# Economic Geology of the Idaho Springs District Clear Creek and Gilpin Counties, Colorado

GEOLOGICAL SURVEY BULLETIN 1208

*Prepared on behalf of the U.S. Atomic Energy Commission*

Moench and Drake—ECONOMIC GEOLOGY OF THE IDAHO SPRINGS DISTRICT, COLORADO—Geological Survey Bulletin 1208

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*By* ROBERT H. MOENCH *and* AVERY ALA DRAKE, JR.

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# ECONOMIC GEOLOGY OF THE IDAHO SPRINGS DISTRICT, CLEAR CREEK AND GILPIN COUNTIES, COLORADO

By Robert H. Moench and Avery Ala Drake, Jr.

## ABSTRACT

The Idaho Springs district is in the central part of the Front Range mineral belt, a northeast-trending zone of porphyritic intrusive rocks and hydrothermal veins of early Tertiary age. From 1860 through 1959 about $65 million worth of gold, silver, lead, copper, and zinc was mined from the veins of the district.

The bedrock of the district is composed largely of conformably layered gneissic rocks and small bodies of granitic and pegmatitic rocks of Precambrian age. The most abundant gneissic rocks are biotite gneiss, granite gneiss, and microline gneiss, which form five large conformable layers. In addition, quartz gneiss, amphibolite, and calc-silicate gneiss occur in small bodies. The granitic rocks form many small bodies that are largely concordant but locally discordant. Three recognized varieties of the rocks, in the order of their emplacement, are granodiorite, quartz diorite, and biotite-muscovite granite. Pegmatitic rocks in part form dikes and sills in and near the granitic bodies, to which most of these small intrusives may be related. One variety of pegmatite, possibly related to the biotite-muscovite granite, has been prospected for its uranium content.

During the Laramide orogeny, in early Tertiary time, the Precambrian rocks were invaded by a sequence of nine varieties of porphyritic intrusive rocks. The older rocks of this sequence, with few exceptions, form irregular plutons, thick dikes, and some thickly lenticular concordant masses, whereas the younger rocks typically form thin, long dikes. All but the youngest member of the sequence were emplaced before the formation of the metalliferous veins.

The gneissic rocks were folded at least twice during Precambrian time. The first deformation produced north-northeast-trending major folds that form the structural framework of the district. This deformation was plastic in character and was accompanied by recrystallization at high temperatures and pressures. The second deformation produced northeast-trending folds and granulation and was most intense in a northeast-trending zone about 2 miles wide in the southeast part of the district. This zone of deformation is part of the Idaho Springs–Ralston shear zone, which extends more than 20 miles northeastward to the front of the range. In the relatively incompetent rocks the second deformation produced many small folds that trend N. 55° E. in a remarkably consistent pattern. The more competent units, however, were not folded but were locally intensely granulated. The second deformation was cataclastic in character and took place at somewhat shallower depths than the first.

Northwest-trending faults of the so-called "breccia reef" system may have originated in Precambrian time, after the later of the two major deformations.

Arching of the Front Range highland during the Laramide orogeny probably produced the regional joint system, and porphyritic intrusive rocks were then emplaced, some as dikes along this joint system and others as plutons. Before the last member of the intrusive sequence was emplaced, regional shear stresses formed a northeast-trending network of faults and locally reopened parts of the northwest-trending faults of the "breccia reef" system.

Laramide strike-slip faults of small displacement are distributed in three principal sets that trend northeast, east-northeast, and east; most of them dip north at medium to steep angles. Individual faults are fairly straight and traceable for as much as 2 miles at the surface. The three principal sets were probably formed almost contemporaneously, for the order of their formation changes from place to place, and their movement patterns are consistent with a single, east-northeast-oriented compressive stress system.

The veins of the district typically fill fault fissures and are similar in structure and mineralogy to those classed as mesothermal by Lindgren. Pyrite, sphalerite, galena, chalcopyrite, and tennantite are the principal ore minerals, and quartz and local carbonate minerals are the principal gangue minerals. The walls enclosing some veins are indistinct and consist of fractured zones as much as 30 feet wide but typically less than 5 feet wide. These fractured zones have been cemented with ore and gangue minerals. Other veins, rarely more than a foot wide, have sharp walls and were formed by the filling of a single fissure. The veins are classified—on the basis of their mineral content—as pyrite veins, pyritic copper veins, pyritic lead-zinc veins, and lead-zinc veins.

Pyrite veins (composed largely of pyrite and quartz) are valued only for their gold content, which is low grade; few veins have been mined profitably. The veins have indistinct walls and show evidence of recurrent movements.

Pyritic copper veins contain abundant pyrite, smaller amounts of chalcopyrite and tennantite, and subordinate amounts of galena and sphalerite. Quartz is the principal gangue mineral; carbonate minerals are sparse. The base-metal minerals typically occupy fractures that cut through quartz and pyrite. The veins are valued chiefly for their content of gold and subordinately for copper, silver, and, locally, lead. Only a few veins have been mined profitably.

Pyritic lead-zinc veins contain pyrite, galena, sphalerite, and subordinate chalcopyrite and tennantite. Quartz is the principal gangue mineral, and carbonate minerals are locally abundant. Some veins are symmetrically banded; the base-metal minerals form the central band and pyrite forms the outer bands. In other veins the ore minerals occupy fractures that cut through the quartz and pyrite. The veins are of value chiefly for their gold and silver, but they also yield lead and copper. The most economically important mines of the district are in veins of this type.

Lead-zinc veins contain galena, sphalerite, and subordinate amounts of chalcopyrite, tennantite, and pyrite. Quartz and various carbonate minerals are the principal gangue minerals. The veins typically have sharp walls and are characterized by massive intergrowths of ore and gangue minerals. Lead-zinc veins are mined chiefly for silver, gold, and lead, and subordinately for copper. Zinc, though abundant, has not always been recovered. The primary ores are richer in lead and silver and poorer in gold and copper than those of other vein types.

Alteration of the wallrocks adjacent to the veins varies both in intensity and in lateral extent. Typically, an inner zone of hard sericitized rock is bordered by an outer zone of argillized rock that grades outward to fresh rock.

The width of the zones does not vary with the width of the veins, and in many places the veins cut across the alteration zones.

A zonal distribution of the ores in the district is indicated by the pattern of vein types. Pyritic copper veins form a broad belt along the west side of the district. This belt grades eastward to a belt of pyritic lead-zinc veins and, farther to the east, to areas where the lead-zinc veins dominate. In general accord with this zonal pattern—away from the zone of pyritic copper veins—silver, lead, and zinc increase in quantity and copper decreases: in the areas of lead-zinc veins, gold decreases and is less systematically distributed.

Because the veins are fissure filling, most ore bodies have been localized by the same factors that controlled the width of the original openings. The major factors are deflections in dip and strike of the fault surfaces, and the amount and type of movement along the faults or fissures. Deflections may be related to intersections of two or more faults, faults and layers of competent wallrocks, and, possibly, faults and fold axes.

A sequence of events, which starts with fracturing and proceeds through wallrock alteration and deposition of pyrite and then of base metals, is postulated as an explanation of the origin of the veins in the district. Districtwide fracturing may have allowed fluids to escape upward and outward from a magmatic source at depth. These fluids altered the wallrocks, and much pyrite was formed by sulfidation of iron that was released from iron-bearing minerals of the wallrocks. At this stage pyrite veins formed throughout almost all of the district, and most fractures became clogged with pyrite and alteration products. During recurrent tectonic movements most pyrite veins were reopened and some new fractures were probably formed. Mineralization then resumed, and the base-metal ores were deposited. The zonal pattern indicates that the source probably centered beneath the central pyritic zone in the Central City district, and that at greater depth it extended southwestward along the west side of the Idaho Springs district.

About 135 mines and prospects in the district have workings that range in length from about 100 feet to several miles; however, many are now partly or completely inaccessible and have not been worked for many years.

During the district's most active period, from the 1860's into the 1890's, the most accessible and locally richer near-surface ores were mined. Mining from 1900 through 1959 was affected by periods of nationwide economic depression, when costs generally were lower and the labor supply greater. In the 1950's costs increased disproportionately to metal values; this increase, combined with deterioration of the mines, inhibited mining activity. These trends can be expected to continue. Unless new low-cost large-tonnage methods can be applied in mining the ores, mining in the district probably will not produce significant quantities of base and precious metals in the future.

## INTRODUCTION

The Idaho Springs mining district (figs. 1, 2) is a segment of the Front Range mineral belt, a northeast-trending zone of intrusive rocks and hydrothermal ore deposits of early Tertiary age (Lovering and Goddard, 1950, p. 72–73, pls. 2, 3, fig. 21). This belt extends a distance of about 50 miles—from the region just northwest of Boulder southwestward across the Front Range.

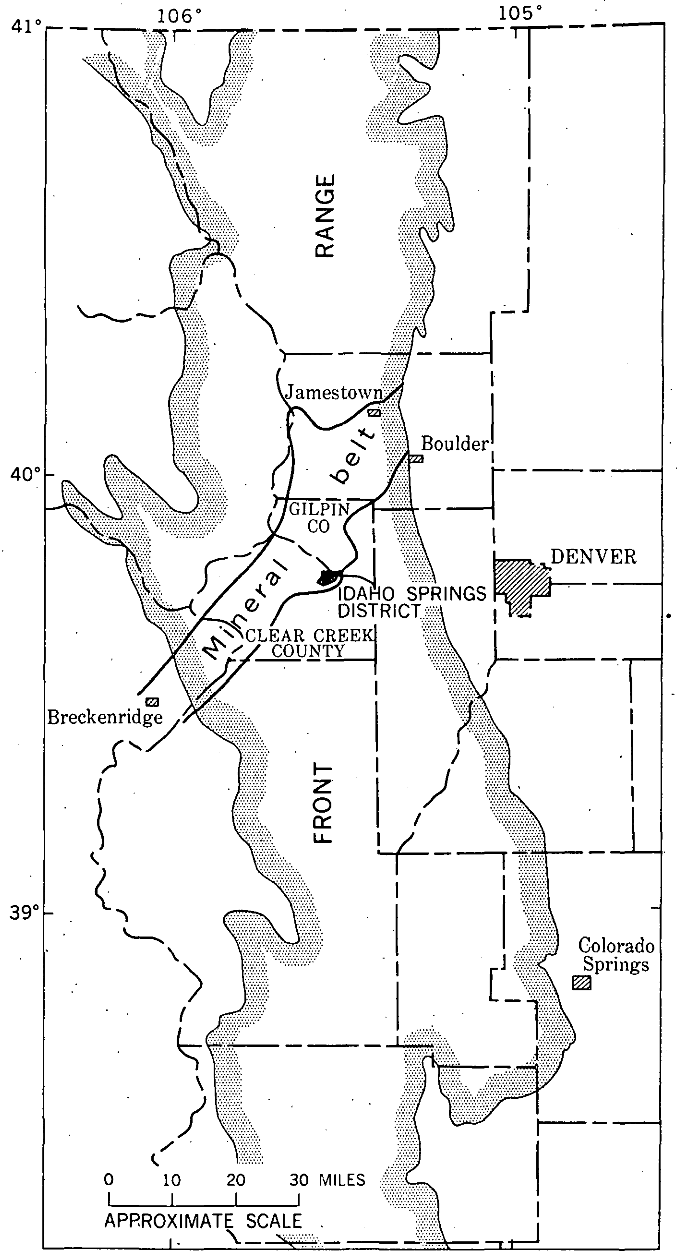


FIGURE 1.—The Front Range, Colo., showing the location of the Idaho Springs district.

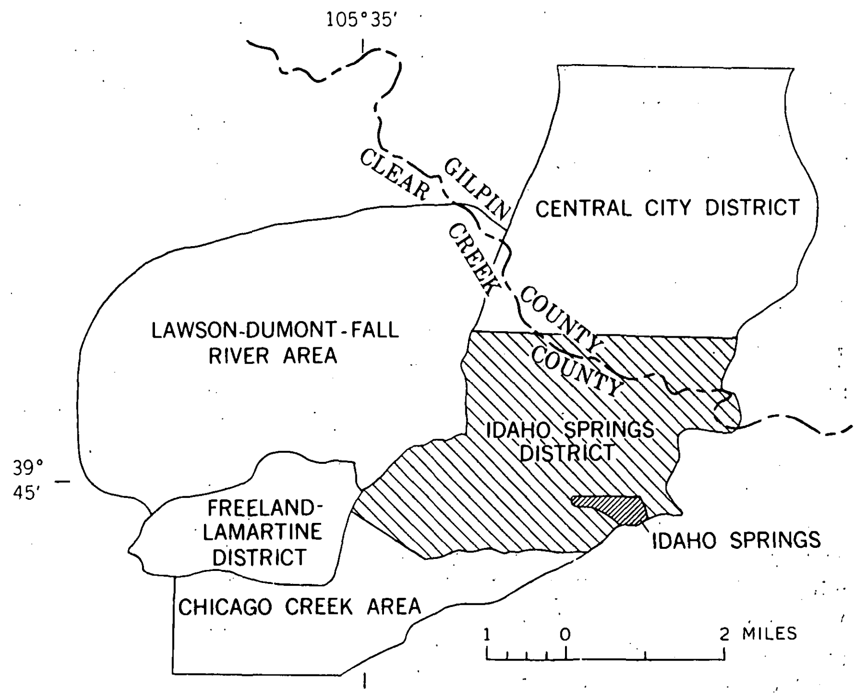


FIGURE 2.—The Idaho Springs district and adjacent mining districts.

From 1859, when placer gold was discovered in Idaho Springs and lode gold was discovered in Central City, through 1959, ores valued at about $200 million were shipped from a 50-square mile area that includes the Idaho Springs and adjacent districts to the north, west, and southwest. The adjacent Central City district, which produced ores valued at more than $100 million, was clearly the leading producer in the mineral belt. Through 1959 ores of a value totaling about $65 million were shipped from the Idaho Springs district. This exceeded the value of ores shipped from the districts to the west and southwest during that time. In most areas the most valuable metal in the ores was gold, in some areas it was silver, and in many areas copper, lead, and zinc were subordinately valuable.

Mining activity in the Idaho Springs and adjacent districts reached its peak in the late 1800's; it declined sharply after 1914, renewed somewhat during the 1930's, and declined greatly during World War II. In the 1950's uranium prospecting stimulated some mining activity. No uranium ore was produced, however, and at the close of the decade only one mine—the Bald Eagle—was being worked for its precious- and base-metal ores.

## GEOGRAPHY

The Idaho Springs district encompasses an area of about 10 square miles in Clear Creek and Gilpin Counties, Colo. Idaho Springs, the principal town in the district, is on U.S. Highway 6 and 40, a segment of a major east-west interstate highway system. Most of the district is easily accessible from this highway and from many good graded roads. State Highway 279 extends from Idaho Springs northward up Virginia Canyon to Central City, and State Highway 103 follows Chicago Creek southward.

The Idaho Springs district lies athwart Clear Creek Canyon; hence, it is characterized by rugged topography. The gently rolling upland surface that extends southward from the Central City district is deeply incised by the canyon and its tributaries. Slopes in the canyon average nearly 35°, and the maximum local relief is about 2,000 feet. Altitudes range from slightly more than 7,500 feet along Clear Creek to 9,925 feet on Pewabic Mountain. Clear Creek, an eastward-flowing tributary of the Platte River, has, as tributaries within the district, Fall River, Trail Creek, Chicago Creek, and Virginia Canyon. Spring Gulch joins Chicago Creek about a quarter of a mile south of Idaho Springs.

The climate of the region is temperate and dry, becoming more rigorous at higher altitudes. The mean annual temperature in Idaho Springs is 43° F., and the mean annual precipitation is 16 inches. Winters are cool, but snow is rarely more than a few inches deep, and, except on shaded slopes, rarely does it remain more than a few days. Summers are characterized by regular afternoon thundershowers and are moderately cool and sunny.

The district was apparently denuded of forest cover during early mining, but since then a new forest has grown locally. South-facing slopes are sparsely timbered, but north-facing slopes are well-covered with conifer and aspen.

## PURPOSE AND SCOPE OF REPORT

The Precambrian bedrock, Tertiary veins and porphyritic intrusive rocks, and all accessible mines were mapped during an investigation of the uranium and associated ore deposits of the district, as part of a larger study in the central part of the Front Range mineral belt. This report describes the general character, structure, and mineralogy of the metalliferous vein deposits of the Idaho Springs district, and summarizes their relations to the Tertiary intrusives and the Precambrian rocks and structure. A report on the Precambrian geology of the district was published separately (Moench, 1964); the Tertiary intrusive rocks have been described by Wells (1960). Sims and others (1963) comprehensively reported on the uranium deposits of the region, including the few small deposits in the Idaho Springs district. Tooker (1956; 1963) studied the altered wallrocks adjacent to the veins in the region. The results of these specialized investigations within the district by Tooker, Wells, and Sims and others are summarized in this report.

## MINES AND PROSPECTS IN THE IDAHO SPRINGS DISTRICT

The mines and prospects whose locations are shown on plate 1 are identified in the following list. Detailed descriptions and maps of individual mines and prospects are not included in this bulletin because of the high cost of publication and the limited interest in them. A separate report entitled "Mines and Prospects, Idaho Springs District, Clear Creek and Gilpin Counties, Colorado—Descriptions and Maps," by R. H. Moench and A. A. Drake, Jr., which presented information supplemental to this bulletin, has been released for open file. The following list also indicates which mines were described and illustrated in the open-file report. Copies of part, or all, of the open-file report may be obtained at cost from the U.S. Geological Survey's Denver library, Federal Center, Denver, Colo., 80225.

*Key to mines and prospects shown on plate 1*

| Opening of mine or prospect | Location on pl. 1, this report (fig. 1, open-file report) | Open-file report | |
|---|---|---|---|
| | | Includes description | Includes illustration |
| Ace of Diamonds shaft | E–II, 13 | -------- | -------- |
| Aduddell shaft | G–I, 3 | × | × |
| Allan shaft | G–II, 11 | -------- | × |
| Alma Lincoln mine (adit level) | D–IV, 2 | × | × |
| Alpha shaft | H–II, 6 | -------- | -------- |
| Alpine adit | C–III, 6 | -------- | -------- |
| Amy shaft | G–II, 15 | -------- | × |
| Annie adit | C–IV, 1 | × | -------- |
| Anoka County adit | C–III, 5 | × | × |
| Argo tunnel | H–IV, 1 | × | × |
| Arizona shaft | G–II, 2 | -------- | -------- |
| Ashland shaft | F–II, 5 | × | -------- |
| Atlantic shaft | E–IV, 1 | -------- | -------- |
| Aurum adit | E–IV, 7 | × | -------- |
| Bald Eagle shaft | E–II, 9 | × | × |
| Bald Eagle Extension shaft | E–II, 5 | -------- | -------- |
| Banta Hill shaft | I–I, 1 | × | × |
| Banty shaft | C–V, 4 | × | -------- |
| Bell adit | G–III, 7 | × | -------- |
| Belle Vue shaft | C–IV, 2 | -------- | -------- |
| Bellman shaft | G–II, 8 | × | -------- |
| Bellman adit | G–II,7 | × | -------- |
| Belman shaft | E–I, 14 | × | × |
| Bertha shaft | G–I, 5 | -------- | -------- |
| Big Chief shaft | B–V, 2 | -------- | -------- |
| Big 51 shaft | C–V, 9 | -------- | -------- |
| Big Five (Central) tunnel | E–IV, 19 | × | × |
| Birtley adit | F–V, 1 | × | × |
| Borealis shaft | E–IV, 15 | -------- | -------- |

*Key to mines and prospects shown on plate 1*—Continued

| Opening of mine or prospect | Location on pl. 1, this report (fig. 1, open-file report) | Open-file report | |
|---|---|---|---|
| | | Includes description | Includes illustration |
| Boreas adit | F–II, 14 | × | × |
| Bourbon adit | C–I, 2 | × | -------- |
| Bride adit | F–II, 11 | × | -------- |
| Brighton shaft | F–II, 10 | × | × |
| Bronaber adit | D–III, 12 | × | × |
| Bryan adit and shaft | E–III, 11 | × | -------- |
| Bulgine shaft | C–V, 8 | -------- | -------- |
| Bullion adit | F–III, 5 | × | × |
| Bullion King No. 3 adit | D–V, 1 | -------- | -------- |
| Bullion King shaft | D–IV, 23 | × | -------- |
| Calvin shaft | F–II, 3 | × | -------- |
| Camp Valley adit | C–V, 10 | × | × |
| Carcasonne shaft | G–I, 7 | -------- | -------- |
| Cardigan mine | D–IV, 25 | × | -------- |
| Carlin shaft | C–V, 6 | -------- | -------- |
| Casino shaft | G–III, 4 | × | -------- |
| Casino adit | G–III, 3 | × | -------- |
| Castleton shaft | D–I, 5 | × | × |
| Central shaft | D–I, 12 | -------- | -------- |
| Centurion tunnel | B–IV, 4 | × | × |
| Champion (Bellevue) shaft | D–II, 2 | × | -------- |
| Champion Dirt No. 1 adit | B–IV, 2 | × | × |
| Christina shaft | E–III, 10 | -------- | -------- |
| Clarissa shaft | E–I, 15 | × | × |
| Clarissa adit (lower) | E–I, 16 | × | × |
| Clarissa adit (upper) | E–I, 13 | × | × |
| Clear Creek shaft | F–V, 3 | -------- | -------- |
| Colfax shaft | G–I, 6 | -------- | -------- |
| Collie adit | B–V, 6 | × | × |
| Columbia shaft | E–I, 1 | -------- | -------- |
| Columbia adit | D–III, 13 | -------- | -------- |
| Columbine adit | D–IV, 22 | -------- | -------- |
| Comstock shaft | C–IV, 8 | -------- | -------- |
| Comstock adit | E–IV, 5 | -------- | -------- |
| Comstock shaft | F–II, 6 | × | × |
| Cornucopia adit | F–V, 4 | × | × |
| Crocket shaft | E–IV, 10 | × | -------- |
| Crown Point and Virginia shaft | E–I, 4 | × | × |
| Crystal adit | F–III, 6 | × | -------- |
| De Lesseps shaft | G–II, 26 | × | -------- |
| Dexter adit | G–II, 27 | × | × |
| Diamond adit | A–V, 1 | × | × |
| Diamond Joe adit | E–II, 2 | × | × |
| Donaldson No. 6 Level adit | C–IV, 15 | × | × |
| Donna Juanita shaft | D–IV, 17 | × | × |
| Donna Juanita adit | D–IV, 18 | × | × |
| Dover prospect | C–III, 8 | × | × |
| Doves Nest shaft | F–II, 7 | × | × |
| Druid shaft | G–I, 2 | × | × |
| Dubuque shaft | C–III, 1 | -------- | -------- |
| East shaft | B–V, 4 | × | × |
| East Hukill shaft | E–IV, 3 | -------- | -------- |
| Eclipse shaft | G–III, 1 | -------- | -------- |
| Edgar adit and shaft | E–III, 13 | × | × |
| Edgardine shaft | G–III, 12 | × | -------- |

*Key to mines and prospects shown on plate 1*—Continued

| Opening of mine or prospect | Location on pl. 1, this report (fig. 1, open-file report) | Open-file report: Includes description | Open-file report: Includes illustration |
|---|---|---|---|
| Edgar Extension adit | F-III, 8 | × | × |
| Edna Fannie adit | G-IV, 2 | | |
| Edward shaft | C-IV, 5 | | |
| Edward adit | C-IV, 11 | | |
| Elkhorn shaft | G-I, 4 | | |
| Elliot and Barber adit | D-IV, 5 | × | × |
| England adit | C-IV, 3 | × | × |
| Enterprise shaft | E-I, 18 | | |
| Esmaralda shaft | G-III, 8 | × | |
| Essex shaft | E-IV, 13 | | |
| Etna adit | G-II, 24 | × | × |
| Eulalie adit | D-IV, 9 | | |
| Eureka-Swansea adit | E-II, 4 | | |
| Fairmount shaft | E-III, 4 | | |
| Fannie shaft | E-I, 12 | | |
| Fanny shaft | E-II, 6 | | |
| Forge Hill (Fairmont) adit | E-I, 17 | × | × |
| Fortune adit | E-III, 2 | | |
| Foxhall tunnel | G-III, 2 | × | × |
| Fraction shaft | B-V, 5 | × | × |
| Franklin No. 73 shaft | H-II, 1 | × | × |
| Franklin No. 87 shaft | H-II, 5 | × | × |
| Free Gold adit | C-II, 2 | × | |
| Freeman shaft | H-II, 7 | × | × |
| Freighters Friend shaft | G-II, 29 | × | × |
| French Flag shaft | H-II, 12 | × | |
| Frontenac adit and shaft | F-I, 2 | × | × |
| G and M (Centennial) adit | E-IV, 9 | × | × |
| Galatea shaft | H-II, 16 | | |
| Galatea (Hudson-Burr) adit | H-II, 14 | × | × |
| Garden shaft | G-II, 4 | × | × |
| Gem shaft | G-II, 13 | × | × |
| Gem adit | G-II, 28 | | × |
| German shaft | D-I, 6 | | |
| German adit | D-III, 1 | | |
| Gertrude shaft | E-IV, 6 | | |
| Gladstone adit | D-IV, 16 | × | |
| Glenella shaft | E-I, 2 | | |
| Gold shaft | I-I, 2 | | |
| Gold Bullion shaft | D-III, 2 | | |
| Gold Dust adit | D-III, 9 | | |
| Gold Dust shaft | E-III, 5 | | |
| Gold Medal (Silver Cycle or Wyoming Valley) adit | | × | × |
| Gold Medal shaft | I-III, 1 | × | |
| Gold Vault shaft | D-II, 8 | | |
| Golden Cloud shaft | E-I, 8 | × | |
| Golden Edge shaft | G-II, 6 | | |
| Golden Edge adit | G-II, 5 | | |
| Golden Hammer shaft | B-V, 1 | | |
| Golden Hammer adit | C-IV, 9 | | |
| Golden Link shaft (Stanley mine) | D-IV, 24 | × | × |
| Golden Link adit (Stanley mine) | D-IV, 27 | × | × |
| Golden Treasure adit | G-IV, 3 | | |
| Gondola adit | C-I, 3 | × | |
| Great American (Big Chief) adit | E-III, 8 | × | × |

*Key to mines and prospects shown on plate 1*—Continued

| Opening of mine or prospect | Location on pl. 1, this report (fig. 1, open-file report) | Open-file report | |
|---|---|---|---|
| | | Includes description | Includes illustration |
| Greenback shaft | B–V, 7 | | |
| Grover Cleveland shaft | D–I, 15 | | |
| Grover Cleveland shaft | E–I, 5 | | |
| Happy Easter (Queen Elizabeth) mine | G–III, 14 | × | × |
| Harpoon shaft | G–III, 13 | | |
| Hayes adit | F–II, 15 | × | |
| Helen adit | C–I, 1 | × | × |
| Highlander claim | | × | |
| Hoosac mine | C–III, 3 | × | |
| Hot Pot shaft | D–I, 8 | | |
| Houston shaft | C–IV, 4 | | |
| Hudson adit | E–II, 7 | × | × |
| Hughes shaft | D–I, 13 | | |
| Hukill shaft | E–IV, 2 | | |
| Hyland shaft | E–IV, 12 | | |
| Idaho tunnel | F–III, 7 | × | |
| Irene adit | D–V, 7 | | |
| Jackson shaft | G–II, 23 | | |
| Jennie Lind No. 1 adit | E–III, 9 | × | × |
| J. L. Emerson shaft | D–I, 14 | × | |
| J. Warner shaft and adit | G–II, 25 | | |
| John L. shaft | E–I, 11 | | |
| John Paul Jones adit | G–IV, 1 | | |
| Jones shaft | G–II, 14 | | |
| Josephine shaft and adit | D–IV, 4 | | |
| Jumbo adit | E–IV, 18 | | |
| Kangaroo shaft | F–II, 12 | | |
| Kelly No. 4 level adit | B–IV, 6 | | |
| Kelly shaft | C–IV, 16 | | × |
| Kentuck adit | E–II, 14 | | |
| Kinda-U.P.R. mine | C–I, 4 | × | × |
| Lafayette adit | E–IV, 17 | × | × |
| Lawrence L. (Philadelphia) mine | C–III, 7 | × | × |
| Lead Belt adit | F–II, 8 | × | |
| Liberator shaft | B–V, 3 | | |
| Little Albert No. 5 adit | B–IV, 1 | × | × |
| Little Annie adit | D–II, 6 | × | × |
| Little Cub adit | D–V, 5 | × | × |
| Little Ella shaft | E–IV, 8 | | |
| Little Emma adit | F–II, 9 | × | × |
| Little Harry adit | D–IV, 15 | | |
| Little Six adit | D–III, 10 | × | |
| Livingston shaft | E–I, 9 | | |
| Loeber shaft | E–III, 14 | | |
| Lord Byron shaft | C–V, 7 | × | × |
| Lost Summit shaft | E–III, 15 | | |
| Lost Vein adit | D–IV, 7 | × | |
| Lower East Lake adit | F–II, 2 | × | × |
| Lower Lake adit | E–II, 10 | × | × |
| Lucania tunnel | C–II, 1 | × | × |
| MAB adit | D–III, 15 | × | |
| M and E adit | D–V, 6 | × | × |

