ABSTRACT

The Blackbird cobalt-copper (Co-Cu) district in the Salmon River Mountains of east-central Idaho occupies the central part of the Idaho cobalt belt—a northwest-elongate, 55-km-long belt of Co-Cu occurrences, hosted in grayish siliciclastic metasedimentary strata of the Lemhi subbasin (of the Mesoproterozoic Belt-Purcell Basin). The Blackbird district contains at least eight stratabound ore zones and many discordant lodes, mostly in the upper part of the banded siltite unit of the Apple Creek Formation of Yellow Lake, which generally consists of interbedded siltite and argillite. In the Blackbird mine area, argillite beds in six stratigraphic intervals are altered to biotitite containing over 75 vol% of greenish hydrothermal biotite, which is preferentially mineralized.

Past production and currently estimated resources of the Blackbird district total ~17 Mt of ore, averaging 0.74% Co, 1.4% Cu, and 1.0 ppm Au (not including downdip projections of ore zones that are open downward). A compilation of relative-age relationships and isotopic age determinations indicates that most cobalt mineralization occurred in Mesoproterozoic time, whereas most copper mineralization occurred in Cretaceous time.

Mesoproterozoic cobaltite mineralization accompanied and followed dynamothermal metamorphism and bimodal plutonism during the Middle Mesoproterozoic East Kootenay orogeny (ca. 1379–1325 Ma), and also accompanied Grenville-age (Late Mesoproterozoic) thermal metamorphism (ca. 1200–1000 Ma). Stratabound cobaltite-biotite
ore zones typically contain cobaltite, in a matrix of biotite ± tourmaline ± minor xenotime (ca. 1370–1320 Ma) ± minor chalcopyrite ± sparse allanite ± sparse microscopic native gold in cobaltite. Such cobaltite-biotite lodes are locally folded into tight F2 folds with axial-planar S2 cleavage and schistosity. Discordant replacement-style lodes of cobaltite-biotite ore ± xenotime (ca. 1320–1270 Ma) commonly follow S2 fractures and fabrics. Discordant quartz-biotite and quartz-tourmaline breccias, and veins contain cobaltite, ± xenotime (ca. 1058–990 Ma).

Mesoproterozoic cobaltite deposition was followed by: (1) within-plate plutonism (530–485 Ma) and emplacement of mafic dikes (which cut cobaltite lodes but are cut by quartz-Fe-Cu-sulfide veins); (2) garnet-grade metamorphism (ca. 151–93 Ma); (3) Fe-Cu-sulfide mineralization (ca. 110–92 Ma); and (4) minor quartz ± Au-Ag ± Bi mineralization (ca. 92–83 Ma).

Cretaceous Fe-Cu-sulfide vein, breccia, and replacement-style deposits contain various combinations of chalcopyrite ± pyrrhotite ± pyrite ± coeval arsenopyrite (not cobaltite) ± arsenopyrite ± quartz ± siderite ± monazite (ca. 144–88 Ma) but mostly 110–92 Ma) ± xenotime (104–93 Ma). Highly radiogenic Pb (in these sulfides) and Sr (in siderite) indicate that these elements resided in Mesoproterozoic source rocks until they were mobilized after ca. 100 Ma. Fe-Cu-sulfide veins, breccias, and replacement deposits appear relatively undeformed and generally lack metamorphic fabrics.

Composite Co-Cu-Au ore contains early cobaltite-biotite lodes, cut by Fe-Cu-sulfide veins and breccias, or overprinted by Fe-Cu-sulfide replacement-style deposits, and locally cut by quartz veinlets ± Au-Ag ± Bi minerals.

INTRODUCTION

Blackbird Cobalt-Copper (Co-Cu) District

Location
The Blackbird Co-Cu district is in the Salmon River Mountains of east-central Idaho and is ~35 km west-southwest of the town of Salmon, Idaho (Figs. 1A and 1B). The Salmon River Mountains are interpreted as an uplifted and dissected plateau with broad ridges at nearly concordant elevations, incised by steep-walled tributaries to Salmon River Canyon, which is ~1.5 km deep. The Blackbird district occupies the central, widest, and most-mineralized part of the Idaho cobalt belt, a northwest-trending belt of Cu-Co mines, prospects, and geochemical anomalies. The Idaho cobalt belt (ICB) is ~55 km long and extends ~27.5 km to the southeast and northwest of its centroid. Broad ridges are deeply weathered to saprolite, and hillside rock outcrops are sparse, except along steep upper slopes and road cuts.

As shown in Figure 1B, the geologically defined Blackbird district is within the legally defined Blackbird Mining District, which is wider but shorter than the Idaho cobalt belt, and includes epithermal silver-gold deposits outside of the Idaho cobalt belt. The Blackbird mine area is in the west-central part of the Blackbird district, where it is elongate northwest.
Regional Geologic Setting

Co-Cu mines and prospects of the ICB and the Blackbird district are hosted in Mesoproterozoic metasedimentary rocks of the Lemhi subbasin of the Belt-Purcell Basin (Fig. 1A). The Lemhi subbasin probably is underlain by cratonized oceanic rocks of the Paleoproterozoic Selway terrane along the Great Falls tectonic zone between Archean terranes of the Wyoming Province and the Medicine Hat block. As shown in Figure 1A, the western Belt-Purcell basin and most of the Lemhi subbasin are within the East Kootenay orogen, which underwent dynamothermal metamorphism and plutonism in Ectasian time.

The Belt-Purcell Basin and the Lemhi subbasin also are within the Cordilleran orogen (Fig. 2A), which underwent orogenesis and plutonism from Late Jurassic through Cretaceous to
The Blackbird district contains at least eight stratabound ore zones, several stratabound prospects, and many discordant veins and breccias—all hosted in metasedimentary rocks of the Lemhi Group. Past production from the Haynes-Stellite, Uncle Sam, and Blackbird mines totals ~2.4 Mt of ore with average grades of 0.82% Co, 2.1% Cu, and 3.3 ppm Au, according to data from Anderson (1943, 1947; Bennett, 1977; and Preen, 2006). The total of production and estimated resources of the Blackbird district was reported by Slack (2013) as 16.8 Mt of ore, averaging 0.735% Co, 1.37% Cu, and 1.04 ppm Au. This estimate included measured, indicated, and inferred resources of eight stratabound ore zones, five stratabound prospects, and one breccia pipe. However, this estimate did not include downward extensions of ore zones that are open down-dip.

Although the Blackbird deposit contains nearly twice as much Cu as Co, unit prices for Co generally are 5 to 10 times higher than unit prices for Cu. Thus, Co is the primary product of the Blackbird district, whereas Cu is a co-product, and Au is a by-product. Slack (2006) suggested that minerals such as monazite and xenotime might also be recoverable as by-products for their rare-earth-element (REE) contents.

Grade-tonnage diagrams by Slack et al. (2013) for 13 Co-Cu-Au deposits in metasedimentary rocks indicate that the Blackbird deposit has a higher average Co grade and contains more Co than any other known deposit of that type. The average Co grade of the Blackbird deposit also is higher than those of most sediment-hosted Cu deposits, and much higher than those of selected Co-bearing volcanogenic massive sulfide deposits (VMS), iron-oxide Cu-Au deposits (IOCG), and magmatic Ni-Cu deposits. Globally, however, most Co is recovered as a by-product from Co-bearing Cu deposits in sedimentary rocks, and from Co-bearing Cu-Ni deposits in mafic-ultramafic rocks (Vhay et al., 1973).

Previous Work

Previous publications on the geology and mineral deposits of the Blackbird Co-Cu district are listed in Table 1, with a brief note about each. Vhay (1948) provided the first geologic maps of the Blackbird district, and Anderson (1947) provided the first summary of the sequence of deposition of hydrothermal minerals there. Geologic maps of the Blackbird Mountain quadrangle by Evans and Connor (1993), the Salmon River Mountains by Tysdal et al. (2003), and the Salmon National Forest by Evans and Green (2003) provide stratigraphic and structural context for the geology of the Idaho cobalt belt and the Blackbird district.

In addition to published work listed in Table 1, a tremendous amount of geologic work has been done by mineral-exploration and mine geologists of several companies, and much of that is documented in unpublished reports, archived in the office of the Geologic Division of the U.S. Geological Survey, in Spokane,
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<td>Ore replaces schist (pC) near gabbro (KT?).</td>
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<td>Ross (1941)</td>
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<tr>
<td>Bookstrom et al. (2014)</td>
<td>Cobaltite-biotite ore is Y-age, polymetallic veins are K-age.</td>
<td>Blackbird Co-Cu</td>
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Washington. Nevertheless, the age, origin, deposit-type, and geologic history of Blackbird Co-Cu deposits have long been controversial (as indicated by notes in Table 1).

Recent articles on the mineralogy, geochemistry, geochronology, fluid inclusions, and isotopic characteristics of the Blackbird Co-Cu deposits are included in a collection of papers on mineral deposits of the Belt-Purcell Basin edited by Box et al. (2012). These studies provide many new constraints and clues to guide interpretation of Co-Cu deposits in the Blackbird district and the Idaho cobalt belt.

This Study

The purpose of this study was to compile and integrate currently available information about the geology of the Blackbird Co-Cu district and its geologic settings through geologic time in the Lemhi subbasin of the Belt-Purcell Basin, the East Kootenay orogen, a regional Grenville-age metamorphic belt, and the Cordilleran orogen. The goal of this paper is to present a regionally to locally consistent history of the geology of the Blackbird Co-Cu district. This study proceeds from Mesoproterozoic to Cenozoic time, from regional to local settings within each time interval, and from observed relative-age relationships and isotopic age determinations to interpretations of their implications for the geologic history of the Blackbird Co-Cu deposits and their host rocks. Names of minerals and rocks are abbreviated as listed in Table 2.

BASEMENT ROCKS

According to Foster et al. (2006), cratonized oceanic-island-arc–type basement rocks of the Paleoproterozoic Selway terrane (ca. 1.86–1.7 Ga) probably underlie most of the southern part of the Belt-Purcell Basin and most of the Lemhi subbasin, which hosts Co-Cu deposits of the Idaho cobalt belt and the Blackbird district (Fig. 1A). According to Harms et al. (2004), these rocks were metamorphosed during the Big Sky orogeny (1.78–1.72 Ga), which contributed to assembly of the supercontinent of Nuna (or Columbia), upon which the Belt basin subsided and filled with sediments and mafic sills.

As shown in Figure 1A, the Idaho cobalt belt and the Sheep Creek Cu-Co deposit are both hosted in Beltian strata that overlie the projected trend of the Great Falls tectonic zone, a broad zone of northeast-trending shear zones (interpreted as thrust faults) in basement rocks between Archean crustal blocks (the Wyoming Province and the Medicine Hat block), according to O’Neill (1993) and Sims et al. (2004).

In north-central Idaho, basement rocks of the Clearwater block consist of Neoarchean orthogneiss (2.67–2.65 Ga) intruded by Paleoproterozoic orthogneiss (1.88–1.84 Ga), according to Vervoort et al. (2015). The location of the boundary between the Clearwater block and the Selway terrane is mostly covered or intruded by younger rocks, so basement rocks beneath the Lemhi subbasin could include rocks of either or both of these terranes.

STRATIGRAPHIC SETTING

Belt-Purcell Basin (Ca. 1470–1250 Ma)

As shown in Figure 3, the lower 12 km of mostly grayish siliciclastic sediments and mafic sills accumulated very rapidly from ca. 1470 to 1454 Ma, during rift-stage subsidence of the Belt Basin. Flows of basaltic Purcell lava erupted at ca. 1413 Ma, after which rates of sedimentary accumulation slowed during sag-stage deposition of the upper 5 km of varicolored strata in shallow-water settings from ca. 1454 to 1250 Ma.

