Geology and Temporal Evolution of Alteration and Au-Sb-W Mineralization, Stibnite Mining District, Idaho

Virginia S. Gillerman
Mark D. Schmitz
Jeff A. Benowitz
Paul W. Layer

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Cover Photos

*Top:* High-grade stibnite mineralization, Yellow Pine deposit. Photo from Midas Gold Inc.

*Middle left:* Yellow Pine pit, July 2015. Photo by V.S. Gillerman.

*Middle right:* Backscattered electron image of “running dog” pyrite, Yellow Pine pit. Image by V.S. Gillerman.

*Lower left:* Yellow Pine mine, 1943. Photo from Bradley Mining Company, courtesy of Jim Collord.

*Lower right:* Bear grass near Fern mine, Stibnite District, Idaho. Photo by V.S. Gillerman.
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Virginia S. Gillerman, Ph.D.
Idaho Geological Survey, Boise Office

Mark D. Schmitz
Boise State University

Jeff A. Benowitz and Paul W. Layer
University of Alaska, Fairbanks
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EXECUTIVE SUMMARY

The Stibnite Mining district in central Idaho hosts the USA’s largest antimony (Sb) resource, as part of a multi-million ounce (>140 metric ton or 4.5 million troy ounce) gold (Au) and tungsten (W) deposit. This study by the Idaho Geological Survey documents timing and character of the Au-Sb-W mineralization in the Stibnite district, which is also known historically as the Yellow Pine district. Midas Gold, the project owner, provided excellent support and resources for the study, and some additional work by Midas, the U.S. Geological Survey and others have been incorporated here to address questions about the ore deposits of this complex, world-class mining district.

Ore is hosted in a series of Cretaceous-age Idaho batholith granitoids and Neoproterozoic to Paleozoic sedimentary roof pendant rocks. Metasedimentary strata range from limestone and impure dolomitic marble to quartzite and argillite. The sedimentary strata were affected by metamorphism and deformation related to Mesozoic orogenetic activity along the western margin of the North American craton. Cretaceous and subsequent Tertiary magmatism followed the folding and thrust faulting. Ore is localized along, and structurally controlled by the regional-scale north-south trending Meadow Creek fault, northeast-trending splays such as the West End fault, and other minor structures. Three principal mineral deposits have been delineated by Midas: 1) the intrusive-hosted Yellow Pine deposit, 2) the largely sediment-hosted West End deposit, and 3) the Hangar Flats deposit along the southern strand of the Meadow Creek fault. Many additional prospects are also present. Each deposit has unique features but also commonalities.

Six new, high-precision U-Pb ages on zircons date the granitic magmatism from 94 to 83 Ma, correlating the plutons with the early metaluminous phase of the Idaho batholith. An 85.3 Ma age on metamorphic titanite indicates contact metamorphism coeval with the felsic magmatism, and two ages of approximately 79 Ma from rutile suggest that cooling closely followed the igneous event. Six new \( {^{40}}\text{Ar}^{39}\text{Ar} \) ages on alteration minerals date the principal hydrothermal activity as Tertiary. A single, weakly mineralized sample from the Yellow Pine deposit provided three homogeneous mineral separates, dating early muscovite at 61.5 Ma followed closely by early potassic alteration at 59.6 Ma. Three samples of hydrothermal potassium feldspar from carbonate-quartz-adularia veins and potassic alteration in gold-mineralized rocks of the West End deposit were dated at 51.7 Ma to 50.8 Ma. The 51 Ma age is interpreted as the period of most intense and highest-grade Au mineralization, and it is temporally distinct from the earlier intrusive-hosted, disseminated, mineralizing event. Previous workers distinguished an early, high-temperature gold-arsenic (Au-As) event followed by an epithermal-style, lower temperature mineralization, and the new dates support this hypothesis, while clarifying that both events were Tertiary in age. A cross-cutting latite porphyry dike exposed in drill core was dated by U-Pb geochronology on zircons at 47.86 Ma. Cooling ages from \( {^{40}}\text{Ar}^{39}\text{Ar} \) spectra on the same dike indicate cooling to below feldspar blocking temperatures (~ 300°C) by about 46 Ma. The multiple porphyry dikes are presumably related to the adjacent Thunder Mountain volcanic complex just east of the district.

Antimony mineralization crosscuts the Au mineralization at the Yellow Pine and Hangar Flats deposits. Antimony is sparse in the West End deposit. Stibnite-carbonate-quartz veins and stibnite-matrix breccias cut pyritic zones, particularly along fractures and cross-cutting dilation zones within the complex fault system. Tungsten, as scheelite, is also late and locally coexists with the stibnite. One sample of stibnite-bearing breccia is clearly intruded by a mafic dike, which has not been dated.

Stibnite is locally intergrown with scheelite dated by LA-ICPMS U-Pb geochronology at 45.6 ± 1.6 Ma to approximately 43 Ma (Wintzer and others, 2016, 2017). Vug-filling scheelite crystals are as young as 36 Ma by this method, indicating very lengthy and likely episodic fluid flow along the active, ore-controlling Meadow Creek fault system.

Gold-bearing intervals in core are associated with sulfidized Fe-bearing minerals, principally biotite, in the quartz monzonite and granites, but also Fe-bearing silicates in the metasedimentary rocks. Biotite has been altered to muscovite with sulfides precipitated along the mica cleavages. Sulfides are also present in quartz-carbonate veins with potassium feldspar envelopes and pervasive secondary potassium feldspar alteration in igneous rocks. In metasedimentary rocks, the metamorphic minerals near Au-mineralized zones and veins are strongly altered.
Gold is principally contained in arsenian pyrite, with arsenopyrite as a common associate later in the depositional sequence. No visible Au is present, but electron microprobe analyses show ore-stage As-bearing pyrites with discrete zones of 200 to 1000 ppm Au. Arsenic contents range up to 10 weight percent in the pyrites. Pyrites show complex zoning patterns with irregular cores, middle zones, and rims, typically separated by periods of dissolution. The highest Au contents are most typical of the middle lamellar pyrite zones and very fine grained pyrites. Textures are suggestive of a long-lived hydrothermal system with repeated episodes of high fluid pressures, fracturing, and subsequent mineral deposition.