*Key to mines and prospects shown on plate 1*—Continued

| Opening of mine or prospect | Location on pl. 1, this report (fig. 1, open-file report) | Open-file report | |
|---|---|---|---|
| | | Includes description | Includes illustration |
| Manhattan shaft | C–V, 1 | | |
| Manhattan adit | C–V, 12 | X | X |
| Martha Perks adit | C–III, 2 | | |
| Mastedon adit | D–III, 11 | X | |
| Maude Munroe mine | D–W, 8 | X | X |
| May Day or Ready Cash adit | | X | X |
| Max shaft | G–II, 12 | | |
| Mayflower adit | E–IV, 16 | X | X |
| May Queen Annex adit | D–III, 4 | | |
| May Queen adit | D–III, 3 | X | |
| McMickle adit | E–IV, 20 | | |
| Merrimac adit | C–IV, 12 | | |
| Metropolitan tunnel | C–III, 4 | X | X |
| Metropolitan adit | F–II, 13 | X | |
| Metropolitan prospect | C–V, 3 | | |
| Miami tunnel | F–IV, 3 | X | X |
| Minnie shaft | D–I, 1 | | |
| Minott shaft | G–II, 22 | X | X |
| MIX adit | F–III, 1 | | |
| M K shaft | D–I, 7 | | |
| Mona adit | E–II, 1 | X | |
| Monte Cristo adit | D–IV, 1 | X | X |
| Moose shaft | H–I, 1 | X | |
| Morgan shaft | A–IV, 2 | | |
| Morning Star shaft | C–IV, 7 | | |
| Morning Star shaft | D–III, 6 | | |
| Morning Star shaft | G–II, 1 | | |
| Mount Etna adit | C–IV, 14 | | |
| Mount Vesuvius adit | C–IV, 13 | | |
| Myra shaft | C–V, 5 | | |
| Nashville shaft | E–I, 10 | | |
| Needham adit | G–II, 18 | | |
| New Bedford adit | D–III, 16 | (1) | (1) |
| New Century adit | E–IV, 11 | | |
| Niagara shaft | C–V, 11 | | |
| Nighthawk shaft | H–II, 15 | | |
| Nonpareil shaft | F–IV, 1 | | |
| Nonpareil adit | F–IV, 2 | X | X |
| No. 11 adit (Alma Lincoln mine) | D–IV, 20 | X | |
| No. 12 adit (Alma Lincoln mine) | D–IV, 21 | X | |
| October shaft | C–V, 14 | | |
| Old Settler adit | A–IV, 3 | | X |
| Old Settler shaft | A–IV, 4 | X | |
| Old Stanley shaft | D–V, 2 | | |
| Oliver shaft | E–IV, 14 | | |
| Oregon shaft | D–I, 4 | | |
| Oro Fino adit | H–II, 2 | | |
| Oro adit | D–III, 14 | X | |
| Ottawa shaft | G–III, 9 | X | |
| Owatonna shaft | F–I, 1 | X | X |
| Patten adit | F–III, 3 | X | |
| Pennsylvania adit | D–IV, 6 | X | |
| Phillips shaft | D–I, 9 | | |
| Phoenix adit | B–IV, 5 | X | X |

See footnote at end of table.

783–337 —65——2

*Key to mines and prospects shown on plate 1*—Continued

| Opening of mine or prospect | Location on pl. 1, this report (fig. 1, open-file report) | Open-file report: Includes description | Open-file report: Includes illustration |
|---|---|---|---|
| Phoenix prospect | C–V, 13 | | |
| Pine Shade shaft | G–II, 20 | × | × |
| Pine Tree shaft | H–II, 3 | | |
| President Hayes shaft | D–I, 3 | | |
| Pride of the West shaft | D–IV, 14 | | |
| Protection adit | E–III, 17 | × | |
| Providence shaft | E–III, 12 | | |
| Quartermaster shaft | F–IV, 4 | | |
| Red Jacket adit | F–III, 4 | × | |
| Red Lyon adits | D–V, 4 | × | |
| Refuge shaft | E–III, 7 | | |
| Reilly(?) shaft | H–II, 8 | | |
| Remington adit | F–III, 2 | | |
| Richmond shaft | D–II, 1 | | |
| Rickard shaft | G–II, 19 | × | × |
| Rio Grande shaft | E–I, 7 | | |
| Road Level adit (Stanley mine) | D–IV, 12 | × | × |
| Rockford tunnel | B–III, 1 | × | × |
| St. Joseph shaft | H–II, 4 | | |
| Salisbury mine | D–IV, 10 | × | |
| Santa Fe shaft | G–II, 9 | × | |
| Seaton shaft | G–II, 16 | × | |
| September adit | C–II, 3 | × | |
| 640 Level adit (Alma Lincoln mine) | D–IV, 19 | × | |
| Shafter adit | E–III, 3 | × | |
| Shafter shaft | E–III, 16 | × | × |
| Ship Ahoy shaft | H–II, 10 | × | |
| Silver Age shaft | H–II, 9 | × | × |
| Silver Age adit | H–II, 11 | × | × |
| Skyrocket mine | D–V, 8 | × | |
| South Lincoln and Ruby adit | D–IV, 3 | | |
| Spear adit | D–II, 7 | | |
| Specie Payment shaft | D–II, 3 | × | × |
| Specie Payment adit | D–II, 5 | | |
| Squaw shaft | C–IV, 10 | | |
| Stanley (Gehrman) shaft | D–IV, 11 | × | × |
| Star adit | D–V, 3 | (1) | (1) |
| Star prospect | E–III, 1 | | |
| Summit mine | E–III, 6 | × | |
| Summit shaft | H–II, 13 | | |
| Sun and Moon shaft | G–II, 3 | × | × |
| Sunnyside adit | A–IV, 1 | (1) | (1) |
| Sunny Side shaft | G–II, 21 | × | × |
| Syracuse mine | D–III, 5 | × | × |
| Telephone adit | D–II, 9 | | |
| Tigris adit | G–III, 6 | × | × |
| Tom Boy adit | G–III, 15 | × | |
| Torpedo mine | D–V, 9 | × | × |
| Transvaal adit | D–III, 7 | | |
| Treasure Vault shaft | G–III, 11 | × | |
| Treasure Vault adit | G–III, 10 | × | × |
| Trio adit | E–I, 6 | × | |
| Tropic (Trojan) shaft | G–II, 17 | × | × |

See footnote at end of table.

*Key to mines and prospects shown on plate 1*—Continued

| Opening of mine or prospect | Location on pl. 1, this report (fig. 1, open-file report) | Open-file report | |
|---|---|---|---|
| | | Includes description | Includes illustration |
| Tropic tunnel | G–III, 5 | × | × |
| Two Brothers tunnel | E–II, 8 | × | × |
| Tyson shaft | C–IV, 6 | -------- | -------- |
| Union adit | B–IV, 3 | × | × |
| United Gold Adit | C–II, 4 | × | -------- |
| U. S. adit | E–IV, 4 | × | -------- |
| Unknown adit | D–I, 2 | -------- | -------- |
| Upper East Lake adit | F–II, 4 | × | × |
| Upper Lake adit | E–II, 12 | -------- | -------- |
| Veto shaft | G–II, 30 | -------- | -------- |
| Victor shaft | D–I, 11 | -------- | -------- |
| Vida shaft | D–I, 10 | -------- | -------- |
| Waltham shaft | F–V, 2 | × | -------- |
| Ward adit | F–V, 5 | -------- | -------- |
| Welch(?) shaft | D–II, 4 | -------- | -------- |
| West Doves Nest shaft | F–II, 1 | -------- | × |
| West Santa Fe shaft | G–II, 10 | -------- | -------- |
| Whale adit (Stanley mine) | D–IV, 13 | × | × |
| Wheatland adit | C–III, 9 | -------- | -------- |
| Wild Rose shaft | C–V, 2 | × | -------- |
| Williams shaft | E–I, 3 | × | -------- |
| Willis Gulch shaft | G–I, 1 | -------- | -------- |
| Windsor Castle shaft | E–II, 11 | -------- | -------- |
| Wolverene adit | E–II, 3 | -------- | -------- |
| Wyandotte mine | D–III, 8 | × | -------- |
| York adit (Stanley mine) | D–IV, 26 | × | × |

[1] In Prof. Paper 371 (Sims and others, 1963).

## PREVIOUS STUDIES

The Idaho Springs district was studied during the early 1900's by the U.S. Geological Survey. Spurr and Garrey (1908) and Ball (1908), during study of the Georgetown quadrangle, mapped the southern part of the district on a scale of 1:62,500 and studied many of the most productive mines. Bastin and Hill (1917) mapped the north half of the Idaho Springs district (scale 1:12,000) during their study of the Central City quadrangle. Lovering and Goddard (1950) summarized the information on the district in their report on the mining districts of the Front Range, and Goddard (1947) prepared a separate brief summary.

## FIELDWORK

Fieldwork for the present investigation was done during the summers of 1953 and 1954. The surface geology was mapped on a scale of 1:6,000 on a special topographic base map prepared by the U.S.

Geological Survey from aerial photographs taken in 1951. The accessible mines were mapped on scales of 1 : 480, 1 : 600, and 1 : 1,200. Approximately 4 man-years were devoted to the study of the district by the writers and their associates. The areas mapped by the writers and those mapped by their colleagues are shown on the index on plate 1.

Concurrently with the study of the Idaho Springs district, other parties of the U.S. Geological Survey were working in the adjacent mining districts (fig. 2). The Freeland-Lamartine district and the Chicago Creek area, southwest of the Idaho Springs district, were studied by Harrison and Wells (1956, 1959). The Lawson-Dumont-Fall River area to the west was studied by C. C. Hawley and F. B. Moore, and the Central City district to the north was studied by Sims, Drake, and Tooker (1963). Specialized studies of the uranium deposits (Sims, Drake, and Tooker, 1963; Sims and others, 1963; Sims and Sheridan, 1964), Tertiary igneous rocks (Wells, 1960), and wallrock alteration (Tooker, 1963) were also made.

## ACKNOWLEDGMENTS

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## HISTORY, PRODUCTION, AND FUTURE[1]

On January 7, 1859, George A. Jackson, a native of Missouri, washed gold from gravels near the mouth of Chicago Creek. This was one of the first discoveries of gold in Colorado and it precipi-

[1] Many of the facts presented here were obtained from reports by Bastin and Hill (1917, p. 67–99) and by Spurr and Garrey (1908, p. 172–174).

tated a rush of prospectors into the region. Placer deposits were mined at first, but rich oxidized ores on the outcrops of the veins were worked as early as 1860. In 1864 the Whale Company was organized to work the Whale lode (now called the Stanley), one of the most productive in the district. Mining activity tapered off during the Civil War because of the shortage of men and materials but resumed with greater fervor shortly thereafter. Added stimulus was given to mining by the completion of the Union Pacific Railroad to Cheyenne in 1867 and by the completion of the Denver Pacific Line between Cheyenne and Denver in 1870. In 1873 the narrow-gage railroad between Denver and Floyd Hill, on Clear Creek, was completed; in 1877 it was extended to Georgetown. Most of the lode mining was confined to the oxidized ores during the first few years of activity. The advent of a satisfactory smelting process for sulfide ore in 1866 permitted profitable lode mining of the unoxidized ores. Though most of the early mining was for gold, silver became increasingly important from 1870 to 1873.

After 1866 the history of the district was closely tied with the development of better milling and smelting techniques and with fluctuations of national economy. Excellent summaries of milling and smelting practices are given by Bastin and Hill (1917, p. 153–163), and by Sims, Drake, and Tooker (1963).

A study of production records provides some insight to the economic factors that affected mining in the district. Records prior to 1904 are incomplete, but the production for the Idaho Springs district was influenced by the same factors that influenced production in all of Clear Creek County (Spurr and Garrey, 1908, p. 174–175). In the county production increased rapidly from $40,500 in 1866 to $2,203,948 in 1874, and more gradually, with some fluctuation, to $3,560,000 in 1894. The financial panic of 1873 caused a decline in the price of silver, and the mining emphasis shifted from the silver-rich veins to those containing a higher proportion of gold. Actual tonnage of ore shipped increased, and the total value of ores mined remained about the same. Apparently the panic of 1893 had a similar effect, because the production for 1894 was greater than that for any preceding or subsequent year. After 1894, production for the whole county gradually declined and, in 1903, was valued at only $1,792,203. This was probably due in part to a simultaneous decline in the price of silver from about $1.00 per ounce in 1892 to $0.60 per ounce in 1902.

Partial production records are available for the Idaho Springs district. From 1904 to 1959 inclusive the ore production from the Idaho Springs district (together with some of the mines outside the district) was valued at $14,987,314 (table 1), but the total production since mining began in the Idaho Springs district was far greater. From 1859 to 1903 the ore production from Gilpin County was valued at about $81 million (Bastin and Hill, 1917, p. 174–175; Sims, Drake, and Tooker, 1963), or about 3½ times the post-1903 production. The same ratio of pre-1903 to post-1903 production applied to the Idaho Springs district gives it a total production valued at about $65 million.

Mining activity in the Idaho Springs district was at its peak before 1900; its decline, which began after 1894, leveled off in 1904 when the Argo tunnel was driven to intersect at depth many of the valuable veins in the Central City and Idaho Springs districts. The veins did not prove to be as rich at depth as expected, however, and the tunnel did not greatly stimulate mining in the district. From 1905 through 1917 the annual production value ranged from $416,282 to $724,309. Production declined sharply because of World War I—from a value of $585,568 in 1918 to $86,957 in 1921. It reached an alltime low of $17,277 in 1924. Production increased again because of the additional labor supply that resulted from the financial crisis of 1929 and because of the increase in the price of gold in 1933 to $35 an ounce.

The value of ores produced in the district was $347,580 in 1934, reached a high of $688,001 in 1940, and dropped to a low of $41,949 in 1944 during World War II. During 1946–50 the annual production ranged from $11,850 to $198,362; during 1951–53 it decreased as a result of the lower base-metal prices. Exploration was stimulated by the demand for uranium in the early 1950's, but it did not result directly in an increase in base- or precious-metal production, nor did it result in any shipments of uranium ore. The Bald Eagle mine consistently produced from 1955 through 1959, however, and it accounted for most of the $328,000-worth of base- and precious-metal production from the district in 1956—the largest annual production of ore after 1941. From 1956 through 1959 production from the Bald Eagle mine gradually declined.

Throughout the history of the district, only a few mines produced most of the ore. Six properties that were discovered before 1874 each reportedly yielded ore valued at more than $1 million before 1900: the Stanley, Specie Payment, Sun and Moon, Frontenac, and Gem mines, and the French Flag–Silver Age–Franklin group. Other important mines, whose production was valued in the hundreds of thousands of dollars, include the Seaton, Aduddel, Druid, Crown Point and Virginia, Fraction, Champion Dirt, Shafter, Edgar, and Bald Eagle.

TABLE 1.—*Ore produced from the Idaho Springs district, 1904–59*

[Source of data: Years 1904–7 and 1952–59, U.S. Bur. Mines (unpub. compilation used by permission; total annual values based on average price of each metal for that year); years 1908–31, 1946–51, U.S. Bur. Mines; years 1932–45, Vanderwilt (1947). Data for all years except 1954–59 may include production from some mines outside the Idaho Springs district]

| Year | Ore sold or shipped (short tons) | Gold, fine (ounces) | Silver, fine (ounces) | Copper (pounds) | Lead (pounds) | Zinc (pounds) | Total value |
|---|---|---|---|---|---|---|---|
| 1904 | 3,702 | 3,639 | 61,302 | 6,302 | 156,548 | -- | $118,200 |
| 1905 | 32,039 | 18,339 | 252,418 | 40,409 | 341,272 | -- | 55,280 |
| 1906 | 30,164 | 16,203 | 185,370 | 108,115 | 500,892 | 1,279 | 509,744 |
| 1907 | 41,981 | 16,708 | 193,983 | 140,665 | 235,227 | 38,339 | 416,282 |
| 1908 | 66,147 | 25,370 | 263,479 | 212,435 | 779,709 | -- | 707,838 |
| 1909 | 52,564 | 18,769 | 182,163 | 176,101 | 867,597 | -- | 530,302 |
| 1910 | 56,605 | 14,712 | 171,248 | 314,759 | 688,114 | 31,978 | 458,703 |
| 1911 | 51,637 | 14,821 | 167,953 | 333,228 | 1,571,755 | 202,193 | 509,358 |
| 1912 | 46,074 | 12,890 | 161,681 | 255,061 | 897,341 | 13,435 | 440,639 |
| 1913 | 53,419 | 13,635 | 214,481 | 309,885 | 1,628,324 | 4,639 | 532,088 |
| 1914 | 62,587 | 18,698 | 193,221 | 250,744 | 1,239,572 | 12,530 | 575,694 |
| 1915 | 58,647 | 18,167 | 236,772 | 341,432 | 1,059,232 | 13,580 | 607,665 |
| 1916 | 51,044 | 14,771 | 231,673 | 409,789 | 2,029,029 | -- | 698,604 |
| 1917 | 40,839 | 10,693 | 260,346 | 424,186 | 1,979,360 | 26,666 | 724,309 |
| 1918 | 33,339 | 9,474 | 203,239 | 293,300 | 1,606,211 | -- | 585,568 |
| 1919 | 11,568 | 3,424 | 89,709 | 100,618 | 640,001 | -- | 223,896 |
| 1920 | 9,267 | 1,476 | 70,800 | 28,022 | 665,075 | -- | 166,050 |
| 1921 | 8,200 | 1,413 | 33,949 | 16,690 | 481,156 | -- | 86,957 |
| 1922 | 6,497 | 1,317 | 36,198 | 3,933 | 510,982 | -- | 92,049 |
| 1923 | 2,535 | 737 | 14,422 | 5,021 | 162,314 | -- | 39,165 |
| 1924 | 1,243 | 276 | 10,373 | 2,680 | 55,363 | -- | 17,277 |
| 1925 | 2,110 | 779 | 13,655 | 12,560 | 95,050 | -- | 37,020 |
| 1926 | 6,500 | 3,515 | 12,503 | 8,564 | 62,000 | 3,800 | 86,897 |
| 1927 | 4,104 | 2,335 | 11,956 | 16,381 | 68,190 | -- | 61,485 |
| 1928 | 1,452 | 638 | 9,429 | 4,424 | 88,707 | -- | 24,479 |
| 1929 | 5,685 | 2,545 | 27,940 | 37,068 | 127,984 | -- | 82,104 |
| 1930 | 23,642 | 4,466 | 27,956 | 13,000 | 123,000 | 14,000 | 111,588 |
| 1931 | 13,326 | 3,137 | 21,638 | 5,055 | 68,750 | -- | 74,126 |
| 1932 | 11,479 | 5,367 | 22,695 | 5,400 | 60,000 | -- | 119,478 |
| 1933 | 12,138 | 4,482 | 23,829 | 9,550 | 103,632 | -- | 105,437 |
| 1934 | 43,162 | 8,485 | 55,914 | 58,300 | 276,100 | -- | 347,580 |
| 1935 | 49,165 | 8,976 | 64,441 | 98,406 | 442,900 | -- | 386,381 |
| 1936 | 55,835 | 6,608 | 73,579 | 45,100 | 481,800 | -- | 314,588 |
| 1937 | 66,168 | 7,376 | 77,122 | 60,800 | 645,100 | -- | 363,218 |
| 1938 | 46,671 | 5,827 | 49,500 | 46,200 | 520,500 | 17,000 | 265,211 |
| 1939 | 50,371 | 10,878 | 68,441 | 129,700 | 693,400 | -- | 473,266 |
| 1940 | 97,507 | 15,449 | 112,344 | 183 200 | 933,900 | -- | 688,001 |
| 1941 | 67,997 | 12,785 | 75,648 | 106,700 | 549,000 | -- | 545,153 |
| 1942 | 39,504 | 6,171 | 30,611 | 56,100 | 193,300 | -- | 257,492 |
| 1943 | 6,664 | 860 | 22,576 | 34,500 | 231,400 | 153,000 | 84,518 |
| 1944 | 2,017 | 226 | 10,658 | 10,800 | 113,600 | 139,600 | 41,949 |
| 1945 | 3,990 | 355 | 11,264 | 11,600 | 155,000 | 202,200 | 58,584 |
| 1946 | 19,639 | 2,781 | 11,683 | 14,400 | 58,600 | 14,000 | 117,203 |
| 1947 | 7,222 | 2,021 | 14,992 | 7,700 | 159,400 | 24,600 | 111,850 |
| 1948 | 5,705 | 2,289 | 10,280 | 3,000 | 174,000 | 129,000 | 138,373 |
| 1949 | 8,126 | 2,311 | 24,961 | 16,000 | 415,000 | 211,000 | 198,362 |
| 1950 | 9,255 | 2,494 | 29,415 | 12,900 | 237,300 | 125,000 | 166,381 |
| 1951 | 3,347 | 987 | 10,215 | 2,500 | 98,500 | 41,000 | 68,898 |
| 1952 | 715 | 485 | 2,376 | 1,500 | 20,000 | -- | 22,708 |
| 1953 | 5,064 | 1,606 | 9,311 | 16,000 | 73,000 | 800 | 78,884 |
| 1954 | 1,382 | 805 | 14,191 | 4,400 | 93,800 | 8,500 | 22,322 |
| 1955 | 2,505 | 214 | 3,951 | 7,100 | 69,300 | 11,600 | 25,485 |
| 1956 | 8,344 | 4,462 | 39,302 | 88,000 | 625,500 | -- | 328,016 |
| 1957 | 9,115 | 4,151 | 34,088 | 80,500 | 685,400 | -- | 296,687 |
| 1958 | 7,402 | 1,282 | 23,868 | 50,400 | 528,800 | -- | 144,595 |
| 1959 | 6,438 | 1,658 | 24,095 | 54,600 | 404,800 | -- | 143,337 |

In value of the total ore output of the district, gold accounts for 59 percent, silver 18 percent, lead 15 percent, copper 7 percent, and zinc about 1 percent. Although sphalerite is almost as abundant as galena in most ores, zinc was not reported for many years.

In the future—unless metal prices increase spectacularly and more efficient means of mining are found—the district cannot be expected to produce significant quantities of precious and base metals. From 1910 through 1959 profits from most mining in the district were marginal. Mining was stimulated somewhat during periods of economic depression or recession, owing to lower mining costs and greater labor supplies. This pattern may continue, but the cost of rehabilitating the old mines may become prohibitive as the district ages. The district has been thoroughly prospected, and new ore bodies are likely to be found only on known veins. Most of the valuable veins contain much unexplored ground. Because the veins are typically narrow and require extensive timbering, however, even the most valuable veins are not generally amenable to low-cost large-tonnage operations that would be required for success under present economic conditions. Even under the most favorable circumstances, it may be impossible to amortize the costs of rehabilitation, exploration, development, and the construction of ore-treatment facilities.

## GENERAL GEOLOGY

The Idaho Springs district is underlain dominantly by gneissic, granitic, and pegmatitic rocks of Precambrian age (pl. 2), which constitute part of the core of the Front Range. These rocks are intruded by numerous small porphyritic dikes and irregular plutons of early Tertiary age and are cut by numerous faults that contain the ore deposits of the district. Some faults possibly originated in Precambrian time, but most formed near the close of the emplacement period of the early Tertiary magma sequence.

Physical character and structure of the Precambrian and Tertiary rocks had a marked influence on the formation of the fault patterns and on the localization of ore bodies. Accordingly, a brief description of the rock types and the structure of these rocks is given in the pages that follow. A more comprehensive report on the Precambrian rocks has been published separately (Moench, 1964). The petrography and structure of the Tertiary intrusive rocks in this district and adjoining ones were reported in detail by Wells (1960).

### PRECAMBRIAN ROCKS

The Precambrian rocks are a generally conformable succession of interlayered gneissic, granitic, and pegmatitic units. The gneissic rocks, which are dominantly metamorphosed sedimentary rocks, are the oldest and by far the most widespread and abundant rocks in the district. These rocks have been invaded by three varieties of granitic rock (from oldest to youngest): granodiorite, quartz diorite, and biotite-muscovite granite. Several types of small pegmatite bodies are interlayered with and locally cut the gneissic and granitic rocks.

The major gneissic rock units are interlayered biotite gneisses, granite gneiss, and microcline-quartz-plagioclase-biotite gneiss, which, for convenience, in most of the text are called, respectively, biotite gneiss, granite gneiss, and microcline gneiss. The biotite gneiss and granite gneiss are intermixed in layers that range from a fraction of an inch to several hundred feet in thickness. The thicker layers can be mapped individually at 1:6,000, but on plate 2 these two rock types are combined so that the major units shown are microcline gneiss and a mixture of biotite gneiss and granite gneiss. within the latter mixed unit, granite gneiss increases in abundance southwestward across the district, apparently at the expense of the biotite gneiss. Thinly layered rocks that consist of roughly equal proportions of biotite gneiss and granite gneiss are termed migmatite. Although migmatite was not noted at the surface, it was discernible in some mines. Small bodies and layers of amphibolite, calc-silicate gneiss, and quartz gneiss are associated with the major units; however, they are not shown on plate 2.

The biotite gneiss and associated minor rocks were assigned to the Idaho Springs Formation by Ball (1906) and by Lovering and Goddard (1950, p. 19–20); the microcline gneiss near Idaho Springs was mapped by Lovering and Goddard (1950, pl. 2) as quartz monzonite gneiss and gneiss pegmatite. The granite gneiss and associated pegmatite were mapped by Harrison and Wells (1956, p. 50–53) in the adjoining Freeland-Lamartine district.

In contrast to neighboring areas, granitic rocks are sparse in the Idaho Springs district. The few bodies of granitic rocks large enough to show on plate 2 are small and appear to be satellite to the larger plutons or batholiths that crop out to the southwest, west, and north of the district. The granodiorite is similar to the Boulder Creek Granite (Lovering and Goddard, 1950, p. 25–27); the biotite-muscovite granite is similar to the Silver Plume Granite from the type locality at Silver Plume, Colo., about 16 miles southwest of Idaho Springs (Ball, 1906). Lithologic names rather than geographic formational names are used here because the stratigraphy of the Precambrian metasedimentary rocks and correlations of the intrusive rocks are not fully established.