A paleocontinental reconstruction by Sears (2007) suggests a predominantly Australian source for rift-stage detrital sediments, transported via grabens across the Siberian craton to the rapidly subsiding Belt-Purcell Basin. Subsidence associated with the eruption of the Purcell lava at 1443 Ma cut off the pathway for delivery of sediments from Australia and opened a new route...
Figure 3. Graphs of stratigraphic thickness (km) vs. age (Ma) for the Belt Basin (after Sears, 2007) and the Lemhi subbasin: Belt age determinations are 1469 Ma for a mafic sill at Plains, Montana, and 1457 Ma for a mafic sill at Paradise, Montana (Sears et al., 1998); 1454 Ma for upper Helena tuff, 1443 Ma for Purcell Lava, and 1401 Ma for Bonner tuff (Evans et al., 2000); 1437 Ma and 1330 Ma for the Garnet Range Formation (Ross and Villeneuve, 2003). Estimated thicknesses of stratigraphic units in the Belt Basin are from Sears (2007), and those for the Lemhi subbasin in the Salmon River Mountains (Yellowjacket Formation to Swauger Formation) are from Evans and Green (2003). Estimated thicknesses of units above the Swauger Formation are from a columnar section for strata of the Lemhi and Beaverhead Ranges by Lonn et al. (2013, p. 206). Mean ages of youngest sets of detrital zircons in samples from key stratigraphic units in the Lemhi subbasin are from Link et al. (2007) and Aleinikoff et al. (2012b). The mean age of 1370 ± 10 Ma for megacrystic granite is from Evans and Zartman (1990).
for transport of sediments to the Belt-Purcell Basin from a southeastern source region, as indicated by the age-distribution profiles of detrital zircons in strata of the upper Belt succession and the Lemhi subbasin.

Ross and Villeneuve (2003, p. 1191) interpreted the Belt-Purcell Basin as “an extensional domain along a collisional/convergent plate margin, analogous to Tethyan sedimentary basins such as the Black and Caspian Seas.” This is consistent with a western (non-Laurentian) source for detrital zircons (dated 1610–1480 Ma) and with evidence for dynamothermal metamorphism and magmatism along the western margin of the Belt-Purcell Basin during the Ectasian East Kootenay orogeny.

**Lemhi Subbasin (Ca. 1454–1370 Ma)**

The Lemhi subbasin constitutes the southwestern part of the Belt-Purcell Basin (Fig. 1A). Prior to 1999, most metasedimentary strata of the Lemhi subbasin in the Salmon River Mountains were assigned to the broadly defined Yellowjacket Formation, which Ruppel (1975) perceived as similar to and correlative with deep-water strata of the lower Belt Prichard Formation in the Belt Basin. By contrast, Winston et al. (1999) suggested that the broadly defined Yellowjacket Formation was similar to and correlative with shallow-water strata of the Ravalli, Piegan, and Missoula Groups in the Belt Basin. However, subsequent dating of detrital zircons by Link et al. (2007) and Aleinikoff et al. (2012b) indicated that grayish metasedimentary strata of the Yellowjacket Formation through Apple Creek Formations are age-equivalent to varicolored, mostly terrestrial strata of the Missoula Group, which overlie strata of the Ravalli and Piegan Groups in the Belt Basin.

As shown in Figure 3, a total thickness of nearly 20 km of siliciclastic strata accumulated in the Lemhi subbasin between ca. 1454 and 1370 Ma, at an average rate of ~238 m/m.y., while average rates of sediment accumulation decreased from ~719 m/m.y. to ~53 m/m.y. in the Belt Basin (Fig. 3). The base of the Yellowjacket Formation is not exposed, so a very thick stratigraphic section, coeval with strata of the lower-to-middle parts of the Belt Supergroup, might be hidden below present levels of exposure.

The lower 11.5 km of strata exposed in the Lemhi subbasin consists mostly of grayish siltite, argillite, and quartzite. Lopez (1981) identified abundant quartz, common plagioclase and orthoclase, and accessory zircon, garnet, tourmaline, and opaque minerals, as detrital constituents of argillaceous quartzite from the lower part of the Yellowjacket Formation.

Bed forms, sedimentary structures, and the grayish appearance of these metasedimentary strata are interpreted to indicate rift-style deposition in shallow-marine settings that were relatively reducing. By contrast, the upper 8.5 km of strata in the Lemhi subbasin are varicolored and were probably deposited in sag-style terrestrial settings that were relatively oxidizing.

**Stratigraphic Units of the Lemhi Subbasin**

The columnar stratigraphic section in Figure 4A illustrates Mesoproterozoic stratigraphic units of the Lemhi subbasin in the Salmon River Mountains. Labels to the left of the column indicate previous stratigraphic nomenclature (according to Evans and Connor, 1993), whereas labels to the right indicate current nomenclature (according to Evans and Green, 2003). Current stratigraphic nomenclature is based on correlations of lithostratigraphic units in the Salmon River Mountains with those mapped previously by Tysdal (2003) in the Lemhi Range. Thus, Tysdal et al. (2003) applied nomenclature from the Lemhi Range to lower, middle, and upper subunits of the Yellowjacket Formation (as previously mapped by Evans and Connor, 1993). Evans and Green (2003) provided lithologic descriptions of these renamed stratigraphic units in the Salmon River Mountains. Age determinations shown in Figure 4A are maximum ages, indicated by the youngest set of detrital zircons in samples from each stratigraphic unit dated by Link et al. (2007) and Aleinikoff et al. (2012b).

Burmester et al. (2013) reported that the Swauger and Lawson Creek Formations are overlain by a very thick quartzite unit in the Lemhi Range and the Beaverhead Mountains. They named this unit the quartzite of Jahnke Lake, which they assigned to the stratigraphic succession of the Lemhi subbasin. This extends strata assigned to the Lemhi subbasin eastward into Montana, as shown in Figures 1A and 2A (Lonn et al., 2013; McDonald and Lonn, 2013).

Unfortunately, there are at least two Apple Creek Formations—the greenish Apple Creek Phyllite of Anderson (1961), and the Apple Creek Formation of the Yellow Lake reference section, established by Ruppel (1975) near Golden Trout Lake (now known as Yellow Lake). A columnar section by Burmester et al. (2013) indicated that the Apple Creek Phyllite overlies the quartzite of Jahnke Lake. However, the Apple Creek Formation of Yellow Lake underlies the Gunsite Formation and contains a banded siltite unit, similar to that of the Blackbird district. Thus, the banded siltite unit of the Blackbird district is here assigned to the Apple Creek Formation of Yellow Lake (as labeled in Fig. 4A).

**Banded siltite unit.** Where it is relatively unaltered, the banded siltite unit of the Apple Creek Formation (of Yellow Lake) in the Salmon River Mountains consists of cyclically repeated couplets, in which each relatively resistant bed of pale-gray siltite is overlain by a recessive bed of dark-gray argillite, thus giving outcrops a rhythmically banded appearance (Fig. 5A). Aleinikoff et al. (2012b) reported an age of 1409 ± 10 Ma for the youngest set of detrital zircons in metasandstone interlayered with biotite phyllite to schist in the upper part of the banded siltite unit in the Blackbird mine area.

Argillic rocks within ~2 km of the megacrystic granite pluton of Big Deer Creek are commonly metamorphosed to dark-brown biotite hornfels with round porphyroblasts of white sodic scapolite (Fig. 5B), and argillite beds near Blackbird ore zones are altered to biotite containing ~75 vol% (or more) of greenish-gray to greenish-black biotite (Fig. 5C). Such biotite is preferentially mineralized in stratabound zones of cobaltite-biotite ore, which consists of disseminated to semimassive concentrations of cobaltite grains in a matrix of biotite ± tourmaline. In the
Blackbird mine area, biotitite beds are concentrated within six stratigraphic intervals in the upper part of the banded siltite unit, and concentrations of cobaltite are preferentially concentrated in biotitite layers within three of these stratigraphic intervals (Figs. 4B and 5C).

Vhay (1948) noted that biotite in and around ore zones is greenish, and he interpreted such biotite as a product of hydrothermal alteration. Lee (1958) reported that greenish biotite from the Blackbird district is unusually rich in ferrous Fe (29 wt%) and Cl (1.1 wt%). Biotite, composed of such greenish biotite, is commonly metamorphosed to biotite phyllyte to schist, but is locally recrystallized to biotitite granofels with randomly oriented biotite plates that sparkle in reflected light. Biotite also is locally altered to chlorite.

Siltite interlayers generally are less biotitized and less mineralized than biotitite layers. Nevertheless, some siltite, which may have had a muddy component, is moderately biotitized, and locally, beds of siltite-to-quartzite beds are partly replaced by tormarine. More commonly, however, siltite is silicified and recrystallized to biotitite granofels with randomly oriented biotite plates that sparkle in reflected light. Biotite also is locally altered to chlorite.

Conn (1990) collected rock-chip samples from strata equivalent to Yellowjacket-Hoodoo unit, the Apple Creek Formation, and the Gunsight Formation, southeast of the Blackbird district. For 111 analyzed samples, he reported median concentrations of 8 ppm Co and 6.5 ppm Cu.

Instead, spidergrams in Figure 5 are interpreted here as manifestations of chemical and mineralogical changes associated with hydrothermal biotitization of argillite to biotitite. Relative to argillite and siltite in banded siltite, biotitite samples are depleted in Sr but enriched in Rb, which can be attributed to hydrothermal replacement of plagioclase by biotite. Also, most biotitite samples are enriched in Y and heavy REEs, but some samples are enriched, while others are depleted in light REEs. This would seem consistent with differential hydrothermal transport and deposition of the chemical constituents of xenotime, monazite, gadolinite, and allanite (identified in Blackbird ore samples by Slack, 2012).

**Gunsight Formation.** In the Haynes-Stellite structural block (Hsb in Fig. 2B), the lower part of the Gunsight Formation consists of quartzite beds (0.2–2 m thick) with thin interbeds of black biotite phyllyte. Northward, this passes into massive biotitic metasandstone, which is predominant in the drainage basin of Little Deer Creek (Figs. 6 and 7). Such biotitic metasandstone is interpreted here as a muddy facies of the Gunsight Formation, in which the muddy component is replaced by black biotite. Around the Sweet Repose prospect, a subfacies of the Gunsight Formation consists of scapolitic metasandstone, interlayered with scapolitic biotitite. Above the Sweet Repose adit, a thick biotite bed is internally folded, and the folds are cut by a quartz vein, containing chalcopyrite and cobaltian arsenopyrite.

In the northern part of the Blackbird structural block (Bs in Fig. 2B), banded siltite grades up section into biotitic siltite to metasandstone (± garnet ± chloritoid), overlain by quartzite with subordinate interlayers of biotite schist (±garnet). These rocks are mapped as quartzite and schist of the Lemhi Group (undivided) in Figure 2B, but on larger-scale maps in Figures 7 and 8, they are mapped as metasedimentary lithologic subunits of the Gunsight Formation.

**Co and Cu Contents of Metasedimentary Country Rocks**

Connor (1990) collected rock-chip samples from strata equivalent to Yellowjacket-Hoodoo unit, the Apple Creek Formation, and the Gunsight Formation, southeast of the Blackbird district. For 111 analyzed samples, he reported median concentrations of 8 ppm Co and 6.5 ppm Cu.