Paleocene Au-As disseminated mineralization at Yellow Pine was roughly contemporaneous with the northern Bitterroot lobe of the batholith but long after emplacement of the southern Atlanta lobe. Early Eocene Au-Sb mineralization at Stibnite was apparently associated with tectonic extension and transtension, active faulting, and uplift that preceded and accompanied the continental volcanism of the Challis magmatic group. The isotopic compositions of common lead (Pb) from the Yellow Pine and West End deposits are slightly different, compatible with the distinct ages determined at the two deposits. Common Pb isotope ratios of the sulfides and alteration minerals require a dominant contribution of Pb from ancient Mesoproterozoic and Archean basement rocks to account for the highly radiogenic common Pb observed at Stibnite.

Mineralization in the Stibnite district combines some characteristics of epithermal, intrusion-hosted, and orogenic Au deposits, as well as Carlin-type, sediment-hosted Au systems in Nevada. Notable features at Stibnite include the abundant and ubiquitous carbonate gangue and simple alteration assemblage (carbonate-quartz-potassium feldspar-pyrite) present in all rock types, plus the post-convergent tectonic timing and the far inboard location of the district. Other Idaho As-bearing Au deposits along shear zones include the Atlanta district, Orogrande district and the Beartrack deposit.

The 61 to 51 Ma ages of mineralization at Stibnite are only slightly younger than the As-rich Main Stage veins at Butte, Montana, but older than multiple Tertiary ore systems in Idaho (approximately 42-45 Ma) and the Carlin-type gold deposits in Nevada (approximately 38 Ma). The Eocene tectonic framework of northwestern North America is unclear, but structurally, the region was transitioning from a compressive environment to an extensional domain. Tectonic models invoke slab collapse, delamination, changes in plate motion, and other hypotheses to explain the extensive melting and magma genesis under western North America during the mid-Tertiary.

Globally, the Stibnite ores share some features with antimony-bearing deposits (Zn-Sb-Sn) in China’s Guangxi region. These include polymetallic mineralization, a post-convergence, extensional setting with scattered porphyry intrusions, and weakly metamorphosed, locally carbonaceous, Paleozoic carbonate-shale host rocks. Gold-antimony districts in Bolivia, Australia, and New Brunswick also show structurally controlled mineralization, hydrothermal carbonate, and an association with black shales and clastic rocks. Carlin-type deposits in Nevada and elsewhere share similar stratigraphic and metal associations (Au-As-Sb-Hg) but the characteristic decalcification alteration of Carlin ores must require different fluid chemistry compared to the abundant carbonate gangue at Stibnite. For Idaho, the Stibnite district lies in a location where multiple magmatic pulses have intersected favorable sedimentary host rocks, and abundant hydrothermal fluid flow has been facilitated by long-lived structural zones. The result is a unique and world-class Au-Sb-W deposit.
CHAPTER 1
OVERVIEW
INTRODUCTION

The Stibnite district, which is part of the larger Yellow Pine mining district, produced nearly a million ounces of gold (Au) plus substantial tungsten (W) and antimony (Sb) since 1900. The district hosts the United States’ largest resource of Sb, a critical mineral resource, and a large Au reserve (over 140 metric tons of Au). Early geologic studies noted the essential characteristics of the complex ores, but the age and origin of the mineralization at Stibnite have been controversial for nearly a century. This Idaho Geological Survey (IGS) study, supported by Midas Gold, Inc., was designed to clarify the geologic setting and age of hydrothermal alteration and mineralization by applying modern geochronological and analytical techniques on carefully characterized samples from drill core as well as surface exposures. As described in the subsequent chapters, our observations of mineral associations coupled with detailed age determinations, isotopic studies, and microprobe analyses of Au-bearing pyrites have documented at least two, and possibly three, overprinted major hydrothermal events at Stibnite. The results show that hydrothermal activity and ore deposition at Stibnite took place in pulses over a 20-million-year interval from early-Tertiary to mid-Tertiary time, largely focused along major structures.

Stibnite is located on the East Fork of the South Fork of the Salmon River (Figure 1-1). Mineralization extends from river level up to the ridgetops over a vertical range of nearly 3,000 ft in outcrop. Gold was discovered around 1900 by prospectors in the Thunder Mountain gold rush (Mitchell, 2000). Early discoveries of the Hermes and Fern mercury (Hg) mines (Figure 1-2) were followed by the recognition of Au and Sb ores. Despite the difficult access and remote location, several mines, including the Meadow Creek mine and Yellow Pine mine, were developed along north-south and northeast-trending faults which cut Cretaceous Idaho batholith and a roof pendant comprised of metasedimentary rocks (Figure 1-2). Tertiary volcanic rocks crop out just to the east, and the presence of two different igneous assemblages in proximity has generated ambiguity and confusion over the igneous association and age of the complex mineralization.