## GNEISSIC ROCKS

The gneissic rocks are divided into three major lithologic units: microcline gneiss (or microcline-quartz-plagioclase-biotite gneiss), biotite gneiss (or interelayered biotite gneisses), and granite gneiss (or granite gneiss and pegmatite). The biotite gneiss and granite gneiss are grouped on plate 2, and together they are conformably interlayered with units of microcline gneiss. If these units are assumed not to be overturned, a stratigraphic succession can be recognized as shown in section B–B′ of plate 2. The lowermost unit, a thick layer of mixed biotite gneiss and granite gneiss, forms the core of the Idaho Springs anticline on the southeast side of the district. This lowermost unit is overlain by a thin and discontinuous layer of microcline gneiss. The microcline gneiss, in turn, is overlain by a thick layer of mixed biotite gneiss and granite gneiss. A thick unit of microcline gneiss is higher in the stratigraphic succession, and it is overlain by a thick layer of mixed biotite gneiss and granite gneiss—the uppermost unit in the district.

In addition to these major units, small bodies of amphibolite, calc-silicate gneiss, and quartz gneiss are exposed, but these small outcrops are not shown on plate 2. All the gneissic rocks are described here without regard to their apparent stratigraphic position.

### BIOTITE GNEISS

Two main varieties of biotite gneiss are recognized: biotite-quartz-plagioclase gneiss and sillimanitic biotite-quartz-gneiss, both of which are locally garnetiferous. These rocks alternate with one another in layers that range from about an inch to several feet in thickness and probably represent the original bedding. In outcrops the biotite gneiss is marked by its dark-gray color, pronounced layering, and tendency to split parallel to the layering. Conformable layers and lenses of granite gneiss, present in most exposures, emphasize the layered appearance of the unit.

Typical biotite-quartz-plagioclase gneiss is fine grained, equigranular, light to dark gray, and is faintly to intensely foliated. The gneiss typically contains quartz and plagioclase in nearly equal amounts and 10–35 percent biotite. The feldspar composition generally ranges from oligoclase to andesine. Foliation is produced by a parallel alinement of biotite and locally by segregation of minerals into light and dark layers.

The sillimanitic biotite-quartz gneiss is light to dark gray and is well foliated; it is flecked with pods and smears of white fibrous sillimanite and has a marked schistose structure. The rock contains abundant quartz, about 20 percent biotite, as much as 30 percent sillimanite (but generally much less), and, commonly, some microcline and albite-oligoclase. Small amounts of muscovite can be found in most specimens, and garnet (almandine-spessartite) is locally abundant.

#### GRANITE GNEISS AND PEGMATITE

Granite gneiss (or granite gneiss and pegmatite) is exposed throughout the district in layers and lenses that range in thickness from less than an inch to several hundred feet. It is generally associated with biotite gneiss, and in many places relatively equal amounts of these two rock types are intimately mixed in thin alternating layers that form migmatite. Granite gneiss, some of which is also associated with microcline gneiss, is most abundant in the southwest corner of the district. Northeastward, layers of biotite gneiss are more abundant, and layers of granite gneiss are less abundant and discontinuous.

The granite gneiss is light colored and contains sparse to abundant wisps, laminae, and layers of biotite gneiss. Excluding the layers of biotite gneiss, most of the unit is nearly devoid of dark minerals and has the composition of a true granite; it contains abundant quartz and microcline and subordinate amounts of sodic plagioclase. The rock is typically fine to medium grained, and the gneissic structure is produced mainly by layers of slightly different grain size as well as by conformable inclusions of biotite gneiss. Locally, the rock is pegmatitic and contains feldspar crystals as much as 3 feet across. The coarse feldspar is white to pink and contains graphic intergrowths of quartz.

#### MICROCLINE GNEISS

Microcline gneiss (or microcline-quartz-plagioclase-biotite gneiss) is exposed in thin discontinuous layers near Idaho Springs and in a major layer that extends northward and eastward far beyond the limits of the mapped area. This extensive microcline gneiss layer wedges near the south margin of the district (pl. 2). The contacts between microcline gneiss and biotite gneiss are typically sharp, traceable for long distances, and provide the best structural "markers" in the district.

The microcline gneiss is a fine- to medium-grained light-gray or light-tan rock that is characteristically thinly laminated and well foliated. Laminae are typically 1 mm or less thick and are in alternating layers that are characterized by abundant or sparse biotite. Biotite-rich layers rarely exceed 1 inch in thickness. The rock contains 25–50 percent quartz, 30–55 percent oligoclase, as much as 35 percent microcline, and typically less than 10 percent biotite.

The microcline gneiss contains many small conformable layers and lenses of amphibolite and also some lenses of biotite gneiss and granite gneiss that are large enough to show on plate 2.

###### QUARTZ GNEISS

Several thin layers of quartz gneiss are exposed along the southeast side of the area, directly north and south of Idaho Springs. The layers are rarely more than about 15 feet thick, but they may be traced as far as 1 mile along their strike.

The quartz gneiss is light colored, fine to medium grained, and typically has a glassy luster. The gneissic structure of the rock results from slight differences in grain size in the laminae parallel to the rock layers. The rock contains as much as 80 percent quartz and some feldspar. Dark minerals, such as biotite and magnetite, are either sparse or absent.

###### AMPHIBOLITE AND ASSOCIATED CALC-SILICATE GNEISS

Amphibolite commonly is exposed along the contact between the microcline gneiss and the biotite gneiss in the form of lenses locally more than 100 feet thick. Amphibolite also forms smaller layers and lenses in the microcline gneiss and, less commonly, in biotite gneiss and granite gneiss.

The amphibolite is a dark-gray to black, fine- to medium-grained rock that contains hornblende and andesine in various proportions and small amounts of quartz. Biotite and pyroxene are common, though rarely are both present in the same specimen. Some varieties of amphibolite are massive and nearly structureless, others are gneissic and laminated. The gneissic structure is produced by alternating hornblende- and plagioclase-rich layers and by the planar orientation of hornblende.

Calc-silicate gneiss locally forms irregular masses or crosscutting veinlike structures in amphibolite. The contacts are ragged and the calc-silicate gneiss appears to be an alteration product of the amphibolite.

The calc-silicate gneiss is mottled, light to dark colored, and fine to coarse grained. It contains calcium-rich garnet, quartz, abundant epidote, and some hornblende, plagioclase, and, locally, clinopyroxene.

###### CALC-SILICATE GNEISS

Calc-silicate gneiss, apparently unrelated to amphibolite, forms large bodies on the north and east sides of Pewabic Mountain. The rock is mottled dark to light, crudely banded, and is fine to coarse grained. It is composed largely of diopside, epidote, quartz, scapolite, oligoclase, microcline, and arfvedsonite ( a sodium-rich amphibole) in various proportions.

#### ORIGIN OF THE GNEISSIC ROCKS

The gneissic rocks, with the exception of the granite gneiss, probably represent a thick succession of sedimentary rocks that were metamorphosed at high temperatures and pressures nearly equivalent to conditions of the upper range of the almandine-amphibolite facies, as defined by Fyfe, Turner, and Verhoogen (1958, p. 230–232). The biotite gneiss probably represents metamorphosed shale and interbedded sandstone, for this unit is marked by alternating sillimanitic and nonsillimanitic layers. The layers have the appearance of beds, and the sillimanite probably reflects the high aluminum content that characterizes most shales. The origin of the microcline gneiss is more debatable. Units of microcline gneiss are conformable with units of biotite gneiss and contain abundant conformable layers and lenses of amphibolite as well as some layers of biotite gneiss. These features are most easily explained as having resulted from sedimentary processes. Conceivably, the microcline gneiss represents metamorphosed arkose. Amphibolite is most abundantly exposed along the contacts between the microcline gneiss and biotite gneiss. This indicates that it was once a sedimentary rock. The composition of the amphibolite suggests formation from impure dolomitic sedimentary rocks; it could also represent metamorphosed basalt. One variety of calc-silicate gneiss probably formed by recrystallization of amphibolite, but the other varieties may represent calcareous sedimentary layers in the predominantly noncalcareous shales and sands that formed the biotite gneiss. The quartz gneiss layers probably represent quartz-rich sandstone beds.

Granite gneiss increases in abundance southwestward across the areas seemingly largely at the expense of biotite gneiss. This would indicate that the granite gneiss formed largely by replacement of the biotite gneisses; the ragged contacts observed in the outcrops support this interpretation. However, the possibilities that the observed regional change in composition represents a sedimentary facies change or that the rock was injected from a magmatic source are not precluded.

### GRANITIC ROCKS

The granitic rock units are intrusive igneous rocks and were emplaced in the following order: Granodiorite, quartz diorite, and biotite-muscovite granite. The granodiorite and quartz diorite occur as small, nearly conformable bodies, and the biotite-muscovite granite forms small sills, phacoliths, and a few thin crosscutting dikes.

#### GRANODIORITE

Several small bodies of granodiorite crop out in the southeastern part of the district. The granodiorite exposed here is fine grained and schistose and is similar to the granodiorite found near the borders of larger bodies in the Chicago Creek area (Harrison and Wells, 1959, p. 12). The rock is dark gray and composed of about 15 percent quartz, 40 percent oligoclase-andesine, 15 percent microcline, 20 percent biotite, and 10 percent accessory minerals—sphene, magnetite, apatite, allanite, and epidote.

#### QUARTZ DIORITE

Quartz diorite forms several small nearly conformable bodies on the southeast side of the district. The rock is similar in geologic occurrence and texture to the granodiorite but is darker in color. The central parts of the larger bodies of quartz diorite are dark gray, medium grained, and equigranular, whereas the margins are typically well foliated. The massive rocks contain as much as 70 percent combined hornblende, clinopyroxene, and biotite, as much as 40 percent plagioclase (andesine), and generally less than 15 percent quartz. Accessory minerals—sphene, apatite, allanite, pyrite, magnetite, and zircon—form as much as 10 percent of the rock. Some specimens of the well-foliated rocks contain as much as 15 percent of untwinned orthoclase.

#### BIOTITE-MUSCOVITE GRANITE

Biotite-muscovite granite forms small sills, phacoliths, and a few thin dikes. Only two bodies of biotite-muscovite granite are large enough to be shown on plate 2, but some small bodies are shown on many of the mine maps. The rock is light tan or gray, fine to medium grained, equigranular to subporphyritic, and is characterized by abundant tabular crystals of feldspar that are as much as 1 cm in length. Near the margins of some bodies most of the tabular feldspar crystals and biotite books are oriented about parallel to the contacts. The biotite-muscovite granite contains approximately 30 percent quartz, 60 percent feldspar, and less than 10 percent biotite and muscovite. Microcline predominates over plagioclase, which is mainly oligoclase.

#### PEGMATITIC ROCKS

Several types of pegmatitic rocks are exposed, but not all are readily distinguishable from the coarser parts of the granite gneiss. With one exception the two rock types are mineralogically similar, containing abundant quartz and microcline, subordinate amounts of plagioclase feldspar, and locally abundant biotite and magnetite. Some pegmatite dikes cut bodies of granodiorite or quartz diorite at various angles; they probably formed late in the cooling history of these rocks. Other dikes are similarly associated with biotite-muscovite granite. Still others cut some of the youngest Precambrian structural features of the district.

A type of pegmatite that contains disseminated uraninite is especially abundant in an area that trends northeast from the mouth of Fall River through Virginia Canyon to Seaton Mountain and perhaps beyond. This rock is of special interest, for an attempt was made to mine it as a source of uranium at the Highlander claim in Virginia Canyon. The pegmatite consists of quartz, microcline, and subordinate plagioclase (oligoclase), and as much as 20 percent biotite. The biotite occurs as randomly oriented books, laths, and knots. The uraninite is a primary mineral in the pegmatite; it is disseminated through the rock but is concentrated mainly in biotite. Molybdenite, galena, and pyrite are rarely present. Additional data on the pegmatite are given by Sims and others (1963, p. 10–12).

## TERTIARY INTRUSIVE ROCKS

The Idaho Springs district contains an intricate network of porphyry dikes and irregular plutons of early Tertiary age (pl. 1). These rocks constitute part of a belt of porphyries that extends northeastward across the Front Range and, together with the Tertiary mineral deposits, constitutes the Front Range mineral belt. Lovering and Goddard (1950, p. 47) inferred that the porphyries of the eastern part of the Front Range are early Tertiary in age. The bases of their inference were (1) the presence of interbedded volcanic rocks in the Upper Cretaceous and Lower Tertiary (Paleocene) Denver and Middle Park Formations, and (2) the relation of porphyry intrusives in different parts of the Front Range to the chronology of Laramide orogenic movements. This age is close to approximate absolute age of 60 million years determined on uraninite from metalliferous veins of the region (Faul, 1954, p. 263), for the veins formed during the waning stage of igneous activity.

Spurr and Garrey (1908), Ball (1908), and Bastin and Hill (1917) made the first comprehensive studies of the porphyries in the parts of the Idaho Springs district dealt with in their respective reports. Lovering and Goddard (1950) utilized these previous reports to aid their investigation of the whole Front Range mineral belt. More recently, Wells (1960) made a detailed petrographic study of the porphyries of the Idaho Springs and adjacent mining districts. With some modifications in terminology, Wells followed the classifications adopted by the earlier workers, and his classification is used in this report.

Wells (1960) described 13 kinds of porphyries, of which 9 are exposed in the Idaho Springs district. These rocks are separable on the basis of color, texture of the groundmass, size, shape, and abundance of the phenocrysts, qualitative and approximate quantitative mineralogy, and the character of fractured surfaces. Because Wells gave complete petrographic descriptions of all porphyries exposed in the region, these rocks are not described in detail here. Their salient characteristics are summarized in table 2.

The Tertiary igneous rocks were emplaced as listed in table 2, from oldest to youngest in ascending order. This sequence shown in table 2 agrees with that of Wells (1960, fig. 58), except that he reversed the emplacement order of trachytic granite porphyry and quartz bostonite porphyry; however, he noted (p. 229) that the intersecting relationships between these rocks do reverse locally. The sequence of intrusion was determined by crosscutting relations and faulting relations, some of which were observed in the Idaho Springs district. All varieties of porphyry except the biotite-quartz latite, the youngest of the sequence, are cut by the metalliferous veins at many places and were emplaced before the veins formed. The biotite-quartz latite, on the other hand, cuts metalliferous veins in many places.

The kinds of intrusive bodies formed by the porphyries changed as time passed. Several of the older porphyries—the albite granodiorite, light-colored granodiorite, and alkalic syenite—form irregular plutons and thick dikes in the northeast corner of the district, whereas the younger porphyries tend to form thin dikes throughout the district. The complex patterns of intersecting dikes in the northeast part of the district (pl. 1) indicate that composite plutons exist at some depth. Quartz monzonite prophyry, which is intermediate in age, forms one large lens and several small concordant lenses on the south end of Bellevue Mountain as well as many dikes throughout the district. Biotite-quartz latite, the youngest porphyry recognized, tends to form small lenticular bodies south of Clear Creek, many of which were emplaced along preexisting veins.

Except for the dikes that have been intruded along the Idaho Springs fault and the biotite-quartz latite dikes that follow veins, the porphyries appear to have been intruded along joints, not along faults. Characteristically, the country rock on either side of a dike has been separated but not offset, except locally where a dike has guided a later fault.

TABLE 2.—*Petrography of the Tertiary igneous rocks*

[Modified from Wells (1960, table 1). × indicates mineral present but variety not identified]

| Rock | Phenocrysts | | | Groundmass minerals | Remarks |
|---|---|---|---|---|---|
| | Potassic feldspar | Plagioclase | Ferromagnesian minerals | | |
| Biotite-quartz latite | ------ | Oligoclase-andesine. | Biotite books and flakes. | Quartz, potassic feldspar, biotite. | 2 varieties: (1) greenish-gray to dark-gray, aphanitic groundmass; 1/16-in. (or smaller) biotite and feldspar phenocrysts form 5 percent of rock, smooth conchoidal fracture; (2) brown-gray aphanitic groundmass; 1/16- to 1/2-in. lath-shaped feldspar phenocrysts form 20–50 percent of rock; sparse 1/16-in. biotite crystals, rough fracture. |
| Trachytic granite porphyry. | × | Albite-oligoclase | Bladed mineral | Quartz, potassic feldspar, sparse ferromagnesian mineral. | Pale-red to reddish-brown aphanitic groundmass and pink to white feldspar phenocrysts and local phenocrysts of dark ferromagnesian mineral; 2 varieties: (1) rounded 3/4-in. (or smaller) phenocrysts form 25 percent of rock; (2) lathshaped and rounded 1/4-in. (or smaller) phenocrysts form 10 percent of rock; both have rough granular fracture. |
| Quartz bostonite porphyry. | Anorthoclase | ------ | ------ | Anorthoclase, quartz | Groundmass aphantic to fine grained, pale red purple to grayish purple where phenocrysts sparse, pale red where phenocrysts abundant; 1/4-in. (or smaller) phenocrysts of pink to gray feldspar, locally altered to gray-green or white clay, form 0–25 percent of rock. |
| Bostonite porphyry | × | Albite-oligoclase | Bladed mineral | Potassic feldspar, sparse quartz, sparse ferromagnesian mineral. | Pink aphanitic to fine-grained groundmass, white 1/4-in. (or smaller) feldspar phenocrysts typically form 15–30 percent (locally, 5 percent) of rock; sparse 1/8-in. (or less) blade-shaped phenocrysts of ferromagnesian mineral. |
| Granite porphyry | × | do | Sparse bladed mineral. | Quartz, potassic feldspar. | 2 varieties: (1) gray to pinkish-gray aphanitic groundmass, sparse 1/4-in. (or less) feldspar phenocrysts; smooth conchoidal fracture; (2) light brownish-gray, fine-grained, aplitic groundmass; almost nonporphyritic; rough fracture. |
| Quartz monzonite porphyry. | Sparse sanidine | Oligoclase, sparse albite. | do | Quartz, plagioclase, potassic feldspar. | Light-gray aphanitic to fine-grained groundmass and sparse to abundant lighter gray, blocky 1/4-in. (or less) feldspar phenocrysts and sparse ferromagnesian mineral, locally nonporphyritic. |
| Alkalic syenite porphyry. | Antiperthite | Albite-oligoclase | do | Potassic feldspar, plagioclase, sparse quartz. | Light-brownish-gray fine-grained groundmass; conspicuous dark-gray diamond-shaped 1/8-to 1/4-in. feldspar phenocrysts form 25–50 percent of rock. |
| Albite granodiorite porphyry. | do | do | do | Quartz, plagioclase, sparse potassic feldspar. | Pinkish-gray aphanitic to fine-grained groundmass; lath-shaped 1/4-in. (or less, rarely 1/2 in.) feldspar phenocrysts form 40 percent of rock; rock of pluton northeast of Sun and Moon mine is brownish gray and garnetiferous. |
| Light-colored granodiorite porphyry. | Sparse anorthoclase. | Oligoclase | do | do | Light-gray aphanitic groundmass and abundant lighter colored euhedral blocky 1/2-in. phenocrysts and 1/8–1/4-in. rounded phenocrysts; commonly contains abundant wallrock inclusions. |

## QUATERNARY DEPOSITS

The Quaternary deposits are composed of alluvium, colluvial creep debris, and talus. Talus is common on the steep slopes below cliffs. Colluvial creep debris is widespread but was mapped only where it completely covers broad areas. The debris sheets rarely exceed 10 feet in thickness, but they effectively cover large areas of bedrock. These debris sheets are composed of a heterogeneous mixture of angular rock fragments and fine-grained material, some of which has moved downhill a considerable distance. Ridges of creep debris as high as 20 feet are common in many gullies that are flanked by debris sheets. The ridges are probably the result of pressures created by the persistent downhill creep on both sides of the gullies. The creep debris sheets may have formed partly in late Pleistocene time because of the more intense frost conditions that prevailed then (Harrison and Wells, 1959, p. 26).

Alluvium covers the floor of Clear Creek Canyon, parts of the valleys of Trail Creek and Spring Gulch, and, locally, the terraces that are well above Clear Creek. The alluvium at the present drainage levels consists of fine to coarse gravels, some of which is locally derived and some of which is derived from several miles upstream. Ball (1908, p. 83–84) noted three sets of terraces near Idaho Springs; these are cut in bedrock at about 160 feet, 55 feet, and 25 feet respectively above Clear Creek. The two higher terraces are capped by about 20 feet of gravel, and the lower is capped by about 5 feet of gravel. The terrace gravels are fine to coarse and contain well-rounded boulders and cobbles.

Some of the gravels on the terraces or on the present valley floors may have been deposited in Pleistocene time by melt waters from valley glaciers, which are known to have existed in the headwaters of Clear Creek and some of its tributaries. There is no evidence of glaciation in the Idaho Springs district other than the gravel deposits.

## STRUCTURE

The structural framework of the Idaho Springs district is outlined by the major units of comformable gneisses, which are folded along northeast-trending axes (pl. 2). During Precambrian time these rocks, which now strike generally northeast, were deformed at least twice, and may have been faulted. The first deformation was pervasive plastic folding that took place at considerable depth at high temperatures and pressures and was accompanied by intense recrystallization and the emplacement of many small bodies of granitic rocks. The second deformation was characterized by intense granulation as well as by folding in a relatively narrow zone.

It apparently took place at a somewhat shallower depth throughout a 2-mile-wide zone, termed the Idaho Springs–Ralston shear zone, which extends at least 20 miles northeastward to the margin of the Front Range (Tweto and Sims, 1963). During the Laramide orogeny the rocks were jointed, intruded by a sequence of porphyritic igneous rocks, and cut by an anastomosing network of faults.

The Precambrian structure of the district was described in detail by Moench (1964), and the joint patterns and Precambrian structure of a larger area were described by Moench, Harrison, and Sims (1962) and by Harrison and Moench (1961). These facets of the geology of the district are summarized here.

### FOLIATION AND LINEATION

All the gneissic Precambrian rocks and some of the granitic rocks rocks are characterized by well-developed foliation and lineation.

Foliation in the gneissic rocks is expressed by compositional layering in the rock and by preferred planar orientation of platy and tabular minerals. With few exceptions both features are parallel, and they are parallel to the major lithologic layers shown on plate 2. This type of foliation probably represents bedding in the original sediments.

Where the rocks were granulated by the younger Precambrian deformation, a cataclastic type of foliation formed. This foliation is characterized by a subparallel mesh of close-spaced fractures that are healed mainly by quartz. The cataclastic foliation typically is parallel to the older foliation described above, but locally it breaks across it.

The granitic rocks commonly have a foliation that is termed a primary flow structure. In the granodiorite, primary flow structure is expressed as an alinement of elongate inclusions parallel to discordant contacts (Harrison and Wells, 1959, p. 12). In the biotite-muscovite granite, tabular feldspar crystals commonly show a similar parallel alinement that has resulted from the flowage of partly crystallized magma.

Granodiorite and quartz diorite also show a secondary metamorphic foliation that is similar in character to the foliation in the gneissic rocks. In the Chicago Creek area the secondary foliation in granodiorite is locally superposed on the primary foliation and is continuous with the foliation in the gneissic rocks (Harrison and Wells, 1959, p. 12).

The gneissic rocks of the district are characterized by many kinds of lineation, but five categories are recognized: (1) the axes of small folds and crinkles; (2) elongate minerals, such as sillimanite and hornblende, and elongate mineral aggregates, such as small pods of granitic material; (3) the axes of boudinage, or "pinch" structures that formed by stretching; (4) rodding (rod-shaped features that result from the rolling or shearing between layers); and (5) slickenside striae.

The lineations have systematic orientations that can be related to each of the two Precambrian fold systems. Small folds, crinkles, and mineral alinements parallel the major folds of the older fold system throughout the district. Also, small warps, crinkles, boudinage, and sparse mineral alinements are oriented approximately at right angles to the major folds and probably formed at a late stage in the older deformation. Within the zone of younger Precambrian folding, small folds, crinkles, and sparse mineral alinements parallel the younger folds. Abundant rodding and slickenside striae are oriented about at right angles to the trends of the younger folds; they formed by the shearing that accompanied the younger folding.

## FOLDS

The gneissic rocks were deformed twice during Precambrian time. The first deformation, which took place at considerable depth in the earth's crust, was pervasive and resulted in major folds that trend sinuously north-northeast. These folds define the structural framework of the district. The second deformation, which took place at somewhat shallower depth, folded the incompetent biotite gneisses and associated rocks along axes that trend N. 55° E. and sheared the more competent microcline gneiss. Folds and shears of the younger deformation, which are superposed on the older Precambrian folds, are largely restricted to the southeast half of the Idaho Springs district (pl. 2). These effects of the younger Precambrian deformation represent part of the Idaho Springs–Ralston shear zone (Tweto and Sims, 1963, p. 998).

The major folds of the older deformation are wide and largely open. Their axes trend sinuously nearly north to about N. 50° E. (pl. 2). The Idaho Springs anticline is the dominant fold in the southeastern part of the district. This anticline is one of the major folds in this part of the Front Range, for it is an asymmetric feature that marks the boundary between a large area of rocks that strike mainly northeast on the northwest side of the axis, and a large area of rocks that strike west to northwest on the southeast side of the axis. Tweto and Sims (1963) interpreted this anticline as an early manifestation of the Idaho Springs–Ralston shear zone. The anticlinal axis trends about N. 60° E. in the southern part of the area and turns to

N. 45° E. in the central part; it plunges gently to moderately northeast. In the southern part of the district, the northwest limb of the Idaho Springs anticline is about 2 miles wide and dips steeply northwest; it is bounded on the northwest by the Trail Creek syncline, a relatively small, open fold (pl. 2). In the central part of the district the northwest limb of the Idaho Springs anticline is about 1½ miles wide and is bounded on the northwest by the Pewabic Mountain syncline—northwest of which are the Bellevue Mountain anticline and the Central City anticline (pl. 2). The four folds in the northwest part of the district are relatively gentle warps in rocks that are grossly flat lying. The Central City anticline, which enlarges northward, is the dominant fold in the Central City district (Sims, 1964).

Although the major older Precambrian folds are open and simple in their gross aspect, some, locally, are closed and overturned. The Bellevue Mountain anticline and Pewabic Mountain syncline, for example, are open warps near the contact between the microcline gneiss and biotite gneiss. Where these warps continue upward into the biotite gneiss, the limbs steepen and the folds overturn to the southeast (pl. 2). This feature reflects the incompetence of the biotite gneiss relative to the microcline gneiss—that is, the biotite gneiss flowed much more readily when it was deformed.

In contrast to the generally open and simple character of the major folds, many small folds—a few tens of feet wide or less—are closed and have axial plane that are subparallel to the rock layering. Many of these are drag folds that have formed by slippage of successively higher layers toward the anticlines. Most of the northwest-bearing small folds are open warps and apparently formed late in the older deformation.

Small folds and lineations that are related to the major older Precambrian folds are of many kinds and are ubiquitous. The axes of small folds, crinkles, and mineral alinements bear nearly north to N. 50° E., averaging about N. 25° E., parallel to the axes of the major folds. The axes of many small folds and boudinage, however, bear northwest about normal to the axes of the major folds. These northwest-bearing lineations were observed mostly in the northwestern part of the district.

The effects of the younger Precambrian deformation are largely restricted to the southeast half of the district. Southeast of the boundary, which is shown on plate 2, the gneissic rocks are pervasively granulated; thin sections of most of the specimens obtained from this zone show a fine network of anastomosing, subparallel fractures. The biotite gneiss within this zone is completely folded as well as granulated, whereas the microcline gneiss is granulated but rarely folded.