In the upper part of the coarse siltite unit and the lower part of the banded siltite unit, Connor (1990, 1991) mapped and sampled a more-or-less stratabound zone characterized by
Figure 5. Photos of and corresponding spidergrams for banded siltite, contact-metamorphosed banded siltite, and biotitized banded siltite: (A) Banded siltite (Ylab) exposure on the southeast rim of Blacktail pit of the Blackbird mine, showing interlayered siltite (sltt) and biotitic argil-lite (arg). Siltite beds pinch and swell, with basal scours ± mud chips, and planar tops. Upward-fining graded beds and upward-concave cross-bedding indicate an upright structural orientation. (B) Dark-brown biotite-scapolite hornfels (Bt hf) interlayered with siltite in banded siltite (Ylab), exposed in road cuts near Deep Creek. This outcrop is within a biotite-scapolite (Scp) aureole around the megacrystic monzogranite pluton of Big Deer Creek (dated 1370 ± 4 Ma by Aleinikoff et al., 2012b). (C) Biotitite (Btt) interlayered with siltite in banded siltite (sltt), exposed along the lower northeast wall of Blacktail pit. Biotitite contains at least 75 vol% of greenish-black biotite (after argillite beds). Erythrite (Er) and chalcanthite (Cth) are weathering products of sparsely disseminated cobaltite and chalcopyrite. Basal scours, cross-bedding, and upward-fining beds indicate upright bedding. (D) Spidergrams for siltite and argillite of the banded siltite unit, compared to a set of spidergrams for marine shale (gray pattern) after Thompson et al. (1984). (E) Spidergrams for biotite-scapolite hornfels after argillite, interlayered with siltite in contact- metamorphosed banded siltite, compared to a set of spidergrams for marine shale (gray pattern) after Thompson et al. (1984). (F) Spidergrams of biotitite after argillite, interlayered with siltite in hydrothermally biotitized banded siltite (stained with erythrite and chalcanthite after cobaltite and chalcopyrite), compared to a set of spidergrams for marine shale (gray pattern) after Thompson et al. (1984).

Figure 6. Photo of megacrystic granite (Yig in Fig. 2B), spidergrams for megacrystic granite and amphibolite (Yim in Fig. 2B), and a photo showing amphibolite clasts in protomylonitic breccia from the Little Deer fault zone (LDf in Fig. 2B): (A) Rapakivi texture in a sample from the megacrystic monzogranite of Big Deer Creek (Yig). A large phenocryst of pinkish K-feldspar (Kfs) is rimmed by white sodic plagioclase (NaPl) in a phaneritic matrix of quartz (Qtz), plagioclase (Pl), and biotite (Bt). An inclusion of biotitic quartzite (Btqtzt) is similar to nearby metasedimentary rocks of the Gunsight Formation. Contact relationships indicate that metasedimentary host strata were folded before granite was emplaced at ca. 1370 ± 10 Ma, according to Evans (1981, 1986), and Evans and Zartman (1990). (B) Spidergrams for megacrystic granite and meta-mafic amphibolite of the Salmon River Mountains compared to spidergrams by Thompson et al. (1984) for continental leucogranites (gray pattern) and a mafic sill, similar to continental flood basalts (dashed line). Chemical data for 92TF samples are from Lewis and Frost (2005). (C) Protomylonitic breccia with amphibolite clasts (am), found downslope from the Little Deer reverse-to-thrust fault (LDf). The cataclastic matrix contains fine- to microfragmental quartz > biotite > amphibolite. Two foliations (S and C) define protomylonitic fabric in the matrix. Quartz grains near clasts are elongate and aligned with flow foliation, which is deflected around amphibolite clasts.
Figure 7. Diagrammatic geologic map and cross section of the Blackbird mine area and its surroundings: (A) Geologic map, showing the Blackbird, Lookout, and Haynes-Stellite structural blocks, and the southwest margin of the megacrystic granite pluton of Big Deer Creek. Sources of information include Vhay (1948), Evans and Connor (1993), Johnson et al. (1998), and Tysdal et al. (2003). Labeled mines and prospects are the Bonanza Copper prospect (bc), Haynes-Stellite mine (HS), Sweet Repose prospect (sr), and Tinkers Pride prospect (tp). Labeled faults are the White Ledge fault (WLF), and the Little Deer fault (LDf). The explanation of map units here also applies to Figures 7B and 8. (B) Geologic cross section of the Blackbird mine area. The line of section (b–c) is shown on the geologic maps in Figure 7A and Figure 8. A line (labeled M) at the 2088 m elevation indicates the deepest level of previous mining (the 6850 level of the Blackbird mine, which was the main haulage level below the stopes). Dashed horizontal lines (labeled E) indicate the deepest levels of exploration beneath the Blackbird mine, the Merle zone, and the Sunshine and East Sunshine zones. Mineral abbreviations from International Union of Geological Sciences. Labeled faults are (from west to east) the White Ledge fault (WLF), Sunshine East fault (SEf), Meadow Creek fault (MCF), Hawkeye Gulch fault (HGF), Little Deer fault (LDF), and Northfield fault (NFF).
Figure 8. Diagrammatic geologic map of the Blackbird structural block, showing ore zones and prospects of the Blackbird mine area: Line b–c is the section line for the diagrammatic cross section in Figure 7B. Labels for ore zones are spelled fully, but labels for selected prospects are abbreviated as follows: Buckeye (bk), Burl (brl), Chelan (ch), East Chelan (ech), Ella (e), Horseshoe (hs), Iowa (ia), Katherine (k), Mushroom (mr), Ridgetop (rt), St. Joe (sj), and Toronto (tor). Labeled faults are (from west to east): the White Ledge reverse-right-lateral fault and shear zone (WLf), Sunshine East normal fault (SEf), East Blacktail normal fault (EBf), Meadow Creek fault zone (MCf), Dandy shear zone (Dsz), Hawkeye Gulch fault (HGf), Northfield fault (NFf), on strike with the Slippery Gulch fault (SGf), and Little Deer reverse-to-thrust fault (LDf). Stippled patterns indicate garnet-bearing rocks. Sparse stipple indicates rocks with small, sparsely scattered garnets; dense stipple indicates rocks with relatively abundant and easily visible porphyroblasts of garnet ± chloritoid. Principal sources of information include Vhay (1948) and Nash and Hahn (1989).
iron-oxide-bearing siltite beds (OZ in Fig. 2B). For 14 samples from the OZ, he reported median concentrations of 21 ppm Cu and 9 ppm Co.

In the upper part of the banded siltite unit and the lower part of the Gunsight Formation, Connor (1990, 1991) mapped and sampled a more-or-less stratabound zone characterized by biotite-rich beds between beds of siltite to quartzite (BZ in Fig. 2B). For 24 samples of biotite-rich rocks from the BZ, he reported median concentrations of more-or-less stratabound zone 19 ppm Co and 5 ppm Cu.

Lopez (1981) had suggested that low-grade syngenetic Co and Cu in metasedimentary country rocks may have been remobilized and concentrated to form Co-Cu ores of the Blackbird district. According to that suggestion, the somewhat elevated Co and Cu contents of rocks from OZ and BZ could be interpreted as precursors to higher-grade Co and Cu deposits of the Blackbird district.

Nash and Connor (1993) interpreted rocks of the OZ and BZ as siliciclastic sedimentary rocks with chemical-exhalative components, derived from submarine hot springs. That hypothesis seems consistent with the presence of rounded pebbles to cobbles of strongly magnetic mudstone in a matrix of moderately magnetic mudstone in diamictite near Tobias Creek in the Lemhi Range (southeast of the OZ). In the Salmon River Mountains, however, rocks of the OZ contain only detrital magnetic mud chips (possibly derived from the diamicite of Tobias Creek), but not of local submarine-exhalative origin in rocks of the OZ in the Salmon River Mountains.

Furthermore, those samples from the OZ that contain somewhat elevated concentrations of Cu (but not Co) are from the vicinity of the Iron Creek Cu-Co prospect area, where semi-massive magnetite deposits are stratabound, but are interpreted here as epigenetic breccia, vein, and replacement-style deposits, hosted in phyllitic biotitic siltite that was metamorphosed to biotite grade before being chloritized, veined, and partly replaced by early hydrothermal pyrite ± magnetite, and later chalcopyrite ± cobaltian pyrite.

Likewise, most samples from the BZ that contain anomalous Co (but not Cu) are from an area labeled BZg in Fig. 2B, where the geochemically anomalous samples are from beds of silicified scapolite-biotite hornfels (after argillite) between siltite beds in banded siltite within the contact-metamorphic aureole of the nearby granitic pluton of Big Deer Creek.

**EAST KOOTENAY OROGENY (Ca. 1379–1325 Ma)**

The East Kootenay orogeny affected rocks of the western margin of the Belt-Purcell Basin and much of the Lemhi subbasin from ca. 1379 to 1325 Ma. McMechan and Price (1982) applied the term “East Kootenay orogeny” to a Mesoproterozoic episode of metamorphism and granitic plutonism in the Purcell Basin of southeastern British Columbia. The southern part of the East Kootenay orogen is imposed on the Lemhi subbasin (Fig. 1A), where Evans and Zartman (1990) noted that plutons of a bimodal gabbro-granite suite (dated 1370 ± 10 Ma) intruded metasedimentary strata that had been previously folded and metamorphosed to biotite grade during an earlier stage of the East Kootenay orogeny.

**Middle Mesoproterozoic Metamorphism (1379–1325 Ma)**

Isotopic age determinations on metamorphic minerals bracket the time span of the East Kootenay orogeny between ca. 1379 and 1325 Ma. A maximum age constraint of ca. 1379 Ma is based on a Lu-Hf age determination of 1379 ± 8 Ma, reported by Zirakparvar et al. (2010) for garnets from lower Belt mica schist exposed in the Priest River metamorphic complex in northern Idaho. A minimum age constraint of ca. 1325 Ma is based on Pb-Pb age determinations on titanite in samples from chloritic rocks beneath the Sullivan Zn-Pb deposit (dated 1325 ± 21 Ma) and from regionally metamorphosed mafic sills in southeastern British Columbia (dated 1329 ± 29 Ma to 1325 ± 25 Ma), according to Schandl and Davis (2000).

McFarlane (2015) reported an age of 1365 ± 10 Ma for monazite in sillimanite metapelite in metasedimentary rocks of the lower Belt-Purcell Supergroup in the Matthew Creek metamorphic zone (MCMZ). McFarlane and Pattison (2000) estimated that peak metamorphism occurred at ~3.5 ± 0.5 kbar and 610 ± 30 °C in the Matthew Creek metamorphic zone. According to depth-pressure-temperature curves by Kern and Weisbrod (1967, p. 90), this indicates that peak metamorphism occurred at a depth of ~12 km in a thermal gradient that was abnormally hot for either continental or oceanic crust. McFarlane (2015) noted that sillimanite in the Matthew Creek metamorphic zone is strongly lineated to the northeast. He suggested that removal of Cretaceous clockwise rotation would restore this into rough parallelism with the basin margin, and that this orientation would be consistent with transpressive to transtensional tectonic settings during East Kootenay orogeny.

In the Lemhi subbasin, Doughty and Chamberlain (1996) reported that migmatic pelites near the base of the Salmon River intrusive complex contain leucosomes dated at 1370 Ma, and they noted that the leucosomes contain metamorphic minerals consistent with metamorphism and partial melting at 6.5 ± 0.5 kbar and 690 ± 50 °C. According to depth-pressure-temperature curves by Kern and Weisbrod (1967, p. 90), this indicates that peak metamorphism occurred at a depth of ~22.5 km in a thermal gradient that was higher than normal for continental crust.

O’Neill et al. (2007) mapped the Great Divide megashear (GDM in Fig. 1A) as a 3–8-km-wide mylonitic shear zone of Mesoproterozoic age with more than 50 km of left-lateral offset. This is consistent with a transpressive to transtensional setting for the East Kootenay orogeny in the Lemhi subbasin as suggested by McFarlane (2015) and with dynamothermal metamorphism, bimodal plutonism, and hydrothermal mineralization during the East Kootenay orogeny in the Lemhi subbasin.