MINING AND EXPLORATION HISTORY

Mitchell (2000) and the Midas prefeasibility study (PFS) technical report (Huss and others, 2014) provide excellent, detailed accounts of the history of the Stibnite mining district. Following early Hg, Sb, and Au mining, the Stibnite district gained more recognition during World War I as a source of Sb and again in the early 1940s, when W was discovered at the Yellow Pine mine. During World War II, the Yellow Pine mine produced 831,829 units of tungsten trioxide (WO$_3$), equivalent to over 16.6 million pounds, during the period 1941-1945 (Mitchell, 2000). Tungsten was an essential metal for the war effort and the Stibnite district was a leading supplier. Approximately 10,828 short tons of Sb were also extracted from the Yellow Pine mine during the same 1941-1945 period. From 1980 to 1997 several small, open-pit Au mines operated in the district. They included the Stibnite mine, West End mine, Homestake mine, Yellow Pine mine, and Garnet Creek mine, as shown on Figure 1-3 plus smaller pits such as the Splay, Midnight, and Northeast pits.

Historical production from Stibnite is estimated at approximately one million ounces of Au, 2 million ounces of silver (Ag), 44,000 short tons of Sb, and 856,250 units of WO$_3$, the equivalent of 13.5 million pounds of W (Huss and others, 2014). In 2004, exploration drilling delineated a 2-million-ounce gold-in-sulfide resource in the river valley bottom below the historic Yellow Pine pit at Stibnite (Gillerman and Mitchell, 2005).

Midas Gold, Inc., began conducting exploratory work in the district in 2009, and by 2011 the company had drilled several hundred holes and completed other geological, engineering, and environmental studies. In September 2016, Midas Gold submitted a mine plan proposal to the Payette National Forest and other state, local, and federal agencies. The project is currently in the permitting stage to develop a new Au mine at the site (Gillerman, 2016, 2017). The Stibnite Plan of Operations Environmental Impact Statement (EIS) documents are available through the Payette National Forest website (U.S. Department of Agriculture Payette National Forest, 2018). Midas has identified three main areas of mineral resources: the Yellow Pine deposit,
the West End deposit, and the Hangar Flats deposit, as well as resources in the historic tailings, and multiple exploration targets (Figure 1-3). Overall indicated mineral resources quoted in the PFS are over 5.4 million troy ounces of Au, 8.9 million troy ounces of Ag, and 155 million pounds of Sb (Huss and others, 2014). Updates of reserves and resources are available from Midas Gold (2018). The proposed plan of operations includes major reclamation and restoration work to be accomplished at the site after the mine closes. Additional drilling was underway during 2017. More information is available from Midas Gold (2018).

**PREVIOUS WORK**

Numerous geologic studies and publications about the Stibnite and Yellow Pine districts exist, and they have been used as a basis for this current investigation. Important historical references include Schrader and Ross (1926), Ross (1934), Currier (1935), White (1940), Cooper (1951), Lewis (1984), Smitherman (1985), and Cookro and others (1987). Previous workers outlined the basic geology of the district but did not include modern geochronology or detailed petrography of the mineralized rocks. In particular, the age and igneous affiliation of the Au and Sb mineralization has been a subject of debate since the early studies in the region. Figure 1-4 shows the location of the Stibnite quadrangle relative to surrounding quadrangles and the more recent IGS geologic maps available for the region. Of interest is DWM-161, the geologic map of the central and lower Big Creek drainage by Stewart and others (2013). The map in DWM-161 shows many regional structures, including the Big Creek graben, a prominent northeast-trending feature located to the northeast of the Stibnite quadrangle. The 1:24,000-scale Stibnite quadrangle...
Figure 1-2. Simplified geologic map of the Stibnite district with major historic mines and pits shown. Geology modified from Stewart and others (2016).
geologic map GM-51 by Stewart and others (2016) was recently published by the IGS as a companion project to this alteration and mineralization study. GM-51 provides a regional lithologic and structural framework for the district and was based on surface geologic mapping. Another new resource is DWM-180, a geologic map of the Burntlog Creek area in Valley County (Stewart and others, 2017). The Burntlog Creek map covers area to the southwest of the Stibnite district.

STUDY OBJECTIVES AND METHODS

Work by the Idaho Geological Survey started in 2012 under a cooperative agreement with Midas Gold with two primary objectives: 1) a new geologic map of the Stibnite 7.5’ quadrangle, and 2) a study of the alteration, mineralization, and geochronology in the district. The first objective resulted in the GM-51 geologic map of the Stibnite 7.5’ quadrangle in Valley County (Stewart and others, 2016).

Field work for the alteration and geochronology study included detailed sampling in the Stibnite and Yellow Pine pits, and less detailed examination and sampling of the West End and Fern mine areas plus other surface exposures in the district. For this study, nearly a hundred surface samples were collected during 2012-2015 (Appendix A-2), principally from the intrusive-hosted ores and wall rocks in the Yellow Pine pit (also known as the Bradley Pit) and the Homestake pit area, and from sediment-hosted mineralization and wall rocks in the Stibnite pit and the Fern mine area. Additional sampling conducted in 2015 was from surface exposures of less-altered rocks on the jeep trail and ridge north of the Fern mine, mineralized zones in the West End and Yellow Pine pits, and the Hermes mine areas. The objective of the sampling was to characterize the lithology, alteration, and mineralization, and to find suitable material for geochronology and other analytical work. Surface sample locations have the format 15VG00X, where the 15VG00X refers to waypoint X taken in 2015. Coordinates of waypoints for surface sample locations are listed in Appendix A-2. An updated Midas Gold site layout map from the 2016 mine plan is in Appendix A-3, and a map and longitudinal section of the historic Yellow Pine and Meadow Creek mine workings is in Appendix A-4.

