

Geology and Geochemistry of the Barneys Canyon Gold Deposit, Utah

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Abstract

Barneys Canyon is a sediment-hosted, disseminated gold deposit located 7 km from the large, gold-rich, Bingham porphyry copper deposit. Host rocks for gold mineralization are the Permian Park City dolomite and siltstone and the Kirkman-Diamond Creek sandstone. The gold deposit is approximately 430 m long, 370 m wide, up to 90 m thick and contains 8.5 million metric tons (t) of reserves averaging 1.6 g/t gold. Intrusive igneous rocks are conspicuously absent. The gold deposit is located on the northern flank of the northeast-trending Copperton anticline, an overturned box fold. A small east-striking, south-dipping thrust fault, the Barneys Canyon thrust fault, with 200 m displacement, repeats the Park City Formation, and north-south-striking steep normal faults form a graben in which the gold deposit is located. The Barneys Canyon thrust fault predates mineralization and the Phosphate normal fault postdates mineralization.

Alteration of the host rocks is similar to other sediment-hosted, disseminated gold deposits but at lower alteration mineral abundances. Kaolinite and illite comprise less than 10 percent of the altered rock. Silicification is minor, barite is rare, and pyrite and marcasite are common, but not abundant. Trace As, Sb, Hg, Tl, and Ba show pronounced increase.

K/Ar age determinations on vein illite from ore grade (1.5 ppm gold) bedding-plane gougelike material yield Jurassic ages.

Fluid inclusion measurements from barite and jasperoid show a mean salinity of 1.7 wt percent NaCl equiv and homogenization temperatures of 130° to 393°C with weak modes at 350 and 230°C, suggesting that two fluids have interacted with rocks at Barneys Canyon. Kaolinite-bearing assemblages formed below 280°C. Pressure correction at hydrostatic pressure is 12°C.

Fluid inclusion measurements, geochronology, structure, trace elements, and distance from Bingham are inconsistent with genesis of the Barneys Canyon deposit as part of the Bingham porphyry copper system. It is unlikely that genesis of the Barneys Canyon gold deposit involved igneous activity.

Introduction

DISSEMINATED gold deposits of the western United States occur in volcanic rocks where they are related to volcanic or subvolcanic activity, near porphyry copper deposits where they are related to the porphyry system, or in sedimentary rocks such as Carlin-type deposits. These Carlin-type deposits typically occur in silty dolomite and limestone (Bagby and Berger, 1985), but siltstone, sandstone, conglomerate, argillite, and interbedded chert and shale host rocks are also known (Percival et al., 1988). Some Carlin-type deposits are spatially associated with felsic to intermediate dikes and other intrusive igneous rocks, but genetic relationships to igneous activity are not clear (Percival et al., 1988).

Recent work on Carlin-type deposits has focused on the geochemical characteristics of the ore-forming fluid using fluid inclusion, isotopic, and geochemical studies (Kuehn, 1989; Bagby and Cline, 1991; Hofstra et al., 1991; Kuehn and Rose, 1995). These studies indicate depths of formation greater than 1 km, precluding formation in the "epithermal" environment. Age determinations on illites from sedimentary and igneous rocks at Mercur and Post-Betze, although inconclusive, support a Mesozoic age of formation and not an age related to Cenozoic igneous activity (Wilson and Parry, 1990; Arehart et al., 1993). Maher et al. (1993) used spatial associations with igneous rocks and geologic arguments to limit the age of sediment-hosted gold mineralization in the Battle Mountain-Eureka trend, Nevada, to pre-middle Miocene. They related the diverse types of gold mineralization to magmatic pulses during the Mesozoic and Cenozoic; however, they concluded that the gold source was not igneous activity.

Based primarily on spatial associations, Sillitoe and Bonham (1990) proposed that Carlin-type deposits were distal products of porphyry-related hydrothermal magmatic systems. Sillitoe and Bonham's (1990) use of Barneys Canyon as an example of a porphyry-related deposit was solely based on its elevated arsenic content and its spatial relationship to the Bingham porphyry copper deposit. Because of relatively simple mineralogy, complex structural settings, uncertain igneous associations, and ambiguous chronology, the genesis of Carlin-type deposits is uncertain and controversial.

The Barneys Canyon gold deposit is located in the Bingham mining district 7 km from the large, gold-rich Bingham porphyry copper deposit, but occurs in sedimentary host rocks with many similarities to Carlin-type deposits. The gold deposit is approximately 430 m long, 370 m wide, up to 90 m thick and contains 8.5 million metric tons of reserves averaging 1.6 g/t with a cut off of 0.5 g/t (Gunter et al., 1990).

The objective of this study is to describe the geologic and geochemical characteristics of the Barneys Canyon gold deposit. The tectonic setting in the Oquirrh Mountains and Bingham mining district is crucial to determining the relationship with Bingham and to constrain the age and genesis of the deposit. The results of detailed mapping, petrography, fluid inclusion studies, clay mineralogy, and alteration studies are combined with K/Ar ages into a geologic description of the Barneys Canyon gold deposit.

Methods

The stratigraphic and tectonic framework of the Barneys Canyon gold deposit was investigated from mapping geology

and measuring stratigraphic sections. The northeast Bingham mining district was mapped at a scale of 1:6,000, and the Barneys Canyon mine was mapped during excavation at a scale of 1:600, resulting in a geologic map at a scale of 1:1,200. The exposed stratigraphy in the mine was measured and described. Stratigraphic and structural ore controls were determined from contoured assay maps and sampling. Hydrothermal alteration was mapped in mine pit exposures to the top of the orebody and in drill core through the full extent of the orebody. Detailed petrographic, X-ray diffraction, electron microprobe, and bulk chemical analysis of core were used to identify alteration and sulfide minerals and paragenesis.

Alteration was mapped in the field using a continuous mapping method similar to that used by Bakken and Einaudi (1986) at the Carlin deposit. The alteration was mapped on five successive benches and an upper haul road during December 1990 and January 1991. Visual estimates of the alteration intensity were supplemented by thin section and X-ray diffraction analysis. The paragenesis of hydrothermal minerals was determined from study of 150 thin sections. Sulfide mineralogy was determined from unoxidized sulfidic ore from 19 core holes, in hand samples and polished thin sections.

Illite-kaolinite ratios and illite crystallinity were determined from X-ray diffraction analysis of $-2\text{-}\mu\text{m}$ -size separates from 129 composite samples. Twenty-foot composites were prepared from 5-ft reverse circulation pulps or core splits through the gold deposit. A 20-ft composite was analyzed from above, within, and below the orebody. Fifty-foot composites were analyzed from holes peripheral to the orebody. Kaolinite and illite peak areas were calculated from X-ray diffraction patterns by multiplying the height of the peak by the width at half height. The illite/kaolinite was calculated by dividing the normalized illite peak area by the kaolinite peak area and then plotted on the cross sections. Illite crystallinity was determined from the illite peak width following the method of Kisch (1980).

The same drill hole composites were analyzed for minor element concentrations using ICP-MS by Chemical and Mineralogical Services (CMS), Inc., of Salt Lake City, Utah.

Trace elements, major elements, mineralogy, illite/kaolinite, illite crystallinity, petrography, and fluid inclusion analyses were determined from mineralized core samples in core hole BC-91, drilled in the heart of the orebody. Mineralized samples from drill hole BC-91 were analyzed for major elements by Bondar-Clegg, Vancouver, B.C., using borate fusion and plasma emission spectroscopy. The modal mineralogy of these samples was calculated using a linear, least-squares fit of rock composition, mineral composition, and mineral abundance (Parry et al., 1980).

Geologic Setting

The Barneys Canyon gold deposit occurs in the northeast part of the Bingham mining district in the central Oquirrh Mountains (Fig. 1). The complex geologic history of the region is reviewed to provide the geologic setting and to constrain the age of structures and mineralization at Barneys Canyon.

The Oquirrh Mountains occur within the area of overlap between the Basin and Range extensional province and the Cordilleran fold and thrust belt (Armstrong, 1968; Burchfiel

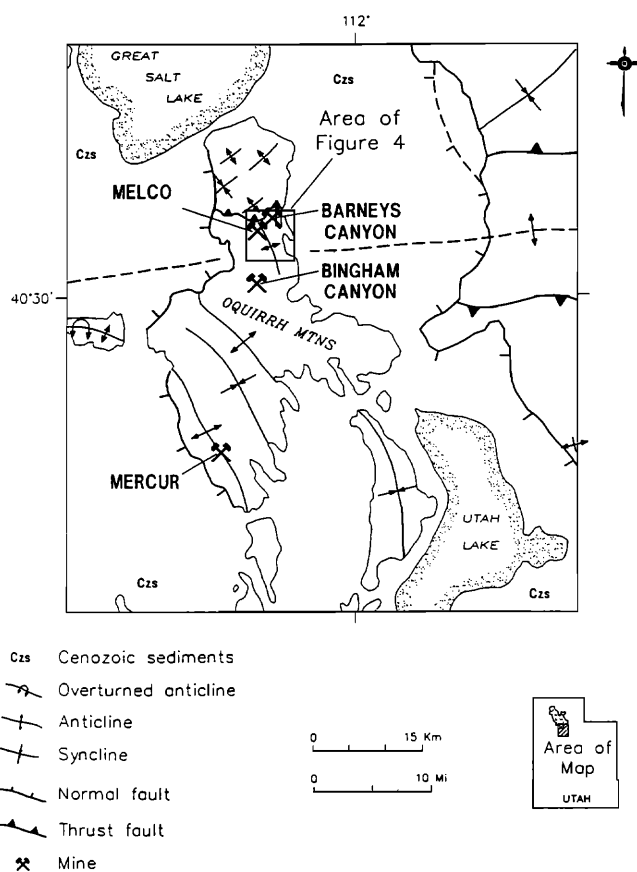


FIG. 1. Structure and location map of north-central Utah after Critten-den (1976), Moore and Sorensen (1979), Tooker (1983), Smith and Bruhn (1984), and Yonkee (1990). Mountain ranges contain Precambrian, Paleozoic, and Mesozoic strata. Area of Figure 4 and locations of producing mines are shown.

and Davis, 1975; Presnell, 1992a). The deformational history of the Oquirrh Mountains includes a Late Jurassic compressional event (Presnell, 1992a; Presnell and Parry, 1995), Cretaceous to Early Tertiary compressional events of the Sevier orogeny (Tooker, 1983), and Tertiary extensional events.

