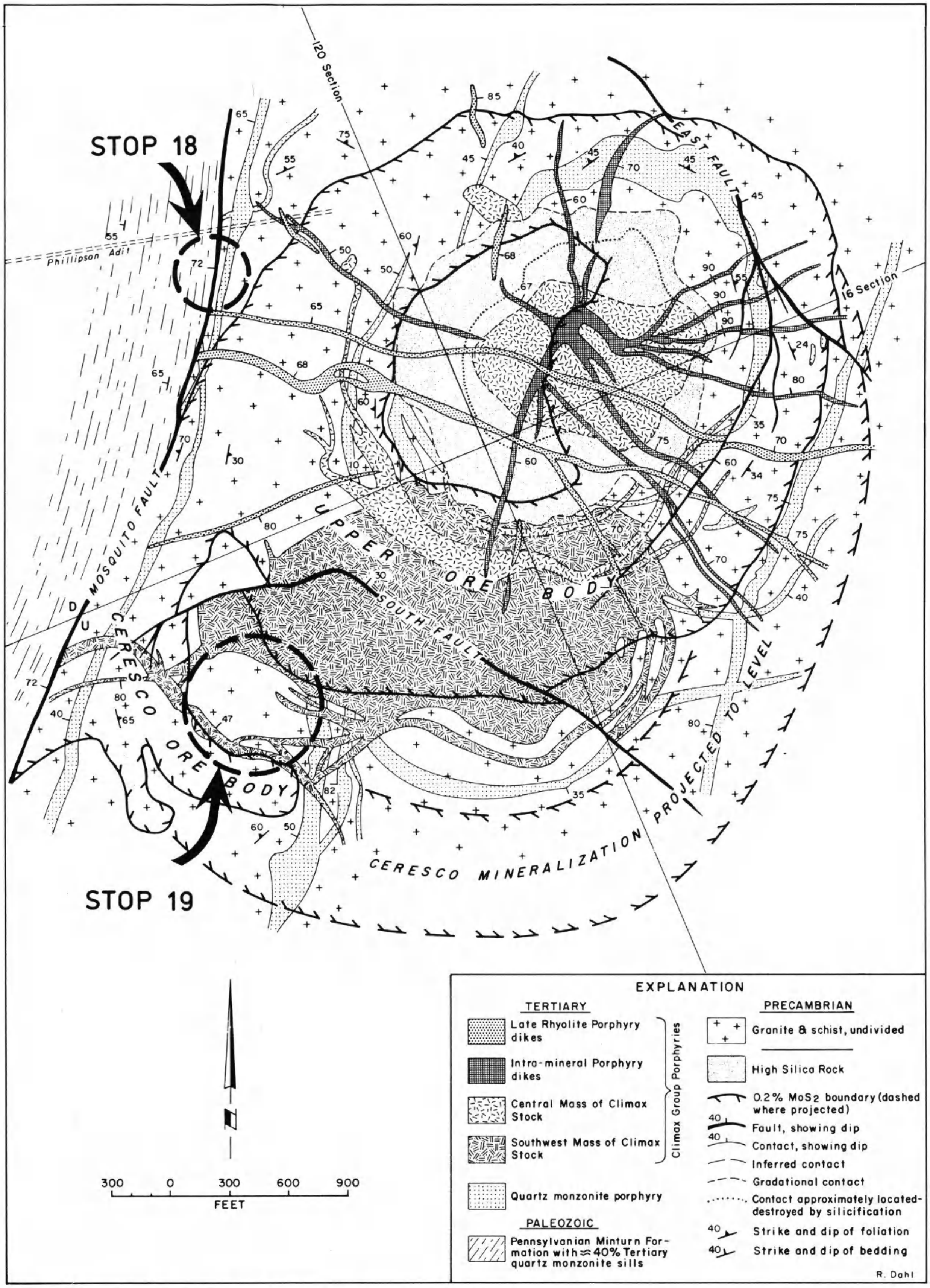

FIGURE 13 Map of the Phillipson level of the Climax mine, showing generalized geology, ore zones, and location of stops. From Wallace and others (1968). Reproduced by permission of The Society of Mining Engineers, Inc.

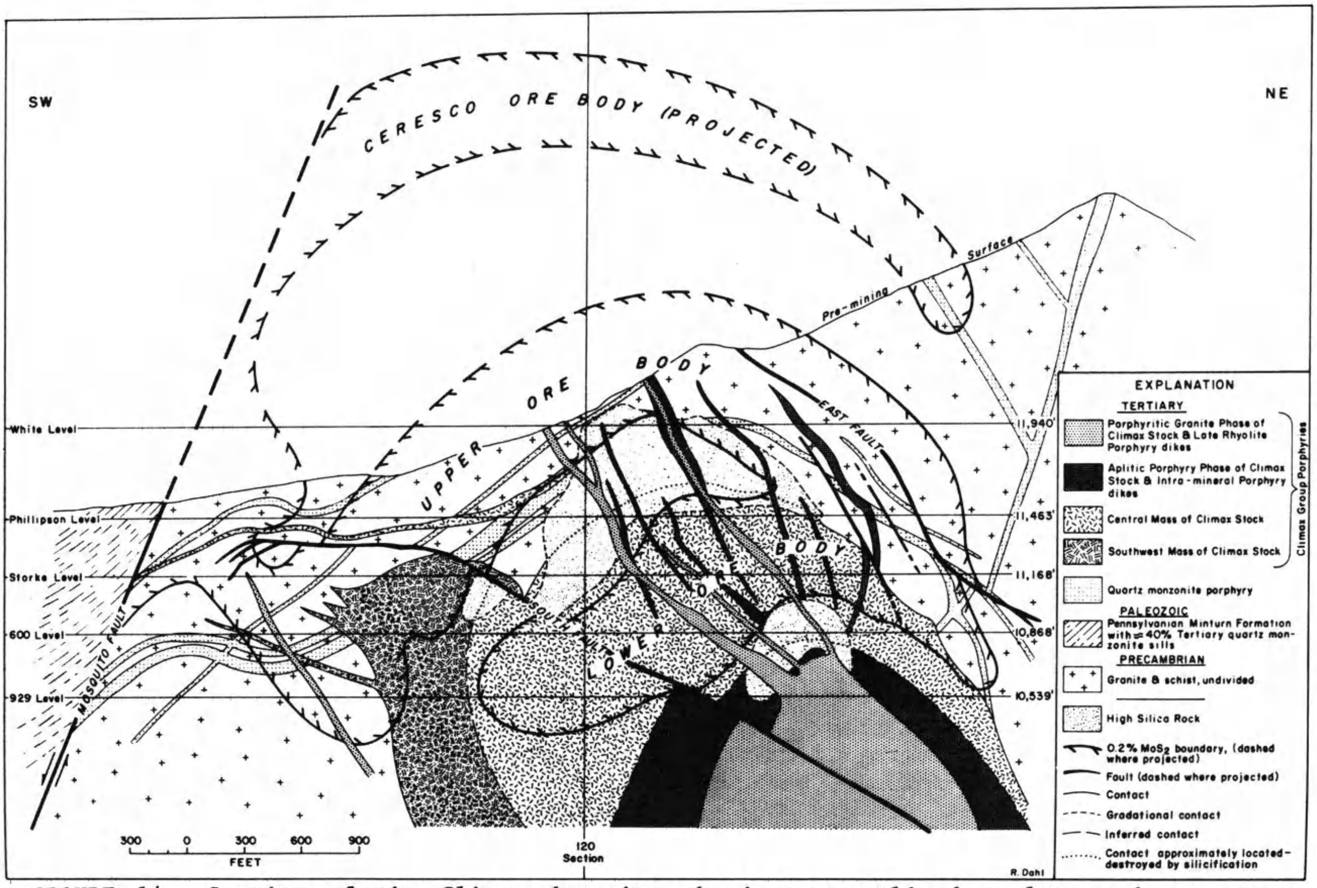

FIGURE 14 Section of the Climax deposit, showing generalized geology and ore zones. From Wallace and others (1968). Reproduced by permission of The Society of Mining Engineers, Inc.

out, a propylitic zone, which forms a halo to the whole system. The outer boundary of the propvlitic zone grades into fresh rock about 2 km east of Climax and 4 km north and south of the mine area; peripheral zones of mineralization are cut off to the west by the Mosquito fault.

Origin of the ores

The repetition of unusual yet very nearly identical magmatic-hydrothermal events in almost the same place indicates the presence of a master reservoir at depth. The cluster of three orebodies, and a period of roughly 3.5 m.y. between mineralizing events for a total time span of 7 m.y. requires a long enduring cupola to concentrate the hyperfusibles and metals that were generated by crystallization within the master reservoir. The size of the cupola is envisioned as being large relative to that of the productive intrusions, either igneous columns or diapirs as suggested by White and others (1981).

Carten and others (1988) have calculated a concentration of 1.3% molybdenum in the upper part of one of the productive intrusive phases of the Henderson deposit system, a geologically similar system about 56 km north-northeast of Climax. This extreme enrichment in Mo as well as F, Rb, Sn, W, Nb, Ta, Sc, and Th is typical of such rocks and systems.

The periodicity of ore-forming events suggests slow cystallization and enrichment by fractional crystallization of the magma in the master reservoir. Final concentration of metals and incompatible elements may represent a second stage of enrichment in which thermogravitational diffusion (Hildreth, 1979), perhaps combined with convection and/or gas-streaming within cylindrical columns or elongate diapirs was the dominant mechanism.

The stops are both within the open pit; they are shown on Figure 13, a geologic and ore-zone plan of the Phillipson level, which is about 30 m below the benches we will visit.

