Epithermal Gold Deposits—Part I

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Chapter H

Epithermal Gold Deposits—Part I

The Goldfield Gold District, Esmeralda and Nye Counties, Nevada

By ROGER P. ASHLEY

The Tonopah Precious-metal District, Esmeralda and Nye Counties, Nevada

By ROGER P. ASHLEY

The Republic Gold District, Ferry County, Washington

By MORTIMER H. STAATZ and ROBERT C. PEARSON

The Comstock Lode Precious-metal District, Washoe and Storey Counties, Nevada

By WILLIAM C. BAGBY and ROGER P. ASHLEY

U.S. GEOLOGICAL SURVEY BULLETIN 1857

GEOLOGY AND RESOURCES OF GOLD IN THE UNITED STATES

DANIEL R. SHAWE and ROGER P. ASHLEY, Scientific Editors
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Epithermal Gold Deposits—Part I

The Goldfield Gold District, Esmeralda and Nye Counties, Nevada

By Roger P. Ashley

Abstract

The Goldfield district is an epithermal precious metal deposit of the quartz-alunite type. Gold ore bodies are associated spatially and temporally with a calc-alkalic volcanic center of early Miocene age Flows, tuffs, and breccias of this center overlap a small caldera of Oligocene age Silicic domes of Oligocene age and porphyritic rhyodacite domes of early Miocene age both intrude the caldera ring-fracture zone. Most of the rocks in and around the caldera are hydrothermally altered, and gold ore bodies are clustered along the caldera margin, hosted both by Oligocene rocks cut by the ring-fracture zone and Miocene rocks that intrude or overlie the ring-fracture zone. The ore bodies formed at relatively shallow depths from meteoric water at about 200 to 300 °C.

The district has produced more than 4 2 million ounces (130 metric tons) of gold and 1 45 million ounces (45 metric tons) of silver, mostly before World War I, and has modest reserves and inferred resources of gold Minor intermittent production continues.

INTRODUCTION

The Goldfield mining district is located in the western part of the Basin and Range province, 250 km northwest of Las Vegas, Nev., and 300 km southeast of Reno, Nev. (fig. H1). It is 40 km south of Tonopah, site of another precious metal district described in a separate article.

Gold was discovered in the northwestern part of the Goldfield district in 1902. The first of the major lodes in the southwestern part of the district was discovered

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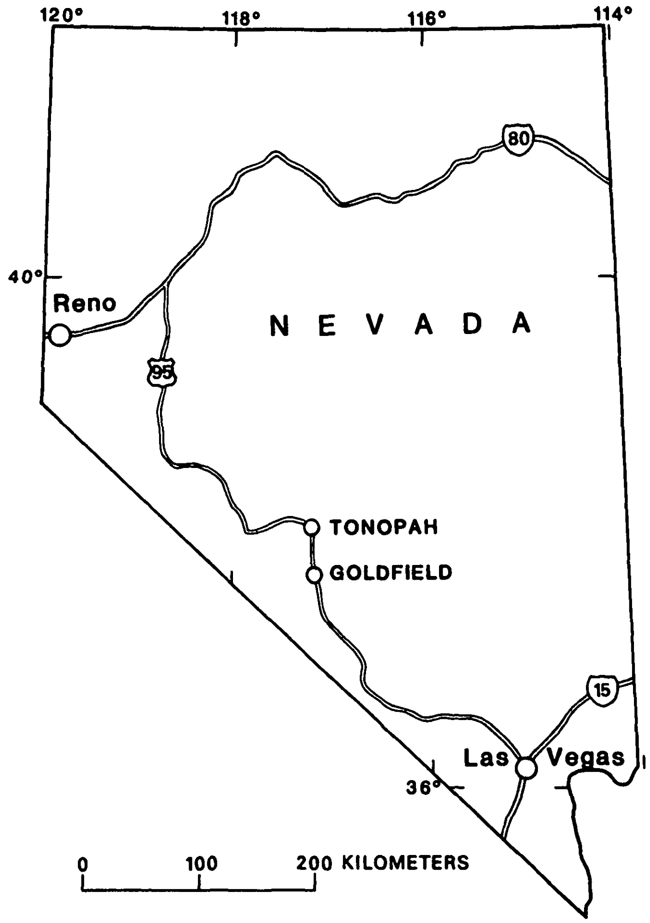

Figure H1 Locations of the Goldfield and Tonopah mining districts, Nevada.

about 6 months later, in June 1903. Production began late in 1903, and new mines were developed rapidly through 1904. The operations of all the major mines except the Florence mine were consolidated as the Goldfield Consolidated Mines Company in late 1906. Local mills were opened and expanded from 1905 to 1909, rail connections were completed in 1905 and 1907, and a 48-km water-supply pipeline was completed from Lida, Nev., in 1907. Peak annual production came in 1910, and production was small after 1919. Total recorded production is 4.19 million oz (130 metric tons) of gold, 1.45 million oz (45 metric tons) of silver, and 7 67 million lb (3,420 metric tons) of copper. The value of production, due almost entirely to gold at $20.67/oz, was about $90 million.

Early in the 1980's, a joint-venture group consisting of Pacific Gold and Uranium, Inc., Noranda Exploration, Inc., and Southern Pacific Land Company carried out a drilling program to test remaining near-surface parts of the main vein system. Blackhawk Mines Corporation, the designated operator, stated the proven reserves as 2,060,000 tons at an average grade of 0.07 oz Au/ton, including approximately 1,700,000 tons to be recovered from three open pits, and the remainder from waste dumps left from previous operations (Blackhawk Mines Corporation, written commun., November 1984). From 1985 through 1987 Blackhawk Mines and a subsequent operator, Dexter Mining, have intermittently operated a small open pit immediately northeast of Goldfield, and have heap-leached both primary ore and waste-dump material. From 1981 to 1984 Blackhawk intermittently operated a small open pit in an ore body located 4 km north-northeast of Goldfield.

Composite level maps for the major Goldfield mines (Goldfield Consolidated Mines Company, unpub. data) indicate that at least 8 million tons (7 million metric tons) of vein rock remain. The grade of this material has not been tested except for samples from the near-surface drilling by Pacific Gold and Uranium-Noranda-Southern Pacific, and sampling of open cuts by the U.S. Geological Survey (Ashley and Albers, 1975). Results to date suggest that the average grade of untested remaining vein rock is not likely to exceed 0 1 oz Au/ton (3 g/metric ton). Identified gold resources thus include measured reserves of about 5 tons (4.5 metric tons), and inferred resources of 16.5–22 tons (15–20 metric tons) (classification of U.S. Bureau of Mines and U.S. Geological Survey, 1980). Because the inferred resources are mostly deeper than about 50 m and are mainly sulfide ore, they presently include unspecified proportions of inferred reserves, inferred marginal reserves, and possibly inferred subeconomic resources. In any future mining, silver is likely to be of minor value and copper probably will not be economically recoverable. In the category of undiscovered resources, additional ore bodies with a few tenths of a ton to about 1.7 tons (1.5 metric ton) of total contained gold likely exist.

The Goldfield deposit is an epithermal bonanza of the enargite-gold or quartz-alunite type (Berger, 1986). Such deposits occur in calc-alkalic volcanic rocks, are associated with the effects of intense acid-sulfate hydrothermal alteration, and typically have significant byproduct copper. Other examples in the United States include Summitville, Colorado; Red Mountain, near Ouray, Colorado; and the Pyramid, Peavine, and Wedekind districts near Reno, Nevada Prominent examples elsewhere in the world include El Indio, Chile; Chinkuashih, Taiwan; and Kasuga and Akeshi, Kyushu, Japan. Goldfield is the largest deposit of this type known in North America, and one of the largest in the world, but it is considerably smaller than the related enargite-bearing massive sulfide deposits described by Sillitoe (1983).

GEOLOGIC SETTING

Volcanic rocks of Tertiary age host the major ore bodies at Goldfield (fig H2). These rocks are mainly porphyritic trachyandesite, rhyodacite, quartz latite, and rhyolite that formed during two episodes of calc-alkalic volcanism, in Oligocene and early Miocene time A small amount of ore in the deepest parts of the vein system formed in argillite of the Palmetto Formation of Ordovician age and quartz monzonite of Jurassic age.

An inferred caldera, about 6 km in diameter, formed in Oligocene time (Ashley, 1974) owing to extrusion of silicic ash-flow tuffs (Vindicator Rhyolite and Morena Rhyolite of Ransome, 1909) Quartz latite air-fall tuff (Kendall Tuff of Ransome, 1909) was erupted from a vent inferred to lie along the ring-fracture zone immediately east of Goldfield. Quartz latite flows followed deposition of the tuff, and the quartz latite that filled the vent formed a dome. These flows (unnamed latite of Ransome, 1909) are exposed 3 km east-northeast of Goldfield, but the dome is seen only in exposures in mine workings 1 km northeast of Goldfield. Rhyolite air-fall tuff was then erupted mainly from a vent located along the ring-fracture zone 2 km north of Goldfield. Rhyolite (Sandstorm Rhyolite of Ransome, 1909) filled the main vent to form a dome. The quartz latite and rhyolite were probably derived from the caldera-forming magma system. The Oligocene rocks were extensively eroded before the second period of volcanism, in early Miocene time.

The lower Miocene volcanic sequence consists of trachyandesite and rhyodacite flows, tuffs, and breccias, with minor quartz latite and basalt (all included in the Milltown Andesite of Ransome, 1909). Rhyolite and rhyodacite intrude this sequence to form numerous domes, and tuff breccias and flows from the vents now occupied by the domes dominate the upper part of the

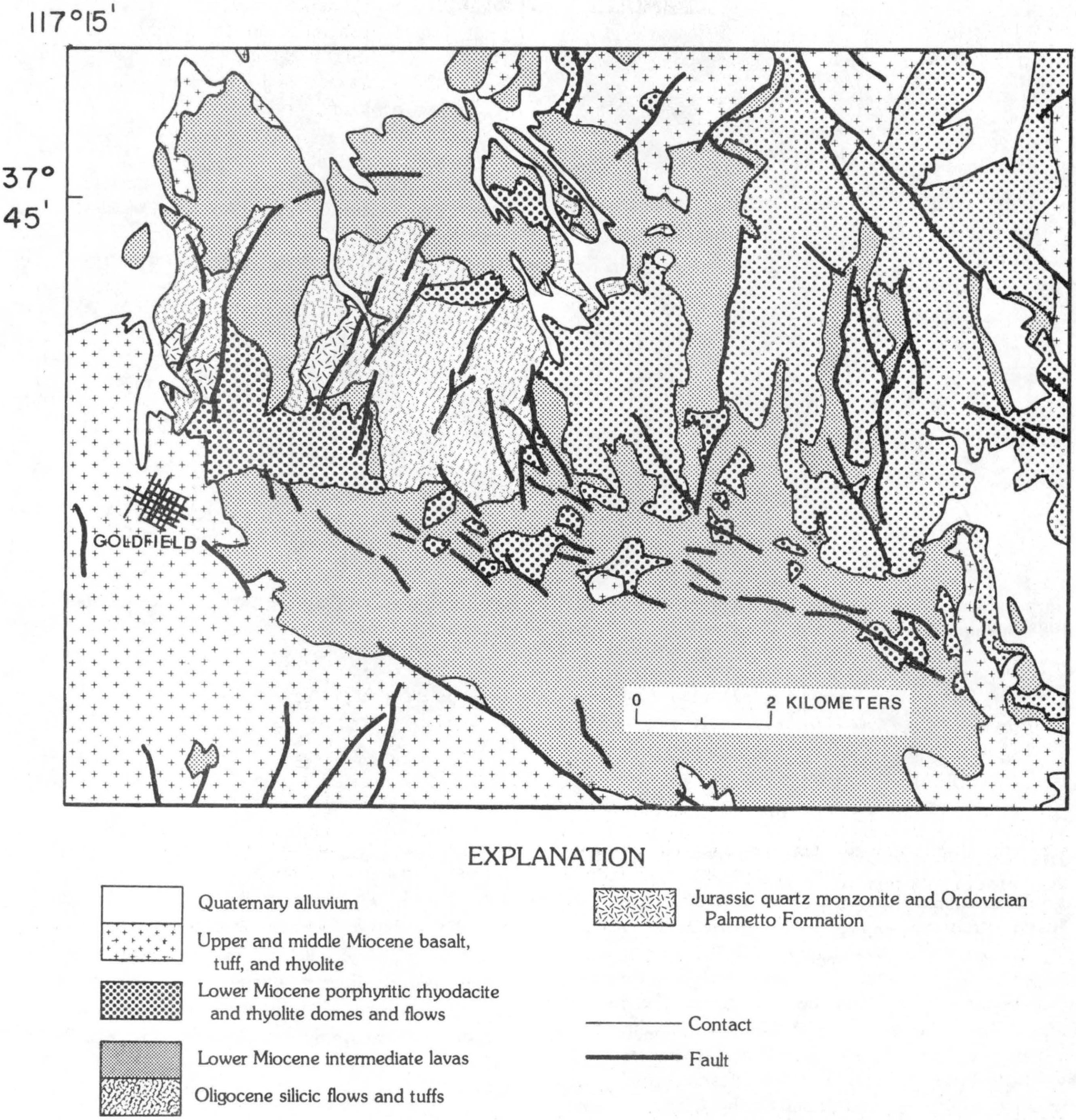

Figure H2. Generalized geology of the Goldfield mining district. Geology by R.P. Ashley, 1974.

lower Miocene section. A cluster of porphyritic rhyolite and rhyodacite domes is located 10 to 15 km east to east-northeast of Goldfield, and several domes of distinctive porphyritic rhyodacite (unnamed dacite of Ransome, 1909) are located along the ring-fracture zone.