The Oquirrh Mountains are the easternmost range in the Basin and Range province, straddle the Uinta arch, and are allochthonous. The entire range is underlain by the east-vergent Oquirrh thrust (Presnell, 1992b). The western projection of the Uinta arch segments the Oquirrh Mountains into southern, central, and northern domains (Presnell, 1992b). The central domain contains the Bingham mining district and the Barneys Canyon gold deposit and is structurally complex because of thrust-belt foreland interaction where the east-west-trending Uinta arch interacted with the east-vergent thrust belt (Bradley and Bruhn, 1988). The central Oquirrh Mountains expose 7.8 km of Pennsylvanian through Permian strata deposited in the fault-controlled, northwest-trending Oquirrh basin. These rocks are part of the Bingham sequence of Swenson (1975a) and are separated from the age-equivalent Rogers Canyon sequence of Tooker and Roberts (1970) by the North Oquirrh thrust (Presnell, 1992a; Fig. 2). The Bingham sequence consists of the Pennsylvanian Oquirrh Group and Lower Permian clastic and carbonate rocks.

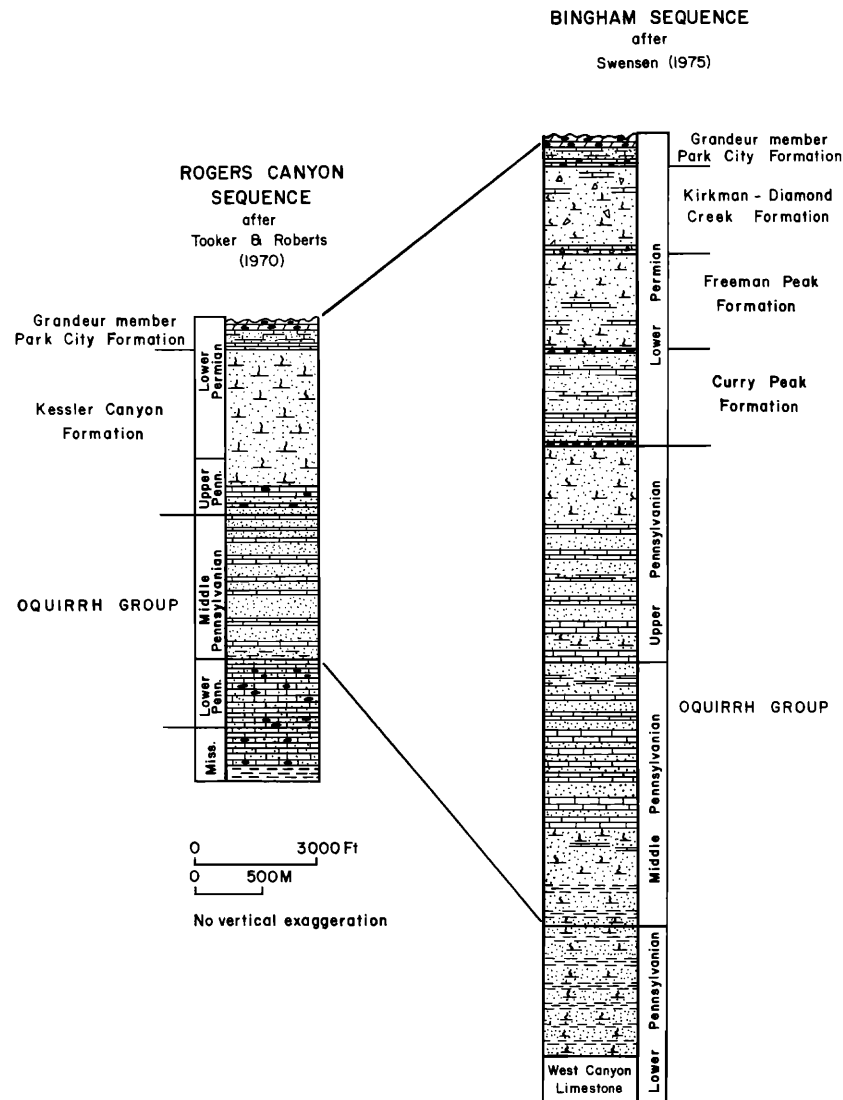


FIG. 2. Stratigraphic columns of the Rogers Canyon sequence and the Bingham sequence after Tooker and Roberts (1970) and Swensen (1975a).

The Bingham mining district includes a large, gold-rich, porphyry copper deposit and spatially associated lead-zinc-silver deposits that are related to the Bingham intrusive igneous complex. The emplacement of igneous rocks at Bingham was controlled by northwest-trending fold axes and northeast-trending faults (James et al., 1961; Moore, 1973; Lanier et al., 1978). The Bingham district contains two major northwest-trending folds, the Bingham syncline and the Copperton anticline, which are separated by the north-vergent Midas thrust fault (Fig. 3). The Bingham stock intruded the axis of the Bingham syncline (Rubright and Hart, 1968).

The majority of the mineralized veins and lodes in the Bingham district have a northeast strike; northwest-trending faults are generally barren and formed after initial movement of the northeast-trending faults (Boutwell, 1905). Mineralization of the northeast-trending faults requires that they were active before or during mineralization which occurred at 39.8 to 38.8 Ma (Warnaars et al., 1978). Presnell (1992a and b)

has proposed that the northeast-trending extensional faults are late Eocene, and the northwest-trending faults are Miocene to Recent. The Bingham stock was emplaced during the late Eocene extensional tectonic event and was partly controlled by preexisting compressional structures.

At Barneys Canyon, Pennsylvanian and Permian strata, dominated by the upper portion of the Bingham sequence (Fig. 2), are deformed by the Copperton anticline, the North Oquirrh thrust, and north- and northeast-trending normal and strike-slip faults (Fig. 4). Strata of the Bingham sequence are overthrust by the Rogers Canyon sequence on to the North Oquirrh thrust. The Copperton anticline is an overturned box fold that is a second mode fault bend fold formed from ramping of an east-verging thrust (Presnell and Parry, 1995). The fold consists of three structural domains: a west limb, a crestal interlimb domain, and an eastern overturned limb. The Barneys Canyon deposit is on the northern flank of the Copperton anticline within the crestal interlimb domain.

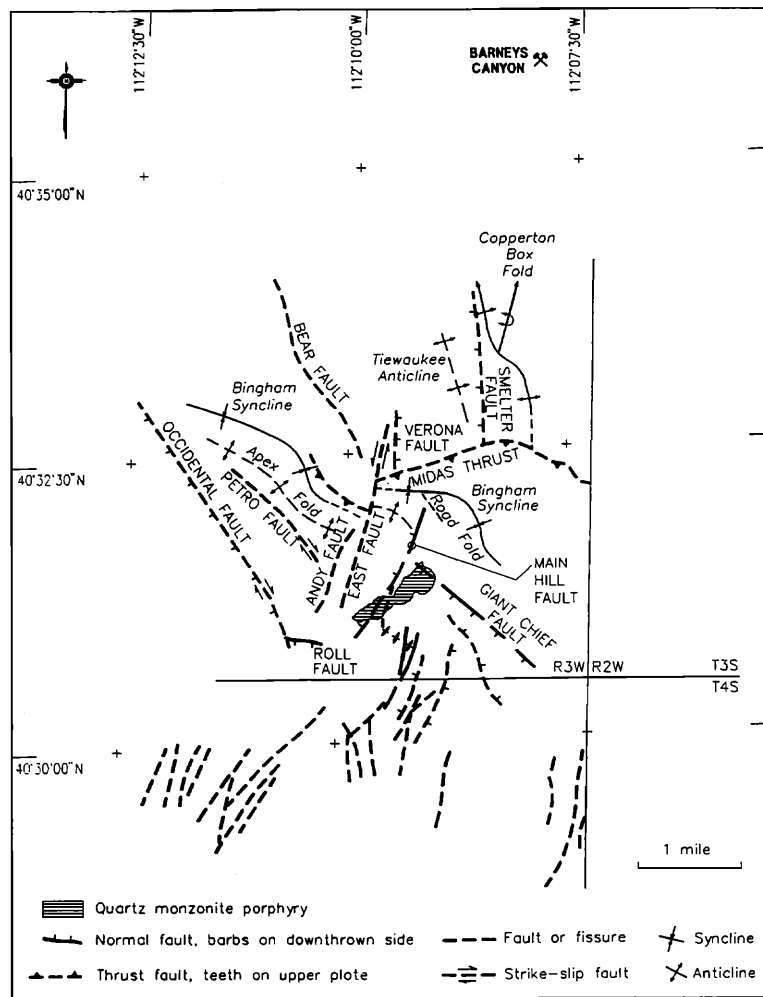


FIG. 3. Generalized tectonic map of the central Bingham mining district after Lanier et al. (1978) and unpublished Kennecott 1989 and 1990 pit maps. The Barneys Canyon mine occurs 5.6 km to the north of the Midas thrust.

Several normal faults have been recognized in the Copperton anticline in underground mapping of the north ore shoot of the Bingham mine. The Smelter fault strikes north, dips 70° and has 610 m of down to the west offset at its southern end near Bingham Canyon. The fault does not cut the Midas thrust and appears to be older (Fig. 3). A Tertiary dike related to Bingham cuts the Smelter fault and shows no evidence of offset (J. Reid, pers. commun., 1991). A north-trending fault which cuts the Freeman Peak-Curry Peak contact in the midst of the Copperton anticline has been correlated with the Smelter fault and extends the fault into the Barneys Canyon area (Fig. 4). The Verona fault is a north-striking, east-dipping normal fault with some sinistral movement on the west limb of the Copperton anticline (Figs. 3 and 4).