**Middle Mesoproterozoic Plutonism (Ca. 1370–1335 Ma)**

McFarlane (2015) reported a zircon U-Pb age determination of 1365 ± 10 Ma for the peraluminous granite of Hellroaring Creek (in southern British Columbia). This pluton intruded an open anticline in low-grade metasedimentary strata of the lower
Belt-Purcell succession. McFarlane suggested that this granite intruded to a relatively high structural level, while peak metamorphism occurred at lower-crustal levels in the nearby Matthew Creek metamorphic zone. He suggested that late-stage pegmatites (1335 ± 5 Ma) were emplaced during transtensional uplift and exhumation of the Matthew Creek metamorphic zone, which he interpreted as a Mesoproterozoic metamorphic core complex.

In the Lemhi subbasin, a bimodal suite of gabbroic and granitic plutons (1370 ± 10 Ma) intruded previously folded metasedimentary strata, according to Evans and Zartman (1990). The Salmon River intrusive complex (Fig. 2B) consists of a large pluton of megacrystic monzogranite (locally metamorphosed to augen gneiss or orthogneiss) underlain by gabbroic sills (mostly metamorphosed to amphibolite). The monzogranitic pluton locally contains bulbous masses of mafic rock, hybrid dioritic rocks, and sparse mafic dikes, the ages of which are all ca. 1370 ± 10 Ma, according to Evans and Zartman (1990), Doughty and Chamberlain (1996), and Aleinikoff et al. (2012b).

**Megacrystic Monzogranite and Metamafic Amphibolite**

Megacrystic monzogranite typically contains large K-feldspar phenocrysts in a coarse-grained matrix of plagioclase, quartz, and biotite (Fig. 6). Evans and Zartman (1990) reported minor primary muscovite in some samples and noted that available chemical analyses indicate a slightly peraluminous composition. Some K-feldspar phenocrysts are rimmed by oligoclase to form rapakivi texture, so megacrystic monzogranite was characterized as A-type rapakivi granite by Schulz (2013), who noted that such granites typically have high contents of incompatible-element contents (such as REEs) but low contents of Co, Sc, Cr, Ni, Ba, Sr, and Eu.

Spidergrams for samples of megacrystic granite (Fig. 6B) generally match those for a set of leucocratic continental granitoid rocks, and a spidergram for metamafic amphibolite is similar to that of a mafic sill belonging to a set of continental flood basalts analyzed by Thompson et al. (1984). Chemical data reported by Lewis and Frost (2005) indicate that five samples of megacrystic monzogranite contain ~6 ppm Co and ~6 ppm Cu, whereas a sample of metamafic amphibolite from the Salmon River intrusive complex contains 51 ppm Co and 65 ppm Cu.

Vhay (1948) and Slack (2012) suggested that the Blackbird Co-Cu deposits are genetically related to the megacrystic monzogranite pluton of Big Deer Creek (which Vhay interpreted as a manifestation of the Cretaceous Idaho batholith, but which Evans and Zartman dated as Mesoproterozoic). That pluton is exposed ~6 km northeast of the Blackbird mine, and it likely dips southwestward beneath the Blackbird mine, as indicated by an associated low-magnetic-potential anomaly that extends toward the Blackbird district (Mankinen et al., 2004). Aleinikoff et al. (2012b) reported isotopic age determinations of 1378 ± 4 Ma for megacrystic monzogranite of Big Deer Creek, and 1370 ± 4 Ma for early hydrothermal xenotime, surrounded by cobaltite in cobaltite-biotite ore. This indicates that early hydrothermal activity in the Blackbird mine area began during or soon after crystallization of the nearby megacrystic granite pluton of Big Deer Creek.

Landis and Hofstra (2012, p. 1189) suggested that volatiles trapped in Blackbird gangue quartz permit models for ore genesis involving evaporative connate brine from metasedimentary strata, heated by and mixed with fluids from the nearby bimodal plutonic complex. With Co and Cu concentrations nearly an order of magnitude higher than those of metasedimentary country rocks or megacrystic monzogranite, metamafic amphibolite would seem a possible source rock for the Co and Cu that are concentrated in Blackbird Co-Cu ore zones. However, metamafic amphibolite has not been found in the Blackbird district, except as clasts in mylonitic breccia along the Little Deer Creek reverse-to-thrust fault (LDf in Figs. 6C, 7, and 8). Nevertheless, abundant clasts of metamafic amphibolite in this fault breccia indicate that metamafic amphibolite was present below the Blackbird structural block before it was displaced upward and eastward along the Little Deer Creek fault.

**Grenville-Age Metamorphism and Mineralization (Ca. 1200–1000 Ma)**

**Late Mesoproterozoic Metamorphism (Ca. 1190–1006 Ma)**

Anderson and Davis (1995) documented geologic and isotopic evidence of a thermal disturbance that affected rocks of the Belt-Purcell Basin, northwestern Canada, and southern Australia, at ca. 1120–1030 Ma (during the time span of the Grenville orogeny along the eastern margin of Laurentia). Vervoort et al. (2005) reported Lu-Hf age determinations of 1149–1006 Ma on cores of garnets in Belt rocks of northern Idaho, and they suggested that these indicate an episode of Grenville-age crustal thickening along the western margin of Laurentia.

Aleinikoff et al. (2015) also reported U-Pb ages of 1160–1050 Ma on metamorphic xenotime from older rocks of the Revett, McNamara, and Garnet Range Formations in the Belt basin (Fig. 3). They noted an apparent absence of pervasive deformational fabrics in the dated samples, which they interpreted to indicate that Grenville-age metamorphism was predominantly thermal, and possibly caused by regional mafic underplating of continental crust.

Grenville-age metamorphism also affected rocks in the Lemhi subbasin, according to Panneerselvam et al. (2012, p. 1185), who reported that “Whole-rock samples of the Apple Creek Formation define a highly radiogenic common lead isochron with a slope corresponding to 1190 ± 60 Ma.”

**Late Mesoproterozoic Mineralization (Stage Y3, Ca. 1200–990 Ma)**

Hydrothermal mineralization accompanied Grenville-age metamorphism of rocks in the Belt-Purcell Basin and the Lemhi subbasin, as indicated by the following age determinations on hydrothermal minerals from the Coeur d’Alene and Blackbird
In that it is concentrated along shear zones that transect previously folded beds of banded siltite. Nevertheless, ore zones also are stratigraphically controlled, in that argillitic layers in the upper banded siltite unit are preferentially sheared, biotititized, and mineralized.

### Mineral Depositional Sequence

Anderson (1947) tabulated and described mineral depositional sequences for two stages of mineralization in cobaltite-biotite lodes, cobaltite-tourmaline lodes, and cobaltite-quartz lodes; and for three stages of mineralization in gold-copper-cobalt lodes. He suggested that all stages of mineralization occurred in association with mafic magmatism in early Tertiary time. Mineral-depositional sequences described by Vhay (1948) and tabulated by Lund et al. (2011) are mostly consistent with those of Anderson (1947). However, Vhay (1948) suggested a Cretaceous age, based on the spatial association of the Blackbird district with the monzogranite pluton of Big Deer Creek, which he mistakenly correlated with the Cretaceous Idaho batholith. Lund et al. (2011) also suggested a Cretaceous age, but on the basis of perceived structural control, related to Cretaceous thrust-faulting.

The mineral depositional sequence summarized in Table 3 is constrained by observed relative age relationships, and calibrated by currently available isotopic age determinations on rocks and minerals associated with Co-Cu deposits of the Blackbird district. Such age constraints bracket an episode of Co-dominant mineralization in Mesoproterozoic time (episode Y), and an episode of Fe-Cu-dominant mineralization in Cretaceous time (episode K). The time span of Mesoproterozoic episode Y (ca. 1370–1000 Ma) includes that of the East Kootenay orogeny along the western margin of the Belt-Purcell Basin (ca. 1379–1325 Ma), and that of Grenville-age metamorphism along the western margin of Laurentia (ca. 1200–1000 Ma). The time span of Cretaceous episode K (ca. 110–83 Ma) corresponds to that of the Sevier orogeny in the Cordilleran orogen (ca. 112–85 Ma).

### MESOPROTEROZOIC COBALTITE DEPOSITS

(Ca. 1370–1000 Ma)

Cobaltite is the principal cobalt-bearing mineral of the Blackbird district. Although cobaltite [(Co>Fe)AsS] forms a solid solution series with glaucodot [(Co>Fe)AsS], the cobaltite-glaucodot series is undivided here, and is simply regarded as cobaltite, which is gunmetal gray with a pinkish-violet tarnish. Major cobaltite is commonly associated with minor quartz + major greenish-black biotite ± minor to major black tourmaline ± minor xenotime ± minor chalcopyrite in disseminated to semi-massive replacement-style deposits, breccias, and veinlets. Textural relationships indicate that early hydrothermal quartz is largely replaced by biotite, which is partly replaced by cobaltite ± minor late chalcopyrite.
**TABLE 3. TECTONISM AND MINERALIZATION IN THE BELT-PURCELL BASIN AND BLACKBIRD Co-Cu DISTRICT**

<table>
<thead>
<tr>
<th>Age (Ma)*</th>
<th>Tectono-stratigraphic features and episodes</th>
<th>Key geologic features</th>
<th>Ore and gangue minerals a</th>
<th>Alteration &amp; metam min sb</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt;1470–1454</td>
<td>Belt basin rift</td>
<td>Very thick sedimentary successions, local mafic sills and flows</td>
<td>Ccp, Py, Mrc, Bn, CoPy, Mg, Cob, Gn, Sp, Tn (Sheep Creek, sed-hosted diagenetic repl deposit in lower Belt strata) a</td>
<td>Brt, Dol, Qtz, Chl, Cal</td>
<td>K 0</td>
</tr>
<tr>
<td>1454–1379</td>
<td>East Kootenay orogeny in western Belt-Purcell basin and Lemhi sub-basin</td>
<td>F1, F2 folds</td>
<td>Ccp, Cc, Bn, Mg, Py, Gn, Sp, Po, Tet, Mag, Hem, CoPy, Cob (Spar Lake, sed-hosted diagenetic repl deposit in Revelt Fm) a</td>
<td>Brt, Scp hf</td>
<td>Y 0</td>
</tr>
<tr>
<td>1379–1325</td>
<td>East Kootenay orogeny</td>
<td>Bimodal plutonic suite</td>
<td>Granit (ca. 1370 Ma), Co-Cu-Ygeochem. (BZg) a</td>
<td>Gbt</td>
<td>Y 0</td>
</tr>
<tr>
<td>1325–1270</td>
<td>Bimodal plutonic suite</td>
<td>F2 folds in Cob ore</td>
<td>Stratabound Cob1—dissem, repl, bx (BB) a</td>
<td>Qtz, bx, gbt</td>
<td>Y 1</td>
</tr>
<tr>
<td>1200–1000</td>
<td>Windermere rift</td>
<td>Within-plate plutons and dikes that cut Cob ore in (BB district)</td>
<td>Qzt, Mag, Ccp, Au (Cu Camp, vn, repl) a</td>
<td>Bt</td>
<td>Y 2</td>
</tr>
<tr>
<td>665–650</td>
<td>Laurentia to Pangea</td>
<td>Rifted margin, Laurentia (?)</td>
<td>Qzt, Cal, Sid, Py, Ccp, Gn, Tet, Hem, Au (Yellowjacket, vn, vnl, dissem) a</td>
<td>Qzt, Chl, Ser, Cla</td>
<td>Y 2</td>
</tr>
<tr>
<td>550–485</td>
<td>Laurentia to Pangea</td>
<td>Laurentia (?)</td>
<td>Qzt, Au-Au-Bi vnl + Ser (ca. 83 Ma) (BB)</td>
<td>Ser</td>
<td>K 1</td>
</tr>
<tr>
<td>155–142</td>
<td>Laurentia to Pangea</td>
<td>Laurentia (?)</td>
<td>Qzt-Py-Au-Ag vnl, dissem (Beartrack) a</td>
<td>Ser</td>
<td>K 2</td>
</tr>
<tr>
<td>142–112</td>
<td>Laurentia to N. America—Cordilleran Orogen</td>
<td>Lull in orogenesis</td>
<td>Mnz (ca. 110–92 Ma), Xtm (ca. 104–93 Ma) (BB)</td>
<td>Grt, Bt, Cld</td>
<td>K 0</td>
</tr>
<tr>
<td>112–85</td>
<td>Sevier orogeny</td>
<td>Metamorphic-plutonic hinterland and fold-thrust belt with F1 folds</td>
<td>Mznz (ca. 112–92 Ma), Xtm, (ca. 104–93 Ma) (BB)</td>
<td>Grt, Sil (SC) a</td>
<td>K 1</td>
</tr>
<tr>
<td>85–55</td>
<td>Laramide orogeny</td>
<td>Basement-cored uplifts</td>
<td>Ccp, Qtz, Py, Po, Sid, CoPy vnl (BB, BP) a</td>
<td>Ms, Chl</td>
<td>K 2</td>
</tr>
</tbody>
</table>