The younger folds are superposed on both limbs of the Idaho Springs anticline. They are small, have steeply dipping axial planes, and are distinctly asymmetric. In contrast to the sinuous northeast trend of the older folds, the younger folds trend N. 55° E. in a remarkably consistent pattern. Only a few folds exceed 100 feet in width, but even the small ones may be traced for long distances. Their plunge is extremely variable; it ranges from nearly horizontal just south of Idaho Springs to steeply northeast in many places to the north. The folds range in shape from structural terraces having nearly flat crests and steep northwest limbs to sharp-crested chevron folds and, locally, to nearly isoclinal folds. East-facing monoclines are locally present on the southeast limb of the Idaho Springs anticline. These shapes depend on the position of the minor folds on the Idaho Springs anticline and their origin may be explained by differential movements in which the northwest sides have been raised relative to the southeast sides (Moench and others, 1962).

Some of the large younger Precambrian folds are shown in sections on plates 2 and 3. Plate 3 was drawn from the map of the Big Five tunnel (Moench and Drake, 1966, fig. 6) and the geologic map of the district (Moench, 1964, pl. 1). The largest younger folds in the Big Five tunnel are exposed in biotite gneiss between the Hudson(?) and Shafter veins (pl. 3). On the tunnel level these folds are sharp-crested chevrons that have intricately crinkled opposing limbs. This is the most common shape of the younger folds in the district. The microcline gneiss, in contrast with the biotite gneiss, has been sheared on northwest-dipping fracture surfaces but has not been folded. This relationship has been observed at several places at the surface where the younger folds in the biotite gneiss were traced to the contact with the microcline gneiss but were not traced farther beyond.

Small folds and lineations of many types are related to the larger younger Precambrian folds. Small crinkles that are parallel to the younger fold axes are abundant in the folded biotite gneiss, and biotite and sillimanite have recrystallized locally along the younger fold axes. In addition, slickenside striae, rodding, and, locally, small folds and crinkles, consistently bear about N. 25° W. or S. 25° E. on the limbs of the younger folds that bear N. 55° E. The slickensides and rodding that bear N. 25° W. and S. 25° E. as well as the small drag folds that bear N. 55° E. were formed largely by slippage of successively higher beds toward the larger anticlines. In many outcrops the lineations of both the older and the younger fold systems are well preserved. The younger folds clearly bend and deflect the older lineations.

## FAULTS

Faults are abundant and closely spaced in the Idaho Springs district and form a complex intersecting pattern. With few exceptions the faults were mineralized with sulfide and gangue minerals in early Tertiary time, forming the veins shown on plate 1. To eliminate distortions of the pattern due to topographic relief, the major veins and faults were projected to a horizontal plane of 9,000 feet (fig. 3).

The faults of the Idaho Springs district formed in two different periods of fracturing. During the first period, probably in Precambrian time (Tweto and Sims, 1963, p. 1001), two prominent and a few lesser northwest-trending faults were produced. The second period of fracturing produced an anastomosing pattern of northeast-, east-northeast-, and east-striking faults. These faults cut all but the youngest of the Tertiary intrusive rocks, most of which were intruded along joints (not along earlier faults), and probably formed in early Tertiary time. A few east- to northeast-striking faults having unusually flat dips, however, may have formed in response to the major Precambrian(?) faulting. In addition, a few short north-trending faults of small displacement appear to have formed later than the second major stage of faulting. Evidence from other parts of the Front Range, however, indicates that some of the north-trending faults may have had their inception during Precambrian time.

Most of the faults are inconspicuous in outcrops, and it is very likely that many more exist than are shown on plate 1. Those that had well-developed gossans were prospected in the early stages of mining in the district and are now marked by rows of prospect pits and shafts. Many faults are narrow, gouge-lined slip surfaces that are subparallel to the layering of the rocks and are not obvious in the outcrops. If it were not for the sulfide ores they contain, the abundance and extent of the faults probably would not have been recognized.

Most of the faults dip northward at medium to steep angles, have dominant strike-slip displacements, and are right or left lateral. These terms refer to the apparent relative movements of the two walls when viewed in plan; right lateral indicates clockwise movement, and left lateral indicates counterclockwise movement. Many faults, however, show evidence of repeated movements, some of

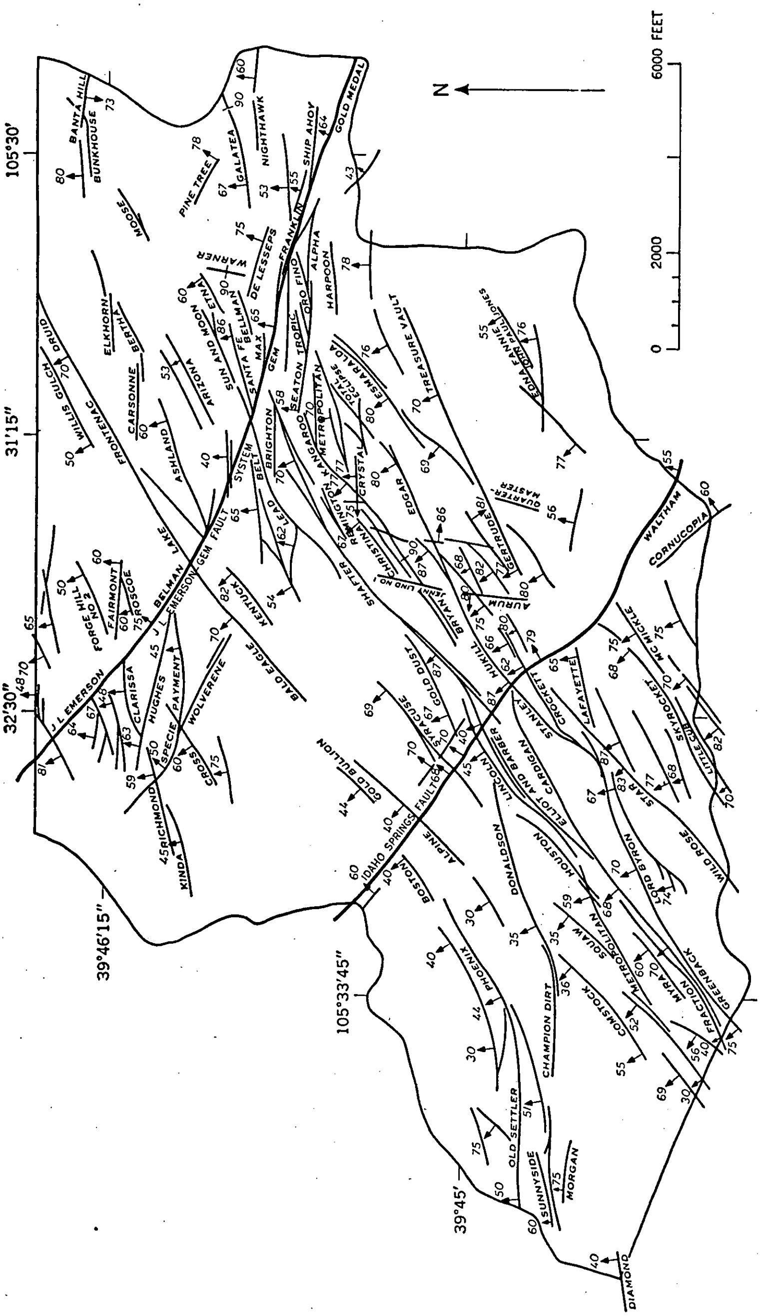


FIGURE 3.—Major veins in the Idaho Springs district projected to a horizontal plane at an altitude of 9,000 feet. Compiled by R. H. Moench.

which were quite different from the dominant relative displacements, and some show evidence of small dip-slip movement.

The faults of the Idaho Springs district may be grouped according to strike into five principal sets, which strike, respectively, northwest, northeast, east-northeast, east, and north. These sets show consistent movement patterns, and some of the sets have other distinguishing characteristics. As stated, the northwest-trending faults are inferred to have formed in Precambrian time and are called Precambrian(?) faults. The other faults formed largely in early Tertiary time and are called Laramide faults.

#### FAULTS OF PRECAMBRIAN(?) AGE

Two major faults—the Idaho Springs fault and the J. L. Emerson-Gem fault system—as well as a few lesser faults strike northwest. These faults are similar in trend and in many of their other characteristics to the faults known as "breccia reefs" or "breccia dikes" (Lovering and Goddard, 1950, pls. 1, 2; Lovering and Tweto, 1953, p. 30). Previously, Lovering and Goddard (1950, p. 79–82) postulated that fractures of this set formed during the Laramide orogeny, though they recognized that some may have followed preexisting zones of weakness. However, data summarized by Tweto and Sims (1963, p. 1001) suggest that these fractures formed in Precambrian time and were reactivated during the Laramide orogeny.

The Idaho Springs fault strikes N. 50°–60° W. and dips steeply north. Regional mapping by Sims (1964) in the Central City quadrangle has shown that the Idaho Springs fault extends a few miles northwest of the Idaho Springs district, but the extent of the fault southeastward is not known. The Precambrian rocks on opposite walls of the fault apparently have been displaced 500–600 feet by left-lateral movements (pl. 2), but several Tertiary porphyritic dikes cross the fault without displacement, and one dike of quartz-monzonite porphyry follows the course of the fault for more than a mile. The Cornucopia vein, which has a left-lateral displacement of about 100 feet, may branch southeastward from the Idaho Springs fault. The Idaho Springs fault shows little other evidence of branching or "horsetailing." The fault is a broken zone, about 50 feet wide, that has been well cemented with silica, some of which is chalcedonic, and that has been sparsely pyritized. Locally, the silicified breccia stands as low walls on the surface, much like the "breccia reefs" described by Lovering and Goddard (1950, p. 79).

The Idaho Springs fault shows evidence of repeated movements. Locally, it reopened in early Tertiary time and was filled by sulfide and gangue minerals, such as those at the Waltham mine (pl. 1).

Still later, minor northwest-trending faults near the Idaho Springs fault displaced the early Tertiary veins, as in the Dover, Lawrence L., and Syracuse mines. These relatively young fractures (faults) are typically gouge-lined and show small right-lateral displacements, opposite to the dominant left-lateral movements on the Idaho Springs fault.

The J. L. Emerson–Gem fault system strikes N. 50°–80° W. and dips steeply northward. Regional mapping by P. K. Sims (oral commun., 1960) has shown that the fault system continues northwest of the Idaho Springs district, and mapping by Lovering and Goddard (1950, pl. 2) suggests that it may join the Floyd Hill fault several miles to the southeast. In the Idaho Springs district the J. L. Emerson–Gem fault system contains several interconnected branching faults (fig. 3) whose total displacement is not known. Except for mines that mark its course, the fault system is inconspicuous at the surface, and it does not appreciably displace the Precambrian rocks. In mine workings the fault system was found to be similar to the Idaho Springs fault, containing as much as 30 feet of broken, silicified, and slightly pyritized rock. It was, however, much more extensively reopened and mineralized than the Idaho Springs fault, for it contains some of the most productive veins of the district.

### FAULTS OF LARAMIDE AGE

Closely spaced Laramide faults form an interconnecting network of fractures that strike east, east-northeast, and northeast. The faults cut all the early Tertiary intrusive rock types except the biotite-quartz latite—the youngest intrusive rock. These faults probably originated about contemporaneously, for they show consistent movement patterns that can be related to a simple compressional (or rotational) stress system. This interpretation is reinforced by the fact that the apparent order of formation of particular sets of faults is reversed from place to place. East-trending faults in the northeast part of the district, in general, are cut by faults that trend northeast, whereas, in the Old Settler mine (southwest corner of the area) the main east-trending fault cuts a fault that trends northeast. Reversals in the sequence have been noted in adjacent areas also. (See Sims and others (1963, p. 19–20); Harrison and Wells (1956, fig. 9).)

The distribution and attitudes of the Laramide faults were influenced somewhat by the Precambrian structure, as may be seen by comparing plate 2 and figure 3. In the central part of the district, most faults strike northeast to east-northeast and dip steeply northwest, subparallel to the layering of the Precambrian rocks. Here, east-trending faults are sparse. Southeast of the Idaho Springs anticlinal axis, where the rocks strike west to northwest and dip north, faults are far less abundant and the fault pattern is less well defined. The rocks in the northern part of the district are nearly flat lying, and most of the faults cut the layering of rocks at wide angles. Faults of the east-, east-northeast-, and northeast-trending sets are abundant here. In the southern part of the district, where the Precambrian rocks progressively flatten to the west, the veins show a parallel flattening apparently controlled by the flat dip of the country rock.

The east-trending faults are sinuous, ranging in strike from N. 70° W. to N. 70° E. A few faults in the northern part of the district dip steeply south, but most of them dip north at medium to steep angles and a few dip north at angles as low as 25°. Offset on the faults is dominantly left lateral—the north wall having moved a few feet west relative to the south wall.

East-northeast- and northeast-trending faults are difficult to differentiate according to strike, for some faults, such as the Lord Byron–Stanley, swing from N. 80° E. to about N. 40° E. Locally, however, a distinction between these sets of faults can be seen (fig. 3). Faults of both sets are nearly vertical in the central part of the district, but, northwestward, the dips flatten to as low as 30° to the north.

Displacements have been observed in many mines and are consistently right lateral. The largest apparent displacement was observed in the Bald Eagle mine, where the Precambrian rocks on the northwest side of the fault have been shifted about 80 feet northeast relative to those on the southeast side. Most of the slickenside striae are subhorizontal, which is evidence in support of the inferred dominant strike-slip movements. Downdip slickenside striae commonly cross the subhorizontal striae, and the latest formed sulfides commonly thicken in pods on the steeper parts of veins. This would indicate that slight normal downdip movement on many faults took place at a late stage of mineralization.

The east-, east-northeast-, and northeast-trending Laramide faults show evidence of repeated movement. Relatively early movements preceded mineralization and produced the channelways for the ore-forming fluid. Later movements accompanied and followed mineralization, for sulfide and gangue minerals are commonly brecciated and cemented by later sulfide and gangue minerals. In many places postmineralization gouge coats the vein walls.

A few nearly vertical north-striking faults that are grouped near the center of the district (fig. 3) probably formed last, for they cut several northeast- and east-northeast-trending faults. Evidently, the north-trending faults also postdate much of the sulfide mineralization, for they are characteristically lined with white clay gouge that contains pulverized sulfides that were probably dragged from the older faults.

### ORIGIN OF THE FAULTS

The extensive through-going northwest-striking faults contrast sharply with faults that trend east, east-northeast, and northeast; they probably were formed much earlier and under a different stress system. The stress system is not understood, but the great extent of the northwest-striking faults (Lovering and Goddard, 1950, pl. 1), indicate that the stresses must have affected much of the Front Range. Although the northwest-trending faults show evidence of repeated movements, their inception was probably during Precambrian time (Tweto and Sims, 1963).

Some flat-dipping faults that strike east to northeast, and some faults that strike north or north-northeast in parts of the Front Range—possibly including the Idaho Springs district—may also have formed in Precambrian time. Evidence for this interpretation was summarized by Sims and others (1963, p. 16–17), but we did not find such evidence in the Idaho Springs district where the north-trending faults cut some of the other Laramide faults and lack any indication of previous movement. The flat-dipping faults differ in no major respects, other than angle of dip, from the other east-, east-northeast-, and northeast-striking faults. Further, many steep faults flatten in depth, and many flat faults steepen; this indicates that the apparently flat and apparently steep breaks do not differ genetically. For these reasons the flat-dipping and the north-trending faults are grouped with the Laramide faults in this report.

Most of the Laramide faults appear to have formed after the emplacement of the quartz bostonite porphyry but before the biotite-quartz latite. In places outside the Idaho Springs district, the east-, east-northeast-, and northeast-striking faults formed in a definite sequence, but the fact that this sequence reverses locally suggests that the three sets of faults all formed over a short period of time. This apparently short period of time and the consistent movement patterns of the fault sets indicate that they are a conjugate shear system. The east-trending faults show left-lateral offset, the east-northeast-, and northeast-trending faults show right-lateral offset, and all three sets may have formed by shear stress related to compressive stress that was oriented east-northeast.

The few north-trending faults formed slightly later than the main network of Laramide faults, and their origin is less certain.

These faults are similar in orientation but are en echelon to the Precambrian(?) Dory Hill fault of the Central City district. (See Sims and others, 1963, fig. 6.) Conceivably, these faults may have formed by a reactivation of the Dory Hill fault during the Laramide orogeny.

The north-trending faults consistently show small left-lateral offset that took place at a late stage of mineralization. Interestingly, the northwest-trending faults in or near the Idaho Springs fault and the J. L. Emerson–Gem fault similarly show late, but right-lateral, offset. This offset is opposite to the dominant movements on the northwest-trending faults. These relations indicate that the late movements on the two fault sets were about contemporaneous. The north-trending and northwest-trending sets have the expected position and movement patterns caused by compression that was oriented north-northwest. These patterns, however, are also attributable to tensional stress that was oriented east-northeast. The tensional stress may have formed subsequent to the Laramide orogeny and seems more probable than a dominant north-northwest oriented compression.

### JOINTS

All rocks of the district contain two or more sets of joints. A study of joints in the rocks of the Idaho Springs and adjacent areas by Harrison and Moench (1961) revealed that many are related to the emplacement and cooling history of the Precambrian and Tertiary intrusive rocks; other joints may be related to the two Precambrian deformations, and still others may be related to the Laramide deformation. The joints believed to be of Laramide age provide much insight to the Laramide deformation. Detailed descriptions of the joints and the methods used to describe them were presented by Harrison and Moench (1961) and are only summarized here.

Harrison and Moench (1961) measured the attitudes of joints in outcrops throughout the area and plotted the data on Schmidt equal-area projections. The projections were then contoured—according to the method described by Billings (1954, p. 111–114) except that Harrison and Moench used the upper hemisphere—to reveal the dominant attitudes of the joint sets. The diagrams and the data on the geologic occurrence of joints were then examined to determine, as far as possible, those joint sets which may be related to the intrusive rocks, those which may be related to the two Precambrian deformation systems, and those, if any, which may be independent of these features. Some of the problems that arise in the interpretation of contour diagrams were discussed by Harrison and Moench (1961).

Harrison and Moench (1961) found that certain joints are largely confined to the intrusive rocks, others vary in attitude and abundance according to variations of the Precambrian structure, and still others, termed the "regional joint system," are nearly ubiquitous and consistent in attitude in areas of contrasting lithology and structure.

### PRIMARY IGNEOUS JOINTS

All the Precambrian and Tertiary intrusive rocks contain primary joints that probably formed during the emplacement and cooling of these rocks. Such joints may extend slightly into the wallrocks, but these joints are largely confined to the intrusive bodies. Tertiary dikes contain longitudinal joints that parallel the dike walls and "ladder" joints that are normal to the walls (Balk, 1937, p. 34–36, 97). The larger plutons of Tertiary porphyry undoubtedly have their own joint patterns, but they were not studied. The Precambrian granodiorite and biotite-muscovite granite have distinctive joints that commonly contain pegmatite or granitic dikes. (See Harrison and Wells (1959, p. 36) for a complete description of the primary igneous joints in Chicago Creek area.)

### JOINTS RELATED TO PRECAMBRIAN FOLDS

Some joint sets can be related to the Precambrian folds. A few joints are normal to the axes of folds of the older deformation—flattening in dip where the fold axes steepen in plunge—and apparently are cross joints that are related to the older fold system. Longitudinal joints, which parallel the fold axes, and diagonal joints, which form a small acute angle to the cross joints, apparently are related to older Precambrian folds. Cross joints that are possibly related to the younger Precambrian deformation strike about N. 35° W., about normal to the trend of the younger Precambrian fold axes; such joints are most abundant in areas that were strongly affected by the younger Precambrian deformation.

### REGIONAL JOINT SYSTEM

Four sets of joints are almost uniformly present in all rocks except the Tertiary porphyries and are remarkably consistent in attitude despite differences in Precambrian structure. These sets, termed the "regional-joint system," are summarized in figure 4, a stereodiagram that was constructed from the regional joint maxima on Schmidt contour diagrams of joints in the Idaho Springs district. The prevalence of these joint sets throughout the Idaho Springs and surrounding areas, the angular relations between the sets, and the fact that they intersect on a nearly common line suggest that they are genetically related to one another. The regional joint system

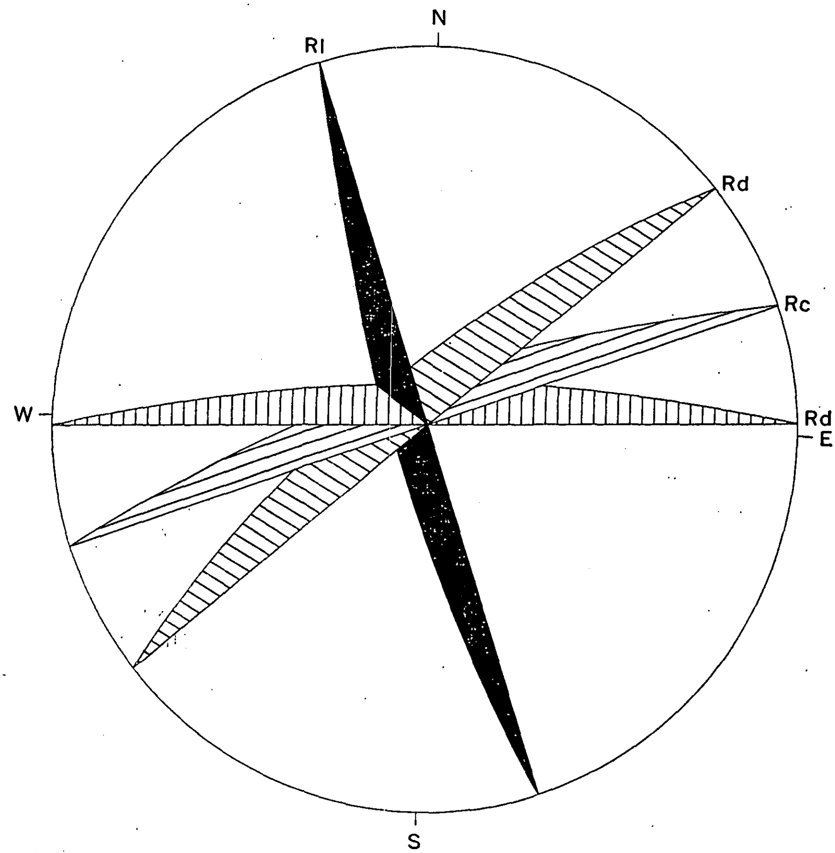


FIGURE 4.—Average attitudes of joints in the regional joint system: Rl, regional longitudinal; Rc, regional cross; and Rd, regional diagonal.

in turn shows systematic relations to the broad archlike form of the Front Range.

The uplift and arching of the Front Range in Paleozoic and Laramide times (Lovering and Goddard, 1950, p. 57–63) are recognized major disturbances after the emplacement of the biotite-muscovite granite to which the regional joint system might logically be related. The Laramide arching was the more pronounced of the two disturbances. This and the fact that the geometric relation of the regional joint sets to the known shape of the Laramide arch is similar to relations shown between joints and folds elsewhere (Harrison and Moench, 1961, fig. 8) indicate that the arch and the regional joint system are genetically related. The arch trends about N. 15° W. near the Idaho Springs district; accordingly, the joint set that strikes N. 19° W. is called the "regional longitudinal set" because it is subparallel to the arch. The arch disappears southward near Cripple Creek, Colo., which indicates that it plunges gently southward. A set of cross joints related to the arch should dip steeply north and strike east-northeast; accordingly, the set that strikes N. 70° E. and dips steeply north is termed the "regional cross-joint set." So-called diagonal joints are common in folded areas. Ideally, two sets are present, in each, the acute angle of intersection is bisected by the cross joints and the obtuse angle of intersection is bisected by the trend of the folds. The regional cross-joint set shown in figure 4 bisects the acute angle between two intersecting sets, which are accordingly termed the "regional diagonal-joint sets."

The regional joint system probably formed as a result of the Laramide arching of the Front Range under compressional stresses that were oriented east-northeast. The two diagonal joint sets have the position of shear fractures for such a stress system and probably originated by shear. The cross-joint set has the position of tension fractures for this stress system and may have originated by elongation or bending of the axis of the arch. Tension fractures may also form approximately parallel to the axial plane of a fold owing to the stretching of layers across the top of the fold, and they may form owing to the release of the principal stress. The regional longitudinal-joint set may have originated in both ways. (See Harrison and Moench (1961) for a review of the joints expected on folds.)

## GEOLOGIC HISTORY SUMMARIZED

In Precambrian time a thick sequence of sedimentary rocks was deeply depressed into the earth's crust; it recrystallized at high temperatures and pressures and was intensely deformed. The rocks were deformed plastically into large, mainly open folds whose axes trend north to northeast. Concurrently, small bodies of granodiorite, quartz diorite, and quartz gabbro were emplaced, recrystallized, and slightly deformed after they solidified. Small bodies of biotite-muscovite granite probably were emplaced late in this early deformation—apparently early enough to form phacoliths in folds of this deformation but too late to be deformed or recrystallized.

After an interval of deep erosion, the rocks were deformed again. This younger Precambrian deformation was largely restricted to the narrow Idaho Springs-Ralston shear zone which extends at least 20 miles northeast and a few miles southwest of the Idaho Springs district. Within this zone, which is about 2 miles wide in the Idaho Springs district, pervasive granulation and folding were superposed on the previously folded rocks.

Probably in late Precambrian time the through-going northwest-trending Idaho Springs fault, the J. L. Emerson-Gem fault system, and some smaller faults formed. A few reltaively flat-lying faults and some steep north-trending faults may also have had their inception at this time, although their main movements came much later.

The Front Range was arched during the Paleozoic, as indicated by the sedimentary record on its flanks (Lovering and Goddard, 1950, p. 29, 30), but this arching did not leave a recognizable record in the Idaho Springs district.

During the Laramide orogeny in Late Cretaceous and early Tertiary time, the Front Range highland was arched and uplifted (Lovering and Goddard, 1950, p. 57–60). In the Idaho Springs district this arching was accompanied by the formation of the regional-joint system and was accompanied or followed by the successive emplacement of nine varieties of porphyritic igneous rock—first, primarily in irregular plutons, thick dikes, and lenticular masses, and later, primarily in long narrow dikes. After most of the igneous activity but before the emplacement of the latest variety, a network of faults formed that trends east, east-northeast, and northeast. The northwest-trending faults were then reactivated. Openings along the faults were filled with ore and gangue minerals, and, as indicated by the brecciation of minerals in many veins, fault movements continued during mineralization. This arching, jointing, and faulting evidently took place under compressive stresses that were oriented east-northeast. Late in the stage of mineralization, a few north-trending faults formed, and renewed movements occurred on some northwest-trending faults. The movement patterns indicate that this late faulting occurred under tensional stress oriented east-northeast.

Since the Laramide disturbance the region has undergone deep erosion. The steep valley walls, narrow canyons, and fast-moving streams are indications that erosion continues today. The coarse alluvium in Clear Creek Canyon may, in part, be outwash from late Pleistocene valley glaciers that did not extend into the district. The continuous downhill movement of loose material is locally indicated by the colluvial creep debris sheets, though these, in part, may have formed in late Pleistocene time (Harrison and Wells, 1959, p. 26).

## ORE DEPOSITS

The principal ore deposits of the district are veins that are valued for their gold content and, to a smaller extent, for their silver, copper, lead, and zinc. The veins are fissure fillings and are similar in structure, mineralogy, and texture to deposits classified as mesothermal by Lindgren (1933, p. 530). Replacement deposits are scarce but have been mined in one area. Placer operations have been

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conducted along Clear Creek. These operations were described by Lovering and Goddard (1950, p. 99–102, 175) and therefore are not discussed in this report. Descriptions and maps of the mines and prospects of the Idaho Springs district are presented in the open-file report (Moench and Drake, 1966) supplementary to this bulletin.