Core logging or inspection and sampling was conducted on ten representative diamond drill holes, locations of which are provided in Appendix A. The ten holes represent a small subset of the over 400 recent core holes from the property. A total of 112 thin and polished sections were prepared. A spreadsheet summary of the thin section samples and simplified mineralogy and features seen in the petrographic examinations is in Appendix B. Appendix B also contains bar graphs of lithology, alteration, and geochemistry logs done by Midas Gold for selected diamond drill holes used in this study. Photographs of the thin-section billets, which each measure 2.5 cm by 4.5 cm, are in Appendix C. Many thin sections were stained for potassium feldspar; a few were stained to distinguish calcite from dolomite. Sample numbers are typically listed as thin-section lab numbers with a format of ABC-002, and an accompanying location number at Stibnite. For drill hole samples, the location is given as the drill hole number and footage. For example, sample 12-306-255 was from drill hole MGI 12-306, at 255 feet downhole.

Midas Gold provided access to surface exposures, drill core, multi-element geochemistry and assays, along with very helpful discussions with their staff. Correlation of the thin section and drill core observations with assay information proved invaluable due to the very fine grain size of the sulfide minerals. Hand specimen and petrographic examination of samples were used to screen which samples to select for age determinations and isotopic or microprobe analyses. For igneous age determinations, several large (2 to 10 kilogram) bulk samples were collected of intrusive rock. Initial results were reported in Gillerman and others (2014).

Igneous rock ages were determined principally by chemical abrasion isotope dilution thermal ionization mass spectrometry (CA-ID-TIMS) at the Isotope Geology Laboratory at Boise State University (BSU) on zircons from intrusive rock. This included preliminary screening with cathodoluminescence (CL) imaging and laser ablation-inductively coupled plasma mass spectrometry (LA-ICPMS) spot dating. Additional details of the U-Pb geochronology methods are in Appendix D. Age determinations by 40Ar/39Ar methods on potassium feldspars and micas associated with hydrothermal alteration and mineralization were completed at the University of Alaska, Fairbanks (UAF) Geochronology Laboratory. The two analytical reports from UAF are in Appendix E.
Figure 1-3. Midas Gold prefeasibility study site map (Huss and others, 2014) showing historical mine features and proposed Midas pits and facilities as of 2014 plan. Proposed open pits are indicated in green. Updated maps from the 2016 mine plan proposal are in Appendices A-3 and A-4.
A lead isotope study of feldspars from intrusive rocks and alteration zones and of common lead in sulfide minerals associated with mineralization was completed at the Isotope Geology Laboratory at Boise State University. The sample list and results are in Appendix F.

Electron microprobe work was completed at the GeoAnalytical Lab at Washington State University School of the Environment with the assistance of Dr. Owen Neill using a JEOL JXA-8500F field emission electron microprobe/scanning electron microscope. Summary spreadsheets of the results of the microprobe analyses are in Appendix G.
CHAPTER 2

GEOLOGIC SETTING
GENERAL GEOLOGY

Lewis and others (2012a) show the general geology of Idaho. The Archean is very sparsely exposed in Idaho, and it is mostly covered by much younger rocks. The oldest extensively outcropping rocks are the Mesoproterozoic metasedimentary and mafic sills associated with the Belt basin (Figure 2-1). In northern Idaho, these Belt rocks of roughly 1.4 Ga age consist of the Prichard Formation through the Missoula Group metasedimentary rocks, and they are locally host to world-class base metal and Ag mineralization in the Coeur d’Alene district. The Neoproterozoic Windemere Supergroup, which is exposed locally in roof pendants in central Idaho and in correlative units in southern Idaho, includes rift-related mafic rocks, diamictites, and marine miogeoclinal rocks (Lund, 2008; Lewis and others, 2012a). Intracontinental rifting during the Neoproterozoic created the new continental edge to western Laurentia, and subsequent sedimentation along this passive margin produced a sequence of Neoproterozoic to Paleozoic continental margin and platform carbonate and clastic rocks which overlie presumed Precambrian basement in southeastern and eastern Idaho (Lund, 2008).

A lengthy period of late Paleozoic to Mesozoic convergence, subduction, and accretion of oceanic rocks and multiple island-arc terranes of the Blue Mountains province to the western edge of North America along the Salmon River suture zone generated voluminous granitic magmas and strong deformation during the Mesozoic (Gaschnig and others, 2010; Lewis and others, 2014; LaMaskin and others, 2015; Gaschnig and others, 2017). That suture zone trends roughly north-south west of McCall (Figure 2-1). Tikoff and others (2017) date the accretion as occurring by Late Jurassic to Early Cretaceous time. Magmas of the Idaho batholith were emplaced along that boundary and eastward, well into the continental crust of Laurentia (Figure 2-1). The batholith has been subdivided into multiple intrusive suites and two main lobes: the Atlanta lobe to the south and the smaller Bitterroot lobe to the north (Gaschnig and others, 2010, 2011, 2017). In addition to the suture zone suite (110 to 100 Ma), the Atlanta lobe includes the border zone suite (92 to 85 Ma), early metaluminous suite (98 to 85 Ma) and the voluminous Atlanta peraluminous suite (83 to 67 Ma) plutons. The Bitterroot intrusive rocks are distinctly younger at 76 to 53 Ma in age (Gaschnig and others, 2017). After the arc-continent collision, the boundary was further modified and compressed with dextral transpression (Tikoff and others, 2017).

The Eocene-age Challis Volcanic Group occupies a large area in east-central Idaho (Figure 2-1). The Challis volcanic rocks and associated sediments unconformably overlie an erosion surface of great relief. This continental volcanic field is part of the Tertiary volcanic flare-up that also affected Nevada and elsewhere in western North America from the Eocene into the Miocene. In Idaho, age constraints from K-Ar and sparse 40Ar/39Ar geochronology show regional Challis volcanism and intrusive activity occurred from approximately 51 Ma to 43 Ma over a large region in central Idaho (Leonard and Marvin, 1982; McIntyre and others, 1982; Moye and others, 1988; Gaschnig and others, 2011; Lewis and others, 2012a). Numerous porphyritic dikes and Eocene-age plutons are also associated with the Challis magmatism over a large area of south-central and east-central Idaho. Though the exact tectonic framework which gave rise to the Challis magmatism is debated, most workers agree that the volcanic rocks were erupted during dominantly northwest-southeast extension which resulted in northeast-striking dike swarms and normal faults, including the trans-Challis fault system (Figures 2-1 and 2-2). The Big Creek graben, which is located just east of Stibnite, is northwest of but subparallel to the trans-Challis system (Figures 2-1 and 2-2). The trans-Challis system may be a continuation of the Great Falls tectonic zone that marks the northern boundary of the Archean Wyoming craton (Kilsgaard and others, 1986; Foster and others, 2006).