STOP 18. Northwestern part of the pit. Carboniferous sedimentary rock of the Belden and possibly the Minturn Formations enclose a Laramide quartz monzonite porphyry

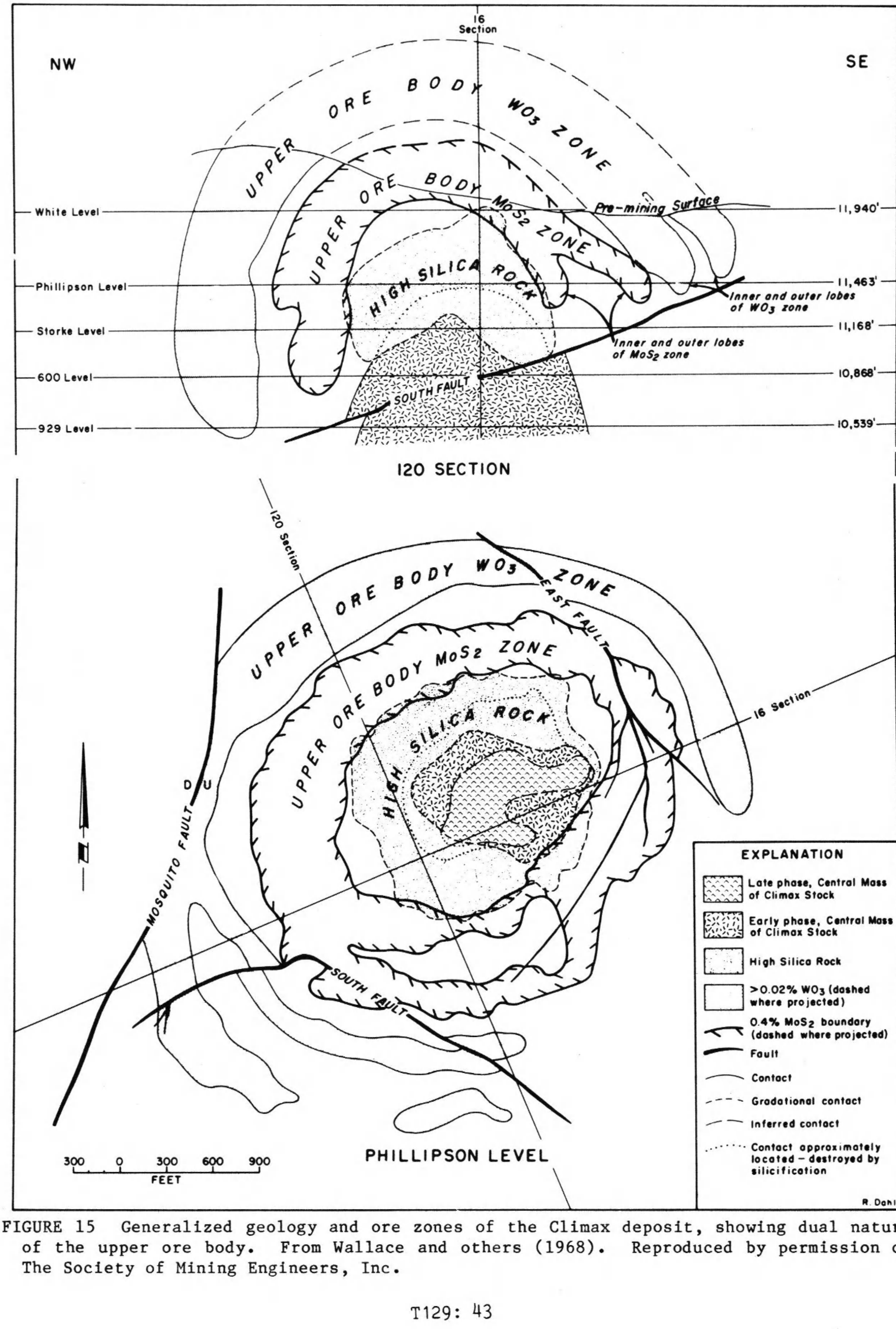

FIGURE 15 Generalized geology and ore zones of the Climax deposit, showing dual nature of the upper ore body. From Wallace and others (1968). Reproduced by permission of The Society of Mining Engineers, Inc.

sill upturned along the Mosquito fault; altered Proterozoic rocks on the footwall side of the fault zone.

STOP 19. Southwestern area of the pit. The large flat area of the bench is underlain largely by barren to low-grade mineralized rock between the hanging wall of the upper ore body and the footwall of the Ceresco orebody. Mineralized zones of both orebodies are exposed along the south and east perimeters of the bench. Rocks exposed are Proterozoic schist and gneiss intruded by several varieties of Middle Proterozoic granite of the Berthoud Plutonic Suite, by Laramide quartz monzonite porphyry, and by rhyolite porphyry of the southwest mass of the Climax stock. Fine-grained clastic dikes of breccia related to lower orebody events cut the southwest mass along the northeast side of the bench. Quartz-pyrite veinlets containing dark brown to black sphalerite (marmatite) with or without sericite (topaz, fluorite, and huebnerite) are in the phyllic zone on the hanging wall of the lower orebody and may be observed in this area where they are superposed on the upper orebody molybdenite zone.

Return to Leadville.

THE LEADVILLE MINING DISTRICT

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INTRODUCTION

History

The Leadville District was discovered by panning for gold in the spring of 1860. By July there were 10,000 miners working the California Gulch placer, and by 1868 the placer was mined out. The first major silver lode ores were discovered in 1874, and in 1875-1876 a second rush of about 40,000 miners took place. Many of the orebodies were mined out within a few decades. Many others closed in the metal price collapses of 1893 and 1918. By 1927 the district was in serious decline. In 1938, ASARCO and Newmont became interested in the district, and initiated the exploration program that culminated in development of the Black Cloud mine (Smith, 1988). The Black Cloud is a major, modern production facility that has been in operation since 1971.

In addition to the ore produced, the Leadville District is significant for the pioneering geological work that was completed there. The first two mining districts in the United States studied in detail by the U.S. Geological Survey were Leadville and Comstock (Nevada), and those two studies stand as landmark contributions to the science of economic geology. The Leadville work was carried out by S.F. Emmons, who published a famous monograph in 1886. Many of Emmons' concepts of ore genesis have withstood the test of time.

Mining and processing

The Black Cloud Mine, the only modern facility in the Leadville District, uses a room and pillar, cut and fill method. Rubber-tired equipment is used in the stopes, and a rail haulage transports ore to the Black Cloud shaft, which is 505 m deep. The ore is concentrated in a flotation mill at the shaft collar. The mill produces a lead concentrate (65% Pb, mill recovery = 87%) and a zinc concentrate (50% Zn), both of which are Ag- and Au-bearing. Copper is not recovered. The coarse-grained tailings are used to backfill the old stopes, the fine-grained tailings are sent to a pond in Iowa Gulch. The concentrates are shipped by rail to smelters at East Helena, Montana and El Paso, Texas. The production rate of the mine is 720 tonnes per day, and total employment is 160 persons.

Production/Reserves

Cumulative production of the Leadville District through 1987 is estimated at 10,700 kg of placer gold, and about 23.8 million metric tonnes of lode ore with an average grade (using metal recovery data) of 0.2% Cu, 3.0% Zn, 4.2%, Pb, 3.65 gpt Au, and 320 gpt Ag. Reserves at the Black Cloud mine as of December 31, 1987 were 805,000 metric tonnes averaging 2.4 gpt Au, 0.20% Cu, 3.95% Pb, 8.09% Zn, and 68.1 gpt Ag (ASARCO 1987 Annual Report).

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GEOLOGY

General Geology

Sedimentary rocks. The Leadville district is located on the eastern flank of the Laramide Sawatch uplift. The ores are hosted by a thin (<200 m) Cambrian to Mississippian sequence of sedimentary rocks which dip 10 to 25^{o} eastward and are exposed repeatedly (Fig. 16) by a series of north-striking, west-dipping normal faults. Most of the ore at Leadville has come from orebodies in the Mississippian Leadville Dolomite.

Igneous rocks. Upper Cretaceous to middle Tertiary igneous rocks intrude the Paleozoic rocks in the district. At least five separate pre-ore igneous rock units are distinguished. These units (the Pando, Lincoln, Evans Gulch, Sacramento and Johnson Gulch Porphyries) are monzonite to quartz monzonite in composition, and occur as and sills, dikes and stocks, that either expand or eliminate the Paleozoic section. The oldest, the Pando Porphyry, was emplaced primarily along the Upper Mississippian unconformity on top of the Leadville Dolomite or synchronously along low-angle Laramide thrust faults. The youngest of the pre-ore Tertiary igneous bodies, the Johnson Gulch Porphyry, forms a stock beneath Breece Hill (Fig. 16) and cuts the Evans Gulch Porphyry. Fission-track dates on apatite and zircon from unaltered Sacramento and Johnson Gulch Porphyries yield 43.9 and 43.1 Ma, respectively (T.B. Thompson, unpublished data). The Johnson Gulch Porphyry has a late phase that resembles the Lincoln Porphyry. Consequently the Johnson Gulch Porphyry was thought to be older than the Lincoln Porphyry, which is 65.6 Ma at a locality about 10 km away from the Leadville district (Pearson and others, 1962). Post-mineral rhyolitic breccia pipes, locally with massive sulfide replacement ore fragments, appear to have been emplaced immediately after formation of the principal Leadville district ore deposits, as they are locally cut by veins with similar mineralogy and fluid inclusion data (Hazlitt, 1985).