Numerous normal faults formed late in the early Miocene episode of volcanism. The most prominent set trends northeast and dips east; bedding and flow contacts in the fault blocks are everywhere west-dipping, about at right angles to adjacent fault planes, and these faults thus are shingle faults bounding a series of rotated blocks. Another set of shingle faults on the south side of the district trends northwest, and the fault blocks dip southwest. Other less well defined fault sets appear to follow the ring-fracture zone, and represent fault movements that propagated upward from the ring-fracture zone through lower Miocene cover rocks, presumably because faults of the ring-fracture system

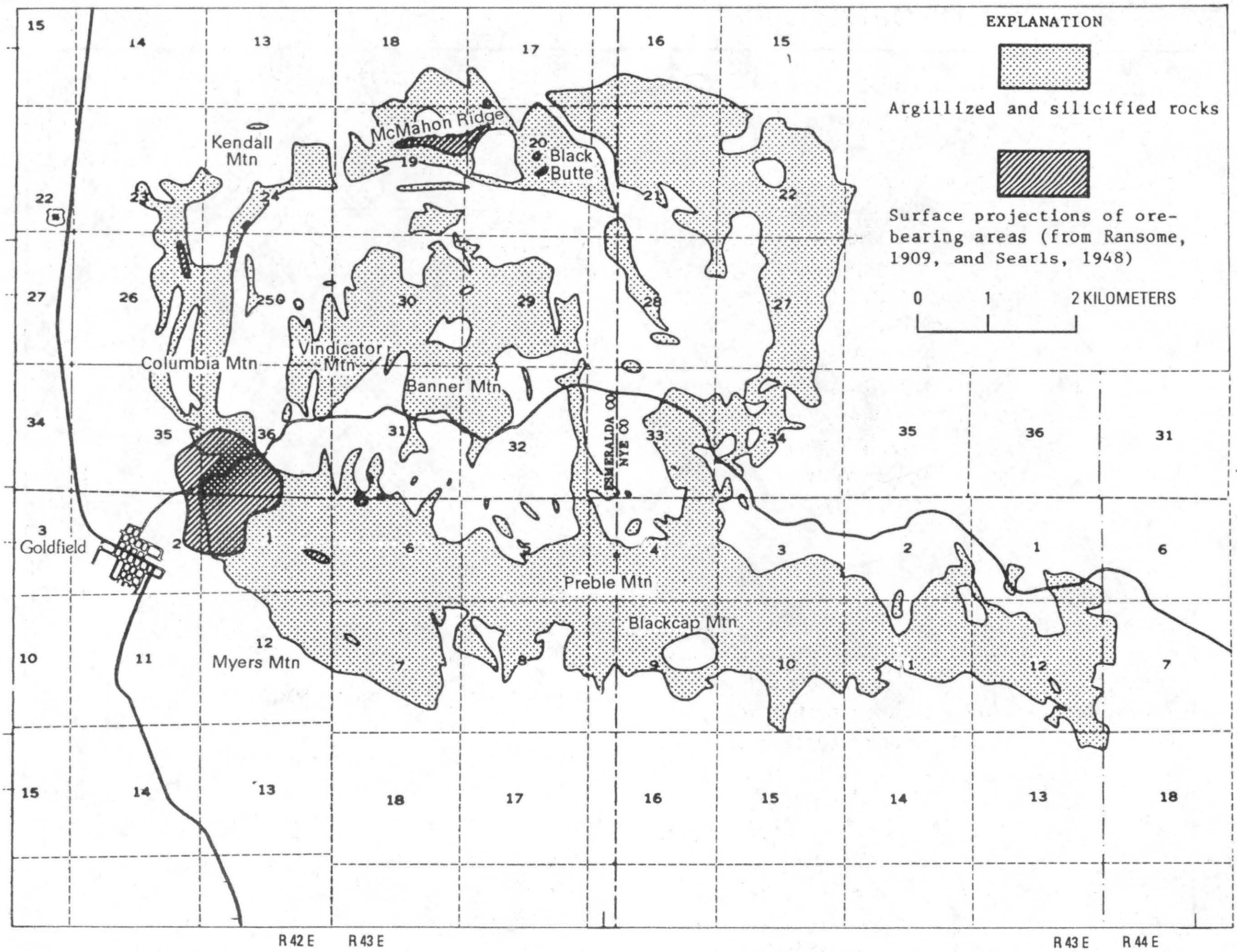

Figure H3. Hydrothermally altered and ore-bearing areas of the Goldfield mining district.

were reactivated. A major fault zone extends from the main ore-productive area, along the south side of the ring-fracture zone, and continues S. 80° E. for at least another 6 km. Individual faults that now define this zone are the same age or nearly the same age as the shingle faults. Distribution of some volcanic units of Oligocene and early Miocene age indicates that this zone is older than early Miocene, and may have formed during or immediately after emplacement of the Oligocene volcanic sequence.

Hydrothermal alteration at Goldfield affected an area of more than 40 km^2 (fig. H3). Much of the terrane in and around the ring-fracture zone is altered. Zones recognized, from the most to the least intense alteration, include (1) advanced argillic (quartz±alunite±kaolinite±pyrophyllite±sericite±diaspore + leucoxene + pyrite), (2) phyllic-argillic (quartz + kaolinite + sericite ±adularia±opal + pyrite), (3) argillic (quartz + montmorillonite + illite±kaolinite±relict feldspar + pyrite), and (4) propylitic (chlorite + albite±epidote±montmorillonite±calcite±zeolite±pyrite). The ore bodies occur in advanced argillically altered rocks. The zone of advanced argillically altered rock can be divided into two subzones, a quartz-rich zone that forms resistant outcrops, surrounded by an outer relatively quartz-poor alunite- or pyrophyllite-rich zone that is poorly exposed. The inner resistant subzone commonly follows fractures or faults to form tabular features that are referred to as silicified zones or silicified ledges (Ransome, 1909; Ashley and Albers, 1975).

Geochronology of Goldfield hydrothermal alteration minerals and host rocks has been studied extensively by Ashley and Silberman (1976). The lower Miocene volcanic rocks range from about 22 to 20.5 Ma. Potassium-argon ages of premineralization and postmineralization units in the early Miocene sequence indicate that hydrothermal alteration and mineralization took place about 20.5 Ma. This age estimate is confirmed by potassium-argon ages of hydrothermal alunite and sericite, and fission-track ages of apatite thermally reset during the hydrothermal activity.

CHARACTERISTICS OF ORE BODIES

Ore bodies occur in silicified zones at several locations along the west and north sides of the inferred ring-fracture system (fig. H3). Most of the ore produced, and nearly all the unoxidized ore, came from the relatively large productive area immediately northeast of the town of Goldfield. The ore bodies throughout the district were irregular sheets and pipes within crudely tabular silicified zones, which constitute the vein systems. The productive silicified zones (veins) have the same appearance and alteration minerals as unproductive silicified zones elsewhere in the altered area. The main vein in the area, just northeast of Goldfield, has a sinuous northerly trace, dips east at steep to moderate angles near the surface, and dips at progressively lower angles with increasing depth. Subsidiary veins occur both in the footwall and hanging wall, and generally dip more steeply; some of the hanging wall veins appear to splay upward from the main vein.

In relatively low grade ore (less than about 1 oz Au/ton), gold apparently is disseminated in the silicified rock; pyrite and lesser amounts of famatinite (tetragonal $Cu_3(Sb,As)S_4$; isomorphous with luzonite, Cu_3AsS_4) are the only prominent sulfide minerals. In high-grade ore, fractures and brecciated areas in silicified rock are filled with quartz, pyrite, famatinite, tetrahedrite-tennantite, bismuthinite, native gold, and local gold-silver tellurides. Spectacular high-grade ore consists of breccia fragments coated with several layers of these minerals, with native gold prominent in one or more of the layers. Two gold samples examined by electron microprobe have fineness greater than 980 (Peter Vikre, written commun., 1981). Gold:silver ratios for most unoxidized ore mined in the district were about 3:1. Much of the silver probably occurs in tetrahedrite-tennantite (Ashley and Albers, 1975). Rich ore from the center of the main vein contained 1.5 to 2 percent Au by weight (440 to 580 oz Au/ton). Ore containing 100 or more oz Au/ton was not uncommon in the district.

The high gold:silver ratios of the ores relegated silver to byproduct status. The only other significant byproduct was copper, derived from famatinite, tetrahedrite-tennantite, and minor amounts of chalcopyrite. Sphalerite occurs in some high-grade ores, but zinc was not recovered. Galena is very scarce, but a small production of byproduct lead is recorded. Tin is reported from some unoxidized ores mined deep in the main vein system (Searls, 1948); it probably occurs in famatinite. In addition to gold, silver, copper, arsenic, antimony, bismuth, and tellurium, high-grade ores show anomalous amounts of lead, zinc, mercury, tin, and molybdenum (Ashley and Albers, 1975). An unidentified selenium mineral was reported by Lévy (1967). Six samples of average-grade ore analyzed by a short-wavelength-radiation spectrographic technique for chalcophile elements (Heropoulos and others, 1984) showed significant amounts of tellurium (100–3,000 ppm) and selenium (70–500 ppm in five samples, <10 ppm in one sample), but only 2 ppm or less thallium (R.P. Ashley and Chris Heropoulos, unpub. data, 1980).

Tellurides were most abundant near the center of the main productive area at relatively shallow depths (about 60 m) in the upper parts of the vein system. Silver and copper content increased downward in the system relative to gold. Significant amounts of sphalerite and galena were found only at the north end of this area. Ore-mineral zoning in the district as a whole is obscured by the oxidized character of ores away from the main productive area. There is some indication that the gold-silver ratios generally decreased away from the main productive part of the area to the north and northeast (Ransome, 1909).

Isotopic and fluid-inclusion studies on Goldfield ores and alteration minerals are limited in scope. Pyrite from ore samples has $\delta^{34}S$ values near 0.0 per mil, indicating a magmatic sulfur source, whereas $\delta^{34}S$ values for several samples of hypogene replacement alunite range from +11.6 to +23.3 per mil, indicating sulfide-sulfate fractionation at temperatures of 150 °C or higher (Jensen and others, 1971). Some vein alunites have $\delta^{34}S$ values near 0.0 per mil, indicating that they were formed from pyrite oxidized during supergene alteration (Field, 1966). Values of $\delta^{18}O$ for a suite of whole-rock samples from several alteration zones, including advanced argillically altered vein material, range from −3.8 to +5.8 per mil (Taylor, 1973). Taylor proposed that Goldfield hydrothermal fluids had the same isotopic composition as those at Tonopah, which he concluded were about −13 per mil, or perhaps somewhat heavier. Qualitative examination of fluid inclusions, which are generally less than a few micrometers in diameter, suggests that temperatures of the hydrothermal system, in the parts presently exposed, were typically 200–300 °C, salinities were low, and boiling was common. Some samples from mine shafts and drill holes at depths of 300 m or more show relatively high salinities (greater than 24 weight percent NaCl equivalent), temperatures in the range 300–400 °C, and again, evidence for boiling. Bruha and Noble (1983) provided data for advanced argillically altered rock collected from a locality 4 km northeast of the town of Goldfield. They described fluid inclusions that show necking, the result of a wide range of apparent homogenization temperatures; they inferred true homogenization temperatures in the range 250–290 °C. Average salinity is 7 percent, and they saw no vapor-dominated inclusions, thus no evidence of boiling.

CONCLUSIONS

The geologic features of the Goldfield district include all the important distinguishing characteristics of the enargite-gold type of deposit (Ashley, 1982); indeed, Bethke (1984) has used the term "Goldfield type" for this class of epithermal deposits. These deposits (1) occur in calc-alkalic volcanic centers, associated with domes of porphyritic rock of intermediate composition; (2) are universally surrounded by prominent areas of acid-sulfate hydrothermal alteration; and (3) have ores that are characterized by native gold and abundant copper sulfosalts, particularly enargite or luzonite-group minerals such as famatinite.

The most important physical ore control for enargite-gold deposits was the presence of permeable zones, including faults, fractures, or permeable beds, which provided channelways for hydrothermal fluids. Major throughgoing structures, however, were not required; breccias and minor, but commonly abundant, faults and fractures associated with dome emplacement and growth were adequate. Equally important was the open space that was available near the conduits late in the episode of hydrothermal activity, because ore minerals were deposited in this open space. Chemical controls of ore deposition presently are not understood, and do not provide any additional ore guides.

The close spatial and temporal association between enargite-gold deposits and volcanic domes indicates that the deposits may have formed as part of the evolution of volcanic centers (Bethke, 1984). Igneous stocks or plugs intruded the volcanic centers to shallow levels, releasing plumes of SO_2-rich magmatic gas that moved upward and mixed with ground water to establish a hydrothermal circulation system in the presence of a high sulfur flux. With declining temperature, SO_2 disproportionated to yield SO_4, and with further temperature decrease, H_2SO_4 dissociated to produce the highly acid solutions responsible for the intense hydrolytic alteration and leaching that characterized the main stage of hydrothermal activity. Deposition of silica near the surface resulted in episodic self-sealing, development of overpressures, breakage, flash boiling, and hydrothermal explosions. Ore sulfides and sulfosalts were universally deposited during a late stage of hydrothermal activity, filling open spaces in zones of advanced argillically altered rock The required open space may have been produced by various mechanisms, including hydrothermal explosions or tectonic breakage, or even by earlier hydrothermal leaching of components of the volcanic host rocks. Little detailed information is available on conditions at the time of ore deposition, so the mechanism and controls of gold deposition are presently unknown. Sulfur must have remained relatively abundant in the fluids to produce the characteristic sulfosalt-rich ore-mineral assemblage (Wallace, 1979).

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The Tonopah Precious-metal District, Esmeralda and Nye Counties, Nevada

By Roger P. Ashley

Abstract

The Tonopah district is an epithermal precious metal deposit of the quartz-adularia type Andesitic and rhyolitic lava flows, ash-flow tuffs, volcanic and intrusive breccias, and volcaniclastic rocks host quartz veins that contain adularia and sericite. Most ore consists of sulfide-bearing quartz-adularia-sericite replacements of wall rock, but some consists of vein quartz with dark bands of ore minerals Silver occurs in acanthite and sulfosalts, and gold occurs in electrum, accompanied by minor amounts of base metal sulfides Host rocks and mineral deposits both are of mid-Tertiary age. Recent stable isotope and fluid-inclusion studies indicate that the ore bodies formed at relatively shallow depths from low-salinity meteoric water at about 240 to 270 °C

The district has produced 174 million ounces (5,420 metric tons) of silver and 1 86 million ounces (57 9 metric tons) of gold, has modest inferred reserves of both silver and gold, and has large hypothetical resources of silver Exploration activity and production, however, have been low since World War II

INTRODUCTION

The Tonopah mining district is located about halfway between Las Vegas, Nev., and Reno, Nev. (fig. H1). It lies in the western part of the Basin and Range province, 170 km east of the Sierra Nevada.

James L. Butler prospected the site of Tonopah in May 1900, and took samples for assay to Belmont, a silver camp 70 km to the northeast (Carpenter and others, 1953). After receiving favorable assay results, Butler returned in August 1900 and staked claims with partners Tasker L. Oddie and Wils Brougher. Production began late in 1900. Through 1901 Butler and his partners divided their ground into lease blocks and granted 120 leases. The Tonopah Mining Company was organized during 1901, purchased the original Butler claims, and took possession from the lessees at the beginning of January 1902. During the next 10 years, more than a dozen other mining companies were organized, and seven mills, ranging from 10 to 100 stamps in size, were built. The main productive period was 1910 to 1930; annual production during this period ranged from 1.9 to 11.6 million oz of silver and 20 to 128 thousand oz of gold per year. Total recorded production is 174 million oz (5,420 metric tons) of silver, and 1.86 million oz (57 9 metric tons) of gold, plus minor copper and lead for a total value of $154 million at prevailing prices (Bonham and Garside, 1979).