*See text and Figures 3 and 4 for sources of age determinations and age ranges summarized in this table.


c. Tectono-stratigraphic features and episodes: (BB)—Blackbird district, (BP)—Black Pine, (BZg)—biotite-zone geochemical anomaly near Deep Creek, (IC)—Iron Creek, (ICB)—Idaho cobalt belt, (N. ID)—northern Idaho, (SC)—Salmon Canyon Copper.


e. Fold-set and cleavage-set abbreviations: F1, F2, F3, F4—numbered fold sets, S1—axial-plane cleavage of the F2 fold set. S2 cleavage is transitional to shear-slip cleavage, and to transposed layering, related to the F2 fold set.


g. Place-name abbreviations: (BB)—Blackbird district, (BP)—Black Pine, (BZg)—biotite-zone geochemical anomaly near Deep Creek, (IC)—Iron Creek, (ICB)—Idaho cobalt belt, (N. ID)—northern Idaho, (SC)—Salmon Canyon Copper.

h. In high grade composite Co-Cu ore zones of the Blackbird district, minerals of the K 0 and K 1 assemblages are superimposed on minerals of the Y 1 and Y 2 mineral assemblages.
Xenotime in Cobaltite Deposits (Ca. 1370–1000 Ma)

Aleinikoff et al. (2012b) reported U-Pb age determinations on xenotime in cobaltite-biotite ore from the Merle ore zone. Backscattered-electron (BSE) images in Figure 9 show cores and rims of selected xenotime grains dated by Aleinikoff et al. (2012b). Most of the samples of xenotime in cobaltite-biotite ore yielded Middle Mesoproterozoic age determinations between ca. 1370 and 1270 Ma—a time span which overlaps with that of the East Kootenay orogeny (ca. 1379–1325 Ma). However, Aleinikoff et al. (2012b) reported two Late Mesoproterozoic age determinations of 1058 ± 11 Ma and 990 ± 12 Ma for grains of hydrothermal xenotime, from the Merle ore zone. This time span overlaps with that of Grenville-age metamorphism along the western margin of Laurentia.

In Figure 9A xenotime, (1370 ± 4 Ma) is surrounded by cobaltite, which is rimmed by cobaltite. In Figure 9B xenotime, (1316 ± 6 Ma) is rimmed by Cretaceous xenotime, (101 ± 2 Ma). Figure 9C shows a grain of xenotime, that contains abundant cobaltite inclusions. That xenotime grain was not dated, but its vaguely splotchy zoning appears similar to that shown in grains of xenotime, that were dated 1316 ± 6 Ma and 1271 ± 7 Ma (Figs. 9B and 9D). The core of the xenotime grain dated ca. 1271 Ma (Fig. 9D) contains a white inclusion, which may be older cobaltite. A surrounding rim of xenotime, resembles that dated ca. 101 Ma (Fig. 9B).

If an episode of cobaltite deposition accompanied or closely followed each episode of xenotime deposition, these xenotime age determinations would indicate multiple stages of cobaltite deposition during Mesoproterozoic time, as follows: stage $Y_0$ (ca. 1370 Ma)—geochemically anomalous Co and Y in silicified scapulitic biotite hornfels (BZg in Fig. 2B), stage $Y_1$ (ca. 1370–1320 Ma)—cobaltite, biotite ore + xenotime, (in strat-abound replacement-style ore that is locally F$_2$-folded), stage $Y_2$ (ca. 1320–1270 Ma)—cobaltite, biotite ore + xenotime, (along axial planar $S_2$ cleavage of F$_2$ folds), and stage $Y_3$ (ca. 1060–990 Ma)—cobaltite, + xenotime, (in discordant quartz-tourma-line breccias).

Stratabound Cobaltite -Biotite Ore ± Xenotime
(Stage $Y_1$, Ca. 1370–1320 Ma)

Textural relationships between xenotime (dated 1370 Ma and surrounded by cobaltite, rimmed by cobaltite,) indicate that cobaltite, is younger than 1370 Ma but older than cobaltite, (Fig. 9A). However, xenotime grains with cobaltite inclusions appear similar to xenotime grains dated 1316 ± 6 to 1271 ± 7 Ma (Figs. 9A–9D and captions). These relationships are interpreted to indicate that cobaltite, probably is older than ca. 1316 Ma, so its approximate age probably is bracketed between ca. 1370 and 1320 Ma.

Cobaltite-biotite ore of stage $Y_1$ is concentrated in strata-bound ore zones of disseminated to semimassive cobaltite, surrounded by greenish biotite ± tourmaline ± xenotime, (Figs. 9E and 9F). Such ore is preferentially hosted in layers of greenish-black biotitite (between layers of siltite) in the upper part of the banded siltite unit. In relatively undeformed cobaltite-biotite ore, cobaltite-rich bands (or layers) appear stratiform, and cobaltite, grains generally appear subcubic to equant but rounded.
However, cobaltite₂-biotite ore is locally folded and lineated, with lines of cobaltite grains (Fig. 9G[a]) that appear point-like as viewed along fold axes (Fig. 9G[b]). Likewise, Vhay (1948) noted that in crumpled zones of tight folds (like the F₁ folds shown in Fig. 9H), high-grade cobaltite-biotite ore is concentrated in cigar-shaped ore shoots that plunge north-northeast (parallel to F₁ fold axes). Thus, cobaltite₁ was deposited before or during F₁ folding, and was locally deformed during F₁ folding, to form cigar-shaped lodes with lineated fabrics that parallel fold axes in crumpled zones of F₁ folds.

**Cobaltite₂-Biotite Ore along S₂ Cleavage (Stage Y₂, Ca. 1320–1270 Ma)**

Cobaltite₂-biotite ore of stage Y₂ commonly follows S₂ cleavage of F₂ folds, as shown in Figure 10A. In crumpled zones of tight F₂ folds, S₂ cleavage is transitional to S₃ shear-slip cleavage, which is transitional to S₄, transposed layering, in which siltite boudins are interlayered with phyllitic biotitite (Fig. 10B).

Cobaltite₂-biotite lodes of the South Idaho ore zone are in the tight, north-plunging hinge zone of the Idaho syncline, where they are preferentially hosted in biotitite phyllite between siltite boudins. The Sunshine and Sunshine East ore zones are on opposite limbs of the nearly isoclinal and west-vergent Sunshine syncline (Fig. 7B), where cobaltite₂ is preferentially hosted in chloritized biotitite phyllite between boudins of sugary-textured quartzite, interpreted here as silicified and recrystallized siltite of the banded-siltite unit (Fig. 10C).

Inasmuch as xenotime₂, dated ca. 1320 to 1270 Ma is similar to xenotime₁ that contains cobaltite inclusions, this is interpreted to indicate that cobaltite₂ probably was deposited between ca. 1320 and 1270 Ma, a time interval that overlaps with late post-metamorphic stages of the East Kootenay orogeny (ca. 1379 to 1325 Ma).

**Quartz-Biotite and Quartz-Tourmaline Breccias ± Cobaltite₂ (Stage Y₃, Ca. 1200–1000 Ma)**

Quartz-biotite and quartz-tourmaline breccias and veins ± cobaltite₂ typically contain partly resorbed remnants of whitish quartz clasts in a matrix of biotitite or tourmalinite (after microbreccia). Cobaltite in such breccias is tentatively classified as cobaltite₂, because these breccias are regarded as later than replacement-style deposits of cobaltite₁, and this supports a logical correlation with hydrothermal xenotime₂ (dated ca. 1058 Ma and 990 Ma by Aleinikoff et al., 2012b). This corresponds to a Late Mesoproterozoic episode of Grenville-age metamorphism in the western Belt-Purcell Basin (Vervoort et al., 2005) and the Lemhi subbasin (Panneerselvam et al., 2012).

**Quartz-Biotite Breccias and Veins ± Cobaltite₂**

Quartz-biotite breccias commonly contain clasts of whitish quartz ± clasts of biotitic siltite and phyllite) in a matrix of biotitite ± disseminated to semi-massive cobaltite₂. Matrix and clasts are either cut by veinlets of cobaltite₂ (Fig. 10D), or partly replaced by semi-massive to massive cobaltite₂ (Fig. 10E). Some quartz-biotite breccias are not notably foliated, but others are well foliated, with aligned lenticular clasts of quartz ± siltite in a matrix of biotitite ± disseminated cobaltite ± siltite assemblage. Some quartz-biotite breccias are notably foliated, and they are elongate subparallel to lenticular clasts, or conform with a subparallel foliation in the matrix. Tabular quartz-biotite veins contain cobaltite books in a matrix of whitish quartz but have not been found to contain cobaltite₂.

**Quartz-Tourmaline Breccias ± Cobaltite₂**

Quartz-tourmaline breccias typically contain whitish quartz clasts in a black matrix of microcrystalline tourmalinite ± cobaltite₂ ± minor chalcopyrite). Dike-like to pipe-like bodies of quartz-tourmaline breccia and stratabound bodies of tourmalin-
ite in quartzite are widely scattered in a broad belt that extends east-southeastward across the south end of the Blackbird district (Fig. 2B). Some of these quartz-tourmaline breccias are enveloped by biotitized host rocks. According to Trumbull et al. (2011), tourmaline of the Blackbird district is aluminous school-dravite with B-isotopic ratios that indicate an isotopically heavy, saline marine-fluid source for the contained boron.

The sample of quartz-tourmaline-cobaltite breccia shown in Figure 10F is from the Haynes-Stellite mine. That sample exhibits at least two generations of breccia, with clasts of the older breccia in the younger breccia. Clasts of both breccias are elongate, streamlined, and similarly aligned, as are cobaltite veinlets, which are deflected around clasts of the older breccia. These relationships are interpreted to indicate multiple upward pulses of high-velocity fluid flux, resulting in multiple episodes of brecciation and tourmalinization, followed by cobaltite deposition in fractures along flow foliation. Thus, quartz-tourmaline and quartz-biotite breccias (± cobaltite) are interpreted to indicate cyclical upward release and forceful injection of overpressured hydrothermal fluid, as modeled by Sibson (2004).

Although most quartz-tourmaline breccias are peripheral to the Blackbird mine area, Vhay (1948) mapped a dike-like body of quartz-tourmaline breccia that strikes north and dips steeply through the South Idaho ore zone (Fig. 8). That quartz-tourmaline breccia appears to cut previously biotitized rocks, containing cobaltite lodes, but it also contains abundant cobaltite veinlets.