REGIONAL GEOLOGY AND METAMORPHISM

The Yellow Pine region is dominated by the Cretaceous-age Idaho batholith; Stibnite is located near the eastern margin of the batholith and adjacent to the western margin of the Challis Volcanic Group (Figures 2-1 and 2-2; Lewis and others, 2012a; Stewart and others, 2016). Within the Stibnite quadrangle, two roof pendants (Stibnite and Missouri Ridge-Sugar Mountain) contain folded and metamorphosed rocks of Neoproterozoic to Paleozoic age (Figure 2-2). Regional metamorphism predates the batholith rocks at Stibnite, as seen elsewhere in Idaho (Bookstrom and others, 2016; McKay and others, 2017). Metamorphism was roughly
Figure 2-1. Simplified geologic map for part of Idaho with major Au districts shown in yellow hammer symbol. Thompson Creek Mo mine is also shown by green box. Faults shown in black. Salmon River suture in yellow dashed line separates the accreted terranes from continental rocks to the east. White box shows location of Figure 2-2, the regional map.
Figure 2-2. Regional geologic map for Yellow Pine region and Stibnite area. Multiple calderas indicated by colored hachured areas within Thunder Mountain volcanic field east of Stibnite. Note Johnson Creek shear zone and Edwardsburg townsite locations. Modified from Stewart and others (2016).
coeval with long-lived deformation along the Salmon River suture zone that marks the boundary of accreted terranes against the western margin of North America (Figure 2-1).

Units within the Stibnite district were metamorphosed to amphibolite facies during a regional metamorphic event (Smitherman, 1985; Stewart and others, 2016). Wintzer and Vervoort (2016) obtained Lu-Hf metamorphic garnet ages of 112.8 ± 7.2 Ma on the Neoproterozoic schist of Moores Station Formation about a mile northwest of the Stibnite district. The schist underlies the stratigraphic section at Stibnite (Stewart and others, 2016). Wintzer and Vervoort (2016) also dated an undeformed aplite from the same area. The aplite cuts the schist and had an age of 99.5 ± 4.0 Ma; they interpreted that as an upper limit to the end of metamorphism and related deformation. These ages are consistent with field observations within the Stibnite district that the granitic rocks appear unaffected by the folding and metamorphism. It is possible that additional contact metamorphism affected some of the strata during intrusion of the Cretaceous granitoids, but overall, there is no field evidence for a steep thermal gradient around the granitoids, suggesting deep emplacement of the granitic rocks.

The eastern boundary of the Stibnite quadrangle is underlain by Eocene Challis volcanic rocks related to multiple calderas, as shown on Figure 2-2. The Challis volcanic pile, which is located east of Stibnite, was termed the Thunder Mountain caldera by Leonard and Marvin (1982) although multiple volcanic centers are now recognized (Stewart and others, 2016). The volcanic rocks were preserved by down-dropping on a series of large normal faults east of the mineralized district, as shown in Figure 2-2. The late, north-northeast trending Big Creek graben cuts the Eocene volcanic rocks and Thunder Mountain caldera complex (Stewart and others, 2013; Stewart and others, 2016).

**DISTRICT GEOLOGY**

The Stibnite roof pendant consists of Neoproterozoic to Paleozoic metasedimentary rocks intruded by several phases of the Cretaceous Idaho batholith (Figures 2-2, 2-3, and 2-4; Stewart and others, 2016). Hypabyssal porphyritic dikes of latite, rhyolite, and basalt or lamprophyre are exposed in the Yellow Pine pit and in drill core, as well as being described from underground mines. The dikes at Stibnite are similar in appearance to dikes of “Challis age” found in much of south-central Idaho.

A recurring and controversial question since the early work of Schrader and Ross (1926) has been the age of the mineralization at Stibnite – is it Cretaceous or Tertiary? The medium- to high-grade nature of the metamorphism has obliterated most sedimentary structures and fossil evidence. Thus, the depositional ages and regional stratigraphic correlations of the metasedimentary host rocks in the Stibnite roof pendant have also been debated. Smitherman (1985) provided the basic stratigraphic column, rock descriptions, and geologic map that have been used in the district by most of the recent mining companies. His stratigraphic terminology was also largely adopted by Stewart and others (2016).