The ore and gangue minerals typically fill openings in the Laramide faults and parts of the older faults that were reopened during the Laramide orogeny. Individual veins are fairly uniform in strike and dip and some may be traced for several thousand feet at the surface. Ore bodies formed where the openings along the fractures were wider than average. Such openings exist where the veins deviate from the average strike or dip and where faults intersect each other. These openings also exist where faults intersect brittle or competent rocks or layers whose structure is different from that of the neighboring rocks. Many veins are composed of several thin subparallel and anastomosing veins in zones as much as 30 feet thick but most commonly less than 5 feet thick. Other veins fill single fissures that may be 2 feet thick locally but are most commonly a few inches thick. In such veins the ore and gangue minerals were precipitated in symmetrical bands according to their relative ages. Where the wallrocks are coarsely brecciated, cockade structures that are characterized by encrustations of ore around the fragments may be seen. Vugs are not uncommon but, more commonly, the fissures are solidly filled.

Most veins show evidence of repeated movements. Early formed ore and gangue minerals were granulated and healed by deposition of later minerals, which, in turn, were broken and recemented one or more times. Subhorizontal slickenside striae are common on vein walls; in many places they are crossed by younger, steeply plunging striae. Major movements between certain mineralization stages are recorded in many veins; this indicates that the successive late-fault movements were virtually synchronous throughout the district.

The wallrocks have been altered, on both sides of the veins, in zones that range in width from 1 inch to several feet. There appears to be no correlation between the width of a vein and the width of the associated alteration zone, nor do all veins occupy the exact center of the alteration zones. None of the veins, however, turn from the altered zones into fresh rock.

Although many minerals in the veins have been identified, the simple assemblage of pyrite, sphalerite, chalcopyrite, tennantite, and galena predominates. Quartz is the principal gangue mineral, although carbonate minerals are abundant locally, and barite and fluorite are sparse.

The veins are classified according to the dominant ore minerals as pyrite, pyritic copper, pyritic lead-zinc, and lead-zinc veins. These vein types are distributed in a crudely defined zonal arrangement that is an extension of the concentric zonal pattern in the Central City district to the north (Sims, 1956, p. 745).

From stratigraphic and structural evidence, Lovering and Goddard (1950, p. 44–47) postulated that the igneous rocks and associated ore deposits of the mineral belt are Laramide or early Tertiary in age. This interpretation harmonizes in general with the many lead-uranium isotope determinations on uraninite from veins of the Central City district (Faul, 1954, p. 263), which gave ages that ranged from 57 to 70 million years.

## MINERALOGY

The principal metallic minerals are pyrite, sphalerite, galena, chalcopyrite, and tennantite. Native gold is sparse but widespread, and gold telluride minerals occur locally. Metallic minerals having more restricted distribution are enargite, pearceite, polybasite, marcasite, wolframite, pitchblende, and coffinite.

The vein minerals known to be present in the district are listed in table 3, in approximate order of abundance within each category. They include primary minerals which formed at depth during hypogene mineralization and secondary minerals which formed near the surface by alteration of the primary minerals.

### PYRITE

Pyrite is uniformly present in the veins of the district and is by far the most abundant metallic mineral. It is sparsely to abundantly disseminated in small grains throughout much of the altered wallrock and, locally, is distributed along tight fractures, well away from the major veins. Pyrite within the veins forms cubes and pyritohedrons as much as 2 inches across, but most of the grains and crystals are much smaller. Very commonly, pyrite is granulated and cemented by quartz and other minerals. Small veinlets of pyrite locally cut other metallic minerals, and, in places, the pyrite coats other minrals and encrusts vugs.

The pyrite in some symmetrically banded veins is of two varieties. That near the center of the veins is typically tarnished and is probably cupriferous whereas that in the outer parts is untarnished and somewhat whiter.

Most pyrite in the district is slightly anisotropic when specimens are viewed under a reflecting microscope. In some specimens the granulated pyrite appears to be more anisotropic than does the unbroken pyrite.

TABLE 3.—*Primary and secondary ore and gangue minerals*

| Primary (hydrothermal) | | Secondary (supergene and postmining) | |
|---|---|---|---|
| **Sulfides** | | | |
| Pyrite | $FeS_2$ | Covellite | $CuS$ |
| Galena | $PbS$ | Chalcocite | $Cu_2S$ |
| Sphalerite | $ZnS$ | Argentite | $Ag_2S$ |
| Chalcopyrite | $CuFeS_2$ | | |
| Marcasite | $FeS_2$ | | |
| Bornite | $Cu_5FeS_4$ | | |
| **Sulfosalts** | | | |
| Tennantite | $(Cu,Fe,Zn,Ag)_{12}As_4S_{13}$ | Pyrargyrite-Proustite(?) | $Ag_3(As,Sb)S_3$ |
| Enargite | $Gu_3AsS_4$ | | |
| Pearceite | $(Ag,Cu)_{16}As_2S_{11}$ | | |
| Polybasite | $(Ag,Cu)_{16}Sb_2S_{11}$ | | |
| **Native Elements** | | | |
| Gold | $Au$ | Gold(?) | $Au$ |
| | | Silver | $Ag$ |
| | | Copper | $Cu$ |
| **Oxides** | | | |
| Uraninite | $UO_2(UO_3)$ | Hydrous iron oxides | |
| | | Wad: hydrous manganese oxides | |
| | | Hematite | $Fe_2O_3$ |
| **Tungstates** | | | |
| Wolframite | $(Fe,Mn)WO_4$ | | |
| **Tellurides** | | | |
| Sylvanite(?) | $(Au,Ag)Te_2$ | | |
| **Halides** | | | |
| Fluorite | $CaF_2$ | | |
| **Phosphates** | | | |
| | | Autunite | $Ca(UO_2)_2(PO_4)_2 \cdot 10\text{–}12H_2O$ |
| | | Meta-autunite | $Ca(UO_2)_2(PO_4)_2 \cdot 2\frac{1}{2}\text{–}6\frac{1}{2}H_2O$ |
| | | Torbernite | $Cu(UO_2)_2(PO_4)_2 \cdot 8\text{–}12H_2O$ |
| | | Metatorbernite | $CU(UO_2)_2(PO_4)_2 \cdot 8H_2O$ |

TABLE 3.—*Primary and secondary ore and gangue minerals*—Continued

| Primary (hydrothermal) | | Secondary (supergene and postmining) | |
|---|---|---|---|
| **Silicates** | | | |
| Quartz | $SiO_2$ | Quartz | $SiO_2$ |
| Chalcedony | $SiO_2$ | Uranophane | $Ca(UO_2)_2Si_2O_7 \cdot 6H_2O$ |
| Coffinite | $U(SiO_4)_{1-x}(OH)_{4x}$ | | |
| **Carbonates** | | | |
| Ankerite | $Ca(Mg,Fe)(CO_3)_2$ | Cerussite | $PbCO_3$ |
| Siderite | $FeCO_3$ | Siderite | $FeCO_3$ |
| Rhodochrosite | $MnCO_3$ | Malachite | $Cu_2CO_3(OH)_2$ |
| Dolomite(?) | $CaMg(CO_3)_2$ | Azurite | $Cu_3(CO_3)_2(OH)_2$ |
| Calcite(?) | $CaCO_3$ | Smithsonite | $ZnCO_3$ |
| | | Bayleyite | $Mg_2(UO_2)(CO_3)_3 \cdot 8H_2O$ |
| | | Calcite(?) | $CaCO_3$ |
| **Sulfates** | | | |
| Barite | $BaSO_4$ | Anglesite | $PbSO_4$ |
| | | Schroekingerite | $NaCa_3(UO_2)(CO_3)_3(SO_4)F \cdot 10H_2O$ |
| | | Betazippeite | $(UO_2)_2(SO_4)(OH)_2 \cdot 4H_2O$ |
| | | Zippeite | $(UO_2)_2(SO_4)(OH)_2 \cdot 4H_2O$ |
| | | Copper sulfate minerals | |
| | | Epsomite(?) | $MgSO_4 \cdot 7H_2O$ |

## SPHALERITE

Sphalerite is a major constituent of many veins and is present at least locally in most veins. It is typically intergrown with pyrite, galena, chalcopyrite, and tennantite, and locally forms large masses. Sphalerite coats and embays pyrite in most specimens and is coated or veined and embayed by other sulfide and sulfosalt minerals. Blebs of chalcopyrite are disseminated through most specimens of sphalerite; they are commonly alined in trains along crystallographic planes of the sphalerite and typically most abundant near the borders of sphalerite grains.

The sphalerite ranges from reddish brown to pale yellow or amber, and, hence, is in part known as "amber jack." As shown in table 10 (p. 88), the darker sphalerite contains about 4 percent of iron by weight, which is considerably less than the maximum iron content of sphalerite from the Central City district (Sims and Barton, 1961; 1962). The darkest sphalerite grains from the Idaho Springs district show marked color zoning, however, and the maximum iron content undoubtedly exceeds the maximum listed in table 10. Light-colored sphalerite generally contains less than 1 percent of iron, and one specimen from the Mayflower adit is nearly pure ZnS (table 10). The light-colored sphalerite is not obviously zoned.

#### GALENA

Galena is abundant in many veins and present in most. It is fine to coarse grained and locally forms cubes as much as an inch across. Some fine-grained galena has been intensely sheared, and the cleavage planes show marked curvatures. Galena commonly encrusts sphalerite; locally, thin veinlets of galena cut through the older minerals.

Some galena contains much silver; for example, galena from the Banta Hill mine contained 81.07 ounces of silver and 0.04 ounces of gold per ton. (Analysis by D. L. Skinner and E. C. Mallory, Jr., U.S. Geological Survey, Denver).

#### CHALCOPYRITE

Chalcopyrite is present in most veins but is seldom abundant. It forms blebs in sphalerite (where it is most abundant in the darker varieties) and also forms irregular blebs in pyrite. It occurs also as veinlets that cut other sulfide minerals—especially tennantite and galena—and as discrete clots and vug fillings in the ore.

#### MARCASITE

Marcasite is scarce in the district. Trace amounts are intimately associated with pyrite in the Bald Eagle mine and the Mayflower adit.

#### BORNITE

Bornite is rarely found. In one specimen a small patch of bornite embayed chalcopyrite, and in another specimen bornite embayed sphalerite but was itself embayed by chalcopyrite. Bastin and Hill (1917, p. 363) noted a small amount of bornite intergrown with galena in a specimen from the Gladstone mine.

#### COVELLITE

Covellite, a secondary sulfide mineral, was found in minor amounts throughout the district in near-surface workings. It embays all the primary base-metal sulfide minerals but preferentially embays galena.

#### CHALCOCITE

Chalcocite, though less abundant than covellite, is similarly distributed in the near-surface parts of many veins. Like covellite, the chalcocite preferentially embays galena. In one specimen covellite embays chalcocite.

### ARGENTITE

Argentite was found only in the Freighters Friend mine; it embays galena.

### TENNANTITE

Tennantite (gray copper) is present in most veins of the district but is rarely abundant, although in the northeastern part of the district it seems to predominate over chalcopyrite. It is myrmekitically intergrown with galena and veined by galena locally, but it veins galena and chalcopyrite as well.

In reflected light the tennantite has a greenish hue that is indicative of the high arsenic content demonstrated by analyses. A sample of tennantite from the Tropic mine contained 11 percent arsenic and 6.3 percent antimony, and a sample from the Kokomo mine, just north of the district boundary on the Frontenac–Druid vein, contained 19 percent arsenic and 1.4 percent antimony (X-ray fluorescent spectroscopy analyses done by W. W. Niles, U.S. Geol. Survey, Denver).

### ENARGITE

Enargite is widely distributed but is rarely abundant. It commonly forms blebs, rods, and plates oriented along crystallographic planes in tennantite, and it also accompanies chalcopyrite or tennantite in veinlets that cut galena and sphalerite. Enargite locally embays chalcopyrite. In one polished section, enargite was myrmekitically intergrown with galena, and in another, it had replaced the galena of a chalcopyrite-galena intergrowth, thereby forming a chalcopyrite-enargite intergrowth.

Under the reflecting microscope the enargite is strongly anisotropic, showing polarization colors of purple, gray, and greenish yellow. Some is gray, but most is pinkish. The pinkish variety may be mistaken for luzonite, a mineral of similar composition but different (tetragonal) from enargite (orthorhombic) in crystallography. (See Skinner, 1960.) X-ray studies indicate that the pinkish variety is true enargite and not luzonite (Paul B. Barton, written commun. to P. K. Sims, 1961).

### PEARCEITE

Pearceite was found in only a few polished sections from the Donaldson, Bald Eagle, Edgar, Alma Lincoln, and Phoenix mines. The mineral is fine grained and is generally associated with galena, though a few patches in pyrite were seen. Pearceite typically occurs as intergrowths with galena and as blebs, veinlets, and embayments in galena.

### POLYBASITE

Polybasite was seen only in specimens from the Freighters Friend, Phoenix, and Alma Lincoln mines from the Mayflower adit. It forms blebs and blades in pyrite and galena and locally embays galena along grain boundaries. One polished section from the Lincoln vein showed a blade of polybasite rimmed by enargite, which, in turn, was myrmekitically intergrown with galena.

### PROUSTITE-PYRARGYRITE

Proustite-pyrargyrite (ruby silver) was not identified in this study. It reportedly occurs, however, in the Seaton mine, where it probably is secondary in origin (Bastin and Hill, 1917, p. 296).

### GOLD

Native gold probably occurs in all veins of the district, but it is not always visible under the microscope. In polished surfaces it was seen as euhedral crystals in pyrite; as blebs in quartz, carbonate minerals, pyrite, chalcopyrite, tennantite, and galena; and as veinlets along contacts between these minerals or as minute veinlets cutting them. In the Treasure Vault mine, native gold, associated with sylvanite(?) coats irregular fracture surfaces and forms small veinlets in the altered wallrocks. Some native gold may be secondary, for Bastin and Hill (1917, p. 139) reported having found secondary growths of the metal in the oxidation zone.

### SILVER

Native silver was reportedly found in the Franklin, Seaton, and other mines (Bastin and Hill, 1917, p. 292, 296). It occurs as wires and vug fillings as a product of secondary enrichment.

### COPPER

Native copper, according to Bastin and Hill (1917, p. 363), was deposited by mine waters in some of the ore of the Gladstone mine.

### PITCHBLENDE AND COFFINITE

Pitchblende is present in small amounts in the Sunnyside mine, the Red Jacket adit, and the Star adit, and is associated with coffinite in the Stanley mine. Radioactive material in ore on the No. 2 level of the Alma Lincoln mine may be pitchblende but it was not positively identified.

In the Sunnyside mine, pitchblende occurs in one of the veins that branches from the main Sunnyside vein, and probably in several others. All these veins contain quartz and pyrite and subordinate amounts of chalcopyrite, tennantite, galena, and sphalerite. At places they are symmetrically banded—quartz and pyrite cover the vein walls, and base-metal minerals form the central band. Small fragments of hard lustrous pitchblende embedded in the outer bands appear to be breccia fragments cemented by quartz and pyrite. In addition, sooty pitchblende covers the slickensided surfaces on the vein walls.

In the Red Jacket adit the pitchblende is sparsely intergrown with quartz and pyrite. In a polished section the pitchblende is hard and lustrous and occurs as highly irregular, delicate branching forms that are aggregated along thin ill-defined stringers.

In the Star adit pitchblende occurs in fractured, but nearly unaltered, granite gneiss a few feet from the Star vein, which itself is not exceptionally radioactive and probably does not contain pitchblende. The pitchblende is associated with quartz, pyrite, wad, meta-autunite, metatorbernite, and uranophane, which fill fractures in the granite gneiss. As viewed in a polished surface, the pitchblende occurs in thin veinlets that have spheroidal outer surfaces; it is intergrown with quartz and pyrite. Some of the pitchblende is hard and lustrous and some is sooty.

In the Stanley mine pitchblende and coffinite, associated with small amounts of pyrite, quartz, and colorless sphalerite, occur in veinlets that cut across the sulfide and gangue minerals of the Stanley vein. The coffinite is finely botryoidal and translucent. In polished surfaces it appears medium gray and is finely interbanded with lighter-gray pitchblende. An X-ray powder pattern shows pronounced sharp coffinite lines and faint, diffuse uraninite lines.

In the Alma Lincoln mine pitchblende(?) occurs in veinlets that crosscut the sulfide and gangue minerals of the Lincoln vein.

### HEMATITE

Hematite was found only in the Donaldson mine and the Cornucopia adit. In the Donaldson mine hematite and covellite replace galena. In the Cornucopia adit trace amounts of hematite are associated with sphalerite.

### HYDROUS IRON AND MANGANESE OXIDES

Hydrous oxides of iron, termed "limonite," are abundant in the oxidized parts of most veins. Limonite has formed in many places since mining operations began. In the Star adit the hydrous oxides of manganese, termed "wad," are associated with metatorbernite, meta-autunite, and uranophane; these materials line fractures in granite gneiss.

### WOLFRAMITE

Wolframite was found in small amounts in the Mayflower adit and in the Collie adit of the Fraction mine. It occurs as euhedral randomly oriented tabular crystals intergrown with early formed quartz. In a simple symmetrical vein, wolframite, quartz, and minor amounts of pyrite form the outer parts of the vein. Wolframite decreases in abundance toward the center of the vein and pyrite increases; both are intergrown with quartz.

The wolframite is probably similar to huebnerite in composition, for in polished surfaces it is gray, has dark red-brown internal reflections, and is only moderately hard. Qualitative visual arc-spectrum tests by J. W. Adams of the U.S. Geological Survey revealed abundant tungsten, iron, and manganese.

### TELLURIDE MINERALS

Telluride minerals reportedly occur in the Treasure Vault, Casino, and Gem mines in the Virginia Canyon and Seaton Mountain areas (Bastin and Hill, 1917, p. 114), and in the Torpedo mine in Spring Gulch (Spurr and Garrey, 1908, p. 380). Except for a small blade of a creamy white strongly anisotropic mineral at the contact of a gold octahedron seen in one polished section of ore from the Lower Lake adit, we found no gold telluride minerals. Available descriptions (Bastin and Hill, 1917, p. 114) suggest that sylvanite is the dominant gold telluride mineral. In the Treasure Vault mine native gold and sylvanite(?) form irregular films on the fracture surfaces near relatively flat veins and veinlets that cut brown-stained altered rock. According to Bastin and Hill (1917, p. 114), sylvanite(?) "occurs in a gangue of blue-gray cherty silica, with which are also associated small amounts of fluorite, ferruginous calcite, and fine pyrite." Other sulfides are present on other parts of the Treasure Vault vein, but their relation to the telluride-bearing ore is not known.

### FLUORITE

Fluorite has been found in the Banta Hill, Moose, Specie Payment, and Treasure Vault mines in the northern part of the district. In the Banta Hill mine, purple fluorite occurs in the Bunkhouse vein, where it is associated with white quartz, galena, sphalerite, rhodochrosite, and pyrite. White and green-and-purple fluorite occur abundantly in association with rhodochrosite, quartz, and sulfide and sulfosalt minerals in ore fragments on the Moose mine dump. In the Specie Payment mine purple fluorite was seen along foliation planes and in tiny fractures in the wallrocks adjacent to the vein. Purple fluorite in the Treasure Vault mine fills vugs in veins composed mainly of chalcedonic quartz and a carbonate mineral and contain sylvanite(?) and sparsely disseminated pyrite.

### QUARTZ AND CRYPTOCRYSTALLINE SILICA

Quartz is the most abundant gangue mineral, and it occurs in a variety of forms and textures. In many veins a gray massive granular quartz containing disseminated pyrite cements broken wallrocks;

the quartz itself shows signs of repeated breaking and healing. In simple symmetrical veins, gray quartz crystals line the vein walls and point toward the vein centers, which, in turn, are filled with metallic and carbonate minerals. In places white quartz crystals coat the earlier formed metallic minerals and project into cavities.

Tan, dark-gray, and black cryptocrystalline quartz is abundant in some ores, especially in those veins that show evidence of repeated opening; it commonly cements brecciated metallic minerals, and some is mammillary in habit.

### URANOPHANE

Uranophane was found in the Star adit in the wallrock several feet from the vein, where it forms small sunbursts along the fractures in granite gneiss.

### URANIUM PHOSPHATE MINERALS

Autunite, to bernite, and metatorbernite occur in the Jennie Lind No. 1 and Shafter adits. These minerals are distributed along joints in the wallrocks and are disseminated in altered biotite gneiss. Metatorbernite and meta-autunite are associated with uranophane and wad as coatings on fractures in granite gneiss in the Star adit.

### PRIMARY CARBONATE MINERALS

Primary carbonate gangue minerals are widely distributed but are abundant only locally. Ankerite is probably the most abundant carbonate gangue mineral, followed by siderite and rhodochrosite; some calcite and dolomite may also be present. Rhodochrosite was observed mainly in lead-zinc veins in the northeast corner of the district. The siderite is white to brown. In the Mayflower adit the white siderite crystals have overgrowths of brown siderite, and the contacts between the two are gradational. In places carbonate minerals vein or embay pyrite but are cut or embayed by sphalerite; in other places they embay or vein sphalerite but are cut or embayed by chalcopyrite, and locally, carbonate minerals vein all the metallic minerals.

The carbonate minerals were identified by indices of refraction.

### SECONDARY CARBONATE MINERALS

The secondary carbonate gangue and ore minerals include nearly pure siderite (refractive index $n_0 = 1.87$), cerussite, smithsonite, azurite, and malachite. Calcite was not identified, but it is probably present locally. Cerussite mixed with covellite commonly replaces galena. Cerussite, intimately mixed with covellite, malachite, and azurite, preferentially replaces galena in some specimens of galena veined by tennantite. The tennantite veinlets were left as septa in the carbonate-covellite mixture. All the secondary carbonate minerals replace and vein the sulfides where the veins are partly oxidized. Carbonate minerals formed during and after mining locally coat the mine walls.

#### BAYLEYITE

Bayleyite, a rare hydrous magnesium-uranium carbonate, was found in the New Bedford adit. It encrusts exposures of biotite-rich pegmatite on the mine walls and does not occur in the New Bedford vein.

#### BARITE

Barite is a sparse gangue mineral and occurs mostly in veins of the Gilson Gulch area. In the Bald Eagle mine tabular barite crystals line vugs in galena.

#### ANGLESITE

Anglesite is rare in the district. It was found at only two localities; at both of these it is associated with secondary siderite.

#### HYDROUS URANIUM SULFATE MINERALS

Schroekingerite, betazippeite, and zippeite—all identified by X-ray powder patterns—were found in small quantities. Schroekingerite was found in the No. 2 level of the Alma Lincoln mine in the form of rosettes that are scattered on the mine wall near the vein, which is radioactive at this locality. Betazippeite occurs at the same locality as small aggregates of crystals along the vein walls. Zippeite was found in the walls of No. 3 level of the Alma Lincoln mine along slickensided surfaces in the gouge that follows the vein walls and locally crosses the veins. In the Diamond Joe adit crystals of zippeite locally coat the back and walls of the drift.

#### COPPER SULFATE MINERALS

Copper sulfate minerals, probably mainly chalcanthite ($CuSO_4 \cdot 5H_2O$), commonly coat the walls of the long-abandoned mines.

#### EPSOMITE(?)

Efflorescences of a delicately fibrous, colorless mineral—probably epsomite—locally coat the walls of many long-abandoned mines.

### CLASSIFICATION OF THE VEINS

Four classes of gold-silver ores—pyritic, galena-sphalerite, composite, and telluride—were identified in the Idaho Springs–Central City region by Bastin and Hill (1917, p. 105–120) and by Lovering and Goddard (1950, p. 173). According to this classification, pyritic ores consist predominantly of pyrite and gangue minerals and contain subordinate amounts of chalcopyrite, tennantite, and other metallic minerals. The galena-sphalerite ores consist predominantly of galena, sphalerite, gangue minerals, and pyrite, and contain subordinate amounts of chalcopyrite and other metallic minerals. The composite ores contain mixtures of pyritic and galena-sphalerite ores; deposition of the pyritic ores in such mixtures was separated from that of the galena-sphalerite ores by a period of fracturing. In the telluride ores much of the gold is contained in silver-gold telluride.

Though most of the veins of the Idaho Springs district fit the classification made by Bastin and Hill (1917, p. 105), we found a modification of that classification which seems more convenient and useful. In many veins in the district there is evidence of composite mineralization, but in some veins the later deposited materials were copper rich, and in others they were lead and zinc rich. This variation provides a basis for further meaningful subdivision. Gold-silver telluride minerals are present in some veins, but they probably do not represent a major class of veins.

In this report we classed the veins into four principal types according to their mineral content: pyrite veins, pyritic copper veins, pyritic lead-zinc veins, and lead-zinc veins. Pyrite veins and lead-zinc veins are mineralogically distinct, but they grade into pyritic copper and pyritic lead-zinc veins through gradual changes of the relative proportions of the definitive minerals. Some veins change from one type to another along strike. The classification of the major vein types, based on the approximate relative abundance of the dominant metallic minerals, is shown graphically in figure 5. Gangue minerals are abundant in all veins, and other ore minerals are present in some or all veins but they do not affect the classifications. The vein distribution of the various classes is shown on plate 1 and in figure 6.

Individual veins are classified on the basis of observations made in mines, inspection of ore materials in the dumps of inaccessible mines, production data, assay data, and previously reported descriptions of inaccessible mines. These data are not uniformly reliable, and, for lack of adequate data, many veins are left unclassified on plate 1.

### PYRITE VEINS

Pyrite veins contain about equal amounts of pyrite and quartz, and some may contain trace amounts of galena, sphalerite, chalcopyrite, and tennantite. Gold values are probably attributable to native gold, and the typically low silver values are probably attributable to admixture of silver with gold and to silver in the sparse base-metal minerals. Typical pyrite veins are characterized by a wide zone of altered and pyritized wallrock that is traversed by many thin anastomosing veinlets of quartz and pyrite. The walls are ill

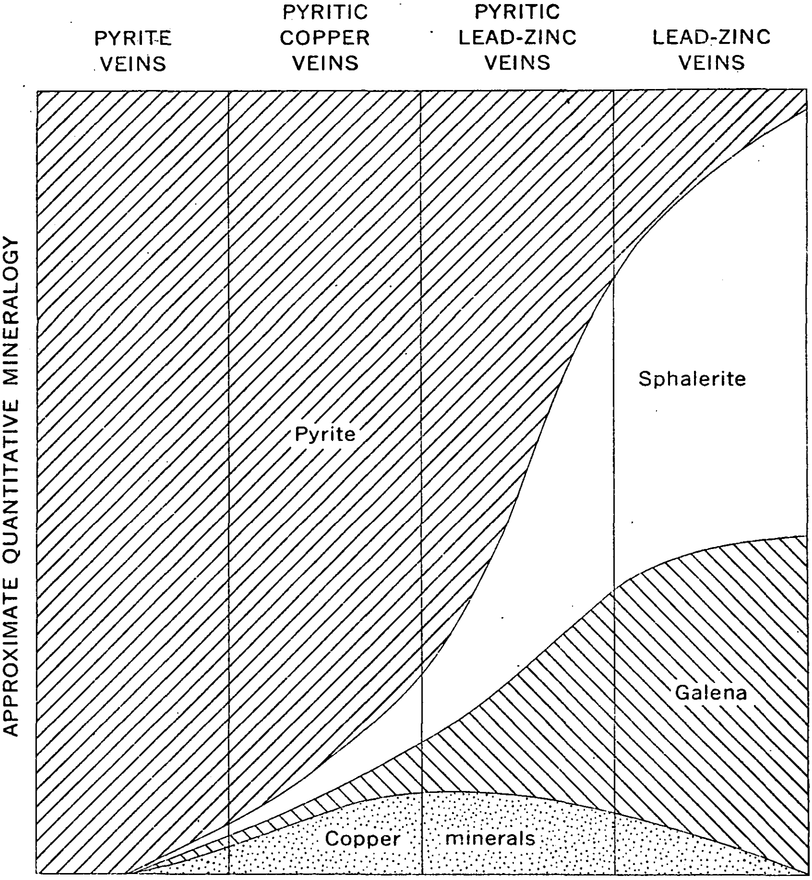


**Figure 5.—Vein classification based on quantitative mineralogy.**

defined and grade into unaltered rocks on both sides. The altered wallrocks are hard and silicified near the center of the zone, and they contain abundantly disseminated fine-grained pyrite. The pyrite in the veinlets is coarser grained and is either massive or disseminated in quartz; large cubes and pyritohedrons are common. Vugs are scarce. The pyrite veins are commonly traversed approximately parallel to their length by many anastomosing striated gouge-lined fractures that contain subhorizontal slickenside striae.