In the northwestern part of the Blackbird Co-Cu district, Vhay (1948) mapped a long quartz tourmaline vein west of the Tinkers Pride prospect, and black tourmalinite also is scattered on the dump of a collapsed adit near the Bonanza Copper prospect. Quartz-tourmaline veins locally contain pyrite and chalcopyrite but have not been found to contain cobaltite.

MAFIC-ALKALIC MAGMATISM (Ca. 530–485 Ma)

Gillerman (2008) and Lund et al. (2010) reported U-Pb zircon age determinations of ca. 530 to 485 Ma for an assemblage of gabbroic to syenitoid plutons in east-central Idaho. Samples from these intrusions have incompatible-element profiles indicative of mantle-derived within-plate magmatism (Lund et al., 2010). An episode of such magmatism occurred east of the rifted continental margin, in association with early subsidence and growth of the Cordilleran miogeocline in Cambrian–Ordovician time, as suggested by Evans (1984) and Lund et al. (2010).

Mafic Dikes of the Blackbird District

Mafic dikes are common in the Blackbird Co-Cu district, and they also occur at other mines and prospects in the Idaho cobalt belt. At least four types of mafic dikes are present in the Blackbird district—gabbroic dikes, amphibole-biotite-plagioclase dikes, biotite-rich dikes ± plagioclase, and ultramafic-breccia dikes and diatremes.

A gabbroic dike that is ~20 m thick trends north to northwest along the western margin of the Idaho ore zone (mg in Fig. 8). That dike cuts F2 folds, and its margins are flow-foliated parallel to its contacts, but its interior is not foliated. Vhay (1948) noted that gabbroic dikes consist of plagioclase (partly replaced by albite and zoisite), clinopyroxene (mostly replaced by dark-brown biotite), and amphibole (mostly replaced by chlorite and clinopyroxene). The outcropping gabbroic dike contains traces of very fine-grained pyrrhotite ± pyrite, and it is cut by a quartz vein. Similar dikes were encountered underground in the Uncle Sam where they cut cobaltite lodes, but are cut by mineralized faults, according to Vhay (1948).

A thinner mafic dike cuts across F2 folds in banded siltite, as shown in Figure 9H. This dike consists mostly of fine-grained intergrowths of plagioclase, amphibole, and biotite. It contains sparse inclusions of similar but coarser-grained rocks with higher proportions of plagioclase. It also contains sparsely disseminated grains and discontinuous veinlets of cobaltian arsenopyrite. Some biotite is altered to white mica, and lenticular vugs are rimmed by limonite-stained chlorite.

Biotitic mafic dikes (Figs. 11A and 11B) are widely distributed throughout the Blackbird mine area, but most are less than ~3 m thick and are too narrow to be represented in Figure 8. Biotitic mafic dikes consist mostly of reddish brown biotite, but some contain small phenocrysts of sodic plagioclase. Although biotitic mafic dikes are variously altered and metamorphosed, they probably were originally lamprophyric. Some biotitic dikes contain inclusions of siltite to quartzite, and one contained an inclusion of calcite-tremolite rock with carbon- and oxygen-isotopic compositions consistent with a magmatic history (either as carbonatite magma, or as limestone that reacted with mafic magma, according to Johnson et al., 2012).

Hahn and Hughes (1984) described ultramafic diatremes and breccia dikes containing fragments of carbonatite, peridotite, pyroxenite, and anorthosite. Drill-core logs indicate that an ultramafic breccia dike cuts cobaltite-bearing quartz-tourmaline breccia at the Conicu prospect (cn in Fig. 2B).

Blackbird mafic dikes have not yielded U-Pb zircon age determinations, because they have not been found to contain zircon (ZrSiO4). However, an effort is under way to find and date baddeleyite (ZrO2), which is more likely to be present in such silica-poor rocks. Until mafic dikes can be directly dated, a probable Late Cambrian to Early Ordovician age is indicated by correlation with nearby intrusions that are geochemically similar and have been dated at 530–485 Ma by Evans (1984), Gillerman (2008), and Lund et al. (2010). Spidergrams in Figure 11C show that patterns of chondrite-normalized concentrations of incompatible elements in mafic dikes of the Blackbird district closely match those of a mafic dike associated with the syenitoid intrusive complex of Deep Creek (dated 485 Ma by Lund et al., 2010). These spidergrams also indicate that Blackbird mafic dikes are enriched in Nb and Ta relative to La, as is typical for ocean-island basalts and alkaline continental basalts, according to Thompson et al. (1984). Thus, Blackbird mafic dikes lack the depletion of
the ICB (as noted by Nold, 1990).

However, lower-grade metasedimentary rocks of the biotite zone that regional zone of garnet-grade metamorphic rocks (Fig. 1B).

northwestern part of the Blackbird structural block are within central Idaho. The northwestern part of the ICB and the upper

cordillera was accreted to the western margin of North America from ca. 155 to 55 Ma (Armstrong and Ward, 1991).

Cordilleran orogenesis affected rocks of the Belt-Purcell Basin from ca. 155 to 55 Ma. According to DeCelles (2004), northwest-southeast compression occurred in the northwestern part of the North American plate from ca. 155 to 142 Ma (during the Nevadan orogeny). This was followed by a lull and westward retreat of subduction-related magmatism from ca. 142 to 112 Ma, and then by the Cretaceous Sevier orogeny, during which there was widespread metamorphism and plutonism in the hinterland (Armstrong and Ward, 1993), and northeast-vergent folding and thrusting in the fold-thrust belt of the Cordilleran orogen (Fig. 2A and Table 3). The Insular Superterran e was accreted to the western margin of North America from ca. 105 to 90 Ma (Giorgis et al., 2008), and the Idaho batholith and associated plutons were emplaced mostly between ca. 90 and 55 Ma (Lewis et al., 2012). Cordilleran orogenesis was then followed by extensional tectonism and bimodal magmatism from ca. 55 to 0 Ma (Armstrong and Ward, 1991).

Regional Garnet Zone

Lewis et al. (2012) outlined a regional zone of garnet-bearing metamorphic rocks in the hinterland of the Cordilleran orogen in central Idaho. The northwestern part of the ICB and the upper northwestern part of the Blackbird structural block are within that regional zone of garnet-grade metamorphic rocks (Fig. 1B). However, lower-grade metasedimentary rocks of the biotite zone characterize the southeastern parts of the Blackbird district and the ICB (as noted by Nold, 1990).

Blackbird Garnet Zone ± Chloritoid

Garnet porphyroblasts are common in metasedimentary rocks of the upper-northern and down-faulted western parts of the Blackbird mine area (Fig. 8), where subhedral porphyroblasts of dark-red almandine are preferentially hosted in layers of biotite phyllite to schist. Eiseman (1988) estimated peak metamorphic temperatures of 400–520 °C, based on biotite-garnet geothermometry.

Chloritoid porphyroblasts are locally present in rocks of the garnet zone, where they are preferentially hosted in layers of biotitic siltstone and biotitic metasandstone. Garnet and chloritoid porphyroblasts generally transect S$_2$ schistosity, and in ore zones, they commonly contain cobaltite inclusions (Figs. 11D and 11E). Some mafic dikes in the garnet zone also contain garnets, so garnet and chloritoid porphyroblasts are younger than cobaltite, S$_2$ schistosity, and mafic dikes.

Age Determinations on Metamorphic Garnet and Biotite (Ca. 151–93 Ma)

Zirakparvar et al. (2007) reported Lu-Hf age determinations of 151 ± 32 Ma for garnets from the Blacktail pit, and 112.8 ± 7.7 Ma for garnets from the Salmon Canyon Copper mine. An additional Lu-Hf age determination of 93 ± 8.3 Ma is reported here for garnets from a schistose biotitic mafic dike near the Tinkers Pride prospect (Fig. 1B) in the Lookout structural block (Fig. 7A).

$^{40}$Ar/$^{39}$Ar step-heating plateau ages of 151 ± 1 Ma and 122 ± 1 Ma also are reported here for metamorphic biotite in schistose biotitic mafic dikes that cut garnet-bearing metasedimentary rocks in the Ram zone (Fig. 8). These step-heating plateau ages are interpreted to indicate when biotite last cooled below ~350 °C. Figure 11D shows relic biotite in retrograde chlorite, which also rims a garnet porphyroblast.

Gradational Base of the Garnet Zone

The base of the garnet zone is gradational, as previously mapped by Vhay (1948) and Nash and Hahn (1989), and as represented in Figures 2B, 7B, and 8. The gradational base of the garnet zone is well exposed on the upper slope of the northeast wall of the Blacktail pit, and it has also been observed in drill core from the Ram zone. On the geologic map in Figure 8, the garnet-in transition is manifested by different versions of the garnet-in boundary, as mapped by different geologists. An outer garnet-in line represents the lower boundary of rocks with small, sparsely scattered garnets, whereas an inner garnet-in line represents the lower boundary of rocks with plenty of easily visible garnets.

As indicated on Figure 8, the gradational base of the garnet zone does not offset ore zones, biotitite-rich intervals, or mapped folds, so it is not a thrust fault (as previously mapped by Tysdal et al., 2003; Lund and Tysdal, 2007; and Lund et al., 2011). Nevertheless, the garnet zone is locally truncated and displaced by post-garnet faults, such as the thrust fault of Little Deer Creek, and the Sunshine East normal fault (Figs. 7 and 8).
The garnet-in transition is interpreted as a gradational metamorphic isograd along the base of a downward-decreasing metamorphic gradient, probably below a hotter thrust plate of higher-grade metamorphic rocks from a deeper crustal level. The observation that biotitic mafic dikes in the garnet zone are schistose, whereas biotitic dikes well below the garnet zone are not schistose, is consistent with the hypothesis that dynamothermal metamorphism was imposed from above by an over-riding thrust plate.

As shown in Figures 1B and 2B, biotite gneiss of the Clearwater orogenic zone lies north of and above the inferred Salmon Canyon thrust fault. This Clearwater thrust plate, which is composed largely of amphibolite-facies metasedimentary rocks, may have been thrust upward and southeastward over the northwestern part of the ICB, causing garnets and biotite to form in its footwall, beginning at ca. 151 Ma. Northeast-plunging F3 folds may also have formed in response to northwest-southeast compression during the Late Jurassic to Cretaceous Nevadan orogeny (ca. 155 to 142 Ma). Such F3 folds are common in metasedimentary rocks of the Haynes-Stellite block (Fig. 2B), and are superimposed on the southwest limb of a major F2 fold (the Idaho syncline), where it is exposed along the upper northeast rim of the Blacktail pit.

Younger garnets (dated ca. 113 Ma) from sillimanite-garnet quartz-biotite gneiss in the footwall of the Salmon River thrust fault at the Salmon Canyon Copper prospect may indicate that Salmon River fault was reactivated during regional northeast-vergent F4 folding and thrusting during the Sevier orogeny (ca. 112 to 85 Ma). Thus, the Salmon River fault may have been reactivated as a lateral ramp, along which the Clearwater thrust plate may have been displaced northeastward during the Sevier orogeny.

The Iron Lake thrust plate probably was thrust northeastward over the northwestern part of the ICB during the Sevier orogeny, and it too may have metamorphosed rocks in its footwall. However, metasedimentary rocks of the Iron Lake thrust plate appear to contain few garnets. This may indicate that quartzitic rocks of the Iron Lake thrust plate were either less chemically favorable to garnet growth, or were less hot than sillimanite-garnet-quartz-biotite gneiss of the Clearwater thrust plate.