Recent detrital zircon work by the Idaho Geological Survey and its collaborators has established that some of the units are indeed Paleozoic in age (Lewis and others, 2014). A new structural interpretation has likewise allowed for better stratigraphic section correlations (Stewart and others, 2016). In addition to metamorphism and folding, the metasedimentary rocks are cut by numerous faults (Figures 1-2 and 2-3). The recent mapping identified a system of Cretaceous-age folds and thrust faults. The bedding sub-parallel fault identified by Smitherman (1985) was re-interpreted by Stewart and other (2016) as an overturned thrust fault, named the Cinnabar Peak fault (Figures 1-2, 2-3, and 2-4). The Cinnabar Peak thrust fault is folded within the northwest-trending overturned Garnet Creek syncline, which underlies the heart of the mineralized area. The mapping identified additional thrust faults and northeast-trending strike-slip and normal faults, one of which appears to offset the Cinnabar Peak fault in a right-lateral sense along Sugar Creek and forms a northern boundary to the mineralized area (Stewart and others, 2016). The contractional deformation probably took place under ductile conditions in a hot, deep-seated environment as deduced from locally tight folding of quartzites and other strong rocks. Stewart and others (2016) concluded that folding likely predated or was synchronous with intrusion of the later stages of the metaluminous suite of the Idaho batholith (100-85 Ma; Gashnig and others, 2010).
Figure 2-3. Geologic map of the Stibnite district with general mineralized resource areas (blue pattern) and exploration prospects (red stipple) shown, plus drill holes used in this study. YP = Yellow Pine; HF = Hangar Flats; WE = West End, all names of Midas’ resource areas. Modified from Stewart and others (2016).
Figure 2-4. Explanation of geologic map in Figure 2-3. From Stewart and others (2016).
Several of the northeast-trending normal faults, such as the Garnet Creek fault, cut through the batholith and roof pendant and localize mineralization within the district (Figures 1-2, 2-3, and 2-4). Because the northeast-trending part of the Meadow Creek fault offsets the overturned northwest-striking thrust sheet and is roughly parallel to the sense of movement, it may have originated as a tear fault in the Cretaceous. It is possible that other northeast-trending faults were localized by Cretaceous or older structures. Cooper (1951, page 162) noted that, "the large displacements on the Fern and West End faults are certainly older than the quartz monzonite intrusion, but smaller movements may have taken place following the quartz monzonite intrusion."

The northeast-striking faults both splay into and are cut by the ore-controlling Meadow Creek fault, the main ore-localizing structure at Stibnite (Figures 1-2, 2-2, and 2-3; Cooper, 1951; Stewart and others, 2016).

The Meadow Creek fault is a major regional structure and subparallel to other north-south structures in western Idaho, including the Johnson Creek shear zone further west and the Salmon River suture (Figures 2-1 and 2-2; Lewis and others, 2012a; Tikoff and others, 2017). The Meadow Creek fault strikes north-south and has a nearly vertical dip. It is actually a fault zone with several strands and styles of deformation. The fault bends and turns northeasterly at the location of the Yellow Pine pit. Recent drilling by Midas confirms that the fault also changes dip in the subsurface in the Hangar Flats area. At the surface in the Hangar Flats deposit, the Meadow Creek fault dips very steeply west but below about 6300-feet elevation, it changes to a steep east dip (Midas Gold Corp., 2018). The north-south trend is inconsistent with either southwest-northeast shortening and contraction or postulated northwest-southeast Eocene extension. The Meadow Creek fault shows both right-lateral and west-side up motion (Cooper, 1951; Stewart and others, 2016). Cooper (p. 163, 1951) described the Meadow Creek fault from its exposure in the Bailey tunnel:

"Where crossed by the Bailey tunnel, the fault is 145 feet wide; it strikes N. 25° E., and the main gouge and shearing planes dip 53° - 90° W. Both slickensides and cleavage indicate that the west, or hanging wall, moved obliquely upward and to the north relative to the footwall. The horizontal component was greater than the vertical."

It is unclear if the Meadow Creek fault might have been rotated, and it appears to have had at least some post-mineral movement (Chris Dail, oral commun., 2012). The Meadow Creek fault has been a locus of hydrothermal alteration and mineralization over a very long time, as documented in this study.

**REGIONAL METALLOGENY**

The Stibnite district and its Au-Sb-W-(Hg) metal suite is unique within Idaho. In general, the Idaho batholith hosts a number of Au deposits along major faults and shear zones (Figure 2-1; Gillerman and Mitchell, 2005). Many of those produced placer Au from largely eroded veins and shear zones of indeterminate age. In contrast are the clearly epithermal Au-Ag and Au ores of the Eocene volcanic-hosted deposits situated in the trans-Challis fault system and the Thunder Mountain volcanic complex located in the area east of Stibnite (Figures 2-1 and 2-2). The historic Dewey mine and the Thunder Mountain mine plus its extension, the Lightning Ridge deposit, were active mines and projects in the late 1980s, and are located near Thunder Mountain, shown near the east margin of the map in Figure 2-2 (Gillerman and others, 1990). A number of epithermal precious metal deposits are also localized in the northeast-trending trans-Challis fault system and graben north of the town of Challis (Figure 2-1; Kiilsgaard and others, 1986). The Stibnite district lies at the west edge of the Challis volcanic field, while the trans-Challis deposits are associated with the northeast-trending normal faults that cut the middle of the volcanic field. Idaho’s largest Au district in terms of historical production is the Boise Basin near Idaho City at the southwest end of the trans-Challis system (Figure 2-1). A quarter of the approximately 2.8 million troy ounces of Au produced was from lode deposits of Au-bearing quartz-pyrite veins that cut Cretaceous and Tertiary intrusive rocks. A northeasterly set of Eocene porphyry dikes are associated with the veins (Kiilsgaard and others, 1997). Whereas, As is a common trace element in the Boise Basin deposits, Sb and W are not described in commercial quantities. However, late-stage stibnite is present locally in the veins, and geochemically anomalous tungsten attracted exploration interest in the 1990s (Anderson, 1947; Christopher Dail, written commun., 2019).

Numerous small mines are also present in the remote Edwardsburg and Big Creek area to the north and
The geologic setting is similar to that at Stibnite with a mix of Precambrian metasedimentary rocks in roof pendants within the batholith, but Tertiary volcanic and intrusive rocks are located nearby. Although north-trending faults are mapped, the ore controls and ore deposit associations are unclear (Figure 2-2). Metals mined include precious metals but also W (Shenon and Ross, 1936; Gammons, 1988).