Since World War II, exploration programs have been carried out at various times in and around the Tonopah district by several companies. As a result of this exploration, a small amount of ore was developed and produced from the King Tonopah mine, at the north edge of the district (Bonham and Garside, 1979). Summa Corporation acquired approximately two-thirds of the formerly productive part of the district in 1969, and rehabilitated the shaft of the Mizpah mine, a major producer in the center of the district. Houston Oil and Minerals purchased Summa's properties in 1977 and carried out a modest exploration program in the central and western parts of the district. Houston Oil and Minerals was subsequently incorporated in Tenneco Minerals, and in 1986 Tenneco sold its Nevada gold properties to Echo Bay Inc. Marginal reserves include 128 metric tons of silver and 1.1 metric tons of gold in dumps and tailings (Saunders, 1984) Unpublished data of Tenneco Minerals suggest that inferred reserves are at least 233 metric tons of silver and 1.9 metric tons of gold, but are unlikely to exceed about 390 metric tons of silver and 3.1 metric tons of gold Tenneco data also suggest hypothetical resources of silver in the range of about 2,200 to 2,700 metric tons, and gold in the range of 9 3 to 14 metric tons (Saunders, 1984; resource/reserve classification of U.S. Bureau of Mines and U.S. Geological Survey, 1980). There is no reason to believe that reserves or resources in any category have Ag:Au significantly different from the ratio of approximately 100.1 typical of past production. Thus any ore remaining in the district

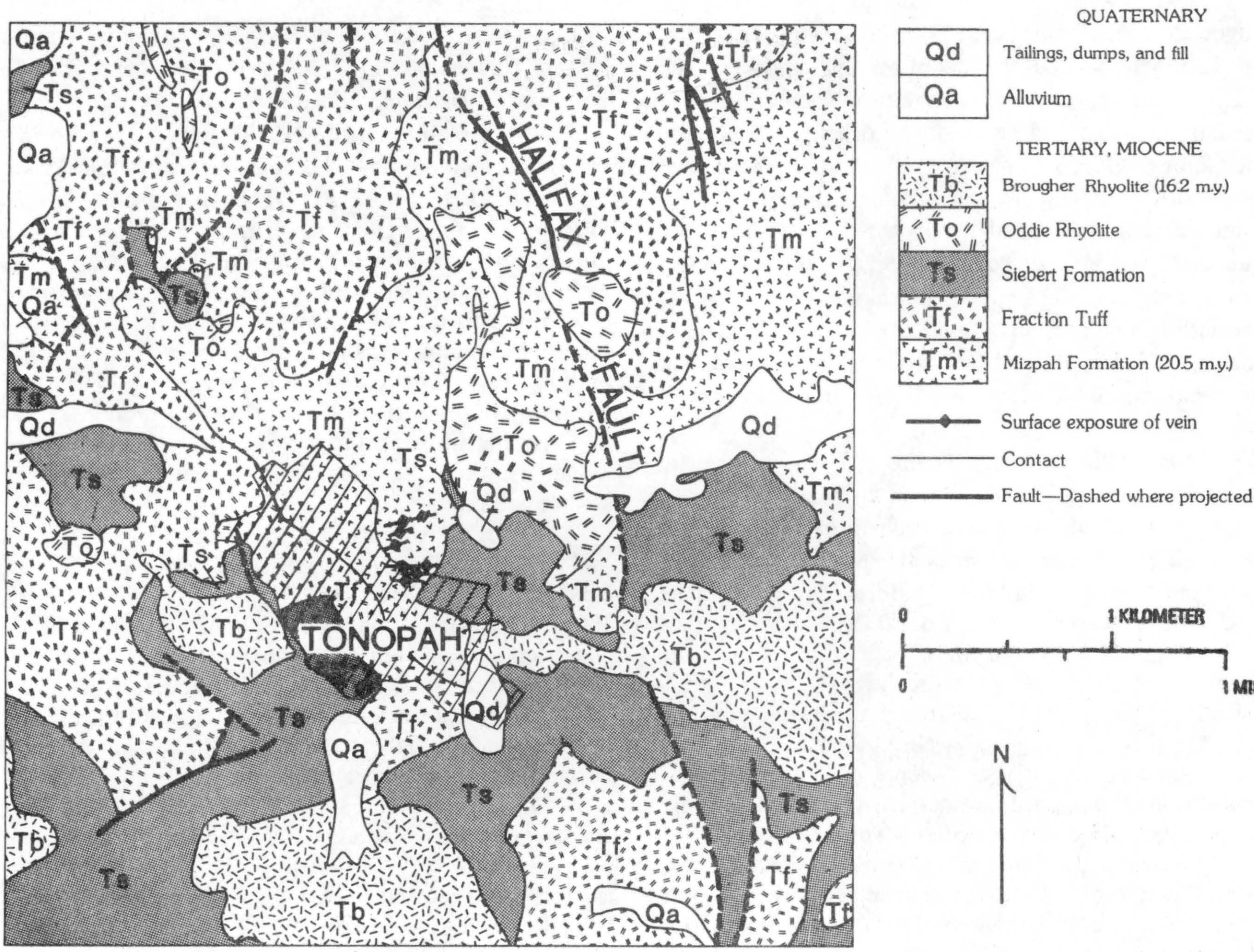

Figure H4. Generalized geology of the Tonopah mining district, Nevada.

has more potential value for silver than for gold, even at present high gold-to-silver price ratios.

The Tonopah district is an epithermal precious metal deposit of the quartz-adularia type (Mosier and others, 1986). The quartz-adularia type may be divided into gold-rich and silver-rich subtypes (Heald and others, 1987); Tonopah is a prominent example of the silver-rich subtype. Some silver-rich quartz-adularia type deposits have significant base metal content, but Tonopah is relatively base metal poor. The only larger deposit of this type known in the United States is the Comstock Lode, a relatively gold rich deposit, which has greater total production of both gold and silver than Tonopah. Other deposits of this type described elsewhere in this series of reports, also relatively gold rich, include Telluride, Colorado, and Oatman, Arizona. Worldwide, the most prominent examples are in Mexico, and all are relatively silver rich; the four largest are Pachuca-Real del Monte, Guanajuato, Guanacevi, and Tayoltita.

GEOLOGIC SETTING

J.E. Spurr visited the Tonopah district in 1902 and 1903, and published a comprehensive report in 1905. The most significant subsequent study is that of Nolan (1935), who mapped the extensive underground workings in 1929–1930. A recent study of Tonopah and the surrounding region by Bonham and Garside (1979, 1982) presents new data on the surface geology and geochemistry; the underground workings were mostly inaccessible by the time of their work. The generalized geologic map (fig. H4) is modified from Bonham and Garside (1979). Kleinhampl and Ziony (1984, 1985) have recently summarized the history, geology, and mineral deposits of the district.

The Tonopah Formation of Nolan (1935) is exposed in the deeper mine workings of the district. The Tonopah consists of silicic ash-flow and air-fall tuff, volcaniclastic rocks, and rhyolite domes and flows of late

Oligocene age. Bonham and Garside (1979) stated that the Tonopah Formation unconformably underlies the Mizpah Formation where the contact between the two formations is exposed north of the district. Kleinhampl and Ziony (1985), however, considered these two formations conformable, and claimed that they intertongue locally. Interbedded porphyritic andesite flows, named the Sandgrass Andesite by Spurr (1915), are also found only in the mine workings. The Tonopah Formation is more than 300 m thick in the underground workings (Nolan, 1935), but its base is not exposed, and the depth to pre-Tertiary basement in the district is unknown.

Most of the ore-bearing veins at Tonopah occur in the Mizpah Formation of Bonham and Garside (1979), originally called the Mizpah Trachyte by Spurr (1911, 1915). In the Tonopah district the Mizpah consists of flows and flow breccia, and the formation reaches its maximum preserved thickness of 600 m, suggesting that the vein system is located on or near a source vent for these volcanic rocks. At most localities the formation is hydrothermally altered, but scarce unaltered material has yielded two K-Ar ages at about 20.5 Ma (Bonham and Garside, 1979). Petrochemical data presently available for the Mizpah are not adequate to determine the petrogenetic character of this volcanism. Data for other volcanic piles of similar age in the region, however, suggest that these rocks are calc-alkalic, with greater potassium content than most calc-alkalic series (Anderson and Ekren, 1968; Ekren and others, 1971, Ashley, 1979).

Two intrusive units, the Extension Breccia and West End Rhyolite, are exposed in underground workings in the western part of the district (Nolan, 1930, 1935). Both are porphyritic rhyolite with abundant lithic fragments from the Tonopah Formation, Mizpah Formation, and Sandgrass Andesite, and pre-Tertiary argillite. The West End Rhyolite is clearly the later of the two, cutting and containing fragments of the Extension Breccia. Effects of hydrothermal alteration are apparently common, perhaps pervasive, in both units. These units are no longer accessible, and thus modern chemical and age determinations are not available.

The Fraction Tuff of Ekren and others (1971), redefined by Bonham and Garside (1979), unconformably overlies the Mizpah Formation. The Fraction is composed of quartz latite to rhyolite ash-flow tuff containing conspicuous moderately abundant lithic fragments throughout. Bonham and Garside (1979) recognized three distinct ash-flow cooling units, and placed the ages of these cooling units between 20.5 and 18 Ma, based on multiple age determinations. They proposed that a source caldera for at least part of the Fraction may be centered in the Divide mining district, located about 7 km south of Tonopah. The north margin of this caldera would lie about 1 km south of the Tonopah district mineralized area.

Bonham and Garside (1979) presented two direct age determinations for Tonopah mineralization, 19.1±0.4 Ma and 18.1±0.7 Ma. They reasoned that the Fraction Tuff is largely if not entirely premineralization in age, based on comparison of the Fraction age range with these ages. Dating of the Mizpah Formation indicates that Mizpah volcanism closely preceded Tonopah mineralization but was significantly older, by perhaps 1–2 million years. The Extension Breccia and West End Rhyolite are products of the most recent premineralization magmatism known in the district, but the difference in age between this magmatism and the mineralizing system is unknown.

The most widespread postmineralization unit in the vicinity of Tonopah is the Siebert Formation of Bonham and Garside (1979) (Siebert Tuff of Spurr, 1905). The Siebert consists mainly of fluvial and lacustrine deposits, including conglomerate, sandstone, and shale. Volcanic rock clasts dominate, and both subaerial and subaqueous silicic tuffs are common. In the vicinity of Tonopah, the Siebert is about 13 to 17 Ma. Another important postmineralization unit in the Tonopah district is the Oddie Rhyolite of Spurr (1905), which forms plugs and dikes that cut units as young as Siebert, and has been dated at about 16.5 Ma. Locally in the Tonopah district it is strongly altered and weakly mineralized. The Brougher Rhyolite of Bonham and Garside (1979) (Brougher Dacite of Spurr, 1905) forms prominent domes immediately south and east of the Tonopah district. Although the domes near Tonopah are 16.2 Ma, nearly the same age as the Oddie, the Brougher is not altered or mineralized.

Few faults are seen at the surface in the district. The most prominent is the Halifax fault, a north-trending, east-dipping normal fault on the east side of the district which shows both pre- and postmineralization displacements. Several other important north-trending premineralization normal faults and many minor faults with the same general trend and various dips were mapped underground by Nolan, but are not recognizable on the surface. The most important premineralization fault is the Tonopah fault, which is found at shallow levels in the central part of the district, and dips at moderate angles to the north and northwest. In the center of the district, the fault surface is nearly flat, and to the east and south it rolls over to low easterly and southerly dips, respectively, before dying out. Much of the ore in the district came from veins localized along the Tonopah fault and in steeply dipping, generally east trending hanging-wall splays above the Tonopah fault. The Tertiary formations of the district show varied but generally moderate westerly dips throughout the terrane west of the Halifax fault, eliminating the possibility that the Tonopah fault is a folded fault surface. Age relationships between faults of various trends are complex, but the north-trending, moderately to steeply dipping faults are generally younger than the low- to moderate-dipping faults and associated steeply dipping splays of the Tonopah fault system.

Several investigators, beginning with Spurr (1905), have described and interpreted the zoning patterns of wall rock alteration at Tonopah. Bonham and Garside (1979) combined an excellent discussion of earlier observations with data of their own; the following summary is from their report. The ore-bearing veins of the district are mainly quartz, with adularia and sericite. They are surrounded by successive replacement alteration envelopes, beginning with a potassium silicate zone having the mineral assemblage quartz + adularia + sericite + pyrite. The potassium silicate zone grades outward into an intermediate argillic zone that is first kaolinite-bearing (kaolinite + quartz + sericite + pyrite), then montmorillonite-bearing (montmorillonite + quartz + kaolinite + sericite + pyrite). The intermediate argillic zone grades outward into a widespread propylitic zone that was divided into several subzones by Bonham and Garside (1979). Combined widths of potassium silicate and intermediate argillic zones reach tens of meters.

CHARACTERISTICS OF ORE BODIES

All ore in the district is localized along faults, and thus the ore bodies, although commonly very irregular in detail, are generally tabular sheets. The productive segments of the veins are confined to a carapace-shaped zone about 200 m thick with an apex at the surface in the center of the district. The carapace extends downward 400–500 m on the east and west ends of the district, respectively, but only 200–300 m on the north and south sides of the district (Nolan, 1935). Much of the ore occurs as sulfide-bearing quartz-adularia-sericite replacements of wall rock, and most of the ore bodies had assay walls (Nolan, 1935). Open-space fillings account for only minor amounts of ore; here quartz with irregular dark bands of ore minerals lines cavities and is overgrown by much more conspicuous coarser grained quartz combs representing a late barren stage of vein filling. Judged from Nolan's block diagrams of the underground geology (Nolan, 1930, 1935), major ore bodies typically reach maximum dimensions of 300–400 m in the planes of the host fault zones, and Nolan (1935) stated that stope widths reach 40 ft (12 m).

The main silver-bearing mineral is acanthite, and silver sulfosalts (polybasite and pyrargyrite) are also important. Gold occurs in hypogene ores as electrum. Sphalerite, galena, and chalcopyrite are generally minor and sporadic, and they were deposited earlier than silver- and gold-bearing minerals. Major gangue minerals are quartz, adularia, and sericite; early rhodochrosite and late barite and calcite are minor gangue components (Fahley, 1981). Oxidized ores mined early in the district's history contained silver haloids and native gold (Burgess, 1911).

Neither mineral nor chemical zonation has been systematically investigated in the district, and such investigations are not possible with the limited sample collections available. Bonham and Garside (1979) found strongly anomalous amounts of silver, gold, and antimony, and weakly to moderately anomalous amounts of arsenic, copper, barium, and lead in hydrothermally altered surface samples from the limited area where veins cropped out. They found strong manganese anomalies accompanied by weak barium and zinc anomalies over blind ore bodies. Arsenic and mercury anomalies occur over some but not all ore bodies.

In recent years studies of oxygen isotope geochemistry (Taylor, 1973) and fluid inclusions (Fahley, 1981) done on the sample suite collected by Nolan have added much to our understanding of the conditions of ore deposition at Tonopah. Taylor found that quartz vein samples and rocks with potassium silicate alteration from the central part of the district have consistently low $\delta^{18}O$ values of about −7 per mil; samples from workings peripheral to the center of the district, including quartz vein material and rocks with propylitic alteration, have values a few per mil higher. Taylor assumed temperatures of 250–300 °C for alteration and ore deposition in this epithermal system, and reasoned from oxygen fraction factors for the quartz-water system at these temperatures that the ore fluid must have been dominated by meteoric water with $\delta^{18}O$ between −10 and −16 per mil. He postulated that an intrusive body beneath the district drove a meteoric-hydrothermal system, and that higher fluid temperatures, or greater volume of fluid upflow, or both produced the lower $\delta^{18}O$ values in the center of the district.

Fahley (1981) divided the main mineralization episode at Tonopah into three stages: an early barren stage, an ore deposition (silver) stage, and a late barren stage. Although the temperature ranges for all fluid-inclusion measurements are large for all three stages (80–180 °C), early barren stage quartz was deposited mostly between 250 and 270 °C, silver stage quartz was deposited at about the same, or perhaps slightly lower temperatures, and late barren quartz was deposited mostly at 240–250 °C. Following the late barren stage, temperatures declined to 140–160 °C, recorded by fluid inclusions in the latest quartz and associated barite and calcite. Salinity of the fluid was low throughout all stages, with typical values of about 1 weight percent NaCl-equivalent. Some samples from all stages showed features interpreted as the result of boiling of the fluid at the time of entrapment. Although wide ranges of homogenization temperatures commonly are observed at individual localities, Fahley's (1981) data suggest that temperatures during the silver stage were highest at depth in the western part of the district rather than in the center of the district. He postulated that fluids in the upper parts of the meteoric-hydrothermal convection system moved upward along the Tonopah fault system and into its hanging-wall fractures; the center of the district was an area of high fluid discharge.