The Melco Gold Deposit

The Bingham mining district also contains the Melco sediment-hosted gold deposit. The deposit occurs on the west hinge of the Copperton anticline. Gold mineralization is localized along steeply dipping, N 55° E-striking fault breccias in

the Kirkman-Diamond Creek formation at the contact with the underlying Freeman Peak formation (Gunter et al., 1990). Alteration consists of argillization, weak silicification, and hydrothermal epidote. Sulfides include pyrite, marcasite, orpiment, and realgar. The Melco orebody has the following trace element abundances based on the mean of seven oxidized samples with greater than 0.55 ppm gold: Ag (0.06 ppm), As (1,520 ppm), Hg (1.89 ppm), Sb (65.35 ppm), Ba (3648 ppm), Tl (120 ppm), Se (14.11 ppm), Te (10.68 ppm), Cu (8 ppm), Pb (11.8 ppm), Zn (28.8 ppm), and Mo (3.98 ppm). The genetic relationship between the Barneys Canyon and Melco gold deposits is not known and will not be addressed in this study.

Geology of the Barneys Canyon Gold Deposit

The gold deposit at Barneys Canyon is within the Permian Park City and Kirkman-Diamond Creek formations that are part of the Bingham sequence. Outside of the Oquirrh Mountains, the Kirkman and Diamond Creek formations are distinct, but in some parts of the Oquirrh Mountains it is not possible to separate them clearly. We have combined them

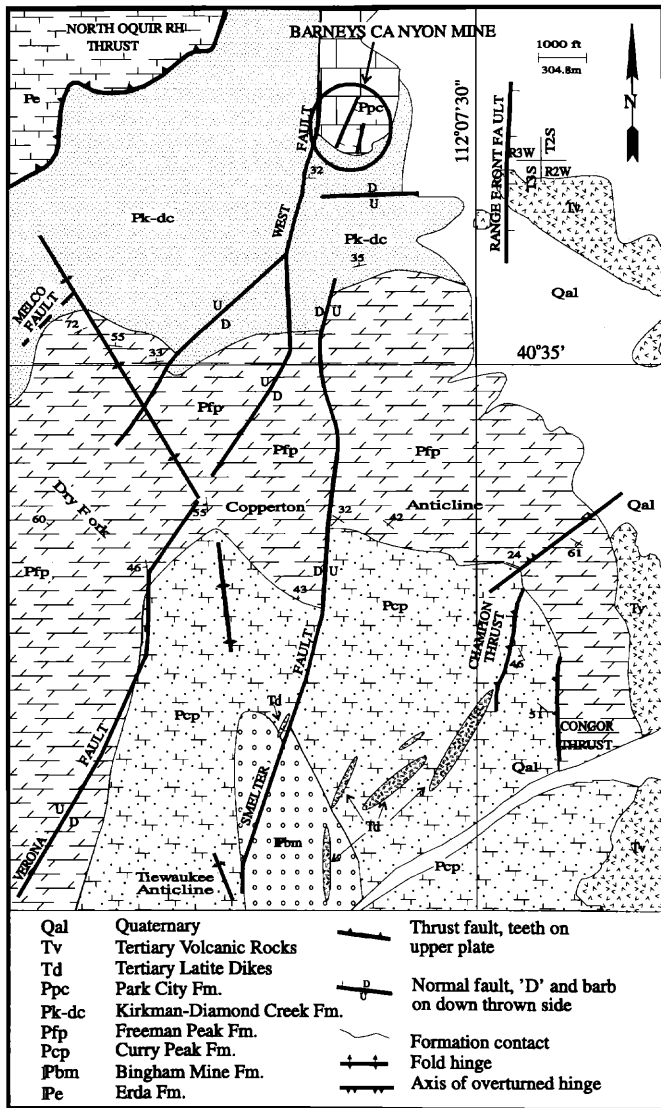


FIG. 4. Simplified geologic map of the northeast Bingham mining district. Modified in part from Swensen (1975b).

into the Kirkman-Diamond Creek formation as suggested by Swenson (1975a). The mine stratigraphy consists of an upper dolomitic siltstone unit, a medial dolomite unit, and a basal sandstone (Table 1). The mine stratigraphy has been repeated by the Barneys Canyon thrust and offset by several north- to north-northeast-striking normal faults and some minor east-west-striking normal faults. Interbedded gouges are common within the mine stratigraphy and are a result of flexural slip during growth of the Copperton anticline (Presnell and Parry, 1995).

The Barneys Canyon thrust cuts up section from the main host dolomite to the northeast where the intercalated cherty dolomite and siltstone are repeated (Fig. 5A). In the mine, the exposure of the thrust is a thrust ramp. Hanging-wall and footwall cutoffs of the phosphate horizon give a minimum displacement of 213 m. Kinematic indicators show that hanging-wall movement is southwest to northeast (Presnell, 1992a). The thrust and the stratigraphy dip 20° to 40° to the

northeast and strike 300° to 310°. A ramping thrust generally dips 20° opposite the direction of transport (Serra, 1977) indicating that the Barneys Canyon thrust was rotated 55° northeast during growth of the Copperton anticline and must predate the Copperton anticline (Presnell, 1992a).

North- to north-northeast-striking normal faults are, from west to east, the West fault, the Phosphate fault, and the East fault (Fig. 5A). These faults form a graben in which the gold deposit is located. Several minor (<1 m of offset) east-west-striking faults occur within the ore zone, and the deposit is bounded on the north and south by east-west-striking faults.

Within the mine area, the West fault has a minimum displacement of 120 m. The central portion of the mapped exposure of the West fault is a single plane that strikes 356°, dips 68° east, and juxtaposes sandstone against dolomite (Fig. 5A). To the north it becomes a 3- to 6-m zone of bleaching and alteration. To the south it branches into two strands. The West fault may correlate with the Verona fault and may be related to the Copperton anticline. North of Barneys Canyon, the West fault cuts a Tertiary conglomerate, but there is evidence of recurrent movement on the fault which does not limit its age to the Tertiary.

The Phosphate fault in the central part of the mine strikes 50° and dips 70° to the east (Fig. 5A). Strike is more northerly at both the northern and southern ends. The fault rock consists of 4 to 6 cm of calcified breccia.

The East fault, in the east-central part of the orebody, strikes 10° to 355° and dips 55° to 70° west. Slickensides and stratigraphic offset show a maximum of 25 m of dip-slip movement which decreases to less than 6 m to the north. However, the lateral offset of map patterns requires sinistral movement (Fig. 5A) and is supported by the presence of two sets of slickensides on the 6440 level. To the east of the East fault, a normal fault (E.E. fault) was recognized within the sandstone and strikes 46° and dips 38° to 45° west.

Several north-south-striking, small displacement faults offset the 32-in. bedding-plane gouge below the bedded chert. Within the center of the deposit, several east-west-striking normal faults have less than a meter of offset. North of the deposit, an east-west-striking fault with 30 m of down to the south offset cuts the West fault and is the youngest fault in the mine area. A fault must be inferred south of the mine because the map distance between the Freeman Peak formation and the Park City formation will not accommodate the 610-m thickness of the Kirkman-Diamond Creek Formation.

Hydrothermal Mineralization

Hydrothermal minerals are less abundant at Barneys Canyon than at other sediment-hosted disseminated gold deposits. Hydrothermal effects on the host rock include decalcification, silicification, and argillization. The distribution of clay, calcite veins, silicification, Fe-Mn oxides, and barite occurrences were mappable in the pit and will be discussed in conjunction with the paragenesis. Oxidation of the deposit is almost complete, and hypogene and supergene mineralization could not be completely differentiated during mapping. For purposes of discussion we have divided our description of hydrothermal minerals into host-rock alteration and gold and sulfide mineralization.

TABLE 1. Barneys Canyons Mine Stratigraphy

Formation	Description	Thickness
Park City Formation	Intercalated cherty dolomite and siltstone: Interlayered gray cherty dolomite and tan to gray cherty siltstone	~200 ft
	Chert-pod dolomite: Dolomite, gray with tan staining abundant 2 to 3 in. chert nodules; clay gouge interbeds; some intercalated massive carbonate	~10 ft
	Oolitic dolomite: Dolomite pale to purplish gray, very oolitic and bioclastic; fetid; large fossil fragment in middle; massive bedded	16 ft
	Dolomitic siltstone: Tan to purple, well-laminated with 2-in. fossil hash at top; phosphate horizon in middle; massive tan dolomitic siltstone at bottom	65 ft
	Main host dolomite: Dolomite, gray to tannish gray, sparry; abundant sponge spicules, oolites, and fossil fragments in top half; chert layers in middle, sandier toward bottom	~80 ft
	Bedded chert: Bedded chert, black; weakly calcareous dolomitic interbeds; some iron oxide staining at contacts; siliceous siltstone at base	6 ft
	32-in. gouge: Yellowish-tan gouge; noncalcareous, brecciated, streaky staining and laminated texture	
	Transition zone: Dolomite, brownish-gray, very sandy (particularly at bottom), blocky to massive bedded, very competent	~66 ft
	Contact: Tan to orange stained, streaky to banded; fine-grained sandy to silty; decalcified in part; sheared to brecciated in part	
	Upper unit: Sandstone, fine-grained bleached gray to white; noncalcareous (within mine) argillaceous with some strong iron oxide staining; bedding-plane gouge and shear zones within; rare black chert	30 ft
Kirkman-Diamond Creek formation	Lower unit: Sandstone, bleached gray-white, decalcified, argillaceous; some strong iron oxide staining along bedding planes and/or shears; abundant lenses of carbonate altered to variably calcareous red-brown finely laminated siltstone	>60 ft

Host-rock alteration

Silicification of host rock is a minor but early form of hydrothermal mineralization. Silicification textures include quartz overgrowths on detrital quartz and small jasperoids. The discovery outcrop of the deposit was a small (1×0.5 m) jasperoid located 100 m above the northwest edge of the orebody (Fig. 6A). Silicification of dolomite is also present along the sandstone-dolomite contact at the southern edge of the deposit and adjacent to the West fault. Several jasperoids were found on the 6440 level in the upper part of the main host dolomite. These jasperoids were smoky gray, veined with quartz, and sometimes brecciated. In the upper levels above the southern part of the deposit, chert is often preferentially replaced or recrystallized. These altered chert bodies appear smoky gray and are cut by quartz veins in contrast to their normal black appearance in unaltered rocks.