The relation between regional lineaments and the location of mineralization in Idaho is complex. Northeast, north-south, and northwest linear structures have been proposed. Gillerman and Mitchell (2005) noted the northerly-trending alignment of Au districts in west-central Idaho. Those districts include Orogrande near Elk City, Florence, Warren, and the Boise Basin with Stibnite slightly displaced to the east. Some of the districts, such as Orogrande and Stibnite, have mappable north-south faults or shear zones that localize mineralization. In addition to the Meadow Creek fault at Stibnite, the Johnson Creek shear zone has a north-south trend from near the town of Yellow Pine north towards Edwardsburg (Figure 2-2). Regional-scale, north-south faults and north-trending valleys, parallel to the Salmon River suture zone, are evident in the present-day topography of west-central Idaho (Figure 2-1; Lewis and others, 2012a). The northeast-trending trans-Challis fault system has been recognized as another important mineralized trend in Idaho (Kiiilsaard and others, 1986). Stibnite shows both north-south and northeast-trending structures within the district and lies near their regional intersection at a very gross scale. The Stibnite district is also part of a north-northwest-trending line of major gold occurrences extending from the Hailey Gold Belt (Camas District) on the southeast through occurrences near Stanley, northwest through the Yellow Pine district, Warren district, Buffalo Hump, and Florence district on the northwest, as evident on the map of Gaston and Bonnichsen (1978).

Molybdenum deposits in the vicinity include a Tertiary occurrence at Quartz Creek just northeast of the town of Yellow Pine along the Johnson Creek shear zone, and the large quartz monzonite type Mo deposit at the Thompson Creek mine near Clayton (Figure 2-1). The pluton has an age of approximately 88 Ma with mineralization being slightly younger at 86 Ma (Marvin and others, 1973; Worthington, 2007).

Antimony is an important and characteristic metal within the Yellow Pine and Hangar Flats deposits at the Stibnite district, having been mined historically at the Meadow Creek mine (Figures 1-2 and 2-3; White, 1940). Antimony veins are also present at Antimony Ridge, just south of the town of Yellow Pine and adjacent to the Johnson Creek shear zone, a major topographic linear (Figures 1-1 and 2-2). Elsewhere in the state, Sb is present in numerous small prospects, but the only substantial production has been from tetrahedrite in the complex quartz siderite sulfide veins of the world-class Coeur d'Alene district in northern Idaho. Though hosted there in Mesoproterozoic metasedimentary rocks, the Ag and Sb may have been introduced, or at least remobilized, during Cretaceous metamorphism, folding, and faulting (Bennett and others, 1989; Leach and others, 1998).

HOST ROCKS OF THE STIBNITE DISTRICT

Rocks in the Stibnite district include both igneous and metasedimentary units, plus Quaternary surficial deposits (Figures 1-2, 2-3, and 2-4). In the core of the mineralized district, both igneous and metasedimentary units are highly altered and structurally disrupted. This chapter will describe the least-altered rock units, including their general compositions and ages, which serve as host rocks for the mineralization within the Stibnite district. Units noted in the subsurface during drill core logging do not correlate precisely with the surface units mapped by Stewart and others (2016) and shown in Figures 2-3 and 2-4.

All of the pre-Tertiary rocks in the roof pendants have been subject to moderate- to high-grade metamorphism and are extensively recrystallized. Compositionally, the metasedimentary rocks and their sedimentary precursors are quite diverse. In general, they contain carbonate units and siliceous to argillaceous strata similar to continental shelf and margin sediments exposed in Neoproterozoic to Paleozoic successions in Nevada and southeastern Idaho (Figure 2-1).

Metasedimentary Rocks

Table 2-1 lists lithologies of the metasedimentary rocks in the Stibnite roof pendant and the interpreted depositional ages. Smitherman (1985) mapped two types of marble (calcite and dolomite dominant), schist,
and schistose quartzite or quartzite schist, calc-silicate (two units), quartzite, and quartz pebble conglomerate. All of these rock units were observed in the drill holes logged and in the surface exposures, but no attempt was made to fully map or correlate the metasedimentary units examined in drill core with the surface stratigraphy. More detail can be found in Smitherman (1985) and Stewart and others (2016).

Geologic mapping and detrital zircon geochronology of selected metasedimentary units confirm a Paleozoic age for at least part of the sedimentary strata in the Stibnite roof pendant (Stewart and others, 2016). The new map interprets the lower quartzite unit as Cambrian and the upper quartzite unit as Ordovician (Kinnikinic or Eureka equivalent), with the base of the section extending down into the Neoproterozoic. By stratigraphic position,