Faulting. Laramide low-angle to steep reverse faults and later normal faults cut the Paleozoic sedimentary rocks (Fig. 16). Tweto (1960) was able to discern a sequence of faulting based on localization of various igneous rocks along the faults or by destruction of the fault by subsequent intrusion. Middle- to late-Tertiary extensional tectonics associated with the Rio Grande rift exposed the up-dip portions of the replacement orebodies, facilitating their early discovery and beneficiation. One graben, referred to as the "down-dropped block" occurs immediately east of Breece Hill and is the site of current mining within the district.

Ore deposits

Orebodies within the main part of the Leadville District developed in two separate events: 1) skarns formed during emplacement of the Breece Hill stock; and 2) sulfide vein and mantos formed later. The skarns are magnetite-rich, and occur around the perimeter of the Breece Hill stock as well as in large blocks within the stock. Calc-silicate minerals have been converted to serpentine by subsequent hydrothermal activity and massive magnetite has been replaced by sulfides in many places.

The second event, also focused around the Breece Hill stock, generated veins and dolomite-hosted massive sulfide replacement deposits (mantos). These ore types have been responsible for most of the production in the district. Surface exposure of the veins and mantos and subsequent physical transport of weathered material led to the development of placers. Only the placer deposits in California Gulch (Fig. 16) survived, because all the other canyons were scoured by Pleistocene glaciers.

Veins. Veins in the district consist of quartz-pyrite-gold within and adjacent to the Breece Hill stock, and quartz-pyrite-base metal sulfides peripheral to the gold-bearing veins. Both types are commonly associated with mantos, where dolostone country rock is cut by the veins (Behre, 1953; Thompson and others, 1982).

Mantos. Mantos at Leadville are massive sulfide replacement bodies composed of pyrite, marcasite, marmatite, galena, chalcopyrite, tetrahedrite, pyrrhotite, electrum, and a variety of trace minerals. The ore is typically 60-100% sulfide, and gangue minerals include quartz, siderite, dolomite and late barite. Some sulfide bodies lack apparent banding, but others exhibit intricate banding. The banding parallels bedding in the host rocks or fractures. The bands are marked by different mineralogies, size variations of mineral constituents, or textural variations, particularly of the iron sulfides.

Metal zoning. Metal zoning in the

vein-manto system ranges between +5 to +10, while δD_{H_2O}= -45 to -70. These values lie within the compositional range interpreted as magmatic water. Late golden barite found throughout the Leadville district clearly indicates the influence of meteoric water ($\delta^{18}O_{H_2O}$ = -18).

Sulfur isotopic analyses ($\delta^{34}S$) of pyrite (1.3 to 3.2), sphalerite (-0.5 to +2.2), galena (-2.4 to +0.7) and golden barite (+7.8 to +15.6) yield a calculated $\delta^{34}S_{\Sigma S}$ = +1.8. Sulfide mineral pairs (pyrite-marmatite and marmatite-galena) yield temperatures within the ranges obtained from pressure-corrected fluid inclusion filling temperatures.

Lead isotope analyses of pre-ore Tertiary igneous orthoclase and galena from Leadville contrast sharply with the Leadville Dolomite lead ($^{206}Pb/^{204}Pb$ = 21.2 to 22.7) and the Precambrian basement rocks ($^{208}Pb/^{204}Pb$ > 36). The galena leads (206/204: 17.46-18.26; 207/204: 15.51-15.59; 208/204: 38.18-38.59) are more thorogenic compared to the orthoclase lead (208/204: 37.9-38.3), but otherwise they are compositionally identical. The data from Leadville define an isochron that intersects the model curve at 1.70 Ga, the age of basement rocks in central Colorado.

Geochronology. The extensive ore-related alteration of porphyritic igneous rocks at Leadville is ideal for fission-track dating methods. Apatite (where not destroyed by the hydrothermal alteration) and zircon from the Johnson Gulch Porphyry yielded an average age of 33.4±5.1 Ma from 15 samples collected within the thermal and alteration halos of the mantos.

Genetic Model. The mineralogical-metal and thermal zoning about the Breece Hill center suggests that a thermal source lies beneath the exposed pre-ore intrusive center. The heating event associated with ore emplacement has been dated by the fission-track method at about 34 Ma. The ore-related magma was derived from a deep crustal source based on the lead isotopic data. Similarly, the sulfur is of magmatic origin. Ore fluids of magmatic origin spread upward along low- and high-angle faults and, to a lesser extent, along bedding planes in the lower Paleozoic rocks. Ore fluids were hot (469° to 310°C), weakly acidic, and reacted rapidly with carbonate rocks to form the mantos. Lower reaction rates between the ore fluid and igneous rocks led to wider alteration halos in igneous rocks than in dolostones. The vertical and lateral movement of ore fluids and their interaction with rocks at depths of 3,600 m led to ore fluid cooling, isotopic exchange, and, on the high levels and margins of the system, mixing with meteoric fluids. Late in the main stage of the mineralizing event, fluidized breccias were emplaced, cutting ore. The breccias were locally veined by late-stage minerals with similar fluid inclusion thermometric data and isotopic compositions. The breccia distribution around the Breece Hill stock and throughout the down-dropped block (Fig. 16), suggests that the entire area beneath the stock and graben were underlain by the magmatic source system. Expulsion of the large volume of ore fluid and fluidized rock column in the breccias resulted in roof collapse with graben formation.

STOP 20. Black Cloud mine (ARSCO-Newmont Mining Company joint venture). The mine has been in continuous production since 1971, processing approximately 720 tonnes a day. Cumulative production to date has been 3 million tonnes grading 2.7 gm Au, 81.2 gms Ag, 3.9% Pb, and 8.0% Zn. Mine tour.

STOP 21 (optional). The North Moyer mine is located on the eastern side of four major northeast-striking mantos beneath Iron Hill (Fig. 17). The composite orebody projection map (Fig. 17) gives the shaft location. Orebodies replaced dolostones as well as being localized along an east-dipping Laramide thrust fault, the Tuscon-Maid fault. Northeast-striking veins were described in the Cord mine beneath Iron Hill by Emmons and others (1927), as well as northeast-striking breccia zones along the Moyer fault. Johnson Gulch Porphyry dikes exhibit a similar strike.

STOP 22 (optional). Regional overview from the Venir mine dump. To the west is the Laramide Sawatch uplift cored by Precambrian rock. Turquoise Lake, dammed by a late Pleistocene terminal moraine and a man made structure, is used to generate electric power. The Turquoise Lake stock (34 Ma), which hosts molybdenum mineralization, occurs midway along the lake. Mine dumps north and south of the lake are from narrow veins of the St.

Kevin and Sugarloaf districts in Precambrian host rocks. The lower part of the valley localizes the Arkansas River within the northern part of the upper Arkansas graben, a northerly extension of the Rio Grande rift. Fluvial and lacustrine sediments of the upper Tertiary Dry Union Formation underlie dissected pediments extending toward the river and were deposited along in the rift valley.

The Venir mine (3,565 m) is developed on four levels with intervening stopes along northeast-striking veins in Johnson Gulch Porphyry and lesser amounts of quartz sandstone and shale of the Pennsylvanian Belden Formation. The mine operated at intervals between 1920 and 1935 "as a small but profitable" mine (Behre, 1953). The ores were narrow (<0.5 m) quartz-pyrite-gold veins.

STOP 23 (optional). The Penn Group mine area has surface exposures of magnetite skarn. These skarns extend southward 600 m along the western margin of the Breece Hill stock. The mineralization is dominantly magnetite, is locally siliceous, and was used as flux in smelting. In the underground workings the magnetite is replaced by sulfides (pyrite, chalcopyrite, marmatite, and galena) and has 19 to 187 gm/tonne Au along one vein and in several mantos. The exposed bodies at this stop are magnesian skarn magnetite at the contact with the Johnson Gulch Porphyry, the dominant igneous rock of the Breece Hill intrusive center. All calc-silicates are replaced by serpentine as a result of post-skarn hydrothermal activity. A post-mineral flow-banded rhyolite porphyry is present along the eastern part of the area. An exposure of the Belden Formation in the eastern part of the area suggests that the magnetite skarn replaced the Leadville Dolomite.

STOP 24 (optional). The famous Matchless and Robert E. Lee mines produced oxidized ores from massive replacement deposits in the Leadville Dolomite sandwiched between irregular sill-like bodies of Pando and Johnson Gulch Porphyries. The orebodies, discovered in 1878, were oxidized and yielded as much as 312.5 kg/tonne. The Robert E. Lee produced 3,346 kg of silver in a 17-hour period in 1878. Both oxidized and sulfide ore can be found on the dump. Late golden barite is present in vugs within the ore and wallrock.

From Leadville our route heads south along the upper Arkansas graben, which breaks the Laramide Sawatch uplift into the Sawatch Range on the west and the Mosquito Range on the east. The Miocene and Pliocene Dry Union Formation, which fills the graben, is disected and, in places, mantled by Quaternary moraine and outwash deposits. The width of the graben marked by the fill varies due to cross faults. In late Eocene and early Oligocene time drainage flowed east cross the site of the graben at the same time that extensive volcanism occurred in the central part of the Sawatch Range.

As we leave Leadville, Mt. Elbert (4,399 m, the highest peak in Colorado) and Mt. Massive (4,394 m) dominate the skyline to the west.