CONCLUSIONS

Tonopah is a prominent example of the quartz-adularia type of epithermal precious metal deposit. Characteristics of this deposit type are described by Berger and Eimon (1983), Heald and others (1987), and Bethke (1984), and summarized by Mosier and others (1986). Discussions of genetic models by Buchanan (1981), Berger (1982), Berger and Eimon (1983), and Bethke (1984), as well as earlier work (White, 1981; Wetlaufer and others, 1979) pointed out the similarities between quartz-adularia-type epithermal systems and active geothermal systems that have been investigated

The hydrothermal origin of the Tonopah deposits has been recognized since the early investigations of the district (Spurr, 1905; Nolan, 1935). The combined results of Taylor's and Fahley's work bring many aspects of earlier models out of the realm of speculation and provide a basis for sound interpretations. Fahley's temperature data show that Taylor's assumptions are largely correct, and point to a specific $\delta^{18}O$ value for the fluids associated with hydrothermal alteration and ore deposition of about −13 per mil. The temperature data also give credence to Taylor's interpretation that water-rock ratios associated with the hydrothermal alteration process were quite high (2:1 or greater). Fahley's observations of boiling in alteration and ore-forming fluids, coupled with a reasonable assumption of prevailing hydrostatic conditions, allow an estimate of depth of ore formation of 300 to 600 m below the paleosurface. The stable isotope data, temperatures, boiling conditions, paleodepths, and uniformly low salinities all indicate conditions very similar to those seen in the explored, near-surface parts of active geothermal systems. Fahley (1981) called on boiling and resulting effects, including temperature drop, f_{O_2} increase, pH increase, and f_{S_2} decrease, to produce ore deposition. Not all samples of silver-stage quartz, however, show evidence of boiling.

Although it seems reasonable that mineralization at Tonopah is genetically related to an episode of intermediate to silicic volcanism represented by flows, flow breccias, and intrusions of the Sandgrass Andesite, Mizpah Formation, Extension Breccia, and West End Rhyolite sequence, the volcano-tectonic setting of the district at the time of mineralization is not clear. The critical problem is the age relationship between various members of the Fraction Tuff and mineralization; the age data upon which Bonham and Garside (1979) based their interpretation that the mineralization postdated the Fraction are not compelling If their interpretation is correct, however, the district may have formed immediately north of a caldera margin. If so, deeper parts of the meteoric convective system were likely localized by caldera-related structures, and some or all of the enigmatic mineralized structures of the district may be related to caldera formation. If mineralization predated the Fraction Tuff, the convective system may have formed in or on the flank of a stratovolcano In this case, the center of the system may have been on the west side or immediately to the west of the district, in the area inferred to host the feeders for the Extension Breccia and West End Rhyolite, as depicted by Fahley (1981). Mineralization at Tonopah clearly predated Basin-Range faulting, which began in this part of the Great Basin about 17 to 16 million years ago (Bonham and Garside, 1979; Ashley, 1979).

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The Republic Gold District, Ferry County, Washington

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Abstract

The Republic district up to 1987 has produced approximately 2,450,000 ounces of gold and 14 million ounces of silver, making it by far the largest gold producer in the State of Washington. Principal production has come from a number of moderate-size veins of limited depth near the town of Republic. Many veins were exhausted in the first part of the 20th century. Consolidation of many properties by Knob Hill Mines, Inc., aggressive exploration by Day Mines, Inc., and since 1981 by Hecla Mining Co., and the rise in the price of gold have kept this district active into 1989.

The ore bodies in the Republic mining district are principally fissure fillings Although the district lies within the Republic graben, most of the graben is not favorable for gold exploration. Localization of the district within the graben probably was dependent upon three essential features. First, and probably the most important, is the presence of the middle Eocene Klondike Mountain Formation that occurs in only a relatively small part of the graben along its west side. The gold veins were formed after deposition of the Tom Thumb Member but before extrusion of the upper basalt member of the Klondike Mountain. Second, the gold veins occur in or are associated with minor faults within the graben that had intermittent movement and that acted as a plumbing system. Third, competent wall rocks, such as flows in the Sanpoil Volcanics, were necessary to allow formation of through-going open fractures that permitted access of mineralizing fluids

INTRODUCTION

The Republic district is in the northern part of Ferry County in northeast Washington. Originally named Eureka, the town of Republic was founded in the district shortly after the first mines were located.

Gold was discovered while the area was a part of the Colville Indian Reservation. When the northern half of the reservation was opened to prospecting on February 21, 1896, all the principal veins were claimed within a few weeks (Full and Grantham, 1968, p. 1482). These included the Knob Hill, Ben Hur, Lone Pine, Mountain Lion, Quilp, Republic, San Poil, Sunrise, and Tom Thumb (fig. H6) (Muessig, 1967, p. 115). Three mills were built during the early years, and considerable ore was handled, mostly unsuccessfully. Mining continued until 1901, when several of the mines, including the Republic (the leading producer at the time), shut down. After the arrival of the railroads in 1902 and 1903, ore shipments were made to smelters in British Columbia and western Washington (Full and Grantham, 1968, p. 1483). From then through World War II (1945), annual production in the district fluctuated markedly. By 1907 it had fallen to approximately 390 oz of gold; it increased in 1908 and by 1911 reached 38,000 oz. Production then declined steadily to 1922 when 4,000 oz was produced. By 1927 production had increased to 15,600 oz; it fell in 1929 and 1930 to 900 oz, and rose in 1944 to 33,400 oz (Muessig, 1967, p. 115–116). During the early history of the Republic district, the Republic, Lone Pine, and Quilp at one time or another were the leading mines in the district. In 1936, however, Knob Hill Mines, Inc., acquired claims in the central part of the district and built a mill to handle low-grade ore from open-pit mining. The open-pit ore was worked out by 1944, and since then the company has operated from underground workings (Full and Grantham, 1968, p. 1483). The Knob Hill Mine has been by far the largest producer in the district, and in some years since the 1940's it has been the only one. Production figures for the district for many years were not released, especially after about 1936. Moen (1976, p. 103), however, estimated the district's production from 1897 to 1975 to be about 2 million oz of gold and nearly 10 million oz of silver. More recently, total production for the district through 1987 is given as 4.4 million tons of ore, having an average grade of 0.557 oz Au/ton and 3.23 oz Ag/ton (Republic Unit staff and Dayton, 1988). Thus, 2,450,000 oz of gold and 14,212,000 oz of silver have been produced. In 1987, record production of 70,095 oz of gold and 341,272 oz of silver was attained, and in 1988, resources in place amounted to 1.2 million short tons grading 0.50 oz Au/ton (Republic Unit staff and Dayton, 1988).

The Golden Promise vein system, a major discovery that will keep the district operating for many years, was found in 1984 about 7,000 ft southeast of the

Knob Hill mine (Republic Unit staff and Dayton, 1988). Although exploration in the Golden Promise area had noted mineralized rock as early as 1963, several periods of drilling were required to delineate large tonnages of ore in at least seven veins 0.6–6 m thick. A shaft was sunk in 1986, and production began soon after from the Golden Promise system.

GEOLOGIC SETTING

The Republic district is within the Republic graben, one of several such north-trending structural features in northeastern Washington (fig. H5). The district is near the west side of the graben near its midpoint. The graben was first described by Staatz (1960), and parts of it have been mapped in considerable detail by Parker and Calkins (1964), Muessig (1967), Staatz (1964), and Moye (1984). The graben has a north-northeast trend, and it extends from southern Canada southward for more than 84 km It ranges in width from 6.5 to 16 km. Vertical displacement on the boundary faults is not known, but some 21 km south of Republic, the middle of the graben is at least 2,140 m lower structurally than the adjacent blocks (Staatz, 1960, p. B304). Similarly, Parker and Calkins (1964, p. 71) noted that the western boundary fault about 29 km north of Republic, where this fault crosses the Kettle River, has a stratigraphic throw in excess of 5,180 m. Erosion, especially of the uplifted blocks adjacent to the graben, has been great, and in most places there is little topographic relief across the graben boundary faults.

The Republic graben began to form in early Tertiary time in a region underlain by plutonic and low- to high-rank metamorphic rocks ranging in age from pre-Permian to Cretaceous(?). The floor of the graben is now covered largely by Tertiary volcanic rocks that were extruded as the graben developed; in part, the floor of the graben is covered by pre-Tertiary metamorphic rocks raised along subsidiary faults, both along the graben margin and in its central part. The adjacent horsts consist mainly of pre-Tertiary plutonic and metamorphic rocks and also of Tertiary intrusive rocks that are coeval with the volcanics in the graben. In the vicinity of Republic, the adjacent horsts are composed of granodiorite of Cretaceous(?) age. Staatz (1964), Parker and Calkins (1964), and Muessig (1967) agreed that the graben formed as a complexly faulted block that dropped, in part at least, in response to withdrawal of magma from depth. Cheney (1980), however, believed that the Republic graben is basically a post-Eocene syncline in which the formerly more extensive volcanic rocks are preserved. As the graben developed, orogenic conglomerate and volcanogenic sediments accumulated in it; these were covered by a thick accumulation of intermediate volcanics, and all were intruded by hypabyssal and plutonic rocks (Muessig, 1962). This sequence has been divided into the O'Brien Creek Formation, Sanpoil Volcanics, Scatter Creek Rhyodacite (in part the intrusive equivalent of the Sanpoil), and the Klondike Mountain Formation. The O'Brien Creek Formation was probably deposited in early Eocene time, the Sanpoil Volcanics are middle Eocene, and the Klondike Mountain Formation is probably also middle Eocene (Pearson and Obradovich, 1977). Extrusion of the voluminous Sanpoil Volcanics into the Republic graben is considered to be one of the primary causes of graben subsidence (Staatz, 1964). Because of their association with the gold deposits in the Republic district, these rocks will be described herein.

O'Brien Creek Formation

The oldest Tertiary deposits in the Republic area are tuff, tuffaceous sandstone, and conglomerate of the O'Brien Creek Formation, which was deposited in a subsiding basin. The formation crops out in the central part of the graben northeast of Republic and along the graben margins just west of the town and about 16 km to the east. At many places along the graben margins the O'Brien Creek has been faulted out or overlapped by the younger Sanpoil Volcanics The thickness of the O'Brien Creek is highly varied, and along the north side of the North Fork Sanpoil River, some 13 km northeast of the Republic district, it is at least 1,280 m thick (Muessig, 1962, p. D57). In the Republic district, however, the thickness of the O'Brien Creek ranges from a meter to about 30 m (Full and Grantham, 1968, p. 1485).

The tuffaceous deposits that form the O'Brien Creek are largely well bedded sandstones that resulted from reworking of air-fall tuff. They consist mainly of plagioclase and quartz grains set in varied amounts of clayey matrix. Chips of gray slate or phyllite are so characteristic of these beds that the chips, like the plagioclase and quartz phenocrysts, must have had a pyroclastic origin, and their presence has resulted in this unit being referred to as "chip-pebble" formation by mine geologists within the Republic district (Full and Grantham, 1968, p 1485). Conglomerate is present at numerous places in the formation, but it is most common near the top and bottom. Full and Grantham (1968) reported an igneous-cemented conglomerate containing rounded pebbles as much as several inches in diameter to be present irregularly at the top of the formation Plant fossils from a water-laid tuff within the O'Brien Creek have been reported to be Eocene(?) (Muessig, 1962, p. D57). A K-Ar age determination on biotite from tuff correlated with the O'Brien Creek is 54.5 ± 1.5 Ma (Pearson and Obradovich, 1977, recalculated using 1977 IUGS constants).

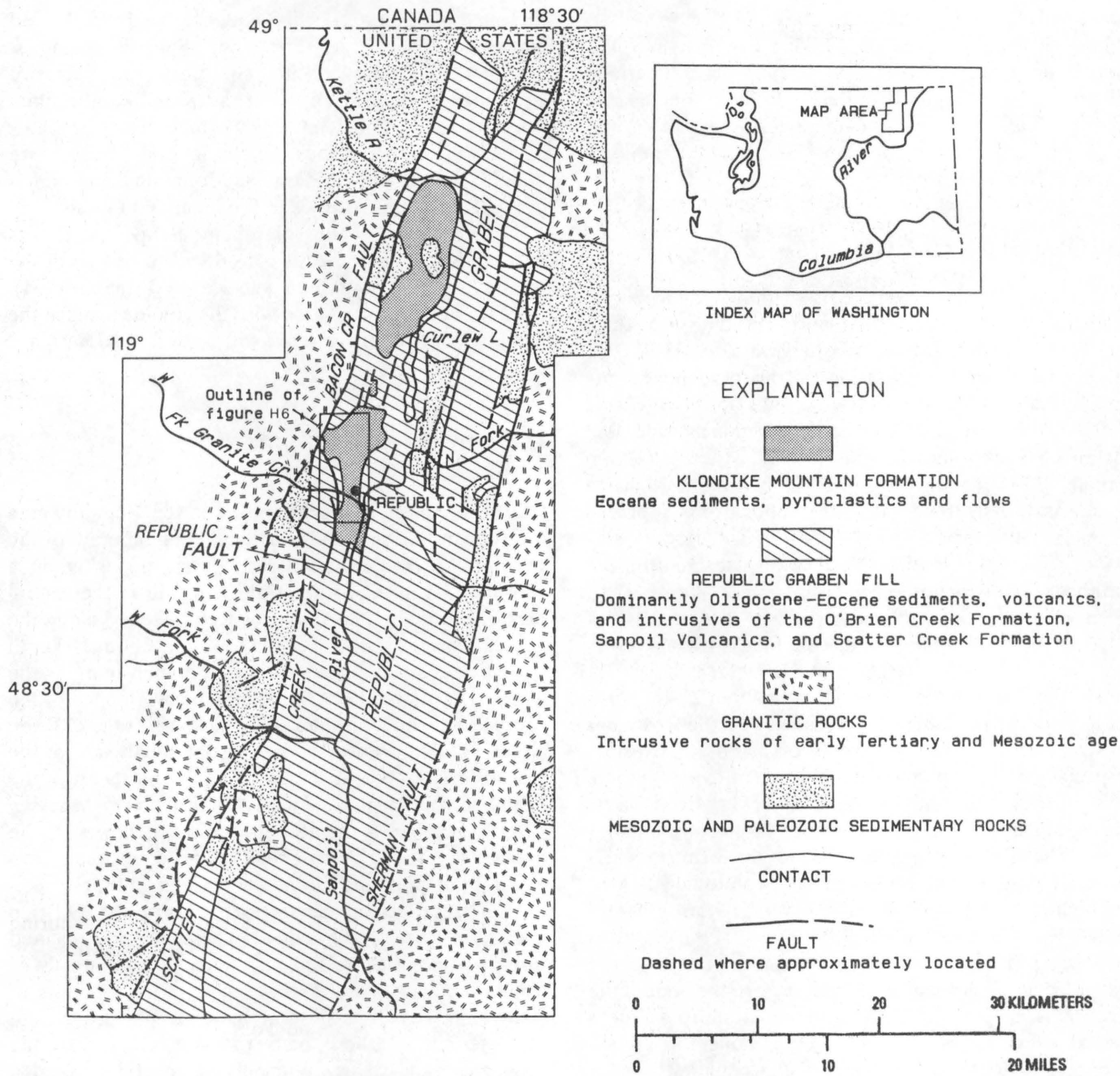

Figure H5. Generalized geology of the Republic graben, Ferry County, Wash. (modified from geologic map of Washington, 1961, by Full and Grantham (1968)).