We saw pyrite veins other than those that were reopened and filled with other ore minerals; only five veins are classified as pyrite on plate 1 and in figure 6. These are: Vein A in the Lower Lake adit, the Idaho Springs fault, the Cornucopia vein in the southeast corner of the district, the Oro vein on the north end of the Lincoln vein, and the May Queen vein. Recorded production from pyrite

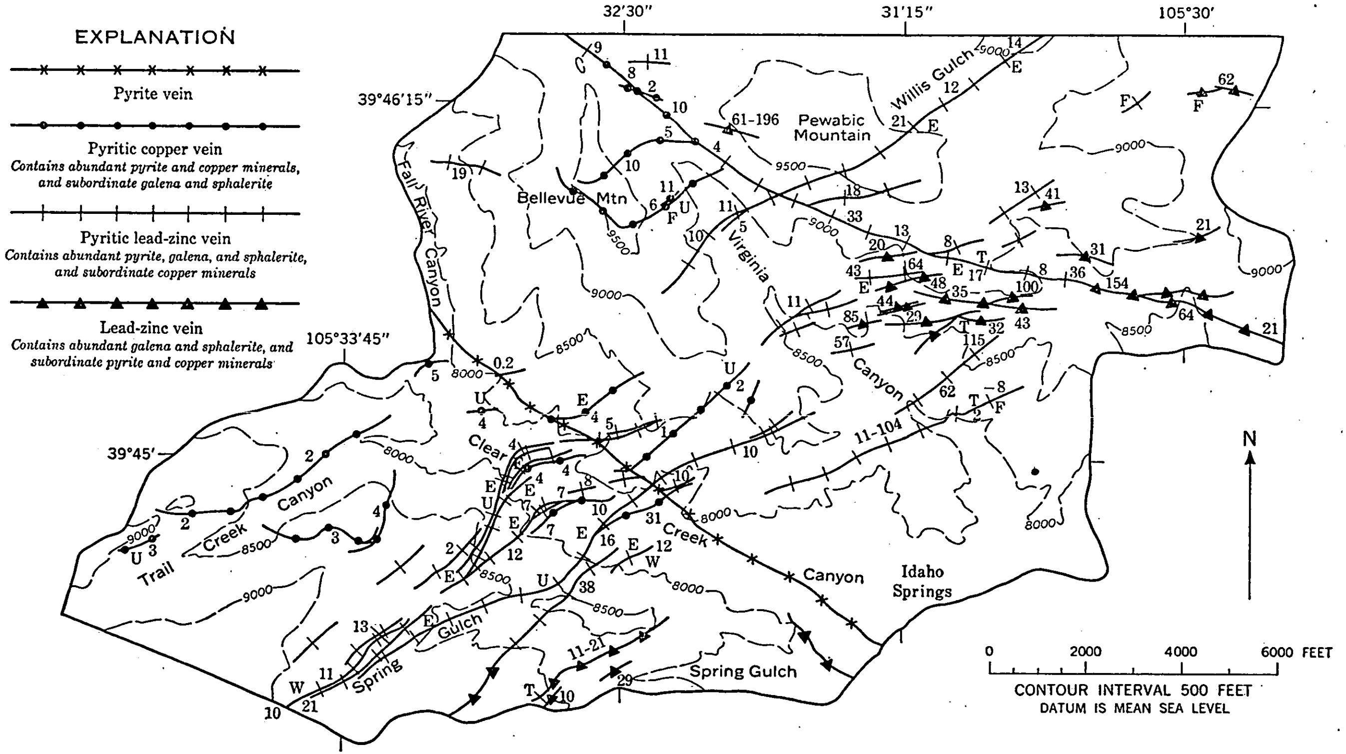


FIGURE 6.—Distribution of different classes of veins of the Idaho Springs district. Compiled by R. H. Moench. Numbers indicate ratio of silver to gold in vein, from production and assay data. E, enargite; T, gold tellurides; W, wolframite; U, uranium minerals; F, fluorite.

veins, primarily from the Waltham mine on the Idaho Springs fault, totals less than 4,000 tons. At current prices this ore would be valued at less than $25,000, 97 percent of which is attributable to gold and the remainder to silver.

The grade of pyrite veins, shown in table 4, is low, and assays in excess of an ounce of gold per ton are uncommon. Base metals are present only in small amounts, and it has not been economically feasible to separate them. Silver is seldom abundant, and silver-gold ratios, calculated from production records and assay data, range from much less than 1:1 to as much as 5:1 (fig. 6).

TABLE 4.—*Tenor of ore from some pyrite veins*

| Location | Type of data | Quantity | Gold | Silver | Copper | Lead | Zinc |
|---|---|---|---|---|---|---|---|
| | | | Ounces per ton | | Percent | | |
| Lower Lake adit (Vein A). | Assay [1] | 4-in. chip sample. | 0.92 | 4.52 | 0.82 | 0.24 | 0.04 |
| May Queen adit. | Production [2] | 5 tons concentrates. | .97 | .20 | ----- | ----- | ---- |
| Oro adit | do | 229 tons crude ore. | .30 | .03 | ----- | ----- | ---- |
| Waltham mine. | do | 3,356 tons concentrates, 43 tons crude ore. | .17 | .24 | ----- | ----- | ---- |

[1] Analyzed by H. H. Lipp, D. L. Skinner, and W. D. Goss of the U.S. Geol. Survey.
[2] Data calculated from production records, U.S. Bur. Mines; published by permission.

## PYRITIC COPPER VEINS

Pyritic copper veins contain abundant pyrite, smaller amounts of chalcopyrite and tennantite, subordinate sphalerite and galena, and local trace amounts of native gold, polybasite, pearceite, and enargite. Quartz is the dominant gangue mineral; ankerite and possibly other carbonate minerals are not uncommon. The walls of pyritic copper veins are somewhat more sharply defined than those of the pyrite veins, and copper minerals are clearly more abundant. Intergrowths of ore minerals cement brecciated quartz and pyrite locally and commonly form lenses and veinlets that cut through the quartz and pyrite. The early formed quartz and pyrite are typically fine grained and have been intensely sheared, whereas the later formed intergrowths of ore minerals are coarser grained, little deformed, and locally exhibit symmetrical banding and cockade structures. Coarse pyrite commonly coats breccia fragments and the walls of lenses and veinlets. The pyrite is, in turn, coated by the base-metal sulfides and sulfosalts.

Pyritic copper veins include some of the most productive veins in the district, such as the Specie Payment, Castleton, and Champion-Trio-Clarissa veins near the head of Virginia Canyon, the Shafter vein on the south slope of Bellevue Mountain, and the Donaldson–Champion Dirt, Phoenix, and Old Settler veins in the Trail Creek area. Several less productive pyritic copper veins are in the Clear Creek area. The Belman vein on the level of the Big Five tunnel is probably a pyritic copper vein also.

Pyritic copper veins are valued chiefly for their gold content, and significantly for their silver and copper content (table 5). According to production and assay records, gold commonly exceeds 1 ounce per ton and copper, 1 percent per ton in ore concentrates. Except as shown in some shipments recorded for individual mines, copper typically exceeds and locally equals lead in abundance. Lead is reported in few of the records, probably because it was not present in commercial amounts. Data on zinc are also incomplete, but, from our observations of the vein mineralogy, lead and zinc are probably about equally abundant in the pyritic copper veins.

TABLE 5.—*Tenor of ore from some pyritic copper veins*

| Location | Year(s) | Short tons [1] | Gold | Silver | Copper | Lead | Zinc |
|---|---|---|---|---|---|---|---|
| | | | Ounces per ton | | Percent | | |
| Castleton mine | 1909 [2] | 185 | 0. 70 | 6. 40 | 2. 80 | ------ | ---- |
| | 1936 [2] | 189 | . 26 | . 57 | . 21 | 0. 04 | 0. 17 |
| | 1939 [2] | 178 | . 44 | 2. 60 | 2. 20 | ------ | . 93 |
| Champion (Bellevue)-Trio mine | 1908–37 [2] | 1, 691 | . 49 | 4. 80 | . 65 | . 64 | . 09 |
| Specie Payment mine | 1902–50 [2] | 53, 238 | . 86 | 5. 10 | . 60 | . 27 | . 04 |
| Donaldson mine | 1922–35 [3] | 6 | . 98 | 5. 95 | 3. 05 | ------ | ---- |
| | 1922–35 [3] | 4 | 1. 66 | 8. 20 | 6. 00 | ------ | ---- |
| | 1922–35 [3] | 2 | . 77 | 8. 25 | 3. 80 | . 20 | . 30 |
| Gladstone mine | 1908–21 [2] | 465 | 1. 20 | 8. 30 | 1. 54 | . 17 | ---- |
| Hoosac mine (Rising Sun vein) | 1906–35 [2] | 314 | . 60 | 2. 10 | . 48 | . 39 | . 04 |
| Phoenix mine | 1927 [3] | 16 | 2. 58 | 14. 30 | 3. 50 | ------ | . 84 |
| | 1928 [3] | 44 | 2. 65 | 24. 20 | 7. 80 | ------ | 2. 60 |
| | 1929 [3] | 20 | 3. 05 | 18. 20 | 3. 79 | . 18 | 2. 76 |

[1] Includes crude ore and concentrates.
[2] Data calculated from production records, U.S. Bur. Mines; published by permission.
[3] Data obtained from records of Idaho Springs Sampling Works; published by permission.

The silver-gold ratios of pyritic copper veins generally range from about 2:1 to about 10:1. These ratios overlap but generally exceed the ratios for pyrite veins (fig. 6). These are the recovered-metal ratios calculated from production totals of individual mines but are generally consistent with assay data of the Idaho Springs Sampling Works and with the assays of the selected specimens we collected.

Exceptional ore shipments and assays may indicate higher or lower ratios; this depends largely upon the base-metal content. Silver-gold ratios generally increase as the base-metal content (particularly lead) increases.

The production total for the Crocket mine showed an exceptionally high silver-gold ratio (31:1) for a pyritic copper vein, and sampling works assay data for this mine are generally consistent with this ratio. The exceptionally high copper and silver content of this vein (Bastin and Hill, 1917, p. 359) suggests that these metals were enriched by secondary processes, or that the vein contains mixtures of primary ores rich in copper, lead, zinc, and silver. The mine is inaccessible, however, and available data are inadequate to favor either interpretation.

Assays of selected samples from the Phoenix and Donaldson mines (table 6) show that gold is associated with both the pyrite-rich and base-metal-rich parts of the veins but that silver is more abundant in the base-metal-rich parts. In addition, gold is associated with the sheared, early (older) pyrite as well as with the unsheared, later (younger) pyrite.

TABLE 6.—*Assays of selected ore samples from the Phoenix and Donaldson mines*

[Analyses made by W. D. Goss, D. L. Skinner, and James Wahlberg of the U.S. Geol. Survey]

| Lab. No. | Material sampled | Gold | Silver | Copper | Lead | Zinc |
|---|---|---|---|---|---|---|
| | | Ounces per ton | | Percent | | |
| | **Phoenix mine** | | | | | |
| D-100873 | Sheared quartz and pyrite | 1.36 | 1.24 | 0.06 | 0.07 | 0.03 |
| 874 | Unsheared quartz and pyrite | .46 | 1.64 | .14 | .11 | .11 |
| 872 | Dominant chalcopyrite and tennantite; sparse carbonate, quartz, and pyrite | 1.34 | 6.50 | 5.74 | .12 | .07 |
| | **Donaldson mine** | | | | | |
| D-217571 | Sheared quartz and pyrite | 0.26 | 0.54 | 0.03 | 0.38 | 0.32 |
| 570 | Unsheared, coarse quartz and pyrite | .80 | 3.92 | .01 | .33 | .23 |
| 569 | Unsheared, coarse quartz and pyrite | .34 | 1.50 | .23 | .39 | 2.02 |
| 567 | Dominant tennantite and chalcopyrite; subordinate galena and sphalerite | .44 | 6.58 | 12.50 | .89 | 1.91 |

## PYRITIC LEAD-ZINC VEINS

Pyritic lead-zinc veins contain abundant pyrite, galena, sphalerite, and smaller but significant amounts of chalcopyrite and tennantite. Small amounts of enargite have been seen in many veins, and trace amounts of wolframite, native gold, and pearceite have been seen in a few. Quartz is the dominant gangue mineral, carbonate minerals are locally abundant, and fluorite and barite are sparse. Ankerite is probably the most abundant carbonate gangue mineral; siderite, rhodochrosite, and dolomite are locally common. The veins are similar in appearance to pyritic copper veins except for different proportions of the ore minerals. The walls of some pyritic lead-zinc veins are gradational and are marked by a zone of broken, silicified, and pyritized wallrock; the base-metal minerals form veinlets and lenses that cut quartz and pyrite and locally cement brecciated quartz and pyrite. Structures of this type, of which the Bald Eagle vein is a good example, were formed by repeated movement and mineralization along the fracture zone. The walls of other pyritic lead-zinc veins are sharp, and pyrite, quartz, and the base-metal minerals are symmetrically banded. Such veins, of which the Lincoln is a prime example, formed during a single stage of mineralization and were uninterrupted by fracturing. Coarse cubic and pyritohedral pyrite typically coats the vein walls in bands as much as 6 inches thick, and the base-metal sulfides and sulfosalts fill the central part of the fissure. Even in the symmetrically banded veins, however, the base-metal minerals locally cement brecciated quartz and pyrite to form veinlets that traverse the coarse pyrite and locally follow the hanging wall or footwall of the main structure. The ores are generally solid; vugs are uncommon.

Pyritic lead-zinc veins are the most abundant and most productive of the four types. The principal mines on veins of this type are the Gem, Sun and Moon, Bald Eagle, Lake (Lower Lake and Lower East Lake adits), Frontenac, Aduddell and Druid mines in the northern part of the district, the Alma Lincoln, Stanley, and Edgar mines in the Clear Creek area, and the Fraction and Lord Byron mines in the Spring Gulch area. Many other mines on pyritic lead-zinc veins have produced significant quantities of ore.

Pyritic lead-zinc veins are valued chiefly for their gold content, and significantly for their silver, lead, and copper content (table 7). Gold is about as abundant as in pyritic copper veins, but lead, zinc, and silver are more abundant; copper is somewhat less abundant, though it may exceed lead content in some shipments. Available data on zinc indicate that it is equal to or exceeds lead in abundance in some ores, such as those from the Fraction and Lord Byron mines, and the Lower Lake adit, and is subordinate to lead in others, such as those from the Stanley and Alma Lincoln mines.

TABLE 7.—*Tenor of ore from some pyritic lead-zinc veins*

| Location | Year(s) | Short tons [1] | Gold | Silver | Copper | Lead | Zinc |
|---|---|---|---|---|---|---|---|
| | | | Ounces per ton | | Percent | | |
| Frontenac mine | 1913 [2] | 52, 716 | 0. 17 | 3. 29 | 0. 84 | 1. 42 | ------ |
| | 1919 | 901 | . 37 | 1. 34 | . 55 | 2. 69 | 1. 08 |
| | 1939 | 2, 967 | . 29 | 10. 17 | 1. 69 | 2. 71 | 1. 46 |
| Gem mine | 1904 [2] | 4, 700 | . 84 | 13. 72 | . 31 | 2. 47 | 1. 39 |
| | 1912 | 17, 872 | . 28 | 4. 39 | . 44 | 1. 77 | ------ |
| | 1922 | 1, 845 | . 42 | 15. 79 | . 04 | 15. 11 | 13. 82 |
| | 1939 | 19 | . 60 | 66. 50 | 1. 69 | 17. 50 | 11. 75 |
| Sun and Moon mine | 1904 [2] | 26, 040 | . 28 | 3. 71 | . 17 | . 70 | ------ |
| | 1906 | 3, 000 | . 30 | 4. 10 | . 80 | 4. 74 | 5. 00 |
| Lower Lake and Lower East Lake adits | 1911 [2] | 715 | . 33 | 1. 94 | . 19 | . 49 | . 36 |
| | 1916 | 2, 338 | . 89 | 7. 98 | . 94 | 1. 44 | ------ |
| | 1943 | 797 | . 07 | 4. 30 | . 30 | 2. 97 | 5. 71 |
| Bald Eagle mine | 1955–59 [2] | 29, 972 | . 38 | 3. 57 | . 44 | 3. 50 | ------ |
| Alma Lincoln mine Lincoln vein) | 1930 [2] | 1, 174 | . 88 | 3. 32 | 1. 04 | 3. 23 | . 55 |
| | 1934 | 2, 474 | 1. 59 | 5. 20 | . 81 | 2. 16 | . 61 |
| Stanley mine | 1910 [2] | 1, 853 | . 90 | 15. 21 | 2. 73 | 2. 26 | ------ |
| | 1913 | 312 | . 39 | 6. 33 | . 30 | 1. 45 | . 48 |
| | 1939 | 58 | . 82 | 10. 22 | . 33 | 4. 03 | 2. 93 |
| Edgar mine | 1925–34 [3] | 16 | . 90 | 26. 11 | 1. 70 | 13. 66 | 5. 42 |
| | | 16 | . 89 | 16. 90 | . 58 | 10. 73 | 2. 60 |
| | | 22 | 1. 72 | 19. 70 | 3. 50 | 2. 00 | 3. 06 |
| Lord Byron mine | 1937 [2] | 12, 703 | . 10 | . 65 | . 10 | 1. 35 | . 39 |
| | 1918 [3] | 21 | 1. 73 | 8. 38 | 2. 50 | 7. 80 | 7. 30 |
| Fraction mine | 1943 [2] | 54 | 1. 15 | 25. 39 | 7. 81 | 17. 00 | ------ |
| | 1920 [3] | 5 | 1. 16 | 18. 25 | 2. 17 | 13. 30 | 14. 00 |

[1] Includes crude ore and concentrates.
[2] Data calculated from production records, U.S. Bur. Mines; published by permission.
[3] Data obtained from records of Idaho Springs Sampling Works; published by permission.

Assays of selected samples from pyritic lead-zinc veins (table 8) showed that gold is distinctly more abundant in the base-metal-rich parts than in the pyrite-rich parts. However, the pyrite-rich part of the symmetrically banded Lincoln vein is richer in gold than is that part of the more complex Bald Eagle vein. Silver content is far greater in the base-metal-rich parts of both veins than in the pyrite-rich parts.

Average silver-gold ratios of pyritic lead-zinc veins, as indicated by production totals of the individual mines, range from less than 1:1 to more than 100:1, but most veins have ratios of 4:1 to as much as 20:1. Yearly production data (table 7) and assays of selected samples (table 8) indicate that similar extremes in the silver-gold ratio may be found within a single vein. For example, the production total for the Alma Lincoln mine indicates an average ratio of about 4:1, but yearly data indicate a range from about 1:1 to 34:1, and assays of selected samples of parts of the vein rich in pyrite and galena indicate a range from about 2:1 to 112:1. Similar rela-

TABLE 8.—*Assays of selected ore samples from the Alma Lincoln and Bald Eagle mines*

[Analyses made by D. L. Skinner, J. E. Wilson, H. H. Lipp, W. D. Goss, and J. S. Wahlberg of the U.S. Geol. Survey]

| Lab. No. | Material sampled | Gold | Silver | Copper | Lead | Zinc |
|---|---|---|---|---|---|---|
| | | Ounces per ton | | Percent | | |
| | **Lincoln vein, Alma Lincoln mine** | | | | | |
| 217535 | Pyrite | 0.02 | 0.46 | 0.02 | 0.26 | 0.12 |
| 536 | Galena and sparse sphalerite | .08 | 9.44 | .41 | 47.33 | 3.05 |
| 537 | Galena, chalcopyrite, tennantite | 4.60 | 23.76 | 3.72 | 30.92 | .16 |
| 538 | Pyrite and quartz | .40 | .78 | .04 | .29 | .08 |
| 542 | Pyrite and quartz, sparse galena and copper minerals | .44 | 4.00 | 1.38 | 1.90 | .20 |
| 543 | Galena, sparse sphalerite and copper minerals | .72 | 31.72 | 1.59 | 67.38 | 2.10 |
| | **Bald Eagle vein, Bald Eagle mine** | | | | | |
| 217575 | Pyrite and quartz | 0.02 | 0.24 | 0.02 | 0.10 | 0.03 |
| 576 | Galena, subordinate galena and copper minerals | .24 | 13.68 | 3.02 | 23.36 | 3.87 |
| [1] 274405 | Pyrite-bearing gouge (2 in.); hanging wall | .02 | .16 | .05 | .34 | .04 |
| [1] 404 | Pyritized breccia cemented by quartz and pyrite (8 in.) | .02 | .20 | .06 | .04 | .03 |
| [1] 403 | Pyritized breccia cemented by quartz and pyrite (5 in.) | .02 | .48 | .08 | .39 | .24 |
| [1] 402 | Coarse-grained galena, sphalerite, copper minerals, carbonate (2.5 in.) | .26 | 21.24 | 3.39 | 22.96 | 15.37 |
| [1] 401 | Fine-grained galena, sphalerite, copper minerals, quartz, carbonate (2.5 in.) | .66 | 12.76 | 1.66 | 9.98 | 7.38 |
| [1] 400 | Quartz and pyrite (7 in.); footwall | .06 | .74 | .30 | .38 | .40 |

[1] Samples taken from section across the vein.

tions can be seen in other veins, such as the Bald Eagle, that have been sampled on a similar basis. The silver-gold ratio, in general, increases with the base-metal content, particularly lead content.

Because the galena-rich or the gold telluride-rich parts of the veins may have been mined selectively, the silver-gold ratio indicated by the production data may not be representative of the vein. The low ratios indicated by production data for the Treasure Vault–Edgardine vein (2:1 and 0.8:1), for example, may be attributable to the selective mining of silver-poor gold telluride ore. On the other hand, the high ratios for the Brighton, Esmaralda, and Crystal veins (43:1–115:1) may be attributable to the selective mining of galena-rich ores. These veins are intermediate between pyritic lead-zinc and lead-zinc veins and have higher proportions of base-metal minerals to pyrite than do veins such as the Lincoln. For this reason the Brighton, Esmaralda, and Crystal veins probably have higher average silver-gold ratios than does the Lincoln vein, and selective mining of galena-rich parts of the former veins would accentuate this difference.

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### LEAD-ZINC VEINS

Galena and sphalerite are the predominant metallic minerals in lead-zinc veins. Pyrite is fairly abundant locally but is distinctly subordinate to galena and sphalerite, and chalcopyrite and tennantite are generally present only in small amounts. Quartz is the dominant gangue mineral, and ankerite, siderite, rhodochrosite, dolomite, barite, and fluorite are more abundant and widely distributed in the lead-zinc veins than in pyritic lead-zinc veins. The walls of lead-zinc veins are commonly sharp, and the wallrocks are typically less altered and less pyritized than are the rocks adjacent to other vein types. In places, lead-zinc veins are well-defined fissure fillings of more than a foot thick; however, they are commonly zones of thin subparallel interconnecting veinlets and lenses, and the sulfide and gangue minerals cement the brecciated wallrock locally. The ores are typically solid, but vugs are more common than in other vein types. Some lead-zinc veins are banded: quartz and carbonate gangue coat the vein walls, and galena, sphalerite, and other sulfides fill the center. Locally vugs are partly filled with clear terminated quartz crystals, crystals of rhodochrosite or other carbonate minerals, and cubes of pyrite.

Several major lead-zinc veins have been mined in the northeast corner of the district, and a few have been mined on the south side of the district. Good examples of the lead-zinc veins are the Seaton and Tropic veins near Seaton Mountain, the Franklin and Gold Medal veins on the east end of the J. L. Emerson–Gem vein system, and the Banta Hill and Bunkhouse veins in the extreme northeast corner of the district.

Lead-zinc veins are valued chiefly for their silver, gold, and lead content and, to a lesser extent, for their copper (table 9). The primary ores are richer in lead and silver than those of other vein types but poorer in gold and copper. Copper is distinctly subordinate to lead, though some shipments may show a content of more than 3 percent. Data on zinc are incomplete, but available assays and production records, combined with observations of vein mineralogy, indicate that zinc is about as abundant as lead.

Silver-gold ratios indicated by production totals for individual mines typically exceed 20:1 and locally exceed 100:1 (fig. 6). Expectably, individual shipments and assay data show a considerably greater range in the silver-gold ratio. The generally high ratios reflect both smaller amounts of gold and larger amounts of silver as compared with other vein types. Available records for the Torpedo mine on the south side of the district indicate an abnormally low silver-gold ratio (10:1) for a lead-zinc vein. Gold telluride minerals have been noted in this mine (Ball, 1908, p. 380), and the low ratio may be due to selective mining of gold-rich silver-poor telluride-bearing ore.

TABLE 9.—*Tenor of ore from some lead-zinc veins*

| Location | Year(s) | Short tons[1] | Gold | Silver | Copper | Lead | Zinc |
|---|---|---|---|---|---|---|---|
| | | | Ounces per ton | | Percent | | |
| Banta Hill mine | 1906[2] | 600 | 0.20 | 8.00 | 3.50 | 11.50 | -------- |
| | 1926[3] | 20 | .30 | 70.80 | -------- | 13.20 | 8.40 |
| De Lesseps mine | 1935[2] | 78 | .28 | 9.05 | 2.04 | 9.05 | 9.09 |
| Gold Medal adit and shaft | 1920[2] | 13 | .21 | 27.00 | .06 | 10.01 | 11.89 |
| | 1921 | 57 | 1.23 | 18.17 | .58 | 5.76 | 4.01 |
| Silver Age adit (Franklin vein) | 1910[2] | 1,453 | .11 | 21.00 | .10 | 2.13 | .13 |
| | 1920 | 1,080 | .13 | 17.90 | .01 | 14.77 | 3.49 |
| | 1950 | 430 | .96 | 35.65 | .79 | 20.41 | 18.13 |
| Forge Hill adit (Fairmont vein) | 1954[2] | 48 | .04 | 7.97 | .42 | 8.75 | 8.85 |
| | 1957 | 156 | .06 | 3.49 | .19 | 3.12 | -------- |
| Foxhall tunnel: | | | | | | | |
| Seaton vein | 1902[3] | 6,000 | 1.50 | 50.00 | 3.00 | 15.00 | -------- |
| | 1913 | 815 | .13 | 6.29 | .19 | 1.98 | 1.71 |
| | 1920 | 189 | .38 | 15.32 | .29 | 9.35 | 12.66 |
| Casino and Total Eclipse veins | 1902 | 403 | .75 | 25.95 | .26 | 6.78 | -------- |
| | 1905 | 200 | .35 | 30.50 | 1.00 | 10.00 | 15.00 |
| Tropic tunnel (Tropic vein) | 1902–35[3] | 648 | .50 | 53.00 | .33 | 6.50 | 5.00 |
| Bullion King mine | 1903[3] | 16 | .18 | 18.80 | 2.80 | 21.50 | 12.00 |
| M and E adit | Unknown[2] | 15.5 | .47 | 13.75 | .14 | 2.30 | 1.58 |
| Skyrocket mine | 1929–34[2] | 141 | .40 | 4.54 | .02 | 4.26 | .88 |

[1] Includes crude ore and concentrates.
[2] Data calculated from production records, U.S. Bur. Mines; published by permission.
[3] Data obtained from records of Idaho Springs Sampling Works; published by permission.