1.4 Large slag pile from Leadville smelter. Water well at smelter bottomed in Dry Union Formation at 600 m.

5.6 Sharp left curve. Lowest gravity low (-338 mgals) in the conterminous United States in valley bottom to right. Late Tertiary valley fill estimated to be 900-1200 m thick; the bedrock surface at less than 2000 m altitude; peaks on west side of valley 4,300-4,400 m.

9.5 Hills at 10 o'clock formed by Dry Union Formation; crest of Mosquito Range to east capped by gently east-dipping lower Paleozoic rocks on the east margin of the Laramide Sawatch uplift.

16.9 Precambrian granitic rock on left. A NE-trending fault separates wide part of graben to north from narrower part to south.

19.8 Shallow canyon in Early Proterozoic granite. A fault 100-200 m west of the canyon wall separates this granite from a thick prism of Dry Union Formation. From here south to Buena Vista numerous north to north-northwest-trending faults cut the Precambrian rocks along the east side of the valley.

29.5- Late Pleistocene morainal complex of Clear Creek glacier. Here the
33.3 graben as defined by the Dry Union Formation constricts to a few hundred meters width due to a series of east-trending faults to the north.

34.5 Late Pleistocene moraine of the Pine Creek glacier. Downvalley notice the scattered large boulders in the outwash deposited during periodic flooding due to breaching of the dam on the Arkansas River by the Pine Creek glacier.

39.8 To left pipeline and pumping station in water diversion system owned by the cities of Aurora (a suburb of Denver) and Colorado Springs.

44.3 Mount Harvard (4,394 m) at 3 o'clock. This peak is on a strongly uplifted horst on the west side of the upper Arkansas valley graben (Shannon and others, 1987). Four miles to the south Mt. Yale (4,326 m) visible at 2 O'clock is part of the same horst.

55.3 Center of Buena Vista.

59.0 Turn left on U.S. 285.

60.7 Turn left on Chaffee County road 304 to viewpoint of Arkansas valley. STOP 25 (optional). The floor of the Arkansas valley here is mantled by outwash gravels from glaciers in the Sawatch Range, which ended near the range front. The gravels overlie Dry Union Formation. Geophysical data indicate that the bedrock slopes sharply westward towards the Sawatch Range and that the valley fill adjacent to the fault bounding the west side of the graben is about 1.2 km thick. Resistivity data indicate step faulting in the basement surface near the east side of the valley.

In the eastern part of the Sawatch Range overlooking the valley are a series of peaks more than 4,300 m in altitude. They occur in a narrow horst and are several hundred meters higher than the Continental Divide farther west in the range. Apatite fission-track dates of 15-23 Ma from the block indicates that it was uplifted in Miocene time (Shannon and others, 1987). Mount Princeton is part of a large pluton about 25 km in diameter emplaced at 36.6 Ma (Ed Dewitt, oral commun., 1988). Numerous dates of a regionally extensive ash flow tuff, the Wall Mountain Tuff, indicate that it is the same age as the Mount Princeton pluton, suggesting that a caldera formed at that site and was immediately obliterated by intrusion of the pluton (Ed Dewitt, oral commun., 1988). Subsequently, other calderas formed along the axis of the Sawatch uplift. The Mount Aetna caldera formed in the Mount Princeton pluton and produced ash flow tuffs that filled valleys leading eastward into South Park.

The white outcrops along the major valley on the south side of Mount Princeton are zeolitized plutonic rock associated with hot spring activity. Water at a temperature of 58°C issues from the bottom of the valley in the altered rocks.

The high ridge to the southeast, south of Trout Creek, is capped by welded tuff and lava flows that filled a paleovalley leading east-northeast away from the Mount Princeton area. Distribution of granites and rhyolite and lamprophyre dikes dated at 30-28 Ma by fission-track determinations on zircon suggests that 28 Ma is a maximum age for the beginning of extensional deformation and foundering of the upper Arkansas valley. Geologic reconstruction suggests 2-4 km of erosion along the crest of the central Sawatch Range since formation of the Mount Aetna caldera.

To the east lower Paleozoic rocks cap the skyline ridge near the east margin of the Laramide Sawatch uplift.

Return to U.S. Highway 285 and turn right. At 62.9 highway turns south down the graben.

South of road junction are roadcuts in gravels of the Dry Union Formation beneath middle Pleistocene alluvium. Sharp hill across the river at 68.2 is Sugarloaf Mountain composed of steeply tilted 29-30 Ma rhyolite. Cross outwash gravels of various ages derived from glaciers of the Sawatch Range. To the right, north to south, are Mount Princeton, Mount Antero, and Mount Shavano.

79.6 Bluffs to right composed of Dry Union Formation.

84.1 To left, just across river, cliffs of Wall Mountain Tuff downfaulted into the graben from an east-trending paleovalley.

85.8 Turn left on Colorado Highway 291 to Salida. Route to Salida is across late Pleistocene outwash terrace; hills on right in Dry Union Formation. Ahead the south end of the upper Arkansas graben defined by west-northwest-trending fault zone.

98.7 Center of Salida (2148 m).

100.4 Turn right on U.S. Highway 50. to lodging.

Fourth Day, Thursday, July

Head west on U.S. Highway 50. Hills to left in Dry Union Formation have numerous vertebrate fossil localities.

6.1 Turn left on U.S. Highway 285 at Poncha Springs. After crossing the West Fork of the Arkansas River the first outcrops are gravel and sand of the Dry Union Formation dipping south into fault, which we cross at 7.9 km. Canyon to south is cut in Early Proterozoic felsic gneiss and amphibolite. East of Salida rocks from the same felsic and mafic volcanic protoliths are less recrystallized and deformed. Volcanic textures and structures are preserved, although the rocks are in the amphibolite facies.

13.2 Oligocene volcanic rocks overlie the Precambrian and are overlain by gravels of the Dry Union Formation.

14.3 Major valley to right was route of a railroad that crossed the Continental Divide at Marshall Pass. Cross fault back into Precambrian rocks.

19.8 Summit of Poncha Pass (2,746 m). Gentle grassy areas underlain by Dry Union Formation. As we descend to the south small hill on left is Precambrian in fault block at the north end of the San Luis valley, a part of the Rio Grande rift which connects to the upper Arkansas graben through a very narrow neck at Poncha Pass. To the left the Sangre de Cristo Mountains are formed by Early Proterozoic metamorphic rocks. To the right rocks of the San Juan volcanic field in and adjacent to the Bonanza caldera form generally wooded slopes.

The San Juan volcanic field is the largest erosional remanent of an extensive volcanic field that covered much of the southern Rocky Mountains. Intermediate composition lavas and breccias that erupted 35-30 Ma from central volcanoes are overlain by about 15 widespread ash-flow sheets erupted from caldera sources. At about 26 Ma, volcanism shifted to a bimodal assemblage dominated by alkali basalt and silicic rhyolite at the beginning of the regional extension that established the Rio Grande rift zone. The San Juan field is one of several areas of Tertiary volcanics along the eastern Cordilleran margin of the North American plate, broadly related to subduction along its western margin (Lipman, 1989).

41.3 Villa Grove. Road to Bonanza.

48.6 Junction Colorado Highway 17; bear right on U.S. Highway 285. At this latitude the Sangre de Cristo Mountains are composed of thick and partly very coarse grained Pennsylvanian and Permian sedimentary rocks deposited at the eastern margin of the late Paleozoic Uncompahgre highland. They are involved in a series of east-verging Laramide thrust faults. The present mountain range is a strongly uplifted horst along the east margin of the Rio Grande rift (Burbank and Goddard, 1937; Tweto, 1978; Lindsay and others, 1983). Holocene scarps cutting late Pleistocene fans show that the uplift of the range continues. The smooth surface of the San Luis valley is composed of Quaternary alluvial and eolian deposits. Geophysical studies and some boreholes show that several horsts and grabens are concealed beneath this surface. The deepest fill, including middle Tertiary volcanic and alluvial deposits is estimated to be 6 km.

From Villa Grove to near Sagauche, mountains on the right are formed by faulted and folded Paleozoic and Precambrian rocks interpreted to be part of the belt of thrusting exposed in the Sangre de Cristos to the east.

63.4 Bluffs at 2 o'clock formed by

Bulldog Mountain fault zone had developed contemporaneously with other mineralized structures in the district, and therefore presented an attractive exploration target for blind vein deposits. In 1963, Homestake Mining Company acquired a land position and explored the Bulldog Mountain fault zone at depth. High grade silver-lead-zinc ores were discovered some 150 m below barren surface outcrops. Homestake's Bulldog Mountain Mine was developed along the Bulldog Mountain vein system, and began production in 1969. Mining continued from 1969 until 1985, when the mine was temporarily shut down and placed on standby status due to a depressed silver market.

Exploration activity in the district has continued intermittently. Since the mid-1970's, efforts have focused upon bonanza vein targets throughout the district and upon disseminated silver deposits in the southern part of the district. In 1983, Homestake Mining Company began exploration efforts in the northern part of the district to search for blind vein deposits along the northern extensions of structures which had provided production in the southern part of the district. These efforts resulted in the discovery of the Northern Amethyst silver/gold vein system 9 km north of the Bulldog Mountain Mine. Limited underground exploration activities identified ore apexing approximately 150 m below barren surface exposures. This work was suspended in 1988 pending improved silver prices.

Modern ore genesis studies of the Creede district began in the late 1950's, and are continuing at the present (Steven and Eaton, 1975; Barton et al., 1977; Bethke and Rye, 1979; Hayba et al., 1985). The most recent summary of Creede ore genesis research is given in Bethke (1988). A complete bibiliography of published studies and articles on Creede is given in Hayba and Conte (1988).