A description by Johnson et al. (1998) of the Mary Ann prospect, which is in the proximal footwall of the Iron Lake thrust plate, did not mention garnets. However, a schistose mafic dike in the Lookout structural block (also in the footwall of the Iron Lake thrust plate) contains large and abundant garnets (dated ca. 93 Ma). A thin-section of a sample from that dike shows a rolled garnet with a crenulated quartz-inclusion train that is offset from crenulated schistosity in the dike. This is interpreted to indicate that deformation to form crenulated schistosity began before, and continued after ca. 93 Ma. Orientations and asymmetries of two complementary sets of crenulations in schistose rocks of the Blackbird structural block are consistent with up-to-the northeast compression, which also is consistent with regional northeast-vergent F4 folds and thrust faults of Cretaceous age in the Cordilleran orogen (Fig. 2A) and the Salmon River Mountains (Fig. 2B).

**IRON-COPPER (Fe-Cu)-SULFIDE DEPOSITS**

**STAGE K1, CA. 110–92 MA**

Fe-Cu-sulfide deposits of the Blackbird district contain minerals of a characteristic suite that includes: early-to-late quartz ± monazite ± sparse magnetite ± locally abundant pyrrhotite + pyrite + late chalcopyrite + siderite ± later chlorite ± cobaltian arsenopyrite (rather than cobaltite). Cobaltian arsenopyrite occurs as silver-white-metallic rhombic prisms with a pale pinkish-silver tarnish (as compared to gun-metal gray cobaltite with violet-tinted tarnish, and silver-white arsenopyrite, which does not tarnish noticeably).

**Fe-Cu-Sulfide Breccia, Vein, and Replacement Deposits**

**STAGE K1**

Fe-Cu sulfide-bearing breccias and veins cut cobaltite lodes, and veins of this assemblage also cut mafic dikes, which cut cobaltite lodes (Fig. 11A). In composite Co-Cu ore zones, Fe-Cu sulfides of this assemblage rim and partly replace cobaltite. Aleinikoff et al. (2012b) reported U-Pb age determinations of ca. 144 to 88 Ma but mainly between ca. 110 and 92 Ma for hydrothermal monazite, and 104 to 93 Ma for Cretaceous xenotime rims on cores of Mesoproterozoic xenotime from the Merle and Sunshine Co-Cu ore zones (Fig. 9B).
Panneerselvam et al. (2012) reported highly radiogenic Pb in pyrite, chalcopyrite, pyrrhotite, and cobaltite (which may have been cobaltian arsenopyrite, rather than cobaltite, as indicated by its accompanying Fe-Cu sulfides). Johnson et al. (2012) reported highly radiogenic Sr in siderite from a quartz-chalcopyrite-siderite vein in the composite Chicago Co-Cu ore zone. These results were interpreted to indicate Cretaceous extraction of Pb and Sr after long-term evolution of Pb and Sr in metasedimentary source rocks of Mesoproterozoic age.

Fluid-inclusion studies by Nash and Hahn (1989) and Landis and Hofstra (2012) indicate that gangue quartz was deposited at temperatures between ca. 375 and 250 °C, probably as temperatures decreased after garnet-grade metamorphism. Fe-Cu-sulfide breccias, veins, and replacement-style deposits are not folded, and their sulfide minerals are not foliated.

**Merle Fe-Cu-Sulfide Breccia**

Stratabound Fe-Cu-sulfide breccia in the Merle zone is chalcopyrite-rich. The breccia contains clasts of quartz and siderite in a matrix of biotitite. The biotitite surrounds remnants of microbreccia with consistently oriented foliation (interpreted as flow foliation). Both matrix and clasts contain minor pyrrhotite and abundant late chalcopyrite (Fig. 12A). Irregularly shaped aggregates of Fe-Cu sulfides curve around the clasts, but also penetrate them. These relationships are interpreted to indicate that biotitized microbreccia and clasts are partly replaced by later ore minerals.

**Dandy Fe-Cu-Sulfide Breccia**

The Dandy Fe-Cu-sulfide breccia zone contains multiple tabular bodies of Fe-Cu-sulfide breccia, which generally strike north and dip steeply, along the Dandy shear zone (Fig. 8). This shear zone appears to truncate a gabbronorite dike and offset the Chicago ore zone relative to the north Idaho ore zone (Fig. 8).

A sample of Dandy breccia, shown in Figure 12B, contains clasts of vein quartz and greenish-black biotitite-phylilitic in a massive-sulfide matrix of intergrown pyrrhotite and pyrite. Thin chalcopyrite veinlets cut the biotitite clast, which also is partly rimmed by lacy intergrowths of quartz and chalcopyrite. Small subhedral crystals of cobaltian arsenopyrite are disseminated throughout the biotitite clast and its pyrrhotite-pyrite matrix. Some cobaltian-arsenopyrite rhombs crosscut chalcopyrite veinlets, and some occur in drusy quartz, which rims and lines open vugs. Minor marcasite follows rough fractures through brittle pyrrhotite and pyrite. Such marcasite may be of supergene origin.

**Quartz-Fe-Cu-Sulfide Veins**

Quartz-Fe-Cu-sulfide veins, veinlets, and pods typically contain various combinations of early quartz ± pyrite ± pyrrhotite ± late chalcopyrite ± cobaltian arsenopyrite ± siderite (Figs. 12C and 12D). Anderson (1947) also reported traces of sphalerite, enargite, galena, and native silver in samples from the Uncle Sam mine (now collapsed). He considered these minerals to be younger than pyrrhotite but older than chalcopyrite.

**Figure 12.** Photos of samples of late Fe-Cu-sulfide breccias, quartz-Fe-Cu-sulfide veins, and composite Co-Cu ores: (A) Chalcopyrite-mineralized breccia from the Merle zone (drill-core sample M7A-174-193 ft, or 53–59 m) contains clasts of hydrothermal quartz and siderite in a flow-foliated microbreccia matrix that is deflected around clasts and is mostly altered to greenish-black biotitite. Chalcopyrite and pyrrhotite replace the biotitized matrix and fill cracks in clasts. Although sulfides follow foliation, the sulfides appear undeformed, indicating that they were deposited after dynamic emplacement of the breccia. (B) Dandy breccia, containing clasts of hydrothermal quartz (Qtz), quartz-breccia (Qtz bx), and biotitite (btt) in a massive-sulfide matrix of intergrown pyrrhotite (Po) and pyrite (Py), locally cut and partly replaced by lacy networks of chalcopyrite (CcP), and also containing sparsely disseminated cobaltian arsenopyrite (CoApy). Open vugs are lined with drusy quartz and rimmed by quartz, containing disseminated cobaltian arsenopyrite. This sample was collected by M.L. Zientek on the 6850 level of the Blackbird mine. (C) Chalcopyrite-quartz vein and veinlet, both along S3 cleavage in F2-folded banded siltite, containing older biotitite ± cobaltite, in F2-folded banded siltite, and biotite ± cobaltite, along S3 cleavage (underground photo by J.F. Slack, 6850 level of the Blackbird mine). (D) Close-up photo showing part of a small quartz-Fe-Cu-sulfide vein, containing chalcopyrite, pyrrhotite, quartz, pyrite, and cobaltian arsenopyrite. This vein belongs to a set of small veins and veinlets that follow and locally transect the margin of a mafic dike in structurally transposed banded siltite. (E) Chalcopyrite-quartz vein and veinlet, both along S3 cleavage in F2-folded banded siltite, containing older biotitite ± cobaltite, in F2-folded banded siltite, and biotite ± cobaltite, along S3 cleavage (underground photo by J.F. Slack, 6850 level of the Blackbird mine). (F) Composite Cu-Cu ore from the Chicago zone (northwest of the Dandy breccia zone). This sample (GHBb08) contains a tabular body of hydrothermal breccia with lenticular quartz clasts in a biotitic matrix with alternating bands of disseminated to semimassive cobaltite that is locally cut, rimmed, and partly replaced by pyrite and chalcopyrite. Various cobaltite-rich bands generally parallel vein walls but are deflected around quartz clasts. By contrast, some large grains and strings of cobaltite grains are elongate nearly perpendicular to depositional banding. Such cross-fabric is attributed to cobaltite recrystallization, possibly during or after Cretaceous metamorphism. Some large grains of recrystallized cobaltite have cores of bright bluish-white safflorite (Saf), possibly exsolved during recrystallization of cobaltite with excess As. Along the right margin of this sample, a discontinuous pyrite veinlet cuts cobaltite but is cut by chalcopyrite, which locally cuts, rims, and partially replaces recrystallized cobaltite. Thus, cobaltite recrystallized before chalcopyrite was deposited to form composite Co-Cu ore. (F) Composite Co-Cu ore from the Blacktail zone. This sample (GHBb16) contains scattered to semimassive cobaltite, much of which is localized along discontinuous biotitic streaks (possibly after argillite) between quartz bands and lenses (possibly after siltite that is silicified and recrystallized) in structurally transposed banded siltite. Cobaltite is present in both biotite and quartz, but is preferentially concentrated along biotitic streaks. Strings of cobaltite grains are locally aligned at ~70° to the banded fabric of the sample. Such cross-fabric is interpreted to indicate recrystallization of cobaltite, possibly during Cretaceous metamorphism. Such recrystallized cobaltite (Cobrect) is locally cut, invaded, surrounded, and partly replaced by pyrrhotite + pyrite and later chalcopyrite, which lack metamorphic fabric. Thus, cobaltite appears to have recrystallized before chalcopyrite was deposited to form composite Co-Cu ore.
Geologic history of the Blackbird Co-Cu district in the Lemhi subbasin of the Belt-Purcell Basin

A

B

C

D

E

F
Quartz-Fe-Cu-sulfide veins cut cobaltite-biotite ore, F2 folds, mafic dikes, and garnet-biotite phyllite. Such veins are relatively undeformed, and they lack metamorphic fabric. This is interpreted to indicate that they probably formed after the peak of Cretaceous dynamothermal metamorphism in the Blackbird district.

Au-BEARING QUARTZ VEINLETS (STAGE K2, Ca. 92–83 Ma)

According to Anderson (1947), Au recovered from the Uncle Sam mine was from electrum-bearing quartz veinlets, which cut quartz-chalcopyrite veins. In the Ram zone, Au-enriched intercepts coincide with intervals containing late quartz veinlets ± bismuthinite ± Bi ± electrum (Au-Ag amalgam). A quartz-magnetite-chalcopyrite vein was prospected for gold at the Anaconda prospect, and a similar quartz-magnetite veinlet was found on a prospect dump near the Sunshine ore zone. Late quartz veinlets probably are of Late Cretaceous age, as indicated by a 40Ar/39Ar age determination of 82.9 ± 1.1 Ma, reported by Lund et al. (2011) for muscovite around a late quartz veinlet.

COMPOSITE Co-Cu-AuORES

Composite Co-Cu-Au ores consist of cobaltite-biotite ore of probable Mesoproterozoic age, in which cobaltite grains are cut, surrounded, or partly replaced by Fe-Cu ore minerals of Cretaceous age ± later Au-bearing quartz veinlets. In composite Co-Cu ores of the Chicago, Brown Bear, Blacktail, Ram, and Sunshine ore zones, Cu grades and abundances of chalcopyrite, pyrrhotite, and pyrite generally decrease with increasing lateral distance from the Cretaceous Dandy breccia zone. Thus, average Cu grades decrease laterally from 2.4% in the Dandy zone, to 1.2% in the Chicago zone, to 0.82% in the Brown Bear zone, to 0.74% in the Ram zone, and to 0.3% in the Sunshine zone. The Blacktail zone, however, averaged an exceptionally high 1.7% Cu, probably because it included supergene-enriched Cu ore that was mined from the open pit.

Examples of Ore from Composite Co-Cu Ore Zones

Chicago Zone

Figure 12E shows an example of composite ore from the Chicago zone (northwest of the Dandy zone). This is a tabular, vein-like body of composite ore, in which early quartz-biotite-cobaltite breccia is bounded by wall rocks of interlaminated micromosaic quartz and biotite ± very fine-grained tourmaline and cobaltite. Within this breccia, alternating bands of disseminated to semimassive cobaltite are surrounded by greenish-black biotite (after microbreccia). Various cobaltite-rich bands generally parallel the breccia walls but curve around elongate quartz clasts, aligned parallel to the centerline of the breccia. This banding probably follows flow foliation in a microbreccia matrix, mostly replaced by biotitite and cobaltite during Mesoproterozoic mineralization.