Sanpoil Volcanics

The Sanpoil Volcanics are mainly a thick sequence of rhyodacite and quartz latite lava flows and breccias. These brown to gray fine-grained rocks are widespread in the Republic graben. At several places south of Republic, the Sanpoil Volcanics are nearly continuous from one side of the graben to the other. The thickness probably varies greatly, but in most places either the upper or the lower contact is not exposed; however, on the east side of Klondike Mountain, about 4.8 km northeast of Republic, Muessig (1962, p. D57) estimated a thickness of approximately 1,220 m. Although the Sanpoil is now confined to the Republic graben and other structurally low areas, the abundance of coeval plutons in adjacent structurally high regions suggests that the volcanics once covered much of northeast Washington and large parts of British Columbia and have since been eroded. The Sanpoil correlates with the Marron Formation in British Columbia, and both units have been determined by K-Ar ages on biotite, hornblende, and plagioclase to be about 52 Ma (Pearson and Obradovich, 1977). Colors of the flows range from drab shades of brown, green, purple, and greenish gray to gray and black. The rocks are strongly porphyritic, containing 30–65 percent phenocrysts in a fine-grained to aphanitic groundmass. Plagioclase is the most abundant phenocryst, but hornblende, biotite, and augite are present in various combinations. Although called andesite by some (Umpleby, 1910, p. 23–26; Full and Grantham, 1968, p. 1487), chemical analyses indicate that nearly all of them are rhyodacite or quartz latite (Muessig, 1967; Staatz, 1964; Pearson and Obradovich, 1977; and Moye, 1984).

Flow breccias are found in a number of places in the Sanpoil Volcanics. They were noted along the Sanpoil River 11–19 km south of the town of Republic by Muessig (1967) and Staatz (1964, p. F39), and they also occur in the eastern part of the Republic mining district. These breccias are generally in depositional units 3–6 m thick that can rarely be traced more than 60 m, but one body 19 km south of Republic is at least 60 m thick.

Air-fall tuff and water-laid tuffaceous beds separate some lava flows, but they rarely crop out. They are most common in the southern part of the graben (Staatz, 1964). The tuffs consist of angular to subrounded rock fragments and crystals in a matrix of devitrified ash.

Within the Republic mining district, the Sanpoil Volcanics have been divided into a lower green andesite member that contains nearly all the veins and an upper purple andesite member (Full and Grantham, 1968, p. 1487). The green andesite member in the western part of the district is 65–90 m thick but thickens to the east and contains pyroclastic as well as flow rocks. Detailed mine mapping has resulted in the division of the green andesite member into three units: upper and lower units that consist mostly of flows, separated by a middle unit that is mostly pyroclastic rocks. As the purple andesite member contains few veins, it has not been well studied in the mining district.

Scatter Creek Rhyodacite

The Scatter Creek Rhyodacite is the hypabyssal-intrusive equivalent of the Sanpoil Volcanics. Large numbers of the intrusives that make up the Scatter Creek are scattered throughout the entire length of the graben, most commonly along the graben-bounding faults. Some also cut the granitic rocks west of the graben's margin. The Scatter Creek cuts metamorphosed Permian sedimentary rocks, Cretaceous granite, O'Brien Creek Formation, and Sanpoil Volcanics, which it resembles and locally grades into. As the Scatter Creek bodies represent feeders to the Sanpoil Volcanics, the Scatter Creek is, at least in part, equivalent in age to the volcanics. In the mine workings of the Republic district, Scatter Creek intrusives cut the O'Brien Creek Formation, but contacts between Scatter Creek and the green andesite member of the Sanpoil Volcanics are commonly indefinite or gradational (Full and Grantham, 1968, p. 1487).

The Scatter Creek Rhyodacite is generally a light-gray to greenish-gray porphyritic rock containing 20–65 percent phenocrysts in an aphanitic or fine-grained groundmass. The most common phenocrysts are white euhedral plagioclase. Hornblende and biotite are the principal mafic minerals. Some intrusives contain a few phenocrysts of quartz. The groundmass contains mostly plagioclase, orthoclase, and quartz. Chemical analyses indicate that the Scatter Creek is rhyodacite or quartz latite similar in composition to the Sanpoil Volcanics (Staatz, 1964, p. F41–F44).

The Scatter Creek Rhyodacite is intruded along some important faults in the Republic mining district, and it contains several important veins (Full and Grantham, 1968, p. 1487).

Klondike Mountain Formation

The Klondike Mountain Formation is not nearly as widespread in the Republic graben as are the other volcanic units. This formation extends for 33 km along the west side of the Republic graben from a point 5 km south of the town of Republic to just north of the Kettle River. The irregular north-trending belt varies in width from about 1.6 to 7.2 km. The Klondike Mountain Formation is separated from the underlying volcanic units by an angular unconformity whose locally high relief indicates that considerable erosion occurred before the Klondike Mountain was deposited in a north-trending synclinal basin that had begun to subside during deposition of Sanpoil Volcanics. The Klondike Mountain is divided into three members: (1) the lower or Tom Thumb Tuff Member, (2) a middle member, and (3) an upper basalt member. The Tom Thumb Tuff Member contains a thin basal unit of coarse pyroclastics, including volcanic breccia, tuff breccia, and conglomerate (Muessig, 1967, p. 71). Most of this member, however, is made up of fine-grained water-laid tuff. In the southern part of the mining district, numerous thin flows are interbedded with the tuff. The thin-bedded tuffs have been referred to as lake beds (Umpleby, 1910, p. 24–25; Lindgren and Bancroft, 1914, p. 141–142) that are noted for their content of plant, insect, and fish fossils. The plant fossils were interpreted by Brown (1959, p. 127) as Oligocene, but K-Ar age determinations indicate that the Klondike Mountain is probably middle Eocene (Pearson and Obradovich, 1977). Wilson (1978, 1979) believed the fish fossils tend to confirm an Eocene age. The thickness of the Tom Thumb Tuff Member ranges from about 580 m northeast of the Knob Hill mine to 1.6 km to the east. The member also thins abruptly to the north and south.

The middle member of the Klondike Mountain Formation consists mainly of coarse pyroclastic rocks containing volcanic conglomerate, volcanic breccia, tuff, tuffaceous sandstone, and tuffaceous mudstone (Muessig, 1967, p. 73). Here and there it contains a few buff to gray latite flows. In the northern part of its exposure, it has thin black glassy flows at its base (Parker and Calkins, 1964, p. 63). The lower part of this member consists largely of volcanic conglomerate and breccia that form massive indistinct gradational layers which probably originated as lahars. The upper part of this member is mainly interbedded tuff and volcanic conglomerate, tuffaceous sandstone, and tuffaceous mudstone. In the Republic mining district, clasts in the volcanic breccia and conglomerate are principally of porphyritic flow rocks. In the northern part of the area, many clasts are also of pre-Tertiary rocks, including quartzite, chert, phyllite, argillite, and greenstone. The middle member varies greatly in thickness, from 915 m thick on the east side of Mount Elizabeth, 16 km northeast of the town of Republic, to 15 m thick about 1.5 km farther north (Parker and Calkins, 1964, p. 63). The middle member in the Toroda Creek graben about 32 km north and northwest of Republic is probably no younger than 48 Ma, or middle Eocene (Pearson and Obradovich, 1977).

The upper member was originally described by Muessig (1962, p. D58) as a brownish-black porphyritic basalt. An irregular basalt intrusive that cuts the lower two members of the Klondike Mountain Formation in the town of Republic is considered to be equivalent in age to flows of the upper member. Mapping in the Republic district (Full and Grantham, 1968, p. 1485) has divided the upper member into a lower andesite unit and an upper basalt unit. Sixteen to 24 km north of Republic, where the upper member is thicker, it is made up of mafic and intermediate flows, and small quantities of locally interlayered tuff-breccia in the upper part. The thickness of the member varies widely. Within 5 km of the town of Republic, it reaches 60 m (Muessig, 1967, p. 71). Sixteen kilometers north of town on Mount Elizabeth, the member is at least 365 m thick, but 4 km farther to the northwest on Granite Mountain it is only 60 m thick (Parker and Calkins, 1964, p. 64).

Late trachyandesite and dacite dikes, sills, and other small intrusive masses cut the other Tertiary volcanic rocks in the Republic mining district including the Klondike Mountain Formation. Muessig (1967) described the trachyandesite as olive drab, microcrystalline, and highly vesicular and amygdaloidal. The dacite is very fine grained and composed of plagioclase and minor quartz; in the mines it has been observed to be younger than the ore (Full and Grantham, 1968, p. 1488).

ORE DEPOSITS

Most ore mined in the Republic district has come from veins in a north-trending belt about 1.6 km wide that extends about 6.4 km north and 3.2 km south of the town of Republic (fig. H6). The veins are along and near north- to northwest-trending faults that are typical of many north-trending faults within the Republic graben, except that only those within the district seem to have guided mineralization. Muessig (1967), whose mapping was based only on surface exposures in a poorly exposed area, believed the principal structure within the district, which he named the Republic fault (fig. H5), to be one that separates the Tom Thumb Tuff Member on the east from Sanpoil Volcanics and Scatter Creek Rhyodacite on the west; the throw was inferred to be down to the east more than 300 m. South of Republic, the Republic fault trends about north and contains the vein at the Republic mine; about 1.6 km north of Republic, it bends to the northwest. Full and Grantham (1968, figs. 6 and 7), however, showed this contact to be depositional and unfaulted, but they did show a fault extending from about 300 to 600 m to the west that they called the Eureka fault (fig. H6). Its trace is in the bottom of Eureka Gulch, where Muessig (1967) also inferred a fault that branched from his Republic fault farther south. The Eureka fault dips steeply northeast and has normal throw of about 60 m. Full and Grantham (1968, fig. 3) showed the Eureka fault to be of pre-Klondike Mountain age, whereas Muessig (1967) interpreted it as a northerly continuation of his Republic fault that offsets the Klondike Mountain. Some of the north-trending faults were intruded by Scatter Creek Rhyodacite, which has been sheared by later movement and mineralized locally, and by trachyandesite dikes that are unsheared and unmineralized.

Although the Eureka fault is mineralized and veins along it have been mined at several places, most of the highly productive veins are along minor fractures in the hanging wall within 610 m of the Eureka fault. Most of these veins trend north to northwest, and some in the southern Eureka Gulch area trend northeast (Full and Grantham, 1968, fig. 8). Some veins appear to be offset by numerous northeast-trending cross faults, but Full and Grantham (1968, p. 1491) stated that movement on these was premineralization. At the Tom Thumb mine in the northern part of the district the veins trend northeast (fig. H6).

The veins are generally simple fissure fillings consisting dominantly of chalcedonic to fine-grained quartz and minor calcite, adularia, fluorite, and ore minerals. Replacement of wall rocks by vein minerals is also evident, particularly adjacent to wider veins, some of which have been mined to widths of more than 15 m. Complex branching of veins in the Knob Hill mine was

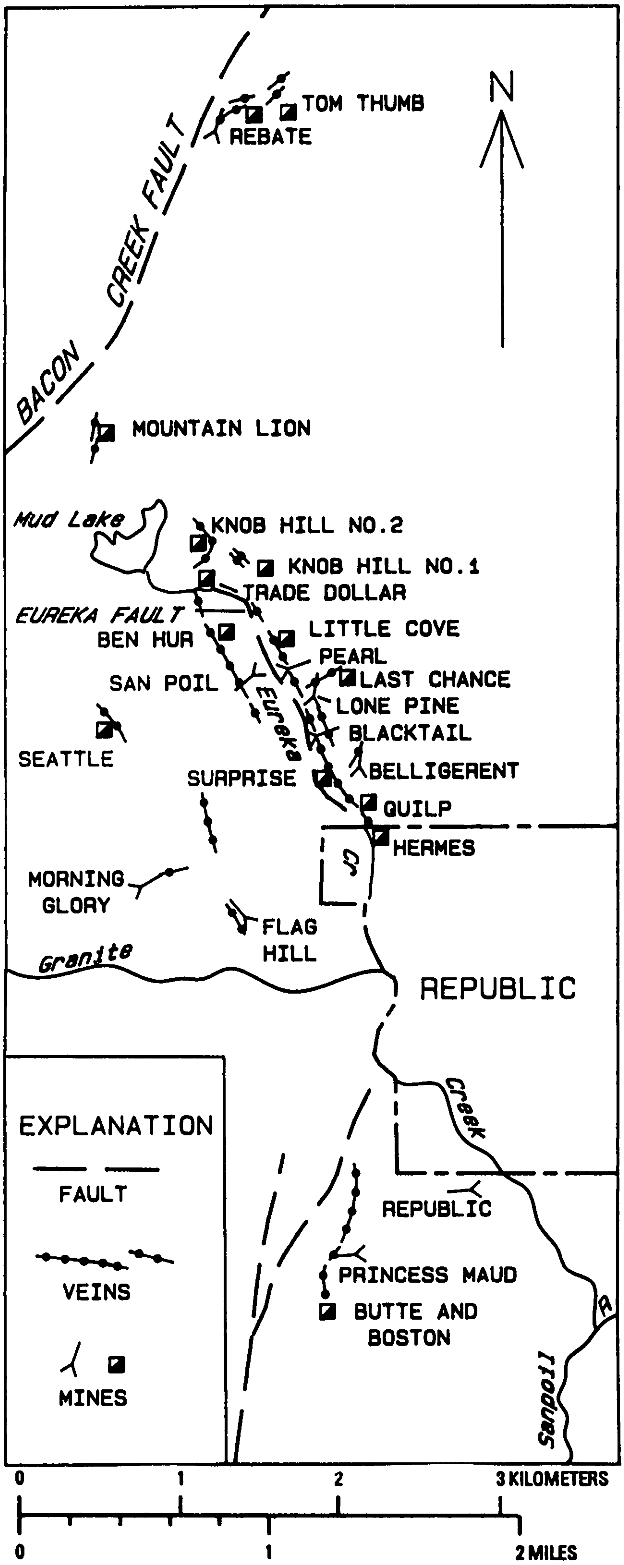

Figure H6 Index map, showing the mines of the Republic mining district, Washington (from Full and Grantham, 1968, fig 4)

illustrated by Full and Grantham (1968, fig. 10). Horses of altered wall rock are present in the wider vein zones, and smaller unreplaced fragments are common in the ore. At least two generations of quartz have been identified. In the richest ore the quartz is dark gray to black, forming laminae or bands arranged parallel to vein walls or to wallrock fragments within the veins. The sulfides and other ore minerals that impart the dark color are mainly pyrite but also include stephanite, naumannite, native silver, and electrum (Muessig, 1967). Moen (1976) also reported native gold, chalcopyrite, tetrahedrite, stibnite, realgar, pyrargyrite, argentite, and umangite (Cu_3Se_2) from the district. Precious metal content of the ore mined from 1902 to 1951 averaged about 0.34 oz Au/ton and 2 18 oz Ag/ton according to Moen (1976), and 0.557 oz Au/ton and 3 23 oz Ag/ton according to the Republic Unit staff and Dayton (1988). However, the gold.silver ratio varied from mine to mine and within mines from 1:3 to 1:12 (Moen and Huntting, 1975, p 55–58).