Mining and processing

The Bulldog Mountain Mine, the only modern mine in the Creede district, was serviced from five levels (approximately 60 m apart vertically). Productive workings extended over 300 m vertically and 2,400 m horizontally. The mine employed a modified underhand cut and fill mining method, and produced at an average rate of 360 tonnes per day. Ore was crushed and processed in an on-site 270 tonne/day flotation mill. Mill head grades averaged 564 g/tonne Ag and 1.9% Pb during the production years, and recoveries averaged 89% for Ag and 82% for Pb. Concentrates were shipped to a refinery in El Paso, Texas, for further processing.

Production/Reserves

Production records for the Creede district are incomplete, particularly for the years 1891-1903. Based upon the best published records and internal Homestake Mining Company records, the Creede district mines produced over 2,381,600 kg of silver, 4,250 kg of gold, 694 million kg of lead, 198 million kg of zinc, and 13 million kg of copper from roughly 4.43 million tonnes of ore. Of this, about 99.9% was produced from vein systems in the southern part of the district. At least five of the eight producing veins were exposed at the surface. Production from the Bulldog Mountain vein system was second only to production from the Amethyst vein system, and totalled (from 1969-1985) 708,750 kg silver and 22 million kg lead from 1.36 million tonnes of ore. Current ore reserves of the Bulldog Mountain mine and Northern Amethyst mine are confidential. Disseminated silver deposits hosted by the caldera moat sediments of the Creede Formation in the southern part of the district contain an estimated 3 million tonnes of ore averaging 181 g/tonne Ag (Smith, 1981).

GEOLOGY

Regional Geology

The geologic setting of the Creede mining district has been discussed in great detail (see, for example: Steven and Ratté, 1965; Steven and Eaton, 1975). For an up-to-date summary of the district's volcanic geology, the reader is referred to Lipman and Sawyer (1988).

The Creede district is located in the San Juan Mountains of southwestern Colorado. The San Juan mountains form the largest eroded remnant of an extensive mid-Tertiary volcanic field (Fig. 19); the volcanic rocks reflect the progressive evolution, rise and emplacement of a large batholith to shallow crustal levels (Steven and Lipman, 1976). The deepest parts of the volcanic pile are comprised of intermediate lavas and associated volcaniclastics at least 2 km thick; the lavas were erupted from stratovolcanoes during the early stages of batholith emplacement. The upper levels of the volcanic field are composed primarily of quartz-latitic to rhyolitic large-volume ash-flow tuffs that erupted from a number of shallow-level magma chambers during caldera collapse. The distributions in the volcanic field of the major ash-flow tuff eruptions and their associated calderas, summarized in Steven and Lipman (1976) and in Lipman and Sawyer (1988), are shown in Figure 19. The Creede district is located in the central portion of the San Juan volcanic field, in a complex set of nested calderas formed by the eruption of 6 major ash-flow sheets between 28.25 to 26.15 Ma (Lipman and Sawyer, 1988).

Ore deposit geology and mineralogy

Most of the production in the Creede district has come from veins filling fractures of a graben extending from the 26.8 Ma Creede caldera on the southeast to the 26.4-26.1 Ma San Luis caldera on the northwest. The major host for ore in the district is the fill of the Bachelor caldera, the 27.6 Ma Bachelor Mountain Member of the Carpenter Ridge Tuff. Mineralization in the northern district is in part hosted by the fill of the San Luis caldera, the 26.1 Ma Nelson Mountain Tuff. Vein mineralization in both the northern and southern district has been dated at 25.1 Ma, approximately 1 million years younger than the last volcanism in the area. (Bethke 1988: Plumlee 1989; Rye and others, 1988; Hayba and others 1985; Bethke and Rye 1979; and Barton and others 1977).

Structure and ore controls. The Creede graben trends roughly N. 15^{o} W.; it is approximately 13 km long and 6 km wide. Four major, throughgoing faults form the framework of the graben. From east to west these include the Solomon-Holy Moses (west dip), the Amethyst (west dip), the Bulldog Mountain (east dip), and the Alpha-Corsair (east dip). Numerous intervening faults, such as the OH and P, form a systematic but complex network of conjugate shears and extension fractures. Most of the production has come from the southern third of the graben.

Through examination of the Bulldog Mountain vein system, mine maps of the Amethyst and OH veins (Steven and Ratté, 1965), surface exposures throughout the district, and underground exposures at the northern Amethyst deposit, Caddey and Byington (unpublished data) have developed a general model for structural activity in the district. They identified five major deformation events, the fourth of which was contemporaneous with mineralization. Structural geometries, fracture patterns, and faulting sequences based on crosscutting relationships all indicate that the Creede graben was structurally active during mineralization; faulting recurred during vein development. During mineralization, the district was subjected to an overall left-handed wrench of oblique-slip style. The four major graben faults experienced repeated left-lateral strike-slip and normal dip-slip displacements. Branching faults such as the OH and P behaved as conjugate shear pairs with right lateral stike-slip displacement.

The location of ore shoots within the veins was controlled by the orientation of the fractures and the types of displacement they underwent (Caddey and Byington, unpublished data). In the southern part of the district, variations in the competency of welding zones within the Bachelor Mountain Member of the Carpenter Ridge Tuff affected the orientation of the fractures and how the fractures opened during mineralization. Fault refraction in section preferentially created steep dips in the competent Campbell Mountain welding zone of the Bachelor Mountain Member, and shallower dips in less competent welding zones above and below the Campbell Mountain. A dominant dip-slip component active throughout most of the southern part of the district during vein mineralization opened the steeper dips and closed the flatter dips, thus localizing most of the ore in the southern district in portions of fractures with Campbell Mountain wallrock. Along the Bulldog Mountain vein system, normal/right-lateral oblique-slip displacement created ore shoot openings along steeper parts of the fault and at strike deflections to the right; the periodic structural movements which created these ore shoots can be correlated (Byington and Caddey, unpublished data) with mineralization stages identified along the vein system by Plumlee (1989). Along the southern Amethyst vein, ore shoots occur on steeper portions of the fault surface and at strike deflections to the left, indicating that normal/left lateral oblique-slip faulting occurred during vein development. Some veins were opened during periods of base metal-rich mineralization, while other veins were opened during periods of silver-rich mineralization. Structural models based upon these principles developed in the southern part of the district played an important role in the Northern Amethyst silver/gold vein discovery.

Wallrock alteration. Wallrock alteration is variably developed depending upon the position in the ore zone. In the southern parts of the vein systems, wallrock alteration in the lower and middle levels of

their high salinities, the southern-recharge brines had the capacity to pick up and transport from depth significant quantities of chloride-complexed metals such as Pb, Zn, Ag, and Cu; in contrast, the dilute northern-recharge fluids could transport much smaller quantities of chloride-complexed metals but greater quantities of bisulfide-complexed metals such as Au. Some mixing of the two fluids may have occurred in the upwelling plume. Boiling occurred at the top of the plume in the hotter portions of the shallow ore zones. The steam and acid volatiles which separated during boiling condensed into dilute groundwaters overlying the system, creating relatively acidic steam-heated waters. In response to the overall north-to-south topographic gradient in the district, the hydrothermal fluids apparently flowed predominantly laterally to the south through most of the shallow ore zones. As they flowed southward, the hydrothermal fluids mixed with the overlying steam-heated groundwaters. Ore deposition occurred in response to boiling and to mixing with the overlying groundwaters capping the system. The carbonate-precious metal ores (Northern assemblage, Fig. 20) in the northern vein systems probably formed through boiling of the dilute northern-recharge hydrothermal fluids. Initial boiling of the southern-recharge hydrothermal brines produced chlorite-hematite-quartz-adularia-base metal mineralogies (part of the OH assemblage, Fig. 20) along the district's central vein systems. Following boiling, the progressive lateral mixing of the southern-recharge brines with the overlying dilute groundwaters produced zoned mineral assemblages, grading from base metal sulfide-rich ores along the central vein systems (part of the OH assemblage, Fig. 20) to barite-native silver-acanthite-base metal sulfide ores along the southern vein systems (Bulldog assemblage, Fig. 20).

187.7 Junction with Deep Creek road; turn left. Cross Deep Creek 2.6 km, 0.2 km farther the junction of the airport road to Creede.

191.0 Turn left on track to pioneer grave at edge of trees or walk up to it from road. Continue on foot up trail to top of hill.

STOP 27. This is one of the travertine knobs which intertongue with the lake sediments of the Creede Formation that fill the Creede caldera moat. The bounding ring fault of the Creede caldera is concealed beneath the moat fill. The resurgent dome of the Creede caldera makes up most of Snowshoe Mountain to the south. The crest of the dome is broken by keystone graben faults that follow Deep Creek, the major N-S drainage transecting the caldera. The composite section of Snowshoe Mountain Tuff is 1.5-2 km thick, with no base exposed. Most outflow Snowshoe Mountain Tuff has been eroded; the only sizeable preserved areas are weakly welded tuffs along ridge crests near South Fork.

From the top of the traverine knob the view northeast is toward the La Garita Mountains on the skyline. The La Garitas, on the Continental Divide (high point, 4,179 m), are the resurgent core of the La Garita caldera, the earliest of the central San Juan cluster. More than 1.5 km of intracaldera Fish Canyon Tuff are exposed, but the top is eroded and the base not exposed.

The view directly north, toward the town of Creede, is into the resurgent core of the Bachelor caldera, the second collapse structure of the central San Juan caldera complex, here exposed in cross-section in the north wall of the younger Creede caldera. The fill of the Bachelor caldera consists largely of variably welded tuff, designated the Bachelor Mountain Member of the Carpenter Ridge Tuff; this member is now recognized as the intracaldera equivalent to rhyolitic outflow parts of the Carpenter Ridge Tuff.