Decalcification of the sandstones results in removal of calcite cement. Decalcified sandstone has vertical contacts with unaltered sandstone, but within the dolomite, decalcification is lithologically controlled and alteration contacts are parallel to bedding. Decalcification of dolomite results in loose dolomite rhombs that megascopically have the appearance of sand. Decalcification is most pronounced within the main host dolomite near or adjacent to the dolomitic siltstone contact and is most intense adjacent to the West and East faults. Bioclastic oolitic units are often the most prone to decalcification.

Argillic alteration is common within the main orebody. Clay minerals are illite, kaolinite, and minor interstratified illite-smectite. Samples from drill hole BC-91 (see Fig. 5C for location) contain a trace to 3.85 percent illite in the ore and 0.37 to 4.21 percent illite in the dolomitic sandstone below the ore. The maximum kaolinite content is 1.67 percent at the bottom of the orebody (Table 2). The illite occurs in veins (Fig. 6B) and as a fine-grained weblike mesh commonly referred to as a felted texture (Fig. 6C and D). The hydrothermal illite is distinguishable from coarse ($>25 \mu\text{m}$) detrital

mica found in unaltered sandstone. Hydrothermal illite veins cut silicification and are in turn cut by calcite veins. Clay content is more pronounced in the sandstones than in the dolomites, possibly because of alteration of detrital feldspar. Clay content in the sandstones is greatest near the West and East faults. Within the dolomites in upper mine levels, clay occurs in pods and patches. Clay veins were recognized just below the thrust on the 6520 level, and dolomites adjacent to the West fault were pervasively argillized. Within the orebody, disseminated to pervasive clay-rich zones are common within the dolomites.

Zoning of illite to kaolinite ratios has been determined from X-ray diffraction analysis of 129 composite samples from two drill hole fences, as well as samples from several holes north of the deposit, all shown in Figure 7. The fences coincide with the geologic cross sections shown in Figure 5B and C.

In the northwest-southeast section, parallel to sedimentary strike (Fig. 5B), the variation in illite/kaolinite does not correspond to the orebody and is only roughly parallel to stratigraphy on the eastern side. The illite/kaolinite can be contoured with subhorizontal zones greater than 2.0 alternating with parallel zones less than 2.0 which suggests that there are alternating zones of kaolinite- and illite-rich alteration that conforms crudely to bedding. In the southwest-northeast section, parallel to sedimentary dip (Fig. 5C), illite/kaolinite variation is roughly subparallel to stratigraphy and crudely mimics the orebody outline. A zone of illite/kaolinite greater than 2.0 within the orebody outlines the orebody shape and narrows adjacent to the East fault. Below this zone and continuing below the orebody is an area that has illite/kaolinite less than 2.0. These variations suggest that the core of the orebody is illite rich and is surrounded by a zone that contains more kaolinite. Drill hole BC-91 shows an illite/kaolinite reaching a maximum value of 2.9 nearly coincident with the maximum gold concentration in the ore (Table 2). This drill hole misses the illite-rich zone where the illite/kaolinite is greater than 10.

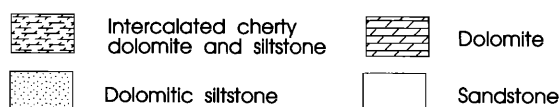
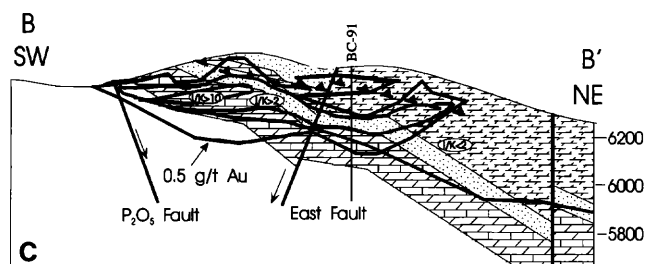
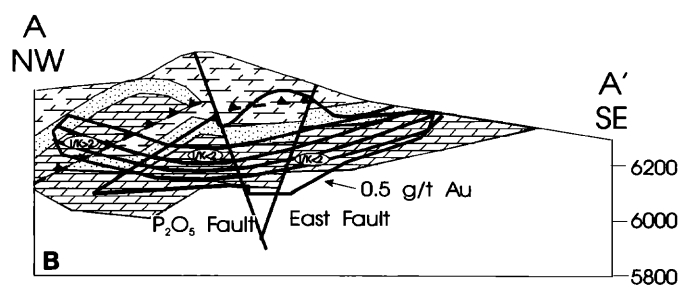
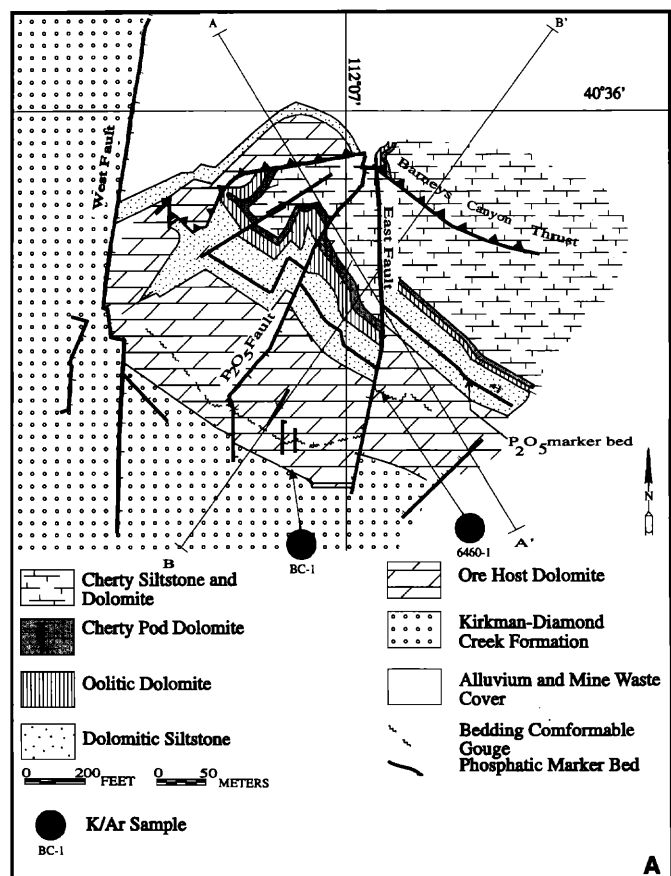


FIG. 5. Simplified geologic map (A) of the Barneys Canyon gold deposit and cross sections parallel to strike (B) and dip (C) as of December 1990. Outline of orebody in strike and dip view based on 0.016 opt Au cutoff in drill hole fences. Grayed line marks areas of illite/kaolinite that are greater than 2.

Illite crystallinity of 0.4 crudely mimics the orebody outline in strike section (Fig. 5B). The crystallinity index increases toward the core of the orebody, particularly within the dolomite adjacent to the East fault. In the dip section, the 0.4 contour outlines the orebody with inward increasing values of the crystallinity index (Fig. 5C). Samples from drill hole BC-91 have crystallinity indices of 0.13 to 0.5. The highest value of 0.5 is coincident with the highest gold value with a second high value in the dolomitic sandstone beneath the ore (Table 2). The illite crystallinity index is inversely proportional to temperature (Weaver, 1961, 1984; Kubler, 1967; Dunoyer de Segonzac, 1970; Kisch, 1980; Duba and Williams-Jones, 1983). The presence of illite with a higher crystallinity index (broader XRD peak) within the ore suggests a lower temperature or retrograde alteration within the ore.

Calcite veins crosscut silicification, kaolinite and illite, sulfides and barite, and are therefore late in the paragenesis. A halo of calcite veins occurs around and above the main orebody. Within the sandstone in the lower exposures of the mine and within the main orebody calcite veins are rare to absent, but are very common in the dolomites of the upper mine levels particularly in the upper plate main host dolomite.

Fe and Mn oxides occur as veins, pods, patches, and pervasive coloration of altered host rocks. Crosscutting relationships show the oxides to be late in the paragenesis. Some oxide pods contain incompletely oxidized sulfides. Iron oxide staining along fractures and joints is ubiquitous throughout the mine. Fe oxides occur adjacent to the West fault, and sandstones adjacent to the East fault on the 6380 level contain intense patches and pods of Fe oxide, some of which contain disseminated sulfides. The abundance of the Fe oxide veins increases towards and within the orebody. Mn oxides occur as dendrites adjacent to the West fault on the 6580 level and next to the East fault on the 6380 level. Drill core shows Mn dendrites to be very common within the orebody.

Barite is not common at Barneys Canyon. Barite veins cut argillically altered sandstone that is in turn crosscut by calcite veins. Barite veins were found within the phosphatic unit of the dolomitic siltstone on the surface at the southwestern part of the deposit. Clots of barite up to 30 cm in diameter were found on the 6580 level at the top of the upper plate, main host dolomite. Barite in thin sections from core samples occurs as isolated grains with carbonate mineral inclusions and a motheaten texture.

The lack of veins and distinct crosscutting relationships in Carlin-type deposits inhibits a thorough and detailed paragenesis. However, a skeleton paragenesis of hydrothermal alteration at Barneys Canyon consists of early silicification, medial argillization, and late barite, oxide, and calcite formation.

Gold and sulfide mineralization

Economic mineralization at the Barneys Canyon gold deposit consists of gold. The gold is believed to be micrometer to submicrometer in size and is only detectable by assay. The gold paragenesis could not be determined, but gold is tentatively assumed to be associated with sulfides that assay as ore.