Although the veins are as much as about 600 m long, the mined ore shoots within them are much smaller (Moen, 1976) Maximum stope length is about 105 m and maximum pitch length is 400 m.

The productive veins in the Eureka Gulch part of the district are essentially all confined to the massive lavas in the lower part of the green andesite member and their intrusive equivalent, Scatter Creek Rhyodacite. Some veins in the district, however, occur in the underlying O'Brien Creek Formation (chip-pebble formation) and in the overlying purple andesite member of the Sanpoil Volcanics and the Klondike Mountain Formation (Full and Grantham, 1968, p. 1489). Disseminated ore has also been produced from volcaniclastic deposits in the lower part of the Klondike Mountain. The decided tendency of the massive lava flows and intrusive rocks to host the veins is ascribed to their competence to hold fractures open during vein formation (Full and Grantham, 1968) The pyroclastic unit that forms the middle part of the green andesite member is much less competent, and veins die out abruptly upon entering it The upper part of the green andesite member also consists of competent lava flows that contain several veins including the Tom Thumb, Rebate, Mountain Lion, and the upper parts of the Republic and Knob Hill veins The recently discovered Golden Promise vein system, however, is considered to be in pyroclastic rock (Republic Unit staff and Dayton, 1988).

The disseminated ore in the volcanic rubble at the base of the Klondike Mountain Formation was about 15 m thick and was mined from an open pit beginning in 1936 by Knob Hill Mines, Inc. This low-grade ore body overlies vein deposits in the Sanpoil Volcanics that were discovered later. These veins do not extend upward into the incompetent rubble, but the mineralizing solutions from the veins were able to permeate the weakly consolidated debris and produce a low-grade disseminated deposit.

Mineralization occurred during emplacement of the rocks of the Klondike Mountain Formation. The Tom Thumb Tuff Member of the Klondike Mountain is the youngest unit that has been mineralized. Unaltered and unmineralized trachyandesite dikes cut the veins at the Mountain Lion mine, and basalt dikes cut veins at other places. The trachyandesite is correlated with lava flows in the upper member of the Klondike Mountain, and the basalt dikes are correlated with the youngest Klondike Mountain lavas. Hence, the mineralization took place after the lower member was emplaced but before the upper member of the Klondike Mountain was formed.

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The Comstock Lode Precious-metal District, Washoe and Storey Counties, Nevada

By William C. Bagby *and* Roger P. Ashley

Abstract

The Comstock Lode district in the Virginia Range of western Nevada was discovered in 1859, and subsequently produced a total of more than 8 million ounces of gold and 190 million ounces of silver. Its most productive period lasted from 1860 to 1880, when large bonanza ore bodies were mined The ore bodies were localized along the fault plane and in the hanging wall of the Comstock fault, an east-dipping normal fault that displaces Miocene andesitic rocks of both the Alta and Kate Peak Formations The ore is dominated by quartz, sphalerite, and galena, with lesser amounts of argentite, polybasite, stephanite, chalcopyrite, pyrite, gold, and electrum Fluid-inclusion filling temperatures for vein quartz range from 240 to 295 °C, with salinities between 2 7 and 3 6 equivalent weight percent NaCl Stable isotope studies indicate that hydrothermal fluids were dominated by meteoric water in most parts of the deposit, except for deep levels which may have been dominated by magmatic water Periods of volcanism, tectonism, and precious metal mineralization overlapped

INTRODUCTION

The Comstock Lode district is located 25 km southeast of Reno, Nev. (fig. H7). The mining camps of Virginia City, Gold Hill, and Silver City are located over the most productive parts of the lode. The Flowery and Jumbo districts flank the Comstock Lode district on the east and west, respectively (fig. H8).

Gold was first discovered in Gold Canyon by John Orr and Nick Kelly in 1850 (Smith, 1943). This discovery led to a period of placer mining along Gold Canyon and Six Mile Canyon. Although most of the best placer deposits were exploited by 1855, a few persistent miners continued working the canyons until 1859, when Peter O'Riley and Patrick McLaughlin discovered the Ophir bonanza in the northern part of the Comstock Lode. In the same year, 1859, James Finney, Jack Yount, and John Bishop discovered gold in place at the head of Gold Canyon. This location later became Gold Hill (Smith, 1943).

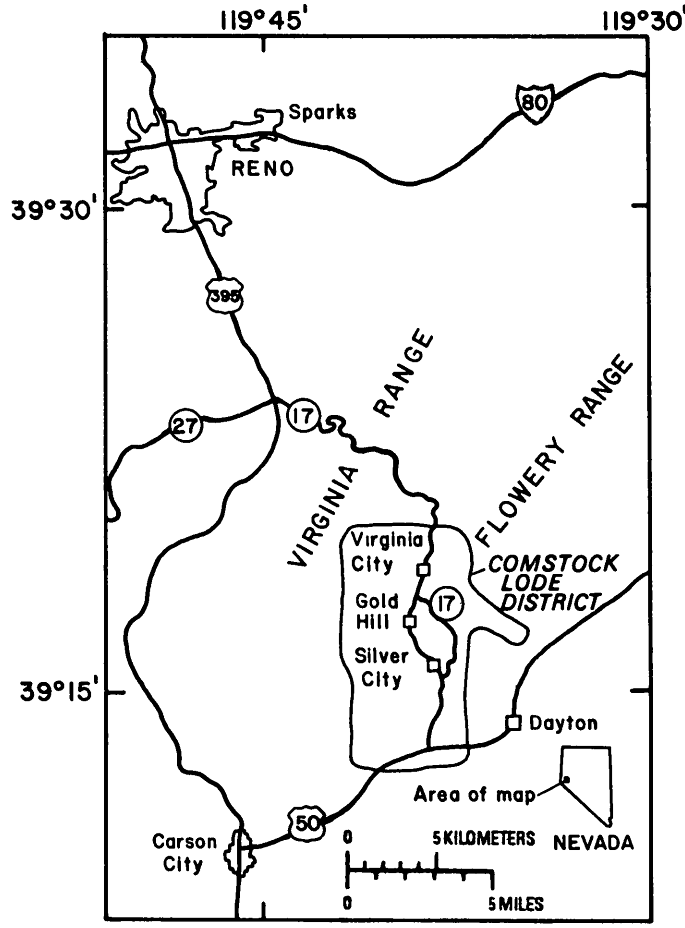
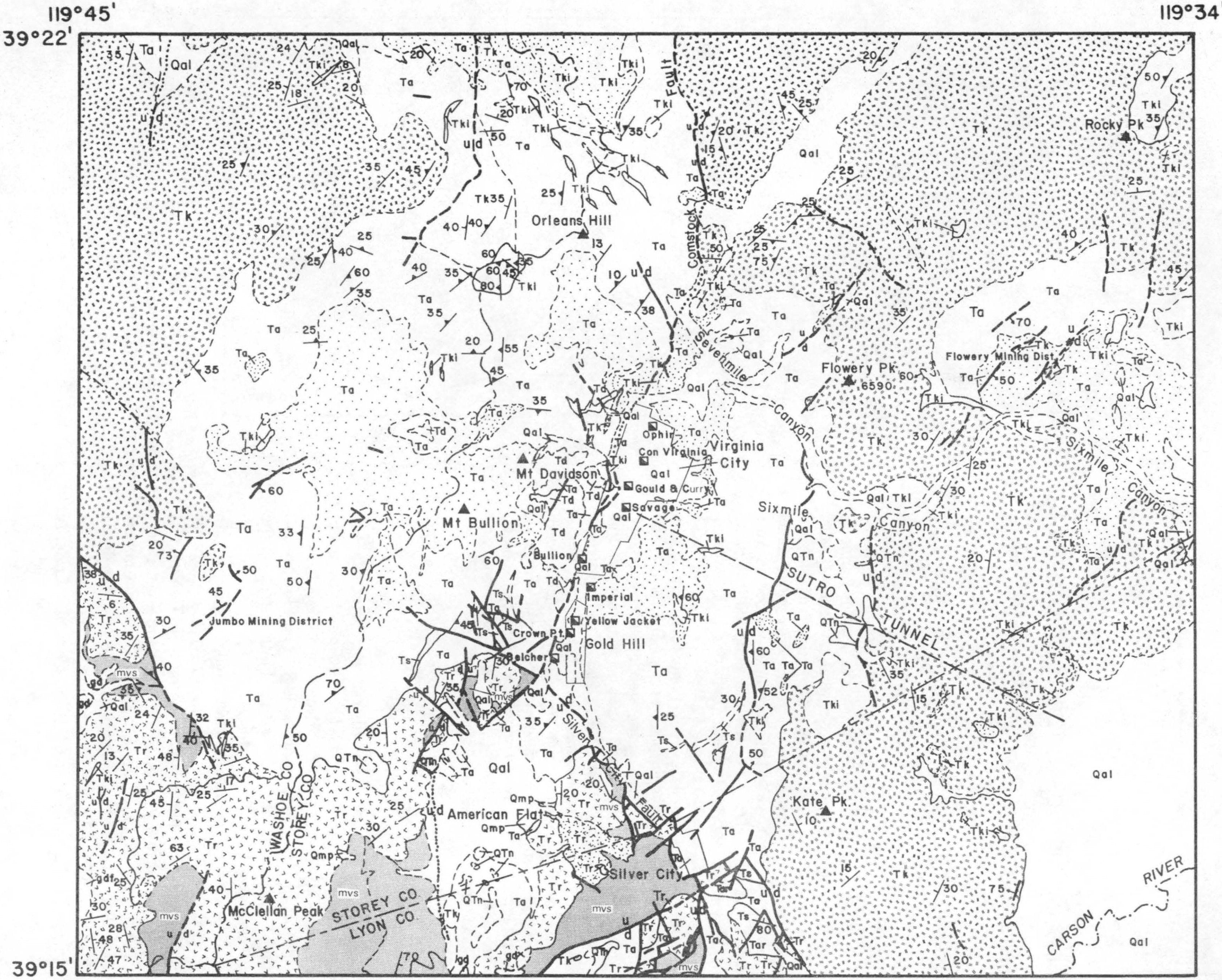


Figure H7 Location of the Comstock Lode district in western Nevada The Comstock Lode refers to the mines that were developed on mineralized rock between and including Silver City and Virginia City

New ore was discovered in the early 1860's, and by 1863 the Comstock Lode had produced about $10 million in gold and silver from near-surface workings (Bonham, 1969). Interest in the Comstock Lode was fired by successful exploration during the late 1860's.


119°45'
119°34'
39°22'
39°15'
Orleans Hill
Comstock Fault
Rocky Pk
Flowery Pk
6590
Flowery Mining Dist.
Sevenmile Canyon
Sixmile Canyon
Ophir
Con Virginia
Virginia City
Mt Davidson
Gould & Curry
Savage
Mt Bullion
Bullion
Imperial
Yellow Jacket
Crown Pt
Belcher
Gold Hill
Jumbo Mining District
SUTRO TUNNEL
Silver City Fault
American Flat
Silver City
Kate Pk
McClellan Peak
WASHOE CO
STOREY CO
LYON CO
CARSON RIVER

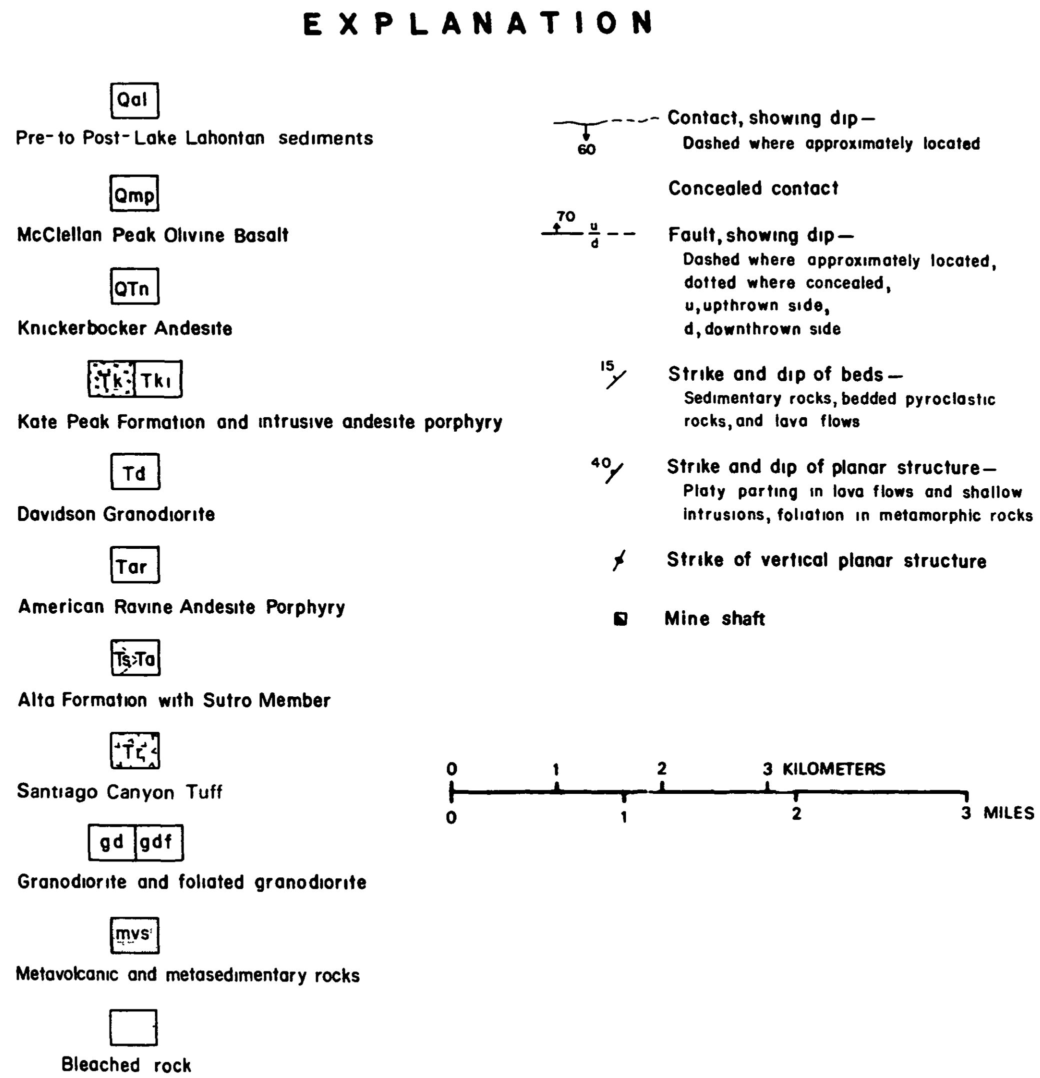


Figure H8. Geology of the Comstock Lode area. Modified from Thompson and White (1964). Con Virginia is Consolidated Virginia mine.

This led to discovery of the big bonanzas in the Crown Point and Belcher mines in 1870 and in the Consolidated Virginia mine in 1873.