The structure of the Bachelor resurgent dome is well-displayed from here. Willow Creek, directly ahead to the north, roughly follows the apical-graben (Creede graben) of the Bachelor caldera. The Creede graben was reactivated following activity at the Creede and San Luis calderas, and the graben structures between these younger calderas localized epithermal veins.

The cliffs to the east of Willow Creek are fluidal rhyolite of the Bachelor Mountain Member. The eastern vein-fault of the Creede graben, the Solomon-Holy Moses vein, runs along the east side of Campbell Mountain, the ridge between the forks of Willow

Creek. The Amethyst vein, the most productive vein in the district runs N. 20° W. along West Willow Creek, through Bachelor Mountain. The Amethyst fault dips west and accomodated most of the normal displacement on the east side of the graben. West of the Amethyst fault, Wason Park Tuff is dropped against the Bachelor Mountain Member. The trace of the east-dipping Bulldog Mountain vein-fault, on the west side of the Creede graben, passes through Bulldog Mountain, the gently rounded hill just west of Windy Gulch. Farther west, just east of Miners Creek, the Alpha-Corsair vein-fault marks the west boundary of the Creede graben.

The distant sawtooth peak just visible to the north is San Luis Peak (4,271 M), the high point of the resurgently uplifted block within the San Luis caldera. Thus, parts of four calderas are in sight.

Two sites of the Continental Scientific Drilling Program, for drilling 0.6-1 km through the Creede Formation in the moat of the caldera, are within view. One is near the airport, on line with the Creede vein system; the other is several kilometers farther west. This drilling is intended to evaluate the connate-fluid environment at depth in the Creede sediments. The high salinities in fluid inclusions from the ores are a distinctive signature for the fluid component inferred to have been derived from the lake sediments, and provides a basis for fluid-flow modeling not possible in mining districts where the fluid components are less distinct compositionally. In addition, the moat drillholes will provide a record of postcaldera events; air-fall tuffs, ash flow deposits, and even lavas may be interbedded with the moat sediments. Drill penetration into the top of the moat-floor Snowshoe Mountain Tuff should also be possible.

Return east on Deep Creek road 0.3 km; turn left on airport road to Creede. Cross Rio Grande bridge and continue ahead past airport.

193.4 Turn right on Colorado Highway 149. Moat fill sediments of the Creede Formation to left.

195.8 Enter Creede (2,694 m). Continue straight ahead through town to theater where lectures on the Creede district will be presented. Continue north to canyon of Willow Creek after lectures.

197.6 Junction of East and West Forks of Willow Creek; bear left on West Fork. Outcrops are Bachelor Mountain Member. Bus stops at bottom of steep grade. Walk 0.3 km up road to dump at the Commodore 5 level. STOP 28. The Commodore 5 level was the main haulage on the Amethyst-OH vein system. The Commodore 4 and 3 levels are marked by buildings higher on the vein.

The Amethyst vein is an epithermal Ag-Pb-Zn vein that filled the Amethyst fault at about 25 Ma. During caldera resurgence at about 27.6 Ma, rhyolitic tuffs of the Bachelor Mountain Member were faulted and brecciated while they were still hot and plastic along an ancestral Amethyst fault zone. The Creede graben was reactivated about 2.5 Ma later, along the same general trends as the ancestral Amethyst, following activity of the Creede and San Luis calderas (Steven and Ratte, 1965; Bethke and others, 1977). The Amethyst fault, as is typical of the Creede graben structures, strikes N. 15-20° W. and dips 50°-70°SW. Some major NW-trending splays, such as the OH vein, were also strongly mineralized. The Amethyst vein is bounded by a clay cap and little exposed at the surface, although sediments of the Creede Formation are mineralized near the Commodore 3 level.

Return to South Fork, Del Norte, and head back towards Garfield If time permits, stop just beyond E5 road 0.5 mi west of the junction of U.S. Highway 285 and Colorado Highway 17 for a view of the San Luis Valley and the Sangre de Cristo Mountains in the afternoon light. STOP 29 (optional).

Night in Salida.

Fifth Day, Friday, July 7.

From Salida head back north up the upper Arkansas graben to south of Buena Vista and go east on U.S.

Highway 285.

As we head east up the Trout Creek valley, Early Proterozoic granitic rock forms the exposures along the road. The hill on the south side of the valley is capped by early Oligocene volcanic rock filling a paleovalley. At 31.8 km cross faults drop the volcanics down so that they form some exposures near the valley bottom.

53.6 Ordovician, Devonian, and Mississippian rocks with a total thickness of about 250 m dip gently east. Upper Cambrian quartzite is absent here. Disconformities separate the formations. Small sharp cliff is Ordovician Harding Quartzite, and northerly, massive ledge is Leadville Limestone.

55.8 Buffalo Peaks at 11 o´clock are capped by early Oligocene welded tuffs and flows filling another east-northeast-trending paleovalley.

Pennsylvanian rocks of the Belden and Minturn Formations underlie the meadows and are poorly exposed.

60.0 Summit of Trout Creek Pass (2,895 m). Descend into South Park.

61.4 Turn right on U.S. Highway 24. Low hills on left are lower Oligocene andesite. Cross poorly exposed Pennsylvanian Minturn Formation to 67.0 km where small cliffs are composed of lower Olicogene Wall Mountain Tuff dipping into the Antero syncline, which is a Neogene fold containing Miocene sandstone, conglomerate, and siltstone of the Wagontongue Formation in its trough.

80.6 Junction with Colorado Highway 9. On right Jurassic sandstone and thin Pennsylvanian-Permian Maroon Formation lies on Early Proterozoic granite. We have crossed a fault that was part of the west margin of the ancestral Front Range uplift. It was active in Permian and (or) Triassic time, since the nearby rocks of the Maroon Formation do not appear to have been deposited next to an active fault. To left, across the South Fork of the South Platte River, Cretaceous Dakota Sandstone dips north in block between two faults. Just east of Hartsel cross the Currant Creek fault, a major fault that extends over 200 km and is called the Ilse fault to the south and the South Park fault to the north. At least part of this fault zone had Precambrian movement along it, some of it was active in the late Paleozoic or early Mesozoic, much of it was active in late Laramide deformation after deposition of the Paleocene South Park Formation, and some movement occurred after extrusion of the Oligocene Thirtynine Mile volcanic field. East of Hartsel large grassy area is underlain by Cretaceous Pierre Shale mantled by Quaternary pediment deposits.

87.5 Arkosic sandstones in bluff to left with flat apparent dip are in the Paleocene South Park Formation. The major deformation in South Park was younger than these deposits, for they are overidden by Proterozoic rocks at the west margin of the Laramide Front Range uplift along the Elkhorn thrust. At 90.7 km we cross the concealed Elkhorn thrust. Hills to left are on the hanging wall of the Elkhorn thrust and are composed of Precambrian rocks with a discontinuous skin of west-dipping Jurassic and Cretaceous rocks. Just beyond to right is Spinney Mountain Reservoir and at 2 o´clock, Spinney Mountain. Southward continuation of the Elkhorn thrust is exposed on Spinney Mountain where Precambrian granite is thrust over Cretaceous Pierre Shale and Paleocene South Park Formation.

93.5 North end of Sulphur Mountain composed of Cretaceous and Jurassic rocks dipping southwest into South Park. Just beyond, cross concealed Neogene Chase Gulch fault. Gravity and borehole data indicate a minimum of 600 m of Neogene fill in a graben east of the fault (Kirkham and Rogers, 1981, referring to an unpublished report by M.E. Shaffer). On skyline to the right is Thirtynine Mile Mountain (3,523 m) composed of Oligocene andesites.

As we start to ascend to Wilkerson Pass exposures of Early Proterozoic granitic and metamorphic rock protrude through the pediment gravels. The metamorphic rocks are mainly sillimanitic biotite schist and gneiss containing a few layers of amphibolite.

107.2 Summit of Wilkerson Pass (2,896 m) is in the Puma Hills. Leave South Park. Pass is controled by a northwest-trending fault zone.

113.6–116.5 Pluton of Middle Proterozoic granite of the Berthoud Plutonic Suite.

123.1 Cross contact of the 1,030 Ma Pikes Peak batholith. Hills on left are on margin of the Lake George pluton, a 8 km diameter texturally zoned pluton of the Pikes Peak batholith intruded by partial ring dikes of quartz syenite to fayalite granite and by a central syenomonzonite stock. The partial ring dikes form the rugged outcrops seen from the road (Wobus, 1976).

124.7 Village of Lake George. From here to 141.2 km the road follows valleys filled with deposits of Wall Mountain Tuff and Florissant Lakebeds. Small ledges here and there along the valley sides are of Wall Mountain Tuff. Florissant Lakebeds, consisting of tuff, shale, arkosic conglomerate, mudstone, lahar, and volcanic conglomerate, fill valleys that existed in early Oligocene time and were dammed by extrusion of the Thirtynine Mile volcanic field (Epis and Chapin, 1975). The Florissant Lakebeds are famous for their fossil flora and fauna. Analyses of the flora shows that this area was only about 900 m above sea level at that time (MacGinitie, 1953); now it is at 2,500 m altitude. Fossilized stumps and logs of Sequoia, a large conifer now growing near the coast of California and on the western flank of the Sierra Nevada in California are exposed in excavations in Florissant National Monument.