Pyrite and marcasite pods and disseminations are the major sulfide minerals in the deposit. Pyrite occurs as individual grains, in veins, and in clusters. Pyrite grains are subhedral to

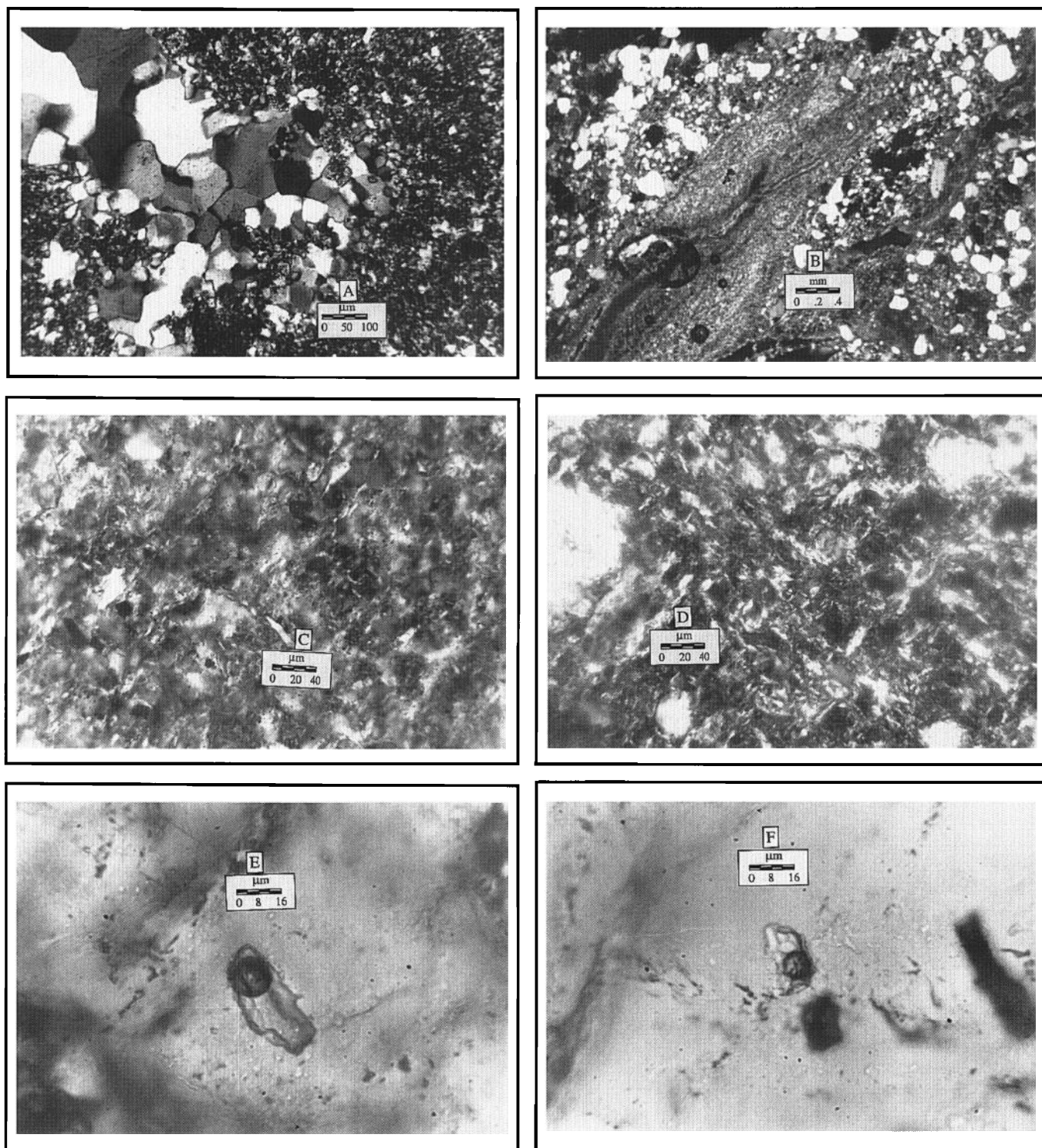


FIG. 6. Photomicrographs of jasperoid, illite, and fluid inclusions. A. Discovery jasperoid. B. Sample BC-I illite vein with calcite in the center of the vein (1.5 ppm Au). C. BC-I illite with a felted texture; illite crystals on edge oriented in different directions. D. Drill hole sample BC 91; 360-ft-depth felted illite texture in vein in siltstone (8.3 ppm Au). E. Fluid inclusion in quartz from drill hole sample BC91, 375-ft depth (1.02 ppm Au). F. Fluid inclusion in quartz from drill hole sample BC-91, 375-ft depth.

anhedral. Marcasite occurs as bladed clusters and aggregates often intergrown with pyrite. Sulfide-bearing intervals in drill core are 0.3 to 3 m thick with concentrations from 2 to 50

percent sulfide. Sulfide-rich zones occur within the orebody, but they are often barren of gold mineralization. Some of the sulfide zones contain carbon.

TABLE 2. Chemical Composition and Mineralogy of Ore Zone Samples in Core from Drill Hole BC-91

	Depth interval (ft)								
	243.5–248.5	264–269	298.5–303	308.8–313.5	342–347	357.5–362	375.5–382	387–392	397–402
Al ₂ O ₃	0.81	1.39	0.34	0.40	0.69	0.78	1.02	1.09	0.22
SiO ₂	26.0	15.7	22.7	51.0	26.2	43.9	59.0	59.4	41.2
CaO ₂	22.3	25.5	24.6	13.4	23.5	18.1	12.5	13.1	24.3
MgO	15.4	17.4	15.8	7.2	14.6	10.5	7.5	6.8	7.2
K ₂ O	0.20	0.41	0.10	0.15	0.13	0.18	0.28	0.26	0.11
Fe ₂ O ₃	0.54	0.91	0.35	0.58	0.38	0.46	0.71	0.71	0.29
CO ₂	33.1	27.1	33.5	16.2	31.4	22.8	16.6	15.6	22.6
S	0.04	0.04	0.05	0.09	0.10	0.07	0.06	0.04	0.10
Au	7.0	5.5	12.6	10.8	6.7	8.3	1.02	0.104	0.027
As	294	371	252	217	248	239	262	265	161
Bi	72	75	40	9	47	32	13	10	<2
Sb	113	135	112	45	102	93	39	38	42
Hg (ppb)	500	1,750	2,000	1,600	1,200	1,700	400	400	105
Tl	15	12	5	10	4	13	94	48	2
Cu	25	32	23	8	23	24	6	4	2
Pb	120	132	106	40	107	76	38	36	43
Zn	173	219	112	99	113	137	94	100	51
Ba	20	20	<20	50	<20	<20	<20	50	<20
W	<10	10	<10	<10	<10	29	<10	<10	<10
Te	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
Dolomite	70.4	72.9	67.7	31.2	65.1	47.9	34.5	30.3	32.6
Calcite	0.1	1.42	4.07	5.55	2.28	0.24	0.60	4.00	16.7
Quartz	25.1	15.7	22.2	53.8	25.4	43.1	57.9	58.3	40.9
Illite	2.3	3.85	1.33	2.14	0.05	0.42	4.20	4.21	0.37
Kaolinite	0.03	0.30	ND	ND	1.67	1.60	ND	ND	0.22
Apatite	1.39	3.13	3.33	4.02	4.40	5.97	2.94	2.36	8.32
Kubler index	0.4	0.4	0.5	0.3	0.13	0.4	0.4	0.5	0.3
Illite-kaolinite ¹	0.6	2.9	1.4	0.7	0.2	0.6	0.1	0.4	0.5
Lithology	Silty dolomite	Sandy dolomite	Sandy dolomite	Sandy dolomite	Sandy dolomite	Dolomite chert	Sandy dolomite	Dolomitic sandstone	Dolomitic sandstone

Major elements in wt percent, rare earth elements in ppm unless noted, ND = not detected

¹ Determined from X-ray diffraction patterns of $-2\text{-}\mu\text{m}$ -size separates

Drilling, north of the deposit downdip from the orebody, encountered pyrite and carbon that were barren of gold mineralization (Gunter et al., 1990). This pyrite may be diagenetic because it is outside the main area of hydrothermal mineralization and alteration.

Arsenic sulfides have a limited occurrence at Barneys Canyon. Orpiment is rarely observed in drill core where it appears as a yellow dusting on the outside of the core. One core sample contains euhedral orpiment crystals. An arsenate mineral, possibly pharmacosiderite, was recognized in sandstone. Most of the hydrothermal pyrite, marcasite, and orpiment was deposited after argillization and before calcite.

Micron and submicrometer gold beads were found by S.A. Williams (pers. commun., 1987) wedged between dolomite grains and imbedded in illite. Williams (pers. commun., 1987) concluded that this was supergene gold because of its paragenetic relationship to iron oxide. E.U. Petersen (pers. commun., 1989) used electron microprobe analysis to identify gold associated with fine-grained quartz and iron oxide.

Gold ore distribution

The orebody at Barneys Canyon trends northwest and dips to the northeast (Fig. 5B and C). The orebody appears nearly strata bound in strike section, but the dip of the orebody is 10° to 15° shallower than bedding.

Nearly 90 percent of the orebody occurs beneath the lower plate dolomitic siltstone (Gunter et al., 1990) within the main host dolomite, but towards the southern end of the deposit, the sandstone becomes the major host. Some of the highest grade mineralization occurs at the contact between sandstone and dolomite. Gold grades increase from background to ore grade at the contact between dolomitic siltstone and the main host dolomite. A simplified blast hole assay contour map (Fig. 8) shows that the gold-grade trends are coincident with the strike of the beds.