Boom years for the Comstock lasted from 1860 to 1880. During this period 5.9 million oz of gold and 143 million oz of silver were produced (Bonham, 1969). Production from the Comstock began decreasing in 1880, but mining continued until about 1955, resulting in total production of 8.3 million oz of gold and 192 million oz of silver. Although copper was first produced in 1905 and lead in 1907, gold and silver continued to be the major products of the district.

GEOLOGIC SETTING

The geology of the Comstock Lode district has been described by Becker (1882), Reid (1905), Gianella (1936), Calkins (1944), Thompson (1956), Thompson and White (1964), and Bonham (1969) Other studies include published topical reports on ore mineralogy (Bastin, 1922; Coats, 1936) and alteration (Coats, 1940; Whitebread, 1976; Ashley and others, 1979), and unpublished mining company reports.

Stratigraphy

The oldest rocks in the Comstock Lode district are pre-Tertiary sedimentary and volcanic rocks that crop out near American Flat (fig. H8). These rocks are considered to be Triassic, though parts of the sequence may be as old as Paleozoic or as young as Jurassic (Thompson, 1956; Thompson and White, 1964; Rose, 1969). Sedimentary rocks include argillite, slate, phyllite, sandstone, and conglomerate derived from volcanic rocks, and minor limestone. Volcanic rocks are basalt and andesite. All these rocks are metamorphosed to at least greenschist facies, and the volcanic rocks are characterized by albite, epidote, chlorite, and amphibole. More intense metamorphism near plutonic contacts produced amphibolite-facies hornfels, schist, marble, and skarn.

Cretaceous plutonic rocks occur as small scattered bodies that intrude the metavolcanic rocks west of Silver City in the Comstock Lode district (fig H8). Several small inliers that are surrounded by Tertiary and Quaternary rocks also appear along the east flank of the Flowery Range. Most of the Cretaceous intrusions are granodiorite, however, the small intrusions in the Comstock Lode district are quartz monzonite porphyry (Thompson, 1956; Thompson and White, 1964, Calkins, 1944).

The oldest Tertiary rocks in the Comstock Lode area are rhyolitic ash-flow tuffs of early Miocene age that crop out on McClellan Peak west and south of Silver City (fig. H8). These tuffs were originally mapped as the Hartford Hill Rhyolite by Gianella (1936). Thompson (1956) referred to these rocks as the Hartford Hill Rhyolite Tuff, based on the occurrence of flattened pumice lapilli. Bingler (1978) abandoned the name Hartford Hill Rhyolite Tuff and separated Thompson's (1956) unit into six separate ash-flow tuffs of regional extent based on different eruptive and cooling histories The rhyolite at Hartford Hill near Silver City was named the Santiago Canyon Tuff after its type section at the mouth of Santiago Canyon (Bingler, 1978). The tuff contains phenocrysts of plagioclase, quartz, sanidine, hornblende, biotite, and sphene in a groundmass of flattened pumice lapilli and glass shards. The formation has a maximum thickness of 300 m and is composed of one simple cooling unit. K-Ar ages of biotite and sanidine from the tuff range from 23 to 20 Ma (table H1).

The Miocene rhyolitic ash-flow tuffs in the Comstock Lode district are overlain by a thick pile of andesitic rocks erupted from vents within the Virginia Range. These intermediate volcanic rocks are divided into two major formations, the Alta and Kate Peak (Gianella, 1936; Thompson, 1956; Thompson and White, 1964). Each formation consists of lavas, tuffs, tuff breccias, and intrusive rocks. Although Thompson and White (1964) regarded tuff breccia as the most common textural type, recent mapping by D.M. Hudson and H F. Bonham (written commun., 1983) in the Virginia Range, the Peavine district (northwest of Reno), and the northern Carson Range indicates that the Kate Peak Formation is dominated by lava. Volcanic conglomerate and sandstone in both formations and shale in the Alta Formation are interbedded with lava and pyroclastic rocks. Each formation is dominated by potassium-rich calc-alkalic andesite (fig. H9); the Kate Peak Formation contains some rhyolite, but the most siliceous rocks in the Alta Formation are dacitic.

The Alta and Kate Peak Formations were originally distinguished in the district on the basis of unconformable relations and the more pervasive alteration of the Alta. Thompson and White (1964), however, stated that structural discordance is small or absent and that the degree of alteration effects varies more from area to area rather than from one formation to another. On the other hand, D.M. Hudson (written commun., 1983) reported that lavas in the lower part of the Kate Peak Formation regionally are strongly altered, whereas lavas in the upper part of the Kate Peak are unaltered. It thus appears that neither unconformable relations nor alteration is an unequivocal guide for distinguishing the two formations. Instead, they are probably best distinguished on the basis of phenocryst content. Thompson (1956) cited evidence suggesting that the two formations intertongue in the Virginia City quadrangle. Potassium-argon ages for the Alta Formation range from 17 Ma to 14 Ma, whereas

Table H1. Potassium-argon ages for rocks and vein minerals in the Comstock Lode district

Rock type	Mineral	Age (m.y.)	Reference
McClellan Peak Olivine Basalt	(whole rock)	1.14	Doell and others, 1966
Kate Peak Formation .	biotite	12.3±0.2	Whitebread, 1976
Kate Peak Formation intrusive rock	biotite	12.4±0.3	Do
Kate Peak Formation intrusive rock	biotite	12.7±0.4	Do
Kate Peak Formation dacite lava	plagioclase	12.8±0.8	Silberman and McKee, 1972
Occidental vein	adularia	12.8±0.4	Whitebread, 1976
Kate Peak Formation lava	hornblende	12.9±0.4	Do
	hornblende	13.7±1.6	Do
Comstock vein .	adularia	13.7±0.4	Do
Kate Peak Formation lava	biotite	13.8±0.3	Do
Kate Peak Formation intrusive	biotite	14.1±0.4	Do
	hornblende	14.9±0.4	Do
Alta Formation	plagioclase	14.4±0.4	Do
Alta Formation	plagioclase	16.5±0.5	Silberman and McKee, 1972
Davidson Granodiorite . . .	biotite	17	(M.L. Silberman and R.P. Ashley, unpub. data, 1976.)
Santiago Canyon Tuff	sanidine	20.5	Bingler, 1978
Santiago Canyon Tuff	biotite	21.8	Do
Santiago Canyon Tuff . .	biotite	22.5±0.7	Whitebread, 1976

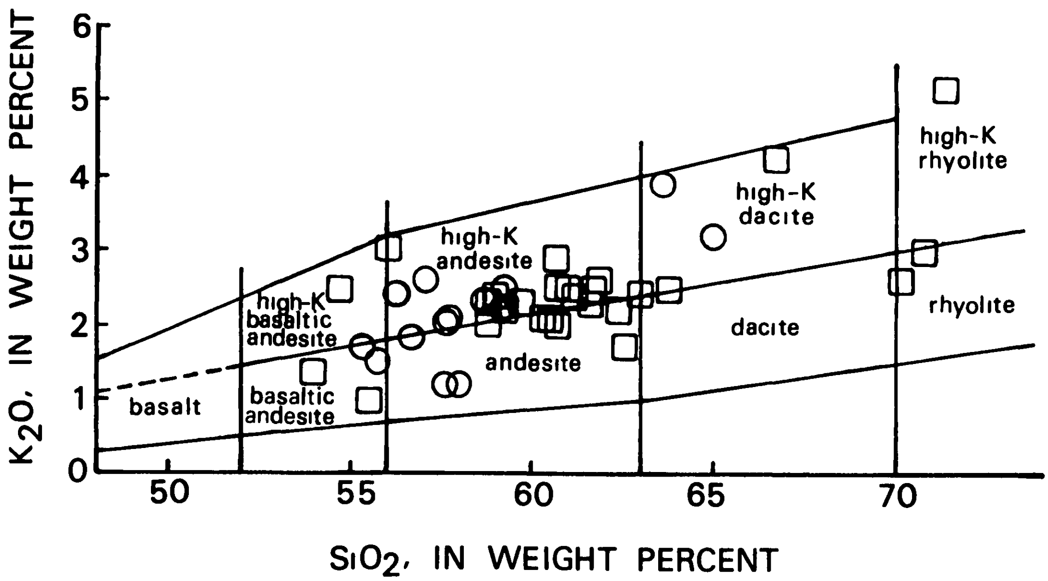


Figure H9. K_2O-variation diagram for samples from the Alta and Kate Peak Formations. Data from Thompson and White (1964), and Whitebread (1976). Classification boundaries from Peccerillo and Taylor (1976). Circles and squares represent Alta and Kate Peak Formation samples, respectively.

ages for the Kate Peak Formation range from 15.3 Ma to 12.1 Ma (table H1). These ages overlap between 15.3 and 14.0 Ma, corroborating Thompson's (1956) evidence for intertonguing of the formations.

The American Ravine Andesite Porphyry crops out in American Ravine and Gold Canyon (near south edge of area of fig. H8). It intrudes the Santiago Canyon Tuff and may also intrude the Alta Formation (Calkins, 1944; Thompson, 1956). Its age thus is probably between that of the Alta and that of the Kate Peak, as the two formations are distinguished south of the Comstock Lode (Calkins, 1944; Thompson, 1956).

The Davidson Granodiorite cuts the Alta Formation on and near Mount Davidson, immediately west of the Comstock Lode (center, fig. H8). The granodiorite is medium grained, contains andesine, quartz, orthoclase, augite, hypersthene, and sparse biotite and hornblende, and has hypidiomorphic-granular texture. Potassium-argon (biotite) and fission-track (apatite) ages of the Davidson Granodiorite suggest that it was emplaced about 17 Ma (M.L. Silberman and R.P. Ashley, unpub. data, 1976). The Davidson Granodiorite is important because it is the only known intrusion in the Virginia Range that has a texture suggesting relatively deep, but still probably hypabyssal, emplacement. Although it is apparently older than the Kate Peak Formation, its composition is similar to that of dacitic rocks in the Kate Peak.

The Alta Formation, American Ravine Andesite Porphyry, Davidson Granodiorite, and Kate Peak Formation are products of regional Miocene igneous activity that spanned at least 5 m.y. All show effects of hydrothermal alteration at many localities. The dominantly intermediate calc-alkalic volcanic pile of the Virginia Range is similar to other volcanic piles in western Nevada that host large epithermal precious-metal deposits (Silberman and others, 1976).

The Alta Formation is overlain by the Knickerbocker Andesite on Basalt Hill, southwest of Silver City (Calkins, 1944). This andesite is virtually unaltered and contains phenocrysts of olivine, plagioclase, augite, and hypersthene. Calkins (1944) considered the Knickerbocker Andesite as younger than the Kate Peak Formation. This age difference was substantiated by Thompson (1956), who stated that dikes of the Knickerbocker cut the Kate Peak.

The McClellan Peak Olivine Basalt is the youngest volcanic unit in the Comstock Lode district. This basalt was originally called American Flat Basalt (Gianella, 1936) but was renamed by Thompson (1956). The unit is distinguished by abundant olivine and augite phenocrysts in a groundmass of plagioclase laths. Doell and others (1966) reported a K-Ar age (on whole rock) of 1.14 Ma for a small remnant of the basalt near Silver City (table H1).

Structure

Pre-Tertiary metavolcanic and metasedimentary rocks are intricately sheared and folded (Calkins, 1944). Cretaceous plutonic rocks are apparently unsheared, suggesting that they were emplaced near the end of Mesozoic deformation that affected the Sierra Nevada region (Calkins, 1944; Thompson, 1956).

Thompson (1956) indicated that structural relief in Cenozoic rocks of the Virginia City quadrangle was caused by a combination of normal faulting, tilting, and warping. The dominant structural features in the Comstock Lode district are the Comstock, Silver City, and Occidental faults (fig. H8), of which the Comstock fault is the largest. The Comstock fault displaces both the Alta and Kate Peak Formations, and is at least 11 km long. Gianella (1936) estimated displacement along the Comstock fault as 1,050 m, based on the relative positions of the locally occurring water-laid strata of the Sutro Member of the Alta Formation. However, Thompson (1956) revised this estimate to 760 m by considering the eastward dips in the footwall strata.

The Comstock fault strikes northeast and is intersected by the northwest-striking Silver City fault just south of Gold Hill (fig. H8). Both faults dip about 45° E. Thompson (1956) noted that this dip is low for normal faults and attributed it to later tilting of the fault planes and of the enclosing strata. Reid (1905) stated that the Silver City and Comstock faults are separate structures and that the Silver City fault is younger than the Comstock fault. Calkins (1944) agreed with Reid's interpretation and suggested that the Comstock fault continues around the western edge of American Flat, southwest of its intersection with the Silver City fault (fig. H8).

Cross faults disrupt the Silver City fault and the southwestern part of the Comstock fault near American Flat (fig. H8). Reid (1905) described cross faults cutting the Virginia City segment of the Comstock Lode. He argued that movement along these faults was responsible for creating the gashes in the hanging wall of the Comstock fault that later served as deposition sites for the great bonanzas. Neither Calkin's (1944) nor Thompson and White's (1964) maps place emphasis on these cross faults. However, continued work in the area suggests that cross faults do in fact exist and that some movement along them may be quite recent (D. M. Hudson, oral commun., 1983).

Alteration

Wallrock alteration in the Comstock Lode district has been considered contemporaneous with vein mineralization (Becker, 1882; Gianella, 1936). Hydrothermal alteration typical of the Comstock Lode district is not limited to the district, however, but extends throughout the Virginia Range (Whitebread, 1976). The lower part of the Kate Peak Formation and older rocks show effects of several types of hypogene alteration, which include propylitic, argillic, and silicic alteration. Zoning patterns defined by different types of alteration are controlled largely by permeable zones, which include faults, fractures, and various fragmental units in volcanic rocks.

Propylitized rocks form a zone of weak hypogene alteration that is transitional to both unaltered and intensely altered rocks. The most characteristic propylitic mineral assemblage is the albite-epidote-chlorite-calcite assemblage (Coats, 1940). Montmorillonite and pyrite are also common, and zeolites are locally abundant. Epidote, however, is not everywhere present. Whitebread (1976) described the following propylitic assemblages in detail: (1) chlorite-montmorillonite-calcite-quartz, (2) montmorillonite-chlorite-calcite-quartz-pyrite, and (3) montmorillonite-calcite-quartz-pyrite. The presence of calcite distinguishes montmorillonite-bearing propylitized rocks from argillized rocks. Contacts are gradational between these propylitic assemblages. Whitebread's work showed that montmorillonite-bearing propylitized rocks are common in the Virginia Range north of and including the Comstock Lode district.

Mineral assemblages in the argillic zone described by Whitebread (1976) include (1) montmorillonite-cristobalite-pyrite, (2) illite-montmorillonite (mixed layer)-quartz-pyrite, and (3) kaolinite-quartz-pyrite. Rocks of the montmorillonite-cristobalite-pyrite argillic assemblage locally contain heulandite or clinoptilolite in the groundmass and along fractures. Original rock textures are preserved generally, and plagioclase and biotite commonly have escaped alteration. Rocks of the illite-montmorillonite (mixed layer)-quartz-pyrite argillic assemblage are more intensely altered. For example, plagioclase and ferromagnesian minerals are mostly obliterated. These rocks may also contain montmorillonite and illite (not mixed layer), kaolinite, mixed layer montmorillonite-chlorite, and chlorite. Rocks of the kaolinite-quartz-pyrite argillic assemblage surround alunitized rocks, and separate them from rocks with montmorillonite-bearing argillic assemblages. The groundmass and phenocrysts of rocks in the kaolinite-quartz-pyrite assemblage are completely altered to kaolinite and quartz. In addition, the texture of the original rock is seldom discernible.