Some cobaltite grains and strings of smaller cobaltite grains are elongate nearly perpendicular cobaltite-rich bands, quartz clasts, and wall-rock contacts. Such cross-aligned cobaltite fabrics are labeled “Cob1,” in Figure 12E, and they are interpreted to indicate recrystallization of cobaltite in response to metamorphism in a post-depositional stress field. Safflorite [(Co, Fe)As2] in the cores of some large cobaltite grains may have exsolved during recrystallization of cobaltite with excess As.

Late pyrite and chalcopyrite are interpreted as products of Cretaceous mineralization, superimposed on Mesoproterozoic cobaltite after it was metamorphosed and recrystallized. A discontinuous pyrite veinlet cuts cobaltite along the right-hand margin of the cobaltite-mineralized breccia shown in Figure 12E, and a lenticular body of chalcopyrite cuts both cobaltite and pyrite along the upper-right margin of the sample. Chalcopyrite also surrounds and partly replaces recrystallized cobaltite grains in the lower-right quarter of the sample. In addition to pyrite and chalcopyrite, some samples from the Chicago and Brown Bear zones also contain pyrrhotite (and some contain associated marcasite, which may be of supergene origin).

Blacktail Zone

Figure 12F shows a sample of a composite replacement-style vein from the Blacktail zone. This vein is bounded by interlaminated siltite and phyllitic biotite (probably after tightly folded and sheared banded siltite). Within this composite vein, discontinuous lenses of quartz (interpreted as silicified siltite) are set in a subordinate matrix of discontinuous stringers and pods of biotite phyllite (interpreted as biotitized argillite). Biotite phyllite preferentially hosts cobaltite, but quartz also contains clusters and strings of cobaltite grains. In the lower-right quarter of the sample, a lens of semimassive cobaltite nearly parallels vein walls, but within it, a cross-fabric is defined by strings of cobaltite grains at −65° to the cobaltite-rich lens. This cross-fabric is interpreted to indicate metamorphic recrystallization of cobaltite (labeled Cob2).

Along the right-hand margin of the sample shown in Figure 12F, a massive intergrowth of pyrite and pyrrhotite cuts cobaltite in quartz. In the upper-left corner of the sample, a pyrite veinlet, which cuts cobaltite in quartz, is cut and partly replaced by chalcopyrite. Pyrite, pyrrhotite, and chalcopyrite lack the secondary metamorphic fabric of recrystallized cobaltite. These observations are interpreted to indicate that cobaltite was deposited and metamorphosed before pyrite, pyrrhotite, and chalcopyrite were deposited.

CENOZOIC HISTORY

Uplift and Erosion

The Cenozoic history of central Idaho is one of regional uplift and erosion, accompanied by Eocene magmatism and
Paleogene to Neogene block faulting. Eocene dikes are mostly quartz-phryic felsites (TF in Figs. 7A and 8). Such dikes weather nearly white, and most of them strike north and dip steeply along the White Ledge fault or shear zone (WLf in Figs. 7 and 8). Most normal faults in the Blackbird district strike nearly north and dip steeply (subparallel to preexistent S3 cleavage). They are classified as normal faults on the basis of features that are offset across tabular zones of fault gouge ± breccia ± slicken-sides. Examples shown in Figures 7B and 8 include the Sunshine East fault (SEf), Meadow Creek fault (M Cf), and East Blacktail fault (EBf). The Hawkeye Gulch fault (HGF) and the Northfield fault (NFf), which projects southward to the Slippery Gulch fault (SGf), probably have longer and more complicated histories but were most recently active as normal faults. Such faults probably were active during Cenozoic time, in response to regional post-orogenic uplift and extensional tectonism.

Weathering and Supergene Enrichment of Composite Ores

Common products of oxidative weathering of composite ores include widespread limonite and even wider-spread manganese wad. Pink erythrite (after cobaltite and cobaltian arsenopyrite), and chalcanthite, tenorite, malachite, and azurite (after chalcopyrite) are common locally (especially near ore zones and prospects). Vhay (1948) reported that depths of oxidation vary from 50 to 100 ft (15 to 30 m) beneath hillsides but are deeper beneath deeply weathered upland surfaces.

According to Vhay (1948), supergene ore minerals, including chalcocite, covellite, cuprite, and native copper were locally deposited just below the vadose zone of oxidative weathering of hydrothermal ore minerals. Supergene vivianite (Fe3P2O8·8H2O), a mineral prized by collectors, was found underground, in vugs within fault gouge. Most supergene-enriched ore was removed during mining before 1963. However, supergene ore minerals continue to precipitate from mine-drainage water beneath stopes that were back-filled with the sand fraction of mill tailings, which are now oxidizing in the vadose zone.

POTENTIAL BY-PRODUCTS

About 3.8 t of Au were recovered from the Uncle Sam mine (Anderson, 1947), but only ~1.6 t of Au were recovered from the larger Blackbird mine. Remaining Au resources of the Blackbird patented claim block have not been estimated for lack of sufficient assay data. However, Kuner and Prenn (2007) estimated that Co-Cu ore of the Ram zone contains ~1 t of Au at an average concentration of 0.45 ppm Au.

Nash et al. (1987) reported that bulk samples of cobaltite ore from the Sunshine zone contain 1–6 ppm of Au as microscopic grains of native gold within cobaltite grains (here considered to be of Mesoproterozoic age). However, Anderson (1947) reported that Au from the Uncle Sam mine was recovered from electrum in quartz veinlets that cut quartz-chalcopyrite veins (here considered to be of Cretaceous age). Thus, minor gold was deposited (but in different forms) during Mesoproterozoic and Cretaceous episodes of mineralization in the Blackbird district.

Slack (2006, 2012) identified xenotime, monazite, gadolinite, and allanite as minor minerals in Blackbird deposits, and he suggested these minerals might be recoverable as by-products for their contained REEs. However, the distribution of these minerals is not sufficiently known to support estimation of REE resources in Blackbird Co-Cu-Au ore zones or mill tailings.

CONCLUSIONS

Co-Cu-Au ore zones of the Blackbird district are classified as Co-Cu-Au deposits in metasedimentary rocks, according to a descriptive model, edited by Slack (2013). Such deposits in the Blackbird district are interpreted as composite, epigenetic, predominantly metamorphic-hydrothermal deposits (± subordinate magmatic hydrothermal input), hosted in metasedimentary rocks of Mesoproterozoic age (ca. 1454–1370 Ma). These Co-Cu-Au deposits underwent multiple episodes of metamorphism ± plu-tonism ± metamorphic-hydrothermal mineralization, as follows:

Mesoproterozoic episode Y, stage Y0 involved F1 folding, biotite-grade metamorphism, bimodal mafic-felsic plutonism (1370 ± 10 Ma), and deposition of hydrothermal xenotime (1370 ± 4 Ma).

Stages Y1 (post-F1, pre-F2 folding) and Y2 (post-F2 folding) involved metamorphic-hydrothermal deposition of replacement-style deposits of cobaltite-biotite ore ± tourmaline ± minor xenotime (ca. 1370–1270 Ma) ± traces of microscopic gold (in cobaltite) ± minor late chalcocpyrite. Such mineralization occurred during and after dynamothermal metamorphism (ca. 1379–1325 Ma) and bimodal mafic-felsic plutonism (ca. 1370–1335 Ma), related to the East Kootenay orogeny along the western margin of the Belt-Purcell Basin.

Stage Y3 involved emplacement of hydrothermal quartz-biotite and quartz-tourmaline breccia, vein, and replacement-style deposits ± cobaltite ± minor xenotime (ca. 1058–990 Ma) ± minor chalcocpyrite. This mineralization probably occurred during regional thermal metamorphism (ca. 1200–1000 Ma), possibly related to regional mafic underplating of the western Laurentian margin, while Rodinia was being assembled during the distant Grenville orogeny (far to the northeast, east, and southeast).

Mafic dikes of probable Cambrian–Ordovician age (ca. 530–485 Ma) were emplaced after Mesoproterozoic deposition of cobaltite-biotite ore, after F2 folding, and after emplacement and mineralization of quartz-biotite and quartz-tourmaline breccias, but before Late Jurassic–Cretaceous garnet-growth, and before deposition of Fe-Cu-sulfide minerals in breccias, veins, and replacement-style deposits of Cretaceous age.

Cretaceous episode K, stage K0 involved metamorphic garnet-growth (151–93 Ma), cooling of metamorphic biotite (ca. 151–122 Ma), and deposition of minor metamorphic monazite (144–110 Ma).

Stage K1 involved deposition of hydrothermal pyrrhotite, pyrite, chalcopyrite, and minor cobaltian arsenopyrite ± quartz
± siderite ± monazite ± xenotime (mostly ca. 110–92 Ma) in association with the Cretaceous Sevier orogeny (ca. 112 to 85 Ma) in the hinterland of the Cordilleran orogen.

Stage K2 involved deposition of hydrothermal quartz ± elec-
tron ± Bi minerals in veinlets ± sericite (ca. 83 Ma) in host rocks around veinlets.

SUGGESTIONS FOR ADDITIONAL RESEARCH

Exploration and Resource Estimation

Deeper drilling will be required to test for downward extensions of ore zones that are open down-dip. Results of such drilling are likely to support significant increases in estimated Co-Cu-Au resources of the Blackbird district.

Mineralogy

Distribution of potential by-products, such as Au and REE-bearing minerals could be better defined through systematic re-
examination and analysis of drill core from within and around intercepts of mineralized rocks.

A systematic analysis of mineral species of the Co-Fe-As-
S solid-solution-series in different types of ore throughout the Blackbird district would support better definition of composi-
tional variations within, and distinctions between such mineral species. This would also better define the spatial distributions and paragenetic relationships of cobaltite, glaucodot, cobaltian arsen-
opyrite, arsenopyrite, and safflorite, relative to those of associ-
ated ore and gangue minerals. Ultimately, this could contribute to optimal extraction and concentration of ore minerals.

Geochronology

A project is underway to obtain Re-Os age determinations on ore minerals of the Blackbird district and the ICB, and an effort also is underway to find and date baddeleyite in a sample from a mafic dike near the Blackbird mine portal.

Geophysics

A magnetotelluric (MT) survey across west-central Idaho by Rodriguez et al. (1996) indicated a very conductive zone, which extends from 10 to 30 km deep, east of the Idaho batholith, beneath an area that contains hot springs and mineral deposits in Precambrian metasedimentary rocks, Cretaceous intrusive rocks, and Tertiary volcanic and intrusive rocks. This conductive zone was interpreted to represent a very thick sec-
tion of sedimentary rocks, containing thermal fluids. More recently, an MT survey across northern Idaho and western Mont-
tana by Bedrosian and Box (this volume) traced a zone of con-
ductive horizons (interpreted as sulfide-bearing strata) beneath lower-Belt strata of the Prichard Formation in the central Belt basin.

A future MT survey across the Lemhi subbasin from the Idaho batholith to the Beaverhead Mountains might indicate whether exposed strata of the Lemhi subbasin (which are time-equivalent to upper-Belt strata) are underlain by a thick section of sedimentary rocks (possibly time-equivalent to lower-Belt strata). If relatively conductive horizons were indicated, these might be interpreted as sulfide-bearing or thermal-fluid-bearing zones, which might have served as source rocks for elements that are concentrated in hydrothermal deposits of the Blackbird Co-Cu district and the Idaho cobalt belt.

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