Assay samples taken in and adjacent to the West fault show that the West fault is an ore-controlling structure, but the blast hole assay map shows that the West fault offsets gold mineralization (Fig. 8). Deep drilling in the bottom of the ore zone shows that gold mineralization is only present adjacent to the West fault. Therefore, the West fault is an ore-controlling structure that was reactivated following gold mineralization. Fault gouge in the Phosphate fault is of a consistently higher grade than the wall rock. The Phosphate fault offsets gold mineralization in that ore is juxtaposed against barren rock of similar lithology (Fig. 8). Deep drill holes do not contain ore in or adjacent to the fault. The Phosphate fault is either a reactivated feeder or it was enriched with gold following the gold mineralizing event. Samples from the East fault contain some gold enrichment within the fault, and

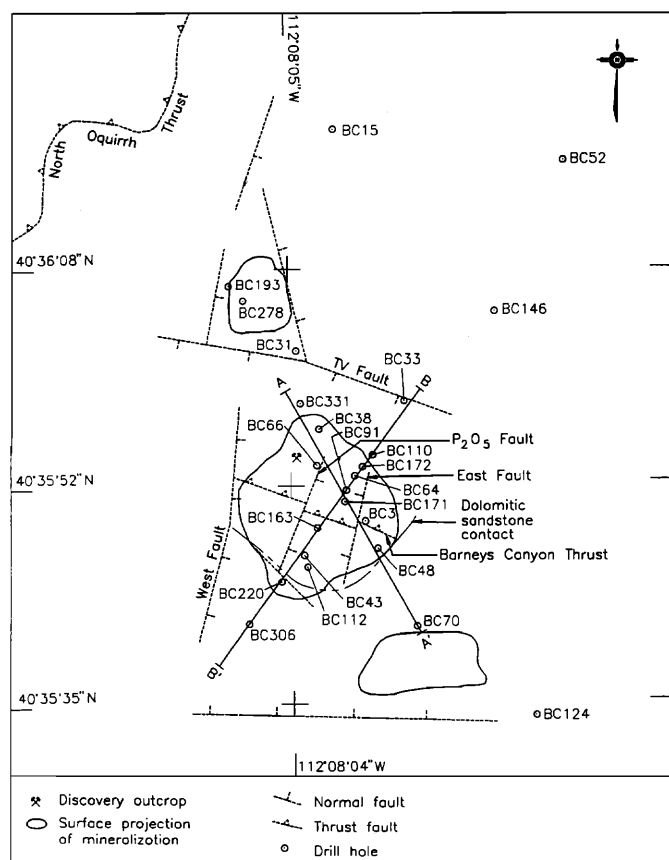


FIG. 7. Simplified geologic map of the Barneys Canyon gold deposit with location of drill holes used in study. The map shows drill holes to the north of Figure 6A.

deep drill holes contain ore at the bottom of the orebody along the East fault. Even though the East fault offsets the main host dolomite, mineralization is continuous across it. In summary, ore-controlling structures at Barneys Canyon trend northerly, but a northeast-striking fault offsets mineralization.

Fluid Inclusion Measurements

Heating and freezing measurements were made of fluid inclusions in six samples of quartz and two samples of barite. The samples of quartz are from patches of quartz in ore-grade transition zone dolomite from within the heart of the orebody. The majority of the quartz-hosted inclusions have dimensions of 10 by 20 μm with liquid-vapor ratios of 2:1 to 5:1 (Fig. 6E and F). Forty-one of the 53 measurements in quartz were paired measurements of homogenization temperature and salinity. The barite samples came from a vein and a sample of massive barite. Barite fluid inclusions range from 10 by 10 μm to 40 by 40 μm . Liquid-vapor ratios are consistent within samples and range from 2:1 to 3:1. Barite inclusions yielded 10 measurements but only three of the measurements were paired.

Using the criteria of Roedder (1984), all of the measured inclusions were primary, two-phase, liquid-vapor inclusions, and none of the fluid inclusion samples showed evidence of boiling. The histogram of fluid inclusion homogenization temperatures for quartz and barite (Fig. 9) shows a weakly

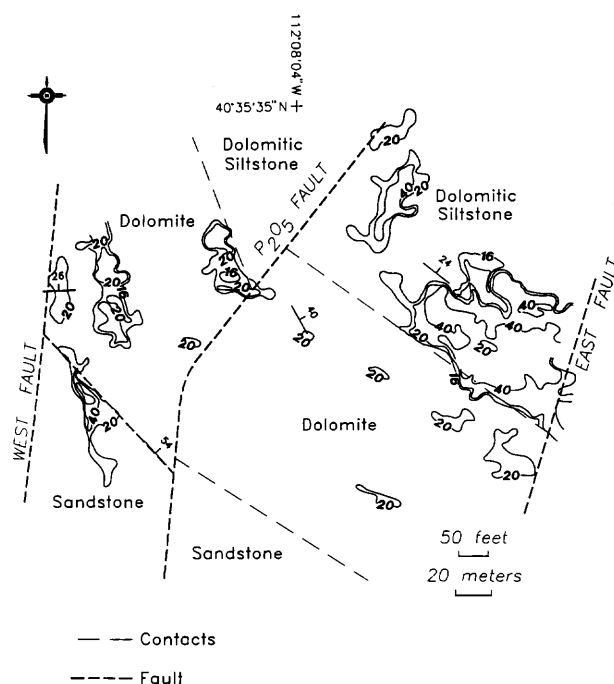


FIG. 8. Simplified blast hole assay contour map. Units are milliounces of gold per ton on the 6480 level at the southern end of the deposit. Contours are based on 16-ft drill hole spacing.

bimodal distribution with modes at 225° and 345°C. The wide spread in temperatures is thought to represent real variations in fluid properties because the inclusions measured were primary and unnecked inclusions. Some individual samples yield the full range of temperatures shown in the histogram. Freezing measurements yield salinities of 0 to 6.7 wt percent NaCl

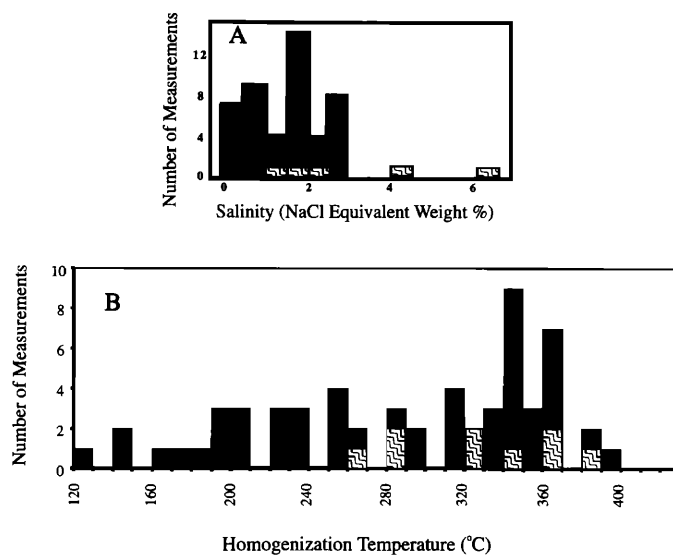


FIG. 9. Upper left insert shows the histogram of salinities (A) (wt % NaCl equiv) of aqueous fluid inclusions from quartz (dark gray) and barite (diagonal pattern) calculated from ice-melting temperatures using the regression equation of Potter et al. (1978). Histogram of homogenization temperatures (B) of aqueous fluid inclusions in quartz and barite from the Barneys Canyon gold deposit.

equiv; however, most of the measurements are between 0 and 3 wt percent with only two measurements above 4.6 and 6.7 wt percent. The mean salinity is 1.7 wt percent NaCl equiv.

A pressure-temperature diagram (Fig. 10) shows the liquid-vapor curve for water and isochores for a 1.7 wt percent NaCl equiv solution. The isochores were constructed over the range of homogenization temperatures shown as points on the boiling curve. Stratigraphic reconstruction of the Mesozoic section suggests removal of 2 km of stratigraphic overburden from Barneys Canyon (Presnell, 1992b). Lithostatic pressure at a 2-km depth would be 515 bars and hydrostatic pressure would be 188 bars. Intersection of these pressures with the isochores yield entrapment temperatures at lithostatic or hydrostatic pressure conditions and pressure corrections at lithostatic and hydrostatic conditions of 35° and 12°C, respectively. The univariant curve for equilibrium of kaolinite, quartz, pyrophyllite, and water, estimated using the thermodynamic data of Hemley et al. (1980), establishes the maximum pressure and temperature of fluids which produced the kaolinite alteration assemblage. The intersection of the reaction curve with the isochore for fluid inclusions homogenizing at 225°C and with a 0.8455-g/cc fluid density gives the maximum pressure (700 bars) and temperature (278°C) of fluids forming the kaolinite alteration at Barneys Canyon.

Minor Elements

The trace element survey of reverse circulation drill hole composites and drill core from BC-91 determined the distribution of gold concentrations and the characteristic trace element suite for Barneys Canyon. Gold concentrations of samples from the orebody in drill hole BC-91 reach a maximum of 12.6 ppm in sandy dolomite and then decline downward to 27 ppb in the dolomitic sandstone of the Kirkman-Diamond Creek formation (Table 2). Mercury reaches a high of 2,000 ppb coincident with the maximum gold value and declines downward to 105 ppb in the dolomitic sandstone. Arsenic reaches a high of 371 ppm above the maximum gold value; below the ore in the dolomitic sandstone, arsenic concentration is 161 ppm. Gold and arsenic are both elevated in the orebody, but samples with the highest gold values do not contain the highest arsenic concentrations. Tellurium is below the detection limit of 0.5 ppm in all samples from BC-91. Bismuth varies from 9 to 75 ppm within the ore, but falls to less than 2 ppm in the dolomitic sandstone; antimony varies from 39 to 135 ppm in ore and remains at 38 to 42 ppm in the dolomitic siltstone (Table 2).

A statistical summary of all the trace element data is presented in Table 3. The original analyses on which this summary is based are available from the authors on request. Modes were visually determined from histograms. Absolute background levels were not determined in this study because all of the Park City Formation at Barneys Canyon is altered to some degree, and it was not possible to analyze unaltered rock for comparison. In order to evaluate enrichment or depletion, the background was assumed to be equal to the mode as suggested by Levinson (1980). Average trace element abundances (Clark values) presented by Levinson (1980) were used for comparison and are also listed in Table

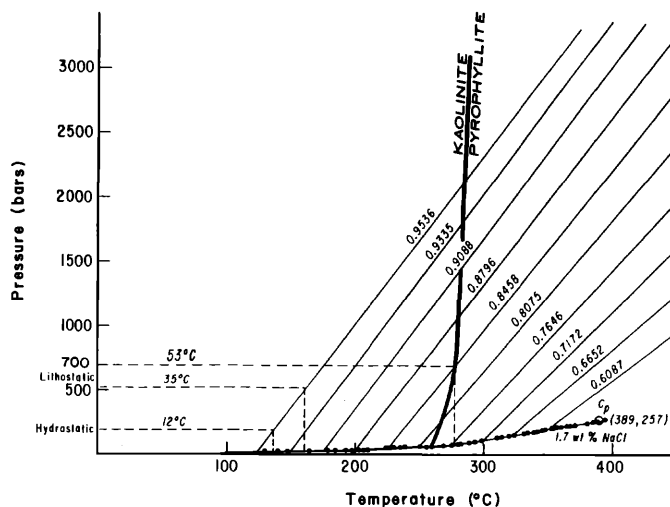


FIG. 10. P-T diagram of the fluid inclusion data with the univariant reaction curve for kaolinite + 2 quartz = pyrophyllite + water from Hemley et al. (1980). The liquid-vapor curve and isochores are calculated for a 1.7 wt percent NaCl equiv solution. The isochores were constructed over the range of homogenization temperatures. Pressure corrections are indicated for hydrostatic (12°C), lithostatic (35°C), and maximum conditions (53°C) from alteration assemblages.