## LOCAL VARIANCE IN THE VEIN ORES

In addition to the common suite of ore and gangue minerals, some veins in the district contain one or more of the following: Enargite, wolframite (or huebnerite), fluorite, gold telluride minerals, and uranium minerals. The known occurrences of these minerals are shown in figure 6 by letters adjacent to the veins that contain them. Because these minerals are usually scarce at each locality, they are easily overlooked and their distributions may be wider than shown in figure 6.

Small amounts of enargite have been found in pyritic copper and pyritic lead-zinc veins. Available data suggest that enargite is confined to a zone slightly less than 4,000 feet wide that extends from the Lord Byron mine in Spring Gulch northeastward across the district to the Druid mine in Willis Gulch (fig. 6). The enargite is associated with the base-metal sulfides and sulfosalts.

Wolframite (or huebnerite) was observed in the Mayflower and Fraction pyritic lead-zinc veins. In both occurrences it is intergrown with early formed quartz near the margins of the veins.

Locally, fluorite is an abundant gangue mineral in the sulfide veins; however, some of the fluorite may be unrelated to the main sulfide mineralization. In the Bunkhouse vein (in the Banta Hill mine) and probably in the Moose vein, fluorite is an abundant gangue mineral and, though its textural relations to other minerals are not clearly known, it probably formed during the main mineralization. In the Treasure Vault mine fluorite is associated with gold telluride-bearing ores, and in the Specie Payment mine it occurs along foliation and joint planes adjacent to the vein. At both localities fluorite may be unrelated to the main sulfide mineralization.

Of the four known local occurrences of gold telluride minerals, three—the Gem, Casino, and Treasure Vault mines—are on and a short distance south of Seaton Mountain; one—the Torpedo mine—is in Spring Gulch. We saw no gold telluride ore, but Lovering and Goddard (1950, p. 174) noted that the major occurrences in the Idaho Springs and Central City districts lie in a narrow north-northeast-trending zone from the Jewelry Shop vein in the Chicago Creek area to the south end of the Dory Hill fault in the Central City district. Most of these occurrences are on or near the south-southwestward projection of the conspicuous fault; however, the Torpedo mine is more than a mile to the west. The gold telluride ores were formed later than the principal sulfide ores, as interpreted by Lovering and Goddard (1950, p. 174, 187); Bastin and Hill (1917, p. 114); and Sims, Drake, and Tooker (1963).

Small amounts of tetravalent and hexavalent uranium minerals are present in many veins of the district or in their wallrocks. Minerals of tetravalent uranium (coffinite and (or) uranite) are known to occur in only three or four mines, but minerals of hexavalent uranium may have formed at other localities—at the expense of coffinite and uranite—under near-surface conditions by oxidation and redistribution. Many occurrences of uranium minerals have been described by Sims and others (1963, p. 103–110). Only two pitchblende occurrences and one coffinite occurrence are described here.

At the Stanley mine coffinite and subordinate amounts of uraninite (pitchblende) are finely interbanded and form finely botryoidal coatings along fractures that cut the dominant sulfide and gangue minerals of the Stanley vein. Small amounts of pyrite, colorless sphalerite, quartz, and a carbonate mineral are interbanded with the coffinite and uraninite. Clearly, these minerals formed later than the main components of the Stanley vein.

At the Sunnyside mine small fragments of hard pitchblende are embedded in quartz and pyrite in a vein that probably branches from the Sunnyside vein. In addition, sooty pitchblende coats slickensided fracture surfaces and may have formed by partial oxidation and redistribution of the earlier formed uraninite. The habit of the hard pitchblende and its textural relations to quartz and pyrite are much like the occurrences in the Central City district described by Sims (1956; Sims and others, 1963). The uraninite of these occurrences probably formed before the main stage of quartz-pyrite mineralization and before the introduction of most of the base and precious elements.

## PARAGENETIC SEQUENCE OF PRIMARY VEIN MINERALS

The possibility that many veins formed during two stages of mineralization—pyritic succeeeded by base-metal veins—separated by a stage of fracturing was recognized by Collins (1913, p. 224–225) and by Bastin and Hill (1917, p. 112–114). Recent studies in mining districts to the south by Harrison (1955, p. 317) and Harrison and Wells (1956, p. 76; 1959, p. 42), and to the north by Sims (1956, p. 746; and others, 1963) have generally confirmed this idea. Three distinct stages of mineralization are recognized. They are, from oldest to youngest, the uranium, the pyrite, and the base metal. The stages were separated by fault movements which probably were nearly synchronous throughout the district. In the Idaho Springs district the early uranium stage is positively recognized only in the Sunnyside mine and possibly in the Red Jacket adit and is not considered significant. Otherwise, the relations described here are consistent with those observed in adjacent districts by other workers.

A distinctive suite of metallic minerals was deposited during the pyrite and base-metal stages of mineralization. The approximate abundance and order of deposition of the ore and gangue minerals that characterize each stage, as represented in three veins, are illustrated in figure 7. Pyrite, as well as a trace amount of marcasite in one locality, was deposited during the first stage; sphalerite, chalcopyrite, tennantite, galena, small amounts of pyrite and, in places, trace amounts of enargite and other metallic minerals were all deposited during the second stage. Quartz is the only gangue mineral of the first stage; it also was deposited during the second stage. As well as can be determined, carbonate minerals and sporadic barite and fluorite were deposited only during the second stage. Gold and silver were deposited during both stages of mineralization, but most of the precious metal was deposited with minerals of the base-metal stage.

The effects of small but repeated movements during mineralization are recognizable in most veins. The most pronounced movement apparently followed the pyrite stage. This movement probably was nearly synchronous throughout the district, for in all veins that contain products of both stages, early formed quartz and pyrite are

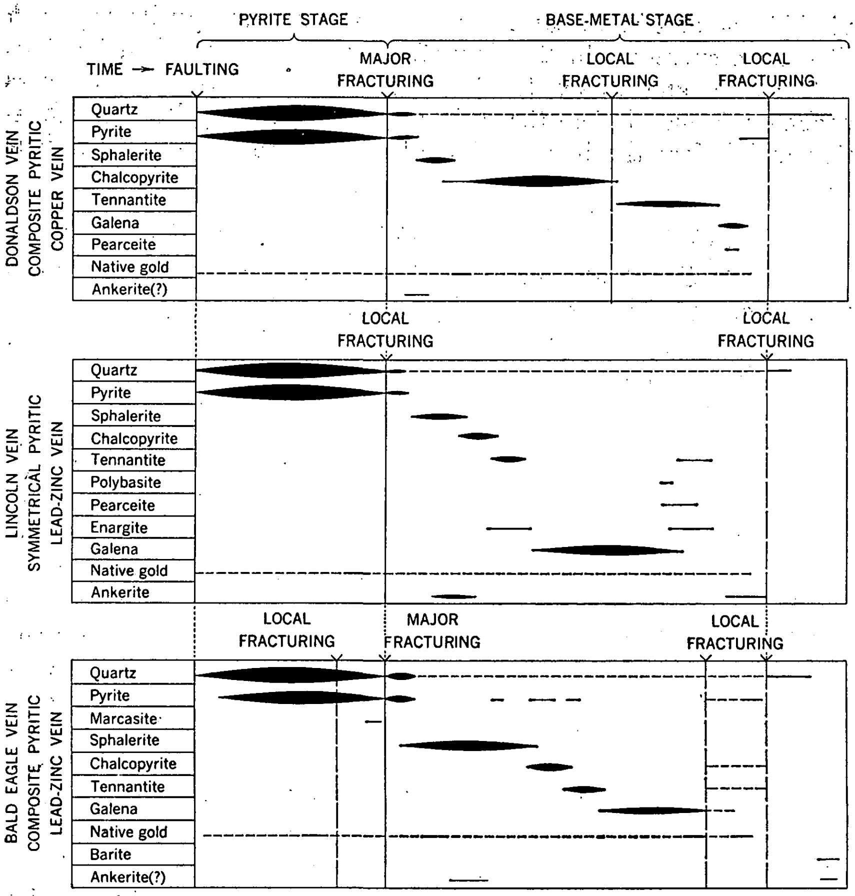


**Figure 7.—Paragenetic sequence of the primary ore and gangue minerals.**

brecciated and cemented by base-metal sulfides and sulfosalts. The symmetrically banded veins, such as the Lincoln, were only locally brecciated, and the movement along them must have been small. In most pyritic copper and pyritic lead-zinc veins, however, the fracturing and brecciation effects at this stage are characteristic. Movement may also have occurred on many pyrite and lead-zinc veins at this time, but if so, the pyrite veins did not receive later base-metal ores, and lead-zinc veins did not receive much of the earlier pyrite.

Fracturing appears to have taken place also near or after the end of the second stage of mineralization, as shown in figure 7. Most veins have been sheared to some extent along their hanging walls or footwalls or along fractures that cut across the ores. Gouge is common and it contains finely ground base-metal minerals as well as pyrite. In places, undeformed quartz, locally of a cryptocrystalline variety, fills vugs in the gouge or fractures that crosscut the base-metal minerals.

With minor changes, the sequences illustrated in figure 7 apply to other pyritic copper and pyritic lead-zinc veins; parts of the diagrams also apply to pyrite veins and lead-zinc veins. For example, the pyrite stage shown in all three diagrams applies to pyrite veins, and the base-metal stage of the Lincoln and Bald Eagle veins applies, with some modifications, to lead-zinc veins. We were able to determine the sequences of deposition by observing ore textures in polished surfaces and comparing them with the polished surfaces described by Bastin (1950).

### PYRITE STAGE

The pyrite stage of mineralization is evident, except in a few lead-zinc veins, in all the veins of the district. During this stage pyrite and quartz were deposited in fissures, and they also replaced altered wallrocks to some extent. Gold was deposited extensively; trace amounts of marcasite were deposited in the Bald Eagle and Mayflower veins, and small amounts of wolframite were deposited in the Mayflower and Fraction veins.

Pyrite and quartz are closely associated and probably formed about contemporaneously, but they do not show consistent paragenetic relations. Both pyrite and quartz are fine to coarse grained and in places have been repeatedly broken and healed by additional quartz and pyrite. In places pyrite is in large cubes and pyritohedrons, but more commonly it is in fine-grained granular masses. Quartz, most of which is gray or white, is, in part, in terminated crystals capped by pyrite; however, most of it is in fine-grained granular masses. Some quartz is interstitial to pyrite grains and some is cut by pyrite veinlets; thus, both earlier and later deposition of pyrite are indicated. Much of the pyrite disseminated through the altered wallrocks probably formed prior to the culmination of the pyrite stage of mineralization.

Marcasite occurs as small blebs in pyrite; therefore, these two minerals probably formed about contemporaneously.

The presence of gold and small amounts of silver in vein matter of the pyrite stage was indicated by assays of selected samples of the pyritic parts of composite veins (tables 6, 8) and by production records of pyrite veins. Very likely much of the silver is contained in the native gold, but data on the fineness of gold are not available. A few octahedral crystals of gold which were seen embedded in pyrite indicate that the gold formed earlier than or contemporaneously with the pyrite that contains it. The gold is more commonly in irregular veinlets or patches in pyrite.

Edwards (1954, p. 136) stated that wolframite generally forms early in any given paragenetic sequence, and the textural relations between wolframite, quartz, and pyrite in the district coincide with this generality. In both the Mayflower and Fraction veins, wolframite (or huebnerite) forms euhedral bladed crystals near the margins of symmetrically banded veins. The blades are randomly oriented and embedded in a matrix of quartz and pyrite, and the abundance of wolframite decreases inward from the margins of the veins. These relations indicate that the wolframite crystallized first, but that it accompanied or alternated with quartz and pyrite in decreasing amounts during the pyrite stage of mineralization.

#### BASE-METAL STAGE

The base-metal stage of mineralization, which followed a period of fracturing, was widespread and produced most of the ore deposits of the district. Galena, sphalerite, chalcopyrite, tennantite, pyrite, and smaller amounts of other metallic minerals were deposited almost entirely as fissure fillings. Only in the Druid mine in the northeast corner of the district is there evidence that base-metal-stage ores have replaced the wallrocks. At many places earlier formed pyrite veins were reopened, fractured, or brecciated, and then were cemented with minerals of the base-metal stage, which process produced composite ores. Gold and silver, particularly silver, were deposited in larger amounts during the base-metal stage than during the pyrite stage. The component minerals of the base-metal stage were deposited, with local variations, in the sequences shown in figure 7.

Minerals of the base-metal stage varied in relative abundance from place to place, which resulted in some veins being relatively rich in copper and some being relatively rich in lead and zinc.

Pyrite of the base-metal stage is clearly subordinate to the copper minerals, galena, and sphalerite. Most of the observed textures indicate that pyrite was the first mineral to precipitate during this stage (fig. 7). Euhedral crystals of pyrite are commonly embedded in sphalerite and other sulfides and sulfosalts; these minerals locally embay the pyrite crystals. Small amounts of pyrite formed later in the base-metal stage. In the Mayflower and Bald Eagle veins, for example, small pyrite veinlets cut sphalerite and, in turn, are veined and embayed by tennantite and galena. In the Bald Eagle vein large breccia fragments of galena and sphalerite are encrusted with pyrite, which, then, is coated with chalcopyrite and tennantite.

Quartz is commonly intergrown with pyrite and formed early in the base-metal stage. This early formed quartz is well crystallized and is commonly in terminated crystals that are coated with sphalerite and other base-metal sulfides and sulfosalts. Later formed quartz is common, however, particularly in veins that show evidence of repeated brecciation. In such veins fine-grained quartz cements fine breccias of base-metal minerals. Some of the late-formed quartz is dark gray and cryptocrystalline.

Sphalerite is one of the earliest and most abundant minerals of the base-metal stage. The sphalerite immediately succeeded a small amount of early pyrite and quartz but preceded other base-metal minerals. Sphalerite commonly embays and locally veins pyrite and is typically veined and embayed by chalcopyrite, tennantite, galena, and other metallic minerals. At places, a small amount of ankerite or other carbonate mineral preceded sphalerite (fig. 7, Donaldson vein). In some veins a small amount of colorless sphalerite fills vugs and fractures in other minerals and was one of the last minerals in the paragenetic sequence. The dominant early formed sphalerite ranges from red-brown iron-bearing sphalerite to amber and colorless sphalerite. Many single crystals are conspicuously zoned. The centers of the zoned crystals tend to be darker than the margins, but the zoning is commonly oscillatory and locally transects crystallographic planes. Some veins, particularly the lead-zinc veins, contain only light-colored sphalerite.

Chalcopyrite generally crystallized after sphalerite and appears to have formed during two generations. The first generation of chalcopyrite is more abundant and is represented by innumerable blebs in sphalerite and by chalcopyrite masses that embay sphalerite and are veined and embayed by tennantite and galena. The facts that the blebs of chalcopyrite in sphalerite are alined along crystallographic planes of the sphalerite and are most abundant near the margins of the sphalerite crystals suggest that this chalcopyrite formed by exsolution as the liquefied sphalerite cooled. The second generation of chalcopyrite is conspicuous only locally and is represented by embayments, veinlets, and vug fillings in galena and tennantite.

Tennantite formed largely after the chalcopyrite but may have begun to form before all the chalcopyrite crystallized. Tennantite commonly veins and embays chalcopyrite, but locally the two minerals are closely intergrown, which suggests contemporaneous crystallization. A small amount of tennantite that veins and embays galena represents a later generation that may be contemporaneous with the late chalcopyrite.

Galena crystallized after most of the sphalerite, chalcopyrite, and tennantite had formed. It fills vugs and open fractures in the earlier formed minerals and markedly embays these minerals. Galena and tennantite and galena and chalcopyrite are locally myrmekitically intergrown, a factor which suggests that, locally, these minerals crystallize together. There is no evidence of more than a single generation of galena.

Enargite is typically associated with chalcopyrite and tennantite and, like them, appears to have been deposited both before and after galena. Enargite of the earlier generation commonly is closely intergrown with tennantite, a factor which suggests that both minerals crystallized together. Veinlets of enargite and associated chalcopyrite of this generation cut sphalerite and pyrite and are veined by galena. Enargite of the younger generation is similarly intergrown with tennantite and closely associated with chalcopyrite. Veinlets of these minerals cut galena. Enargite is also myrmekitically intergrown with galena; contemporaneous crystallization is therefore suggested. One specimen from the Brighton mine showed a myrmekitic intergrowth of galena and chalcopyrite in which part of the galena had been selectively replaced by enargite to form an intergrowth of enargite and chalcopyrite with an inherited myrmekitic texture.

Pearceite and polybasite are associated with galena, and the observed textures indicate that these minerals crystallized during or shortly after the formation of galena. In the Donaldson vein, pearceite is intergrown with galena, and a few patches (probably of replacement origin) were also seen in pyrite. Pearceite in the Lincoln veins occurs as blebs, veins, and embayments in galena. One specimen was observed to have a blade of polybasite rimmed by enargite, which, in turn, was myrmekitically intergrown with galena. In one specimen from the Mayflower mine, polybasite embayed galena along grain boundaries.

Native gold is widely distributed and shows a variety of associations and habits, which suggest that it formed during many stages in the paragenetic sequence. The native gold occurs as euhedral crystals in pyrite and as blebs and veinlets in pyrite, chalcopyrite, tennantite, galena, sphalerite, quartz, and carbonate minerals.

Barite is sparse and does not show a consistent position in the paragenetic sequence. In the Bald Eagle vein tabular barite, whose crystals line vugs in gelena, probably was one of the last minerals to form; in other veins it formed earlier.

Ankerite and other carbonate minerals formed during many stages in the sequence. Most ankerite seems to have formed early—before, during, or shortly after the deposition of sphalerite—but some did form later.

## ZONAL DISTRIBUTION OF VEIN TYPES

A zonal distribution of the ores in the Central City district has been recognized by Sims (1956, p. 745; 1960) and by previous workers (Lovering and Goddard, 1950, p. 170, 173, pl. 9; Bastin and Hill, 1917, p. 115; Collins, 1904, p. 480). There, a roughly concentric zonal pattern is shown by the distribution of pyrite veins in a central core area that is surrounded by a zone of transitional veins (equivalent to pyritic copper and pyritic galena-sphalerite veins in the Idaho Springs district) and bordered at places by a peripheral zone of galena-sphalerite veins. The pattern is also defined by carbonate gangue that increases in amount radially from the core area, by barite that increases from the peripheral zone outward, and by average silver-gold ratios that increase radially. Details of this zonal pattern are given by Sims, Drake, and Tooker (1963) and by Sims and Barton (1962). The zonal pattern in the Idaho Springs district is less well defined and is not concentric. It appears to be an extension of only the southeast side of the pattern of the Central City district.

### AREAL ZONING

A zonal distribution of the ores in the Idaho Springs district is indicated by the pattern of vein types (figs. 6, 8). Pyritic copper veins are dominant in a zone on the west side of the district; this zone is bordered on the east by a belt of pyritic lead-zinc veins, and farther to the east and south by areas of mainly lead-zinc veins. Also, the J. L. Emerson–Gem vein system cuts across the areal zonal pattern, and its mineral composition changes in general accord with this pattern. The various zones overlap considerably (fig. 8).

Unlike the Central City district the Idaho Springs district has no recognizable central core of pyrite veins, but pyritic copper veins are abundant in a broad zone that extends from the Trail Creek area northeastward to the head of Virginia Canyon, approximately on the southwest projection of the central zone of the Central City district. Galena and sphalerite are subordinate to copper minerals in this zone and do not form large masses. Pyrite is clearly the dominant sulfide, and quartz is the dominant gangue mineral; carbonate minerals are sparse.

A broad irregular belt of pyritic lead-zinc veins extends from the head of Spring Gulch northeastward across Clear Creek and Virginia Canyons to Pewabic Mountain. A few veins of this type are exposed near the extreme northwest corner of the district. The areas of these veins are probably equivalent to most of the intermediate zone of the Central City district. In Clear Creek Canyon the belts

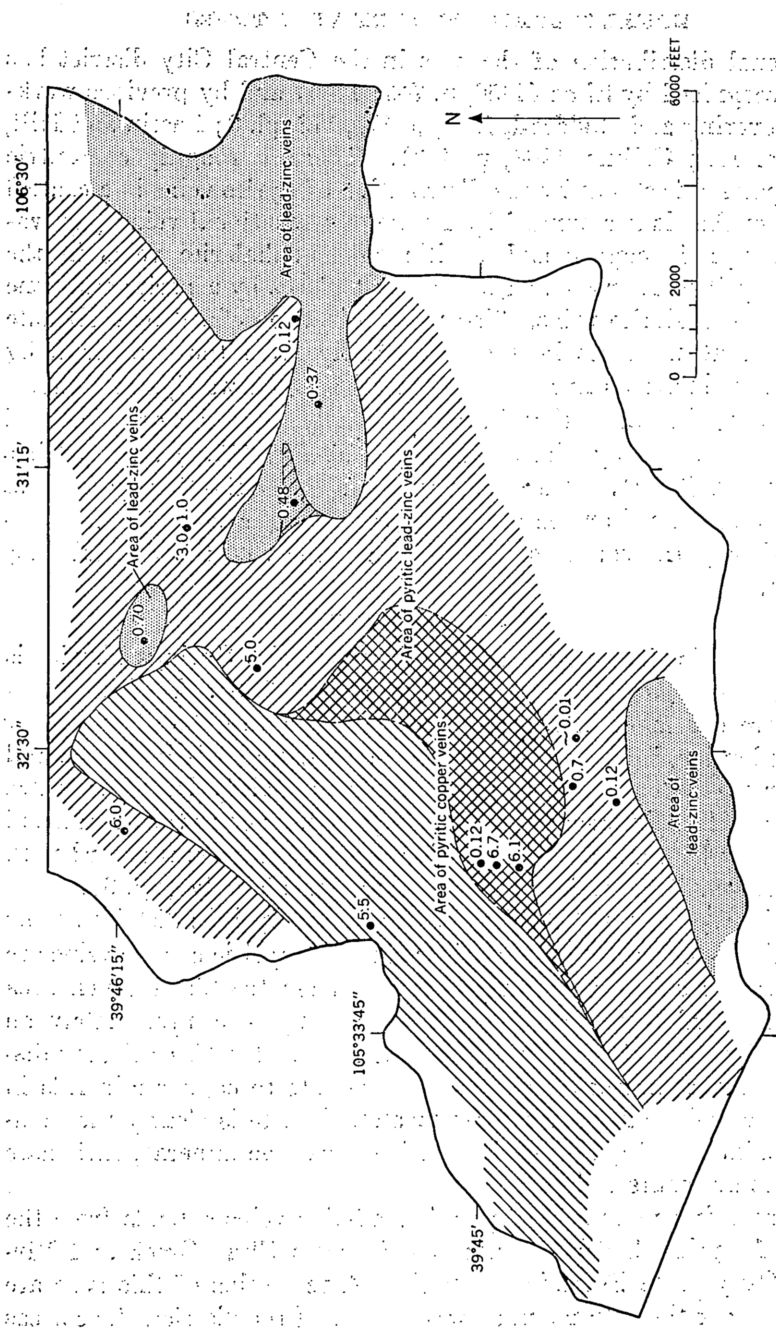


FIGURE 8.—Hypogene mineral zones, sample localities, and molecular percentage of FeS in sphalerite, Idaho Springs district.

of pyritic copper veins and pyritic lead-zinc veins overlap a width of half a mile, and the overlap of pyritic lead-zinc and lead-zinc veins is evident in the northern part of the district. With these exceptions the pyritic lead-zinc veins define the character of this zone. Pyrite is abundant but generally less so than in the belt of pyritic-copper veins; galena and sphalerite exceed copper minerals in quantity. Quartz is the dominant gangue mineral; carbonate minerals are locally abundant.

Lead-zinc veins are numerous in two areas to the east of the belt of pyritic lead-zinc veins—one in Spring Gulch, the other near the northeast corner of the district (figs. 6, 8). These areas of lead-zinc veins overlap areas of pyritic lead-zinc veins, and a single lead-zinc vein is present on the west side of Pewabic Mountain near the boundary of the zones of pyritic copper and pyritic lead-zinc veins. Pyrite is distinctly less abundant in the areas of lead-zinc veins than it is in zones to the west, and galena and sphalerite are the dominant ore minerals. Carbonate minerals as well as quartz are abundant in the gangue, and small amounts of barite—rarely present in the other zones—have been found.

The northwest-trending J. L. Emerson–Gem vein system seems to have influenced the character of the ores in the local areas. With the exception of the Comstock vein segment on the south slope of Pewabic Mountain, this vein system changes progressively southeastward from a pyritic copper vein at the head of Virginia Canyon to a pyritic lead-zinc vein and, then, to a lead-zinc vein on the east side of the district. On Seaton Mountain, where it is called the Gem vein, it is a pyritic lead-zinc vein as are most veins directly to the north. An area of lead-zinc veins lies directly to the south, however, and the zone boundary appears to be deflected along the northwest-trending vein system.

A pyrite vein occupies the northwest-trending Idaho Springs fault throughout the length of the fault in the Idaho Springs district. This pyrite vein crosses the zonal pattern of the district, but it does not appear to have markedly affected the mineral content of the nearby east- to northeast-trending veins. The Cornucopia vein in the southeast corner of the district is probably a split from the footwall of the Idaho Springs fault and is very similar in character and mineral content to the Waltham vein—a pryrite vein along the Idaho Springs fault.

### DEPTH ZONING

Evidence of depth zoning in the district is not conclusive. True hypogene depth zoning would reflect the changing character of the ore-bearing solution as it rose along a fracture or vertical changes in the pressure-temperature environment of the solution, whereas apparent depth zoning might reflect reopening of fractures after the pyrite stage and precipitation of ore minerals in definite shoots that are controlled by the vein structure. Data on the changes in veins with depth are scanty; therefore, hypogene depth zoning cannot be distinguished from apparent zoning related to ore-shoot structure. In addition, as Collins pointed out (1930, p. 252–267), the maximum depth of mining in any particular vein may be governed by cost increases as well as by geologic changes in depth. Lovering and Goddard (1950, p. 175) noted that the primary ores of the Central City–Idaho Springs area extend without noteworthy change from an altitude of about 7,000 feet to about 9,700 feet. Nowhere in the Idaho Springs district has mining extended below an altitude of about 7,200 feet, which is about 400 feet below the level of Clear Creek. This maximum depth, achieved in the Donna Juanita and Stanley mines directly below Clear Creek, was probably influenced more by water problems than by the character of the ore.

Meager evidence of hypogene depth zoning can be seen in the Belman vein, a part of the J. L. Emerson–Gem vein system. This vein, at the surface, is classed on plate 1 as a pyritic lead-zinc vein. This classification is uncertain, for the known production is small and may not have come from the near-surface workings. The classification coincides, however, with the records of the Blue Bell, a vein about 500 feet northwest of the Belman shaft and on part of the same vein system. Observations by Bastin and Hill (1917, p. 283) on the level of the Big Five tunnel, more than 1,400 feet below the surface workings, and by Moench on the Sayre level, about 1,200 feet below the surface workings, indicate that the vein at depth is copper rich. On the Sayre level the vein contains abundant pyrite, chalcopyrite, and tennantite, and very little galena and sphalerite. These relations suggest that galena and sphalerite decrease and copper minerals increase in abundance downward. The inferred mineralogic change is consistent with the areal zoning pattern and suggests that the copper-rich zone broadens downward.

### RELATION OF SILVER-GOLD RATIOS TO ZONAL PATTERN

As previously described, silver-gold ratios in ore of the Idaho Springs district increase progressively, with local exceptions, from pyrite veins through pyritic copper and pyritic lead-zinc veins to lead-zinc veins. Because these vein types are distributed in a zonal pattern, the silver-gold ratios show a similar zonal pattern, much like the pattern noted originally by Collins (1904, p. 480). Figure 6 shows a broad area of low silver-gold ratios (5:1 or less) between Bellevue Mountain and Trail Creek, and from the mouth of Fall

River southeast to the Elliot and Barber vein. Higher ratios (5:1–20:1) are distributed around this area, and the highest ratios (more than 20:1) are confined largely to the areas of lead-zinc veins.