Silicified rocks occur as two mineral assemblages: (1) alunite-quartz-pyrite, and (2) alunite-jarosite-quartz-pyrite (Whitebread, 1976). Alunite preferentially has replaced feldspar phenocrysts. Silicified zones formed along fractures and represent the most intensely altered rocks in the hydrothermally altered areas. These rocks form resistant craggy outcrops owing to the abundance of quartz. D.M. Hudson (written commun., 1983) noted that in the hanging wall of the Comstock Lode, alunite-quartz altered rocks are zoned outward to an assemblage of quartz and kaolinite, pyrophyllite, and (or) diaspore that is followed outward by an assemblage of quartz and sericite, illite, and (or) montmorillonite, ending in propylitized rocks.

Mineralogical changes associated with supergene alteration were generally minimal in silicified and argillized rocks, even though most such rocks contain pyrite that has been oxidized to limonite. Rocks with progressively weaker hypogene alteration effects show progressively greater supergene changes. However, the exact proportion of hypogene minerals versus supergene minerals is ambiguous. Whitebread (1976) noted that pyrite-bearing propylitized rocks have been affected by supergene alteration and are thus converted to bleached rocks with the quartz-montmorillonite-illite assemblage, an assemblage that is indistinguishable from oxidized argillized rocks.

Age Relations

Radiometric age data are provided in table H1 for vein material from the Comstock and Occidental faults, along with data for volcanic rocks in the Comstock Lode district. Adularia from the Comstock fault yielded a K-Ar age of 13.7±0.4 Ma (Whitebread, 1976). This age is bracketed by the Kate Peak Formation (see table H1). Likewise, adularia from the Occidental vein is dated at 12.8±0 4 Ma (Whitebread, 1976). This age also falls within the age range for the Kate Peak Formation. These data indicate contemporaneous volcanism, faulting, and mineralization during the Miocene.

ORE DEPOSITS

Ore bodies associated with the Comstock fault were grouped into three spatial zones by Becker (1882): the Bonanza, Central, and Gold Hill zones. The distribution of ore bodies in these zones is shown in the longitudinal section in figure H10. The ore bodies are tabular along strike but appear as irregularly shaped bodies when projected onto the longitudinal section. The great bonanza of the Consolidated Virginia mine formed in an open fracture in the hanging wall of the Comstock fault whereas other bonanzas occurred along the Comstock fault (fig. H11).

Becker (1882) described the great bonanza of the Consolidated Virginia and California mines as a group of three ore bodies. Each ore body was composed of crushed quartz and fragments of wallrock. These crushed zones were cut by rare narrow veins containing stephanite. The whole mass of crushed quartz carried argentite and gold with a moderate amount of pyrite. Clays were present but apparently were not prominent in the Consolidated Virginia and California bonanza.

Bastin (1922) noted that the Comstock ore minerals are fine grained; metallic minerals are commonly less than 1 mm in diameter but do attain diameters of 5–10 mm. The ores have a granular texture (the sugar quartz of Becker, 1882), and "banding" (layering) is rare. The most abundant minerals in nearly all specimens studied by Bastin (1922) were quartz, sphalerite, galena, chalcopyrite, and pyrite. An approximate paragenetic sequence, based on Bastin's (1922) study, shows that quartz and sphalerite crystallized throughout the period

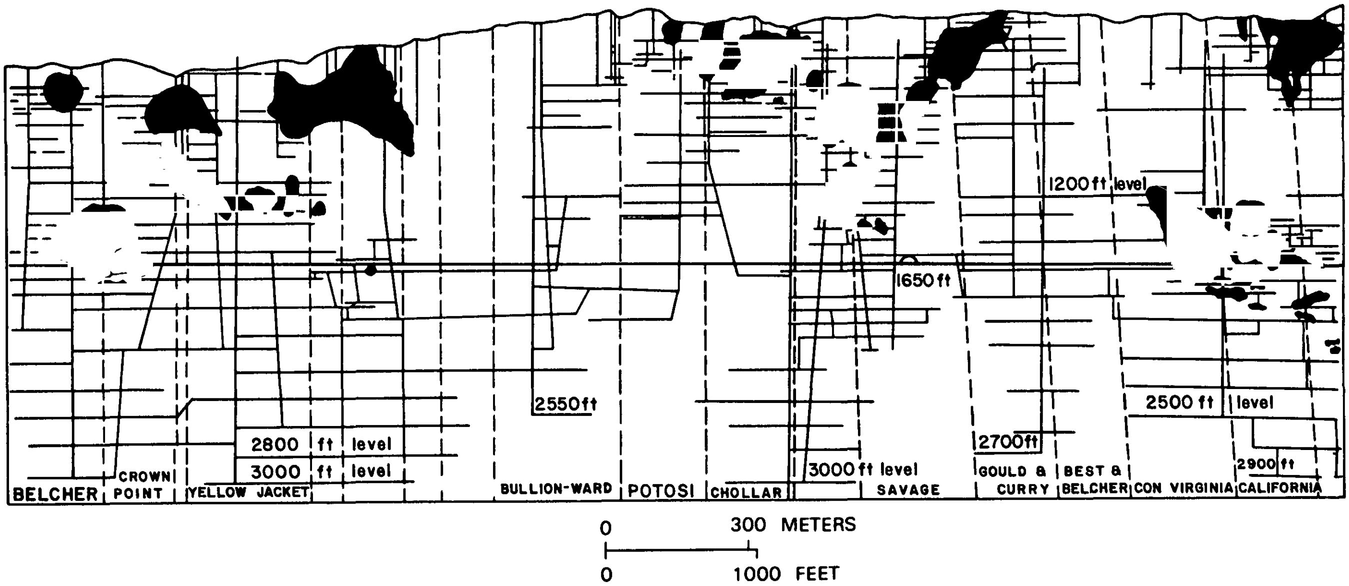


Figure H10. Projection of major ore bodies in the Comstock Lode onto the plane of the Comstock fault. Underground mine workings are simplified The dashed vertical lines separate different mines along the Comstock Lode Modified from Becker (1882).

of mineralization. Calcite, galena, chalcopyrite, pyrite, argentite, gold, electrum, and polybasite all occur interstitially between quartz and sphalerite. Aguilarite, a silver selenide, replaced the base-metal sulfide minerals and was, in turn, replaced by argentite, electrum, and stephanite (Coats, 1936).

Supergene enrichment has been minor, but it has been recognized in the shallow parts of the lode (Bastin, 1922). Supergene minerals include covellite and chalcocite that replaced chalcopyrite, and native silver that replaced argentite. Supergene enrichment occurred only to a depth of about 130 m, and Bastin (1922) concluded that such enrichment had a negligible effect on tenor of the ore.

Taylor (1973) collected samples of andesite from the Alta Formation and Davidson Granodiorite along a traverse parallel to the Comstock fault and analyzed these, and three samples of vein quartz, for their oxygen isotope compositions. His data show that the rocks are locally depleted in ^{18}O. Taylor (1973) concluded that the hydrothermal fluids in the Comstock Lode district contained a major component of meteoric water. O'Neil and Silberman (1974) determined the isotopic composition of hydrogen and oxygen in vein quartz from the Savage, Dayton, and Consolidated Virginia mines. They obtained a sample from the 1200 level of the Consolidated Virginia high-grade bonanza that contained primary sulfide minerals, sulfosalts, and quartz. The quartz in this sample contained both primary and secondary fluid inclusions The primary inclusions had filling temperatures of 295 °C with 3.1 weight percent equivalent NaCl, whereas the secondary inclusions had a filling temperature range of 240–258 °C with 2.7–3.6 weight percent equivalent NaCl. This sample also had substantially different isotope values (D = –68.5; ^{18}O = +8.9 per mil) compared to other samples (D = –87.5 to –132.5 per mil; ^{18}O = +1.1 to +3.2 per mil). O'Neil and Silberman (1974) concluded that the deep sample from the Consolidated Virginia formed from water that contained a large magmatic component (about 80 percent), whereas other samples from shallow parts of the system were dominated by meteoric water.

There are no published systematic geochemical studies of the Comstock Lode ore deposits. However, Cornwall and others (1967) and Whitebread (1976) examined surface rock and soil samples in the Virginia City quadrangle. The results of these studies showed that gold and silver anomalies delineate areas of known major production. On the other hand, mercury anomalies occur away from known deposits. Cornwall and others (1967) suggested that mercury is zoned both vertically and laterally away from silver. Whitebread (1976) noted that, although arsenic and mercury show the distribution noted by Cornwall and others (1967), highly anomalous values are limited to fracture fillings and quartz veins in the hanging wall of the Comstock Lode. Thus, Whitebread (1976) concluded that ore-related trace elements are most abundant near productive areas where they are localized along structures but not dispersed far into country rock.

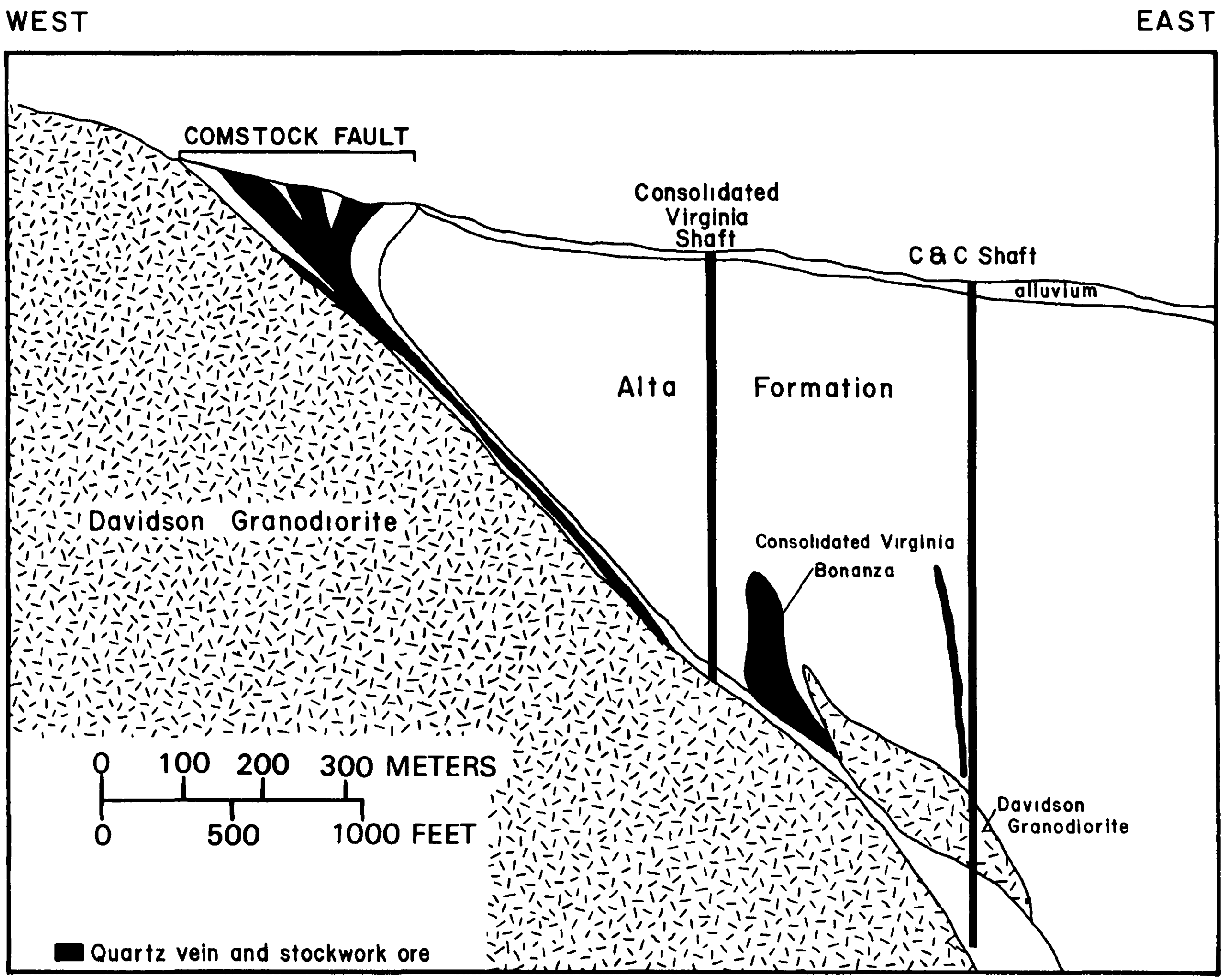


Figure H11. Schematic cross section of the Comstock Lode at the Consolidated Virginia bonanza. Modified from Becker (1882) and Reid (1905).

CONCLUSIONS

The Comstock Lode silver-gold deposits are classified as epithermal bonanza deposits in quartz veins hosted by volcanic rocks. This type of deposit was exploited in many of the famous early mining camps of western North America including: Tonopah, Nevada (Spurr, 1905), Oatman, Arizona (Ransome, 1923), and Creede, Colorado (Emmons and Larsen, 1913, 1923) in the United States; and Tayoltita, Durango (Smith and others, 1982), Pachuca-Real del Monte, Hidalgo (Geyne and others, 1963), and Guanajuato, Guanajuato (Wandke and Martinez, 1928) in Mexico. Most of these deposits have low gold.silver ratios. For example, Comstock and Tonopah have Au:Ag=1:40 and 1:100, respectively. Gold-rich deposits include Oatman, Arizona (Au·Ag=1:1), and Hayden Hill, California (Au:Ag=3:1) (Nolan, 1933).

Features of the Comstock Lode similar to those of other epithermal deposits in Tertiary volcanic terranes of western North America include: (1) localization along a normal fault, (2) significant hanging wall mineralization, (3) temporal and spatial association with igneous activity, (4) both precious- and base-metal sulfide mineralization, and (5) fluid inclusions indicating a temperature range of 200–300 °C for meteoric-dominated hydrothermal solutions in the shallow parts of the deposit. On the other hand, the Comstock Lode ore bodies have a prominent characteristic that is different from other epithermal deposits: the bonanzas are zones of crushed quartz with sulfide minerals that show little of the continuous banding typical of other deposits.

Genesis of the Comstock Lode can be considered in terms of geothermal models (Buchanan, 1981; Berger and Eimon, 1982; Henley and Ellis, 1983). These models suggest that heat, provided by igneous activity, drives a convecting hot-water system. The hot water flows through conduits provided by fractures, and minerals are deposited in the open spaces of fractures and breccias. Albers and Kleinhampl (1970) suggested that intrusion and uplift in the volcanic vent area of the Comstock Lode district were responsible for the heat and structural conduits necessary for ore deposition in the Comstock Lode. The ages of the volcanic and intrusive rocks in the district bracket ages of vein minerals (table H1). These relations indicate that convecting geothermal fluid, driven by igneous activity, flowed through the Comstock structures. Vein minerals were deposited in the structures in response to changing physical and chemical parameters, including cooling, boiling, and changes in fluid chemistry owing to wallrock/water interactions.

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Note added in press:

A recently published article by Vikre (1990) provides excellent extensive details from fluid-mineral relations in the Comstock Lode that greatly clarify the depositional history of the silver-gold ores

Vikre, P G, 1990, Fluid-mineral relations in the Comstock Lode Economic Geology, v 84, p 1574–1613

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