3. When a Clark value was not given, the average crustal abundance was used.

The epithermal suite of Ag, As, Sb, Hg, Tl, and Ba shows a lognormal distribution. For each of the elements, except Ag, the mean is greater than the corresponding Clark value. This suggests that the Barneys Canyon gold deposit is enriched in As, Sb, Hg, Tl, and Ba. Most of the values for Ag are at or below the detection limit, and the mean is well below the Clark value. As, Sb, Hg, Ba, and Tl show pronounced concentrations and appear to be characteristic trace elements at Barneys Canyon.

The distribution of the base metals are lognormal to normal and appear to show some increase relative to Clark values but their low concentrations do not characterize the deposit. The porphyry-related elements Te, Bi, W, and Sn have lognormal distributions. Only W shows significant increase relative to its Clark value and Te does not. Many of the other trace elements show increases relative to their Clark values but due to their low concentrations they do not appear to be characteristic of the deposit.

Geochronology

Two <4- μ m-particle-size separates from two illite-rich, bedding-parallel veins were dated by the K-Ar method. Sample BC-I was collected from a 2- to 4-cm-thick, white gouge-like material at the sandstone dolomite contact (Fig. 5A). The sample contained illite, quartz, and carbonate with a minor amount of kaolinite. The illite texture in thin sections consists of illite flakes interwoven in an irregular, unoriented fashion within an irregular vein (Fig. 6B and C). The sample also contained 1.5 g/t gold and elevated As, Ba, Sb, Hg, and Tl and is from within the main orebody (Fig. 5a). Sample 6460-I was obtained from a bedding-parallel, veinlike, illite concentration in the lower part of the main orebody (Fig.

TABLE 3. Statistical Summary of Trace Element Concentrations

Element	Mean	Median	Max	Min	Std. dev.	Mode	Clark value ¹
Ag	0.14	0.1	0.7	0	0.14		1.00
As	137	66	2,655	0	293	70	2.50
Sb	8	1	803	0.2	68	1.5	(0.2)
Hg	1.3	0.3	19	0.02	2.6	0.7	0.05
Cu	28	14	1,090	1	94	37	15.0
Pb	27	8	390	1	49	14	8.0
Zn	104	99	431	7	59	106	25.0
Mo	5	3	80	1	8	3	1.00
Co	2	2	5	0.25	0.775	1.36	4.00
Ni	21	20	211	0.30	18	21	12.0
Cd	1.4	1.2	13	0.01	1.4	0.9	0.10
Tl	30	12	998	0.01	88	33	(0.45)
Bi	0.2	0	17	0	1.5		(0.17)
Te	0.04	0	0.40	0	0.08	0.02	(0.001)
Se	2	1	95	0	8	3	0.08
Cr	105	89	371	0	66	99	10.0
Mn	255	210	1,520	26	208	125	1,100
U	15	3	990	0.15	89		2.00
Th	1.5	1.3	4.2	0.20	0.76	1	2.00
Ga	1.2	0.95	5.2	0.01	0.82	0.87	0.06
Ge	0.12	0.10	0.65	0	0.10	0.02	0.10
V	22	20	74	0	10	20	15.0
In	0.3	0.05	15	0	1.43		0.02
Sr	57	47	32	0.01	49		500
Ti	131	86	850	0.01	125		400
Ba	331	145	5,580	2.5	623	187	100
W	4	0.7	146	0	16	5	0.50
Sn	3	0.4	260	0	22		4.00
Sc	2	2	5	0	1	2	5.00

All data in ppm

¹ Average abundance in limestone from Levinson (1980); values in parentheses are crustal averages

5A). The two illite samples yielded ages of 147 and 159 Ma, respectively (Table 4).

Discussion

The Barneys Canyon gold deposit has the geologic characteristics of a Carlin-type gold deposit according to the descriptions and criteria presented in Bagby and Berger (1985) and Percival et al. (1988). The deposit contains micrometer to submicrometer size gold, pyrite, marcasite, and rare arsenic sulfides hosted in carbonates and clastics. Host-rock alteration

is similar to other Carlin-type deposits consisting of silicification, decalcification, and argillization, but alteration minerals are less abundant. Fluid inclusions at Barneys Canyon are simple, two-phase inclusions with temperatures and salinities typical of other Carlin-type deposits but with a noted lack of observable CO₂ which is common at Carlin (Kuehn and Rose, 1995). The Barneys Canyon characteristic trace element suite of As, Sb, Hg, Ba, and Tl is typical of Carlin-type deposits as well. The trace element suite is not typical of epithermal deposits related to igneous activity which have a high silver content (Silberman and Berger, 1985) and is also not typical of porphyry-related deposits which often have Te contents in the ppm range (McCarthy and Gott, 1978; Alvarez and Noble, 1988). Barneys Canyon is associated with extensional and compressional structures including an anticline similar to other deposits of the Carlin type.

This study was not able to determine unequivocally the relationship of Barneys Canyon to the Bingham porphyry copper deposit. Even though the Bingham system contains more gold than any other known porphyry copper deposit (Sillitoe, 1988), most of the gold occurs in the chalcopyrite and bornite zones (Jones, 1992) and is directly associated with the main stage of copper mineralization. Copper-gold skarns occur in the Carr Fork mine and are also associated with main-stage mineralization (Atkinson and Einaudi, 1978). High-grade quartz, pyrite, and sericite gold mineralization overprints the skarns at Carr Fork and contains anomalous As, Sb, Hg, and Tl but is crosscut by quartz-molybdenite veinlets suggesting main-stage Bingham associations as well.

TABLE 4. K/Ar Ages of Illite in the Oquirrh Mountains

Sample no.	⁴⁰ Ar _{total}	⁴⁰ Ar	%K	⁴⁰ Ar (ppm)	Age (Ma)
Barneys Canyon					
BC-I	0.775	3.85	0.0445	4.598	159 ± 6
6960-I	0.850	6.28	0.0666	7.491	147 ± 5
Southern Oquirrh Mountains illite veins (Wilson and Parry, 1990)					
LT-3	0.6755	2.352	0.02053	2.816	122 ± 5
FM-3	0.8970	2.721	0.03766	3.246	189 ± 7
FM-9	0.7950	3.068	0.03088	3.660	170 ± 6
O-3	0.7965	2.726	0.03088	3.252	156 ± 6
CC-2	0.4970	1.333	0.01715	1.590	177 ± 7
MLT-11	0.7620	2.164	0.02291	2.582	147 ± 6

Age equation and constants used are presented in Faure (1986)

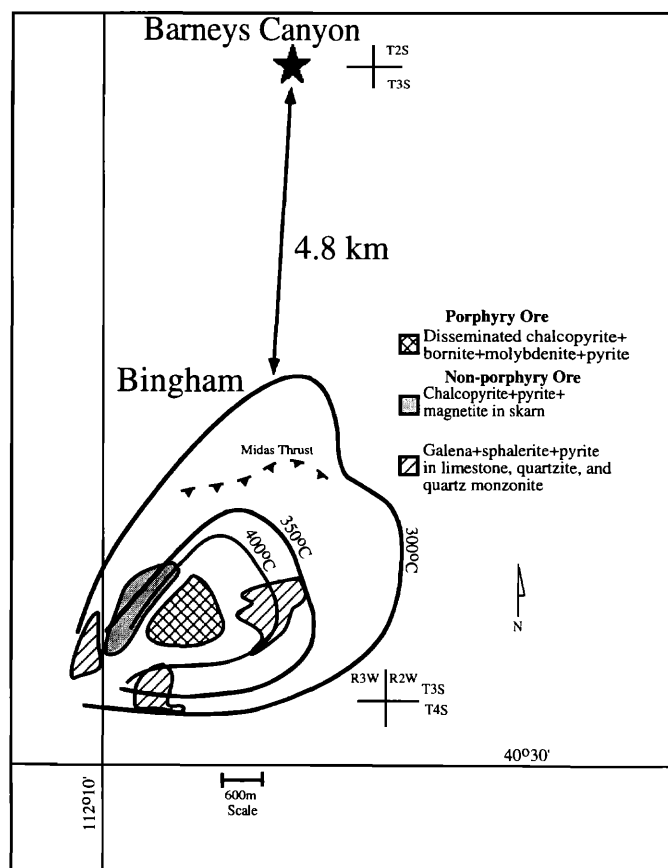


FIG. 11. Simplified temperature contour map of the Bingham hydrothermal system based on fluid inclusion data of Bowman et al. (1987) and Roeder (1971) and alteration of Lanier et al. (1978) and mineral zonation of Atkinson and Einaudi (1978). The 400°C contour is the limit of potassic alteration and the mean fluid inclusion homogenization temperatures from within the potassic zone. The 355°C contour is the limit of propylitic alteration and the mean homogenization temperature from within the propylitic zone. The 300°C contour represents the limit of the Pb-Zn halo and the average homogenization temperature from within the Pb-Zn halo.

Although Barneys Canyon lies on the edge of an erratic gold-arsenic geochemical halo (2–50 ppb gold in soils; G.H. Ballantyne, C. Phillips, J. Maughan, and T. Smith, pers. commun., 1995), for reasons discussed below, we do not think that there is a genetic link between the Bingham porphyry copper deposit and Barneys Canyon.