132.5 Turn right on Teller County Road 1 at village of Florissant. Just to the south about 25 km^2 of land has been put in the National Park system to protect a sample of the Florissant Lakebeds. Scattered outcrops of Pikes Peak Granite along the valley sides.

148.2 Valley to right with road. Contains exposure of lahar in the lower part of the Thirtynine Mile Andesite. As we head up valley to the southeast Pikes Peak Granite on the left and granite of the Berthoud Plutonic Suite on the right. Valley follows fracture zone. Between here and Cripple Creek the late Eocene erosion surface has been uplifted about 400 m along north-trending faults.

155.1 Major valley of Spring Creek on right.

157.5 Spring Creek pluton of the Pikes Peak batholith on the right composed of clinopyroxene-fayalite-quartz syenite, a member of the sodic differentiation trend in the batholith (Barker and others, 1975).

159.5 Crest of hill; Mt. Pisgah at 3 o'clock is a plug of 29 Ma phonolite of the Cripple Creek volcanic center. Major part of the volcanic center lies ahead east of the town of Cripple Creek and is marked by numerous mine dumps.

162.3 Center of Cripple Creek (2,893 m), county seat of Teller County. Much of the town was built between 1891 and 1910 after the discovery of gold.

THE CRIPPLE CREEK DISTRICT

Tommy B. Thompson
Dept. of Earth Resources, Colorado State University, Fort Collins

INTRODUCTION

History

The Cripple Creek District was discovered by a cowboy prospector, Bob Womack, in 1891, and within two years, a rail system connected the remote ranching country with Colorado Springs in order to service the producing mines. Output from the district peaked in 1900 when 30,100 kg Au were produced (Lindgren and Ransome, 1906). Total production from the district has exceeded 750,640 kg gold. The gold:silver ratio of production has ranged between 0.2:1 and 20:1 with the bulk of values >10:1.

Mining and processing

Most of the ore has come from narrow veins averaging 31 g Au/tonne and hosted in sillimanitic biotite schist and gneiss containing a few layers of amphibolite.

107.2 Summit of Wilkerson Pass (2,896 m) is in the Puma Hills. Leave South Park. Pass is controled by a northwest-trending fault zone.

113.6–116.5 Pluton of Middle Proterozoic granite of the Berthoud Plutonic Suite.

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Mining and processing

Most of the ore has come from narrow veins averaging 31 g Au/tonne and hosted in granite or in diatreme breccias. Wallrocks are not extensively altered, and, thus, did not require much support other than timber stulls in the stopes. In local zones, the veins form a sheeted zone in which mining widths required some timber square-sets. In general, many large stopes are open without any evidence of timbering. Mining was generally accomplished by overhand stoping (Lindgren and Ransome, 1906).

Present day operations are from open pits utilizing power shovels and 50-ton capacity truck haulage. This production is lower grade, with gold recovery from cyanide solutions sprayed on heap leach pads. Gold is recovered from the leach solutions by reaction with activated carbon produced from coconut shells. The gold is removed from solution onto the carbon in a 5-compartment carbon column or in 4-stage portable carbon tanks. The gold-barren solution returns to a second holding pond for further spraying onto the ore. Gold-loaded carbon is transferred to a carbon-stripping facility at the old Carlton Mill, where gold is stripped from the carbon and recovered by electrowinning onto steel wool. The gold-coated steel wool is finally melted to produce a gold-silver ore. Cripple Creek doré typically ranges from 50% to 80% gold and is shipped to a commercial refinery for gold-silver separation.

Production/Reserves

Exploration in the Cripple Creek district in recent years has focussed on bulk-tonnage, open pit mineralization. There are three pits that have produced low-grade (<3 g/tonne Au) ore: Globe Hill, Ironclad (Victor Project), and Portland (Fig. 22). The first two contain some reserves, but the Portland pit reserves were exhausted in September, 1988. The latter has had the most recent mining of the three listed. The Portland pit produced about 10,000 tonnes per day with a waste:ore ratio of 4:1 for approximately 10 months.

Two areas have yielded ore from underground mining during 1988: the Proper and Mineral Hill sites. These orebodies are higher grade and are processed through the Carlton mill at a rate of 150 tonnes per day.

GEOLOGY

General Geology

The Cripple Creek district is within and adjacent to a 27.9 to 29.3 Ma nested diatreme-intrusive complex, surrounded by Proterozoic granite (1.03 Ga) and older metasedimentary and igneous rocks (Thompson and others, 1985). Volcanic flows, subvolcanic intrusives and phreatomagmatic breccias occur in three coalescing basins. The igneous rocks are phonolites, syenites, trachytes and alkali basalts. Detailed mineralogical data, whole-rock chemistry and initial $^{87}Sr/^{86}Sr$ ratios (0.70391 to 0.70475) show that the entire suite is genetically related (Birmingham, 1987). Subsidence of the diatreme rocks is indicated by (1) a thick fluvial-lacustrine sedimentary sequence in the eastern sub-basin, (2) the presence of carbonaceous debris, ripple-laminated rocks and dessication cracks in sedimentary rocks at depths more than 300 m below the present surface, (3) the fracture systems near the diatreme sub-basin margins that reflect basement rock influence, and (4) flat-dipping veins near intrusive bodies or small breccia bodies.

There are several indicators that fluidization was important in the emplacement of the Cripple Creek diatreme breccias: (1) accretionary lapilli occur in several discordant breccia masses, (2) the matrix in the diatreme breccias has been sorted to the point that little clay-size particles remain, (3) the breccia is bimodal in regard to size of fragments and matrix, but matrix dominates, and (4) discordant hydrothermal breccias occur as steep- or flat-dipping bodies with fragments of vein minerals in them.

Ore Deposits

Cripple Creek is known for its vein ores; however, they are not typical banded quartz veins that form bold outcrops. Instead they formed in sheeted zones with only minor quartz and base metal sulfides, fluorite, adularia, carbonate, hematite, gold and telluride minerals. The ores are vuggy, in many places the result of country rock dissolution. The veins occur in radiating and concentric bodies near the outer margin of the diatreme complex, cutting both the Tertiary breccias and Proterozoic granitic rocks, or along complex fracture zones, overlying basement ridges beneath the diatreme complex (Koschmann, 1949; Koschmann and Loughlin, 1965). Major veins exhibit remarkable vertical continuity, extending to more than 1,000 m below the present surface.

Bulk tonnage deposits consist of (1) mineralized tectonic and hydrothermal breccias or (2) fracture-controlled ores dispersed adjacent to the vein systems.

These deposits are much lower grade than the vein bodies (more than 30 grams per tonne Au in veins as compared to 2 gpt Au in bulk tonnage bodies). There is no recognized vertical zonation in the vein systems or bulk tonnage deposits.

Hydrothermal Alteration. Vein-related hydrothermal alteration occurs in a narrow selvage that extends outward no more than five times the vein width. Secondary K-feldspar, roscoelite (V-mica), pyrite and dolomite occur within an inner zone adjacent to the veins whereas an outer zone contains sericite, montmorillonite, magnetite, minor secondary K-feldspar and pyrite (Thompson and others, 1985). There is no expansion of the alteration zones in the upper level mine exposures.

The bulk tonnage deposits exhibit similar alteration assemblages, but with more overprinting by argillic assemblage products.

Fluid Inclusion Studies. Fluid inclusion studies on vein minerals have documented the presence of early saline fluids (33 to greater than 40 equivalent weight percent NaCl) with the higher salinities in the upper 300 m of vein exposures at the Ajax mine and at temperatures near 300°C. The fluids were boiling and contain CO_2. Later fluids were less saline (0 to 8.3 equiv. wt. percent NaCl), and exhibit progressively lower filling temperatures (<200°C)(Dwelley, 1984; Thompson and others, 1985). Similar low temperature-low salinity fluids are present in matrix minerals of the hydrothermal breccias (Thompson and others, 1985). The latter also exhibit strong evidence of boiling.

Stable Isotope Studies and Age Determinations. Investigations in progress are focussed on stable isotope studies (O, H, and S); however, no data are yet available. Additionally, fission-track dating of hydrothermal apatite has been initiated. Hydrothermal adularia and biotite associated with vein deposits have been prepared for argon-argon dating.

Genetic Model. The genetic model presented here is preliminary since no isotopic data are yet available. Similarly, a precise age of ore deposition is not possible at this time.

Physical relations between ore and host rock suggest that the ore deposits formed shortly after diatreme development and igneous activity ceased. There are some alkali basalt dikes that are only weakly altered compared to other wallrocks. The evolution of ore fluids from high temperature-high salinity to low temperature-low salinity would suggest that the ore fluids initially were magmatically-derived with later dilution from meteoric influx. Complexing of gold with aqueous Te species has been suggested for telluride deposits associated with alkaline rocks (Saunders, 1986). The large CO_2 content in the fluid inclusions was responsible for deep level ore fluid boiling, oxidation of the fluids, and precipitation of gold and gold-silver tellurides throughout a large vertical interval. Restriction of ore fluid degassing in some areas led to overpressures with resultant hydrothermal breccias at shallow levels. There is a genetic tie between the veins and hydrothermal breccias as vein material is commonly found in the hydrothermal breccias.

STOP 30. The contact between the Cripple Creek diatreme breccias and Proterozoic granitic rock is exposed just west of the Carlton Mill. Note that the rock is argillized and contains pyrite or its oxidation products. To the south are the heap leaching pads of the Cripple Creek and Victor Gold Mining company.