The silver-gold ratio of a vein is largely controlled by the ratio of pyrite to base-metal minerals, particularly galena. Assay data given in tables 6 and 8 show that the amounts of silver and the silver-gold ratios are distinctly higher in the base-metal-rich parts of the zone than in the pyritic parts. With few exceptions the amount of silver present is closely linked with the amount of lead.

Silver appears to be contained largely in galena, but some occurs in rarely observed pearceite and polybasite, and some is probably in tennantite. A separate of galena from the Lower East Lake mine assayed 27.3 ounces of silver per ton, and galena from the Banta Hill mine assayed 81.07 ounces of silver per ton, whereas sphalerite separated from the same ores assayed only 1.73 and 2.12 ounces of silver per ton, respectively. Silver readily substitutes for other metals in tennantite and is assumed to be present in this mineral. Pearceite and polybasite are silver sulfosalts, and, though rarely observed, they may be widespread and may account for much silver. They were observed in specimens or polished surfaces of at least three high-silver ores.

These relations suggest that the zonal distribution of silver-gold ratios is controlled by the distribution of base-metal sulfides and sulfosalts, particularly galena, because the affinity of silver for galena appears to have been great. Pearceite or polybasite may have formed where the concentration of silver in the solutions exceeded the ability of available galena and tennantite to accommodate it. This possibility is strengthened by the fact that pearceite and polybasite apparently formed contemporaneously with and shortly after galena.

## ORE BODIES

Because veins are not uniform in thickness or in grade throughout their length, only parts of them have been worth mining. These mined parts or ore bodies constitute much less than half the known extent of most veins. Most ore bodies are tabular; they are much thinner normal to the vein walls than they are broad (strike length) or high (dip length). Some ore bodies are roughly equant in outline; others have a distinct long dimension. Ore bodies range in size from those having largest dimensions of several hundred feet to small bodies, more aptly termed "pods" or "lenses," having largest dimensions of only a few feet. A large ore body mined from the Lincoln vein, for example, has a plunge length of about 1,500 feet; it was stoped almost continuously from the lowest level to the vein apex about 600 feet above. The vein within the ore body or ore shoot ranges from 3 to 24 inches in thickness and averages about 10 inches in thickness. Outside the main shoot the vein is generally less than 6 inches thick. Most ore bodies have well-defined or sharply gradational walls; the valuable elements—gold, silver, copper, lead, and zinc—are concentrated in the vein minerals and not in gouge or altered wallrock.

Unique ore shoots have been mined in the Druid mine (Collins, 1913, p. 229). They are pipelike in form and have a strike length of 10–30 feet, a thickness of 1–10 feet, and extend to considerable depths. The principal ore shoot extended from the surface to the 300-foot level. The shoots do not occur on the principal veins in the mine, but on a vein that is thin or indefinable between the shoots. The shoots have irregular walls, extend several feet into the biotite gneiss, and apparently were formed by replacement of the wallrocks.

## LOCALIZATION

Because most veins in the district ore fissure fillings, most ore bodies formed in the relatively open parts of faults. Openings in the faults are related to deflections in dip and strike of the fault surfaces and to the amount and type of movement along them. Deflections commonly result from changes in the physical or structural character of the rocks cut by the faults. At places thick open breccias have formed at vein intersections and at vein splits. With the possible exception of the pipelike shoots at the Druid mine, there is no evidence for a chemical control of ore deposition related to different wallrocks.

Many of the most productive ore bodies are in veins that cut the most competent Precambrian rocks or that cut the gneissic structure of these rocks at a wide angle. The competent rocks—those that support openings along fissures—are the microcline gneiss, large pegmatite bodies, granite gneiss that is relatively free from layers of biotite gneiss, and some Tertiary intrusive rocks. The productive J. L. Emerson–Gem vein system, for example, crosses predominant microcline gneiss at a wide angle to the foliation. The Bald Eagle–Lake–Frontenac vein system, along steep fractures in relatively flat-lying rocks, was most productive at depth, where it cuts microcline gneiss, and not near the surface, where it cuts biotite gneiss. The Lincoln vein is subparallel to the gneissic structure of the wallrocks, but there the microcline gneiss supported a clean fissure far enough to be filled by ore and gangue minerals. The productive Edgar vein, however, is subparallel to the foliation of relatively incompetent biotite gneiss.

The primary factors that localized ores throughout the Front Range mineral belt have been summarized by Lovering (1942); these coincide with our findings, which were limited by the inaccessibility of many of the larger mines.

### ORE BODIES ALONG DEFLECTIONS IN STRIKE OR DIP OF VEINS

Deflection in strike or dip of veins is the most commonly observed factor of localization in the district. Almost all fractures have irregularities in strike and dip, and slight movements of opposing walls produce open spaces if the rocks that are cut are sufficiently competent. Thus, right-lateral movements along northeast-striking faults produce openings on the relatively east-striking parts, and left-lateral movements produce openings on the relatively north-striking parts. Similarly, normal movements produce openings on the steep parts of a fracture, and thrust movements produce openings on the flat parts. When such openings are filled with ore and gangue minerals, an ore body may be formed.

The main ore shoot in the northwest-striking Brighton vein is on an east-striking part, and several small ore bodies in the east-striking Fairmont vein (in the Forge Hill adit) are on the northeast-striking parts. Both veins are in left-lateral faults.

The localization of ore in the neartheast-striking Bald Eagle vein is related to two stages of movement. Right-lateral displacements, totaling about 80 feet, preceded, accompanied, and possibly followed the pyritic mineralization, and a small normal displacement accompanied the later base-metal mineralization. Accordingly, the major ore bodies of pyritic and base-metal ore are largely on strike deflections to the east, and small lenses of base-metal ore occur on the steep parts. In addition, cross fractures, filled with base-metal minerals above and below the flatter parts of the vein, are sufficiently numerous locally to permit mining of the wallrock. Such cross fractures probably formed as a result of the normal movement.

The largest stopes on the Edgar and Fraction veins, both northeast-striking veins that show right-lateral offset, are on the east-striking parts.

The largest stope on the Lower East Lake vein, which strikes northeast and shows right-lateral displacement, is on a strike deflection to the east. The deflection occurs where the vein passes from relatively incompetent biotite gneiss into more competent pegmatite. Thus, the change in rock type refracted the fissure, and the more competent pegmatite maintained the open space that was formed by faulting.

Ore bodies in the Stanley vein are not related to strike changes, but almost all stopes below the Whale level are on the relatively

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flat parts of the vein. Here the wallrocks were a major factor in the localization, for the dominant wallrock below the Whale level is relatively competent amphibolite and pegmatite, whereas above that level the rock is largely incompetent biotite gneiss. This change in rock type refracted the fracture much the same way as in the Lower East Lake mine, and the competent rocks supported the open spaces that were produced by faulting. Open spaces along the relatively flat parts of the fracture could have been produced by an updip component of movement.

An ore shoot in the Phoenix vein is on the intersection of the vein with a dike of Tertiary quartz bostonite porphyry. The vein steepens and thickens upon entering the dike, and two stages of movement (left lateral followed by normal) widened the fractures. Gold-bearing pyrite ore was deposited after the first stage of movement, and copper-rich gold-bearing ore was deposited after the second stage. Together, they form an ore body.

#### ORE BODIES AT VEIN INTERSECTIONS

Ore bodies have formed locally where two veins join or cross one another. The rock along such intersections may be more brecciated than elsewhere along the fractures, thereby providing added open space. Also, where two veins join, at least one is deflected in dip or strike along the intersection, and movements produce openings as described previously. Finally, an ore body may be formed by a junction of two thin veins that individually would not be minable.

Ore bodies at vein intersections are present in the Banta Hill, Sun and Moon, Stanley, and Donaldson mines. The principal ore body in the Banta Hill mine is a shoot that plunges steeply southwest at the intersection of the Banta Hill and Bunkhouse veins. In the Sun and Moon mine, a large ore shoot plunges northeast along the intersection of the Sun and Moon and the Moon–William Penn veins. In the Stanley mine, large ore bodies have been mined at the intersections of the Joker and Crocket veins with the Stanley vein (Spurr and Garrey, 1908, p. 348). In the Donaldson mine, two large ore shoots were formed by vein junctions (Spurr and Garrey, 1908, p. 338).

#### ORE BODIES RELATED TO ROCK COMPETENCY

Many fractures are deflected in strike and dip where they cross the contacts between relatively incompetent and competent rock types. Open spaces that result from fault movement may be larger in the more competent rock types. Examples of this fracture deflection in the Stanley and Lower East Lake mines have been described. The more competent rock types, though shattered, still maintain these openings.

At places ore bodies occur in the more competent rock types even though the vein is not appreciably deflected in strike and dip. Such ore bodies are attributable to the fact that the more competent rocks maintain permeability along the fracture, because they tend to form coarse breccia when they are sheared, whereas the less competent rocks may form microbreccia and gouge that tends to clog the openings. The larger stopes in the Camp Valley–Lord Byron mine, for example, occur where the vein cuts microcline gneiss and granite gneiss and not where it cuts biotite gneiss. The edge of the stope approximately follows the contact between the biotite gneiss and the microcline gneiss, but the vein is not obviously deflected at the contact.

#### ORE BODIES AT THE INTERSECTIONS OF VEINS AND FOLD AXES

Veins are subparallel to the foliation at places in the wallrock, and ore bodies occur where the vein crosses the foliation along or closely parallel to the axis of a fold. The fold may deflect the strike and dip of a vein, thus providing openings related to differential movements, or it may produce a zone of splitting or "horsetailing" that effectively widens the mineralized zone.

Definite examples of the aforementioned type of localization were seen at several places, but most are too small to be mined. Ore bodies in the Happy Easter, Mayflower, and Edgar mines are localized at the intersection of veins with folds. The Happy Easter vein, subparallel to the foliation of the enclosing biotite gneiss, is stoped where it intersects a fold axis that plunges northeast. The large stope near the portal of the Mayflower mine is about on the intersection of the vein with a fold axis that plunges northeast. The vein is 5–12 inches wide and fills a single fracture within the stoped area, but farther beyond it splits into several thin anastomosing veinlets. Ore in the Edgar vein probably is localized where it intersects a series of folds near the northeast end of the main level.

### WALLROCK ALTERATION

The wallrocks adjacent to the veins are altered to varying degrees and distances from the veins. The alteration envelopes typically contain an inner zone of hard, sericitized rock, and an outer zone of argillized rock that grades outward to fresh rock. The envelopes range in thickness from a few inches to several feet and are generally thicker in the western part of the district than in the eastern part. Typically, the envelopes are thicker than the enclosed veins, but there is no detailed correlation between the thickness of the veins and thickness of associated envelopes. Tooker (1956, 1963) has intensively studied the altered wallrocks in the Idaho Springs and adjacent mining districts, and the results of his study and of our observations are summarized below.

In most veins supergene alteration was superposed on hypogene alteration, and it is not possible everywhere to distinguish the products of the two. Limonite, covellite, and some other minerals are obvious products of supergene alteration in the veins, but some clay minerals in the wallrocks may have formed by both hypogene and supergene processes. In the Idaho Springs district the obvious products of supergene alteration in the veins are typically within 350 feet of the surface but have been recognized at a depth of 1,200 feet. Below 350 feet most veins and associated alteration envelopes probably owe their principal character to hypogene processes.

Four zones in the alteration envelopes were recognized by Tooker (1963, p. 8–9): (1) Fresh rock farthest from a vein, (2) weakly argillized rock, (3) strongly argillized rock, and (4) sericitized rock adjacent to a vein. From fresh rock toward zone 2, the rock gradually softens as clay minerals replace plagioclase feldspar and hornblende; in zone 3, clay minerals replace some of the biotite. Clay minerals of all major groups—montmorillonite, illite, kaolinite, and chlorite—are present in the two argillic zones. The boundary between the argillized rock and the inner, sericitized rock (zone 4) is typically sharp, and the sericitized rock is characteristically hard and bleached and commonly contains disseminated small grains of pyrite. Biotite is absent in zone 4, and the clay minerals have been converted to sericite. Quartz of the original rock has been crystallized in zone 4, but potassium feldspar is not visibly altered. The argillic or the sericitic zones are seldom absent.

Most of the pyrite in the wallrocks and possibly much that is actually in the veins probably formed by sulfidation of iron derived from the wallrocks. The mafic minerals (magnetite, hornblende, and biotite) in the wallrocks are altered and the iron released from them is more than adequate to account for the pyrite in the sericitic zone. Some of the vein quartz may also have been derived from the wallrocks, for the abundance of silica decreases markedly from fresh to argillized rock and is rarely completely regained in sericitization (Tooker, 1956, figs. 2, 3; 1963, p. 42–43). Several elements that ordinarily are present in only trace amounts in the fresh rocks are depleted in the altered rocks, and some may have been reprecipitated in the veins.

Most of the alteration apparently preceded the formation of the pyritic and base-metal ores. The envelopes adjacent to ore bodies are no wider than those adjacent to the narrow or barren parts of the veins, nor do the width ratios of the argillic and sericitic zones vary with the width of a vein. This suggests that parts of the fissures that were open during alteration were constricted or closed during the pyritic and base-metal mineralization. In addition, sulfide-bearing veins commonly show minor deflections in strike and dip that are not followed by the alteration envelopes. In the back of the Houston drift in the England mine, for example, a 3-inch symmetrically banded pyritic lead-zinc vein locally crosses sharply from the center almost over to the margin of the alteration envelope and back without a corresponding deflection of the envelope.

Tooker postulated (1956, p. 360; 1963, p. 96) that districtwide hydrothermal alteration by mixed meteoric and hypogene water preceded ore deposition. Solutions of the later pyrite stage may have intensified the earlier transformations, but solutions of the following base-metal stage had little effect on the wallrocks. Sims and others (1963) postulated that wallrock alteration culminated in the pyrite stage of mineralization.

## SUPERGENE ALTERATION

The veins of the Idaho Springs district have been altered by meteoric waters and by air to various depths below the surface. Above the natural water table (prior to interruption by mining activity), pyrite has been oxidized, in varying degree, to hydrous iron oxides, other sulfides have been partly removed, and gold locally has been enriched. Below the water table, silver and copper, leached from the oxidized zone, were reprecipitated as secondary argentite, proustite-pyrargyrite(?), chalcocite, and covellite and thus formed enriched sulfide ores. The character of enriched ores and the processes of enrichment were described by Bastin and Hill (1917, p. 137–152) and by Lovering and Goddard (1950, p. 88–90, 175–177).

The surface workings of the Fraction and Donaldson mines indicate that residual gold enrichment in the oxidized zone was probably widespread there. The Old Settler and Hukill mines (E–IV, 2, pl. 1) are among the oldest in the district and were worked initially for their residually enriched oxidized gold ore.

The depth of the oxidized zone is highly irregular. It is greater on ridges and mountaintops than in valley bottoms, and, except where the water table has been markedly disrupted by mining activity, it roughly conforms to the present water table. The maximum depth of the oxidized zone in the upper levels of the Kinda–U.P.R. mine, on the west side of Bellevue Mountain, is at least 100 feet below the surface; in other mines that are well above the valley bottoms it ranges from 150 to 200 feet; and in mines near Clear Creek depth of the zone is generally less than 20 feet below the surface.

The zone of secondary copper enrichment extends to considerably greater depths. In the Stanley mine secondary covellite, which embays other sulfides, was found in a specimen obtained from the Road level at a depth of 1,200 feet, directly below Spring Gulch. In the Mayflower mine covellite was found at a depth of 200 feet and in the Edgar mine, at 300 feet. In the lower three levels of the Alma Lincoln mine, where the zone of copper enrichment is fairly well defined, covellite and some chalcocite are widespread as deep as 350 feet below the surface. In the upper workings, which are entirely within the zone of copper enrichment, covellite mixed with cerussite and preferentially replaced galena. Chalcocite also embays galena. Covellite and chalcocite are sparse in all specimens, and enrichment of this type probably did not significantly change the ore value.

Secondary silver minerals—argentite, proustite-pyrargyrite, and native silver—have been found in near-surface workings of many mines in the northeastern part of the district (Bastin and Hill, 1917, p. 146), but we did not see any. They have not been found in other parts of the district, but records of early production from shallow ores suggest that there was some silver enrichment.

## GENESIS OF THE VEINS

The Idaho Springs district is an integral part of the Front Range mineral belt, which is characterized by temporally and spatially associated porphyries and ore deposits of Laramide age. In accord with most other workers in the Front Range, we consider the porphyries and vein deposits of the district to be genetically related. Direct evidence of a genetic relation between the ore deposits and a particular porphyritic rock type is lacking. Instead, the ores and porphyries are considered to have originated from a common source, as suggested by Lovering and Goddard (1950, p. 75–76). Although the available data suggest that at least some of the iron and quartz in the veins was extracted from the wallrocks during alteration, the other elements apparently were introduced from magmatic sources.

## SEQUENCE OF ORE FORMATION

The ores of the district are products of a sequence of events that included (successively) fracturing, wallrock alteration, pyritic mineralization, and base-metal mineralization. Mineralization occurred after the emplacement of all but the youngest intrusive rock—biotite-quartz latite. Districtwide fracturing may have lowered pressures above the source, presumably a magma chamber, thus allowing fluids to escape upward and outward from the magma. This fluid, which probably mixed with downward-moving meteoric waters, was considerably hotter than the enclosing wallrocks and was out of equilibrium with those rocks. As a result, the rocks were altered. According to Tooker (1956, 1963), the rocks were first argillized to various clay minerals, and the clay minerals, under intensified conditions, were converted to sericite, and quartz was recrystallized. As the alteration proceeded sericitization progressed outward into previously argillized rocks, and argillization progressed farther into fresh rock. Iron-bearing minerals in the walls were destroyed, and pyrite was formed in substantial abundance by sulfidation of iron released from these minerals. At the culmination of this stage, pyrite veins formed throughout most of the district. The veins then may have become at least partly clogged with quartz, pyrite, and clay minerals; as a result, mineralization largely ceased.

Recurrent districtwide tectonic movements reopened most pyrite veins and mineralization then resumed. At this stage base-metal ores were deposited. Composite copper-rich or lead-zinc-rich ores formed in the reopened pyrite veins, symmetrically banded ores formed in those pyrite veins that previously had not been completely clogged, and lead-zinc-rich ores formed in new fractures or in unfilled old fractures.

Although it is generally agreed that base-metal mineralization followed pyritic mineralization, the age relations between the copper-rich and lead-zinc-rich ores have not been established. In the Chicago Creek area, as postulated by Harrison and Wells (1959, p. 43), copper-rich ores were deposited before lead-zinc-rich ores, in accord with a sequence of fracturing. In the Central City district, however, as suggested by Sims and Barton (1961, 1962) and Sims and others (1963), there was only one stage of base-metal mineralization, which probably followed a period of districtwide refracturing, and the different kinds of base-metal ores owe their main characteristics to different pressure-temperature conditions. If the copper-rich and lead-zinc-rich ores formed in sequence, as postulated by Harrison and Wells (1959, p. 43), one would expect to find successive bands of such ores in the symmetrically banded veins, such as the Lincoln and Houston veins, which apparently remained open from the pyrite stage through the base-metal stage. Also, one would expect the copper-rich parts of some composite veins to be cut by lead-zinc-rich parts. Such relations were not observed, though they could easily have been missed because access to the mines was limited.

### HYPOGENE ZONING

The zonal pattern of the ores of the Idaho Springs district is part of a regional pattern and is an extension of the southeast side of the concentric pattern recognized in the Central City district (Sims and others, 1963; Sims and Barton, 1961; 1962). In contrast with the Central City district, a central zone of pyrite veins is not recognized in the Idaho Springs district; also, the intermediate zone of the Central City district is subdivided into areas of pyritic copper and pyritic lead-zinc veins. The area of lead-zinc veins on the east side of the Idaho Springs district is equivalent to the so-called peripheral zone of the Central City district to the north. Although a central pyritic zone is not recognized, the belt of pyritic copper veins is alined with the central pyritic zone of the Central City district. The zonal pattern near Idaho Springs is imperfect, however, because the boundaries are highly irregular and the different zones overlap one another (figs. 6, 8).

To explain the zonal pattern, we postulate, in general accord with Sims and coworkers, that altering and ore-bearing solutions emanated from a magmatic source in two distinct stages. First, after districtwide fracturing the juvenile solutions moved outward from the source, altering the wallrocks, and forming pyrite veins throughout most of the district; this process decreased in intensity with distance from the source. The veins, with few exceptions, became clogged with alteration products, quartz, and pyrite; mineralization generally ceased. Second, the solutions then changed in character at the source, and after renewed districtwide refracturing, they escaped and deposited minerals of the base-metal stage. Chalcopyrite and tennantite were abundantly deposited in parts of the district closest to the source; proportionately more galena, sphalerite, and carbonate gangue minerals were deposited farther from the source. The resulting zonal pattern may reflect decreasing pressures and temperatures away from the source. This postulate is based on the assumption that recurrent fracturings affected the entire area synchronously, and that most fractures had access to the same source; these assumptions cannot be tested by the data available.

In view of the many irregularities of the zonal pattern, this concept of its origin is oversimplified. Many of the irregularities, however, are attributable to the following described structural relations and do not necessarily conflict with the concept.

The persistent large northwest-trending J. L. Emerson–Gem fault and the Idaho Springs fault must have been principal channelways for the ore-forming fluids. High-intensity conditions apparently were attained along the Idaho Springs fault well away from the inferred source, for this fault is characterized throughout by a wide zone of altered and pyritized rock. High-intensity conditions probably were also attained along the Cornucopia vein on the southeast side of the district; the vein occupies a fault that is similar in character to the Idaho Springs fault and probably is a branch of that fault. The deflection of the zonal pattern adjacent to the J. L. Emerson—Gem vein system (figs. 6, 8), indicates that high-intensity conditions were attained as far southeast as the Gem and Freighters Friend veins but not in the lead-zinc veins directly to the south. Solutions of this stage may have spread northward from the fault and produced the pyritic parts of the veins there, but they did not strongly affect the veins to the south.

The fractures that now contain the lead-zinc veins in the Seaton Mountain and Spring Gulch areas may have escaped intense alteration and pyritic mineralization because of their remoteness from the inferred source or, possibly, because they did not exist at that time. The Fairmont vein, which is much closer to the inferred source, possibly did not form until some time after the stage of alteration and pyritic mineralization; accordingly, it was mineralized only with lead-zinc-stage minerals. The early solutions also apparently failed to gain access to the fracture occupied by the Comstock vein, a lead-zinc vein in the middle of the J. L. Emerson–Gem vein system.

The reason for the overlap of zones of pyritic copper veins and of pyritic lead-zinc veins is not yet understood. A possible explanation for the overlap is that the pyritic copper and pyritic lead-zinc veins formed successively and at a time when changing tectonic forces caused some fissures to open as others closed; however, evidence in support of this is lacking. It is also possible that pressure-temperature conditions differ from one vein to another, as postulated by Sims and Barton (1962) for certain veins in the Central City district. As the copper-rich ores are generally closer to the inferred source, they may have formed under higher pressure-temperature conditions.

### ENVIRONMENT OF ORE DEPOSITION

Geomorphic evidence and the character and distribution of sediments of Tertiary age indicate that the ores of the central part of the Front Range probably formed at depths ranging from 1 to 2 miles (Sims and Barton, 1962). These depths are only estimates, for the amount of cover removed since mineralization in Tertiary time cannot be measured directly. At an assumed depth of 1–2 miles, the rock temperatures probably would not have exceeded 150°C unless the geothermal gradient was extreme. In the Idaho Springs district the degree of alteration of the wallrocks suggests that the ore-forming fluids were considerably hotter than 150°C. Further, the rocks are altered most extensively and intensively in a broad zone along the west side of the Idaho Springs district. This indicates that the fluids may have been hotter or in larger supply there.

Sims and Barton (1961; 1962) inferred, on the basis of sphalerite geothermometry (Kullerud, 1953; Barton and Kullerud, 1958) and fluid-inclusion studies, that mineralization in the adjacent Central City district occurred at temperatures ranging from at least 620°C to about 150°C. The higher temperature estimate may be greatly excessive, however, for kaolinite, which cannot exist at much above 410°C (Carr and Fyfe, 1960), is a major constituent of the argillized wallrocks only a few inches away from the veins. A more recent study of the iron-zinc-sulfur system also indicates that 620°C is an excessive estimate (P. B. Barton, Jr., oral commun., 1964).

Sphalerite from the Idaho Springs district was sampled on the basis of a reconnaissance and was analyzed for its iron content to determine if the thermal pattern in the district was similar to the inferred pattern in the Central City district. However, in view of current (1964) uncertainties in the use of sphalerite as a geothermometer, these data are not interpreted in this report but are presented in the hope that they may prove meaningful when the iron-zinc-sulfur system is more clearly understood. The distribution of samples is shown in figure 8; the analytical results are shown in table 10. Sphalerite was handpicked from the ore specimens. It contained no visible pyrite. Most of the sphalerite contains blebby chalcopyrite, but because this chalcopyrite probably formed by exsolution, the iron that the chalcopyrite contains probably dissolved in the sphalerite when the chalcopyrite formed. Many of the

TABLE 10.—*Iron content of sphalerite*

[Analysts (Fe): W. D. Goss, Nos. 623-635, by colorimetric method; E. C. Mallory, unnumbered and No. 631, by volumetric method; P. R. Barnett, Nos. 611, 612, by quantitative spectrographic analysis]

| Location | Lab. No. | Fe (weight percent) | FeS (molecular percent) |
|---|---|---|---|
| Lawrence L mine (Boston vein) | 289623 | [1] 3.1 | 5.5 |
| Alma Lincoln mine: | | | |
| Elliot and Barber vein | 624 | [1] 3.6 | 6.1 |
| Lincoln vein | 625 | [1] .07 | .12 |
| South Lincoln vein | 626 | [1] 3.9 | 6.7 |
| Stanley mine | 627 | [1] .42 | .7 |
| Mayflower adit | 628 | [1] .007 | .01 |
| Star adit | 629 | [1] .07 | .12 |
| Lucania tunnel (vein 17) | 630 | [1] 3.5 | 6 |
| Two Brothers tunnel (Bald Eagle vein) | 632 | [1] 2.9 | 5 |
| Brighton mine | 633 | [1] .28 | .48 |
| Seaton mine | 634 | [1] .21 | .37 |
| Freighters Friend mine | 635 | [1] .07 | .12 |
| Banta Hill mine | -------- | [1] .26 | .45 |
| Forge Hill tunnel (Fairmont vein) | 246611 | .40 | .70 |
| Lower East Lake adit | 612 | .60 | 1 |
| Do | 289631 | [1] 1.54 | 3 |

[1] Reported as $Fe_2O_3$, recalculated to Fe.

sphalerite crystals are zoned. Marked color zoning was seen in the course of separation, particularly in the samples that contain the most iron. Samples were not taken from separate zones; instead, an attempt was made to extract the darkest material from the ore. Polished sections of ore samples from which the sphalerite specimens were taken indicated that the sphalerite formed in the sequence shown in figure 7.

In terms of molecular percent FeS, the iron content of the analyzed sphalerite samples from the Idaho Springs district ranges from almost 0 to 6.7. The samples that contained more than 3 molecular percent FeS were collected in or near the area of pyritic copper veins; those having lower FeS content were collected from the areas of pyritic lead-zinc veins and of lead-zinc veins. The iron content of sphalerite does not decrease systematically from the center of the vein outward, however, for high and low values were obtained on sphalerite from neighboring veins. The highest assay, for example, was obtained on sphalerite from the South Lincoln vein, whereas sphalerite from the neighboring Lincoln vein contains little iron (table 10).

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