Even though Barneys Canyon is within the Bingham mining district, it is 7 km from the center of the Bingham porphyry system. A simplified temperature contour map (Fig. 11) shows that Barneys Canyon is located 4 km north of the 300°C contour and that alteration at Bingham extends 3.5 km from the center of mineralization. The size of the hydrothermal system at Bingham is comparable with that of other porphyry systems which generally extend 3 to 4 km from the center of mineralization, e.g., 2.5 km at Santa Rita, New Mexico (Nielson, 1970); 4 km at Sierrita-Esperanza, Arizona (Titley et al., 1986); and 3 km at Red Mountain, Arizona (Corn, 1975). Calculations by Jaeger (1959) and Cathles (1977) indicate that the peak temperature will have reached ambient temperature within about 1.5 pluton diameters. Bar-

neys Canyon is at least 7 pluton diameters north of the Bingham quartz monzonite porphyry. This reasoning suggests that Barneys Canyon is related to a separate hydrothermal system. Intrusive rocks are notably absent from Barneys Canyon; the nearest is a small latite dike 3 km to the west near the crest of the Oquirrh Mountains. Neither the regional aeromagnetic map of Mabey et al. (1964), nor detailed geophysical surveys by Kennecott show any magnetic anomalies attributable to an intrusion at or near Barneys Canyon.

Bingham was emplaced in an extensional tectonic regime controlled by northeast-striking extensional structures and preexisting compressional folds. Although the ore-controlling structures at Barneys Canyon are normal faults, they may not have formed in an extensional tectonic regime. The West and East faults have a northerly strike and can be related to the Copperton anticline by correlating them with the Verona and Smelter faults, respectively. A Tertiary dike related to Bingham crosscuts the Smelter fault constraining the age of the fault as pre-Bingham. The Verona and Smelter faults are subparallel to the axial planes of the Copperton and Tie-waukee anticlines and may have resulted from the folding process. The extensional ore-controlling structures at Barneys Canyon may have formed in a compressional regime during growth of the Copperton anticline. The extensional events recorded at Bingham are reflected in the offset of ore along the northeast-striking Phosphate fault and the ore offset, interpreted as reactivation, along the West fault.

The highest fluid inclusion homogenization temperatures in Carlin-type deposits are generally not interpreted as the temperature of gold mineralization. Peak homogenization temperatures in fluid inclusions at Barneys Canyon are 393°C in quartz and 380°C in barite (Fig. 9). Fluid inclusion homogenization temperatures from the Carlin, Nevada, deposit reach 350°C in late stage 2 or early stage 3 quartz, 365°C in barite, and 308°C in sphalerite (Radtko et al., 1980). Moderate salinity H₂O-CO₂ fluid inclusions homogenize at temperatures as high as 280°C in main gold ore stage mineralization at Carlin (Kuehn and Rose, 1995), and homogenization temperatures of 303°C in quartz and 353°C in barite have been recorded at Mercur, Utah (Jewell and Parry, 1988). Kaolinite-bearing mineral assemblages must form below 278°C. Marcasite may precipitate at a temperature as high as 240°C (Morrowchick and Barnes, 1986). Orpiment and realgar, rare at Barneys Canyon, melt at 312° and 307°C, respectively, and have solubilities that are so high above 200°C as to preclude their stability (Heinrich and Eadington, 1986) even though α -AsS is stable up to 265°C. The lack of correlation of gold and arsenic values within the ore at Barneys Canyon suggests that gold and arsenic are not coeval, though erratic variation among the elements As, Au, Ba, Hg, Sb, and Tl are common in Carlin-type ore (Wilson and Parry, 1995; Ashton, 1989). The marcasite, kaolinite alteration, and arsenic sulfides suggest formation at less than 250°C.

The paragenesis of gold is not known nor do we know the age of gold mineralization at Barneys Canyon, but the geologic setting and K-Ar ages of vein illite do constrain the age of mineralization. The age of the Copperton anticline is a maximum age for the deposit. The vein illites dated at Barneys Canyon occur within bedding-plane gouges formed by flexural slip during growth of the Copperton anticline (Pres-

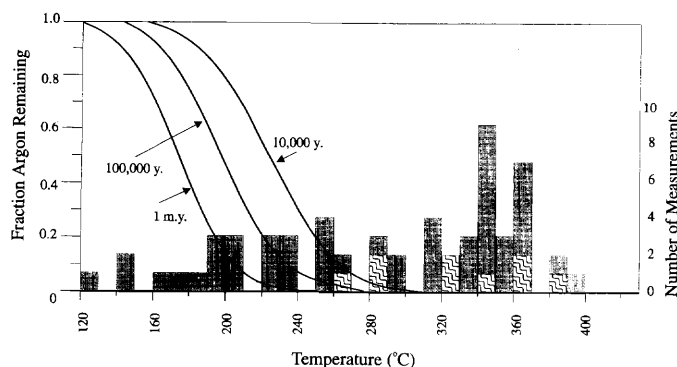


FIG. 12. Argon diffusion curves for illite with a maximum particle size of $4\ \mu\text{m}$ calculated from the equations in Dodson (1973) for heating times of 1 million, 100,000, and 10,000 years. Histogram of the fluid inclusion temperature from quartz (dark gray) and barite (diagonal pattern) are shown for comparison.

nell and Parry, 1995). Because they are folded by the Copperton anticline, the illite-rich gouges give a minimum age for anticline formation. The dated illite veins are gold bearing and contain the Carlin-type characteristic trace element suite, therefore, they give a maximum age for gold mineralization.

The dated illites are from veins (Fig. 6B and C), but the possibility of repetitive hydrothermal events requires evaluation of diffusional loss of ^{40}Ar and resetting of the illite ages. Argon loss is a function of particle size, time, and temperature. The fluid inclusion and illite crystallinity data suggest that the Barneys Canyon gold deposit experienced a complex thermal history, but the duration of the hydrothermal system at Barneys Canyon is not known. Silberman (1985) suggests that porphyry copper systems have durations of 1 m.y. or less except at Bingham, Utah where the lifetime is 2 m.y., and precious and base metal systems have lifetimes of 0.5 to 1.5 m.y. Arribas et al. (1995) estimate that the lifetime of the Far Southeast porphyry and Lepanto epithermal hydrothermal system in northern Luzon, Philippines, is 300 ka. Diffusion curves for argon loss were calculated for durations of 1 m.y., 100 ka, and 10 ka for illite with a maximum particle size of $4\ \mu\text{m}$. Comparison of Ar diffusion curves with fluid inclusion homogenization temperatures (Fig. 12) shows that complete resetting of illite particles would occur at temperatures above 300°C if heated for 10 ka, 280°C if heated for 100 ka, and 240°C if heated for 1 m.y. The empirical resetting curve of Hunziker (1987) also indicates that argon would be completely lost from $4\ \mu\text{m}$ and smaller illite grains at 250°C . Figure 6C and D show illite up to $20\ \mu\text{m}$, but the K-Ar age measurements were made on $<4\text{-}\mu\text{m}$ -size separates. Detrital illites, if incorporated in the dated samples, would contain only Ar generated since thermal resetting. Dated illite samples are veins which have hydrothermal textures and paragenetically occur between quartz and barite precipitation. The gold orebody is associated with argillic alteration and high illite/kaolinite supporting the contention that the dated illites formed during hydrothermal mineralization at Barneys Canyon. Based upon these arguments, we believe that the maximum age of the hydrothermal mineralization at Barneys Canyon is Jurassic. The K-Ar ages in Table 4 together with the

Ar loss curves of Figure 12 suggest that partial resetting may have occurred.

The Oquirrh Mountains also contain several other Carlin-type deposits 25 km southwest of Bingham in the Mercur district. Illite veins in the Mercur district and illites from altered rocks in the Mercur deposits also yield Jurassic K-Ar ages (Table 4) and based on detailed geology and clay mineralogy are thought to represent the age of gold mineralization in the southern Oquirrh Mountains (Wilson and Parry, 1990, 1995). Due to similar deposit type and hydrothermal illite age we believe that Mercur and Barneys Canyon may have formed in the Jurassic by the following scenario: orogenic fluids driven by Jurassic tectonism leached gold from metalliferous Mississippian shales; these gold-bearing fluids migrated along active decollements and subsequently deposited gold when they encountered meteoric waters in the crests of anticlines.

Summary and Conclusions

The Barneys Canyon gold deposit is a sediment-hosted, disseminated gold deposit within the northeast part of the Bingham mining district. The host rocks are Permian Park City dolomite and siltstone, and Permian Kirkman-Diamond Creek sandstone. These units occur on the north flank of the Copperton anticline. The Park City Formation was repeated by the Barneys Canyon thrust which predates anticline formation. All units at Barneys Canyon were offset by several north-south-striking normal faults and minor east-west-striking normal faults. Hydrothermal mineralization consists of alteration of the host rock by decalcification, silicification and argillization, and deposition of pyrite, marcasite, minor arsenic sulfides, and gold. The deposit has a characteristic trace element suite of As, Sb, Hg, Ba, and Tl. Fluid inclusion homogenization temperatures from quartz and barite and the lack of pyrophyllite in the alteration assemblage give a maximum pressure and temperature of 700 bars and 278°C , respectively, of fluids which produced the argillic alteration. The deposit has the structural, hydrothermal mineralization, trace element, and fluid inclusion characteristics of a Carlin-type gold deposit. K/Ar ages of hydrothermal illite-rich, bedding-plane gouges give a minimum (Jurassic) age for the Copperton anticline and a maximum age of gold deposition.

Although inconclusive, spatial, structural, and thermal arguments allow for Barneys Canyon to have formed prior to and unrelated to the Bingham porphyry copper deposit or any other igneous system.

Acknowledgments

Financial assistance, access to the mine and drill cores, advice, guidance, and encouragement provided by Kennecott is gratefully acknowledged. We thank R.C. Babcock, J.W. Hammitt, and P. Smith of Kennecott for their support and encouragement. We thank R. Ramsey, W. Gunter, G. Austin, and G. Slothower at the Barneys Canyon mine for access, support, and encouragement. Geoff Ballantyne made numerous helpful suggestions that improved the manuscript. We thank Kennecott for permission to publish this manuscript, but the views expressed are those of the authors. We also thank E. Petersen, P. Wilson, R. Lambert, and W. Nash for valuable assistance with laboratory measurements. Comments

and suggestions of three *Economic Geology* reviewers clarified and improved the manuscript. The figures were drafted by P. Onstott.

May 25, 1994; November 1, 1995

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