STOP 31. The Cresson mine dump contains a unique collection of alkali basalt-dominated breccias developed in a very late-stage small diameter diatreme. The breccias are also intruded by pre-mineral alkali basalt dikes. Ore within the Cresson rocks occurs in open cavities, and most of the ore mined was direct smelting ore. Between 1903 and 1930 1,925,350 tonnes averaging 26 gpt Au were produced from the mine. The Cresson basaltic diatreme breccia records the last major igneous event in the district prior to the mineralizing event.

STOP 32. This stop will allow mine dump sampling of wallrocks and vein material found in the Ajax mine. Veins and alkali basalt dikes strike north-northwesterly. Altered and fresh wallrock, of Proterozoic granite and Tertiary phonolite, and phlogopite basalt porphyry can be found. Vein materials include quartz, purple fluorite, pyrite, celestite, calaverite, and traces of base metal sulfides.

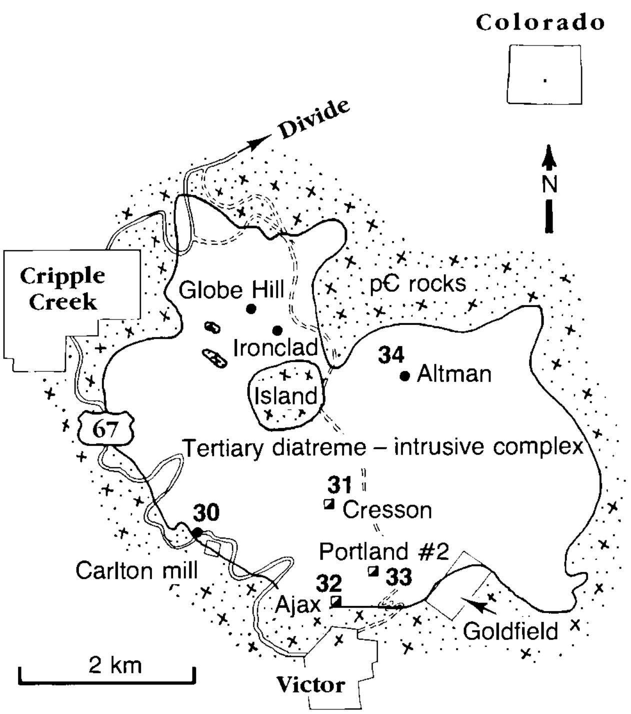

FIGURE 22 Index map of Cripple Creek district showing stops.

STOP 33. The Portland open pit exploited dispersed gold mineralization adjacent to veins of the Portland mine. Wallrocks include diatreme breccias cut by dikes of phonolite, trachyte porphyry, and alkali basalt. Argillic alteration has been overprinted on very finely crystalline K-feldspar alteration (flooding). Bedded pyroclastic rocks are present locally near the top of the mine walls; near the eastern limits of the pit, accretionary lapilli were found in the pyroclastic rocks. Localization of ore along faults and second order structures is an important control on mineralization.

STOP 34. The Altman site has exposures of multiple stages of trachyte porphyry and basalt porphyry dikes. These cut the Tertiary Cripple Creek Breccia. The principal rock unit in the area is a trachyte porphyry. The Pharmacist vein system strikes northeasterly and dips 65-67^{o} to the north. Lower grade dispersed gold mineralization occurs in the wallrocks and is being exploited in the open pit. Wallrocks immediately adjacent to the vein are potassically altered; however, later argillization has been superimposed on these rocks.

From Cripple Creek the route heads north across the Pikes Peak batholith. To the east the high country surrounding Pikes Peak (4,297 m), apparently is a block upfaulted during Neogene time. In places the road follows the Oil Creek fault zone, one of the faults along which this uplift occurred.

287.2 Edge of a gentle grassy upland is underlain by gravel that is not precisely dated but contains 29 Ma rocks from the Cripple Creek volcanic center. Farther east the gravel has been displaced by Neogene fault movements, some of which occurred along the Ute Pass fault zone. At 11 o'clock a conical hill is a plug of lower Oligocene andesite rising above a well developed erosion surface cutting the Pikes Peak Granite. To the right is the flat crest of the Rampart Range formed by the same erosion surface displaced by Neogene faulting.

192.9 Turn right on U.S. Highway 24 at Divide. Just before Woodland Park at 204 km we cross the Ute Pass fault zone and enter the Manitou Park graben. Pennsylvanian and Permian Fountain Formation and lower Paleozoic rocks are preserved in the graben, showing that the east margin of the ancestral Front Range uplift lay to the west. Miocene gravel occurs on the crest of the Rampart Range east of the graben, and Neogene faulting had the net effect of displacing those gravels 370 m in a sense opposite to that of the Laramide movements responsible for the preservation of the Paleozoic sedimentary rocks in Manitou Park (Taylor, 1975). The Ute Pass fault zone contains clastic dikes of Cambrian sandstone. Such dikes are common in fault zones in the eastern Front Range from here for as much as 65 km to the north, and a few are found another 30 km north. Where the dikes occur in fault zones within the ancestral Front Range uplift, from which early Paleozoic rocks were removed in late Paleozoic time, fracturing allowing the downward injection and preservation of the Cambrian sandstone must have occurred earlier in the Paleozoic. The occurrence of a few diabase dikes and some alkalic intrusions in the region dated as early Paleozoic suggests a regional but weak extensional event at that time (Larson and others, 1985). Opening along faults may have allowed unlithified sediments to get into the fissures. The sediments were lithified and shifted around by later fault movements during late Paleozoic and (or) Laramide events (Scott, 1963).

From Woodland Park the route descends to Colorado Springs and the east margin of the Rocky Mountains through a valley controlled by the Ute Pass fault zone.

224.3 Lower Paleozoic rocks overlie Pikes Peak Granite and are only about 70 m thick.

229.0 Turn right and make hairpin turn and cross Business Highway 24. Continue ahead on Garden of the Gods road. Right turn followed closely by left turn. At 229.8 enter Garden of the Gods Park administered by the City of Colorado Springs. Scenic red rocks formed by erosion of Pennsylvanian and Permian Fountain Formation dipping gently south off the south end of the block forming the Rampart Range. Unlike at Red Rocks Park here we are far enough from the margin of the ancestral Front Range that the thin early Paleozoic section is preserved beneath the Fountain Formation.

232.3 Pass from Fountain Formation upsection through Cretaceous Niobrara Formation turned up steeply and cut by strands of the Rampart Range fault. Dakota Sandstone forms a sharp ridge.

233.5 Turn left on 30th street. On right Pierre Shale capped by middle Pleistocene alluvium. Large quarry in lower Paleozoic rocks high on slope to left.

236.8 Turn left into Flying W Ranch; 0.7 to parking area. Red Fountain Formation and white Lyons Sandstone (Permian) are almost vertical south of the Flying W. This is along the Rampart Range fault, which is the major mountain front structure to the north but dies out to the south.

Sixth Day, Saturday, July 8.

Colorado Springs to Denver on Interstate Highway I-25. From the low area at the city of Colorado Springs (1,881 m) underlain by Pierre Shale the route rises stratigraphically in the flat-lying rocks of the Denver basin and climbs in altitude as we go north. Exposures are sparse. At the base of the scarp north of Colorado Springs is a roadcut in sandstone and carbonaceous shale of the Laramie Formation and to the north one of andesitic sandstone of the Upper Cretaceous Denver Formation. The Denver underlies the Upper Cretaceous and Paleogene Dawson Formation, which forms the crest of the scarp north and northeast of the city, and the bedrock from here until we descend into the Denver area. Here, near the margin of the Front Range uplift, the Dawson Formation is an arkose derived from the Pikes Peak Granite. To the north the route is on middle Pleistocene alluvium.

(Distances starts at EXIT 146 on Highway I-25)

12.3 Air Force Academy stadium to left. In the distance the prominent building is the chapel. Nearby are academic buildings and dormitories.

16.1 To the left at the edge of the mountains is a white outcrop of Dawson Formation. The Rampart Range fault cuts out Paleozoic and Mesozoic rocks along the margin of the Denver basin. The fault apparently dies out in the Pierre Shale southeast of the Garden of the Gods without joining the Ute Pass fault, which forms the mountain boundary south of Colorado Springs. Surface exposures do not permit an accurate determination of the dip of the fault. Seismic study by industry is said to show that it dips moderately west and consequently is a thrust fault. Southwest of Colorado Springs surface exposures show that there the Ute Pass fault is a thrust dipping about 45^{o} west.

23.8 Top of Monument Hill (2,286 m). Divide between Platte River and Arkansas River drainages. This is the highest area of the high plains province and has a ponderosa pine forest like that of the lower slopes of the Rocky Mountains to the west.

North of Monument Hill many of the higher hills are capped by remnants of Wall Mountain Tuff, which was derived from the area of the Mount Princeton batholith in the Sawatch Range.

56.3 Exit 181, Castle Rock. The cliffy hill after which the town was named is at 1:30 and is capped by Oligocene Wall Mountain Tuff overlain by Castle Rock Conglomerate, a proximal facies of the White River Formation of the high plains of western Nebraska. It contains fragments of Wall Mountain tuff and fossils of early Oligocene mammals.

At the bottom of the descent to the lower area occupied by the city of Denver andesitic sandstone of the Denver Formation is exposed below the Dawson Formation in a roadcut. The remainder of the trip to Denver and Stapleton Airport is on Quaternary alluvium of various ages locally overlain by eolian deposits.

Acknowledgments

We appreciate the cooperation of Chain O'Mines, Smuggler Consolidated Mines, Climax Molybdenum, ASARCO, Homestake Mining, Mesa Unlimited Partnership, Coca Mines, and Texas Gulf in hosting this field trip on their properties.

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