<table>
<thead>
<tr>
<th>Unit</th>
<th>Name</th>
<th>Thickness (approx.)</th>
<th>Age (probable)</th>
<th>Brief description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ouq</td>
<td>Upper quartzite</td>
<td>400 m</td>
<td>Ordovician</td>
<td>Light gray orthoquartzite.</td>
</tr>
<tr>
<td>Ohm</td>
<td>Hermes marble</td>
<td>100 m</td>
<td>Ordovician</td>
<td>Light brown dolomitic marble. Recrystallized with tremolite.</td>
</tr>
<tr>
<td>O₇mq</td>
<td>Middle quartzite</td>
<td>80 m</td>
<td>Ordovician</td>
<td>Light gray quartzite.</td>
</tr>
<tr>
<td>O₇mm</td>
<td>Middle marble</td>
<td>160 m</td>
<td>Ordovician</td>
<td>Gray massive calcareous marble with argillaceous laminae. Local anthophyllite, phlogopite, biotite, diopside.</td>
</tr>
<tr>
<td>O₇mmq</td>
<td>Quartzite in middle marble</td>
<td>25 m</td>
<td>Ordovician</td>
<td>Quartzite.</td>
</tr>
<tr>
<td>O₇us</td>
<td>Upper calc-silicate</td>
<td>125 m</td>
<td>Cambrian?</td>
<td>Gray-green calc-silicate marble with laminae. Quartz, calcite, biotite, plagioclase, scapolite, wollastonite, diopside, tremolite, muscovite, magnetite. Local serpentine.</td>
</tr>
<tr>
<td>Elq</td>
<td>Lower quartzite</td>
<td>180 m</td>
<td>Cambrian</td>
<td>Gray, coarse-grained quartzite with schist. Local sedimentary structures.</td>
</tr>
<tr>
<td>E₇qpc</td>
<td>Quartz pebble conglomerate</td>
<td>100 m</td>
<td>Cambrian/Neoproterozoic</td>
<td>Impure quartzite and conglomerate with flattened, rounded pebbles. Unconformity at base and karst at contact of marble below.</td>
</tr>
<tr>
<td>Zfm</td>
<td>Fern marble</td>
<td>40 m</td>
<td>Neoproterozoic</td>
<td>Medium gray, recrystallized dolomitic marble, local tremolite.</td>
</tr>
<tr>
<td>Zlcs</td>
<td>Lower calc-silicate</td>
<td>300 m</td>
<td>Neoproterozoic</td>
<td>Gray to green calc-silicate rock and marble. Ribbon laminated. Quartz, biotite, calcite, muscovite, epidote, tremolite, actinolite, anthophyllite, diopside, vesuvianite.</td>
</tr>
<tr>
<td>Zqs</td>
<td>Quartzite-schist</td>
<td>130 m</td>
<td>Neoproterozoic</td>
<td>Interbedded gray quartzite and mica-quartz-feldspar schist. Local Al-silicates.</td>
</tr>
</tbody>
</table>

Table 2-1. Summary of Stibnite stratigraphy (after Smitherman, 1985; Stewart and others, 2016).
Stewart and others (2016) determined the middle marble as Cambrian or Ordovician, and below the Hermes marble, which they correlate with the Ordovician Ella Dolomite exposed at Bayhorse, Idaho.

A few useful generalizations can be made about the origin and depositional setting of the original sedimentary facies. Most of the metamorphosed units display interbedded compositional layering and variations within a single outcrop and even within a single thin section. The initial sediments contained varying amounts of detrital sand, silt, mud, and carbonate, including significant (25 to 75 percent) amounts of dolomite. The compositions, particularly in the impure carbonate and calc-silicate rocks, are similar to rocks of the Paleozoic miogeoclinal shelf and slope sequences in Nevada and Utah (Lund, 2008; Dail and others, 2015; Dail and others, 2016). Although highly deformed, the Stibnite strata were deposited with unit thicknesses of tens to hundreds of feet, also typical of the western North American Paleozoic miogeoclone. Due to the metamorphic recrystallization, recognizable fossils are rare to non-existent in the Stibnite roof pendant with only one previous report that was not confirmed (Lewis and Lewis, 1982). One sample of gray marble float, most likely from the Hermes marble, was collected along the Sugar Creek road. It shows branching fossil-like structures reminiscent of bryozoans (Figure 2-5). In thin section, the Sugar Creek float sample and a marble sample from the Stibnite pit contain curving, arcuate, ovoid-like structures with calcite walls and interiors of dolomite, as is typical of diagenetic alteration in platform carbonate rocks with biogenic structures. However, some of these ovoids are scapolite crystals in cross-section view or the ovoid is infilled with scapolite, phlogopite, and other silicates (Figure 2-6). Abundant phlogopitic mica, scapolite, amphibole and other minerals in the marbles suggest an original silty component to the calcareous or dolomitic sediment.

In general, the marbles are recrystallized and typically have scapolite, phlogopite, tremolite, or anthophyllite along with talc, serpentine and other fine-grained minerals (Smitherman, 1985). The white “matchsticks” visible in fresher outcrops and in thin sections seem to be mostly scapolite, but some could be tremolite. The
first-order birefringence of the scapolite indicates a composition close to marialite, the Na- and Cl-rich end-member. Interbedded calc-silicate and marble horizons are common, with layering on a scale of centimeters or millimeters (Figure 2-7). Metamorphic minerals include scapolite, tremolite-actinolite, anthophyllite, phlogopite, biotite, talc, clinopyroxene, titanite (sphene), plagioclase, andalusite(?), sillimanite(?), wollastonite, chlorite, and epidote, plus some pyrite and oxides. Smitherman (1985) provides more comprehensive descriptions of the lithologies and minerals. Garnet skarn is principally confined to one area near the backfilled Garnet Creek pit, which was not available for study, and is also present as bedding replacements south of the Fern mine area and at the Rabbit prospect (Christopher Dail, written commun., 2019). Preservation of the compositional layering during deformation and metamorphism is evidence of the lack of significant fluid flow and mass transfer during the metamorphic event, in contrast to later mineralizing events.

In calc-silicate units outside of the mineralized area (such as along the jeep trail northwest of the Fern mine at waypoint 15VG024A), coarse unmineralized pyrite was seen intergrown with metamorphic amphibole, probably of actinolite composition (Figure 2-8). This indicates that at least some pyrite (but not necessarily Au) formed synchronous with the metamorphic minerals. As discussed in Chapter 6 and Appendix G, electron microprobe analysis of this pyrite outside of the mineralized area showed the pyrite is barren of ore metals and associated trace elements.

Cretaceous metamorphism of the sedimentary rocks was critical to developing the physically competent host rocks which later failed under brittle conditions to provide the abundant fracturing, veining, and brecciation that is characteristic of the Au-Sb mineralization. A small-scale example of the effect of varying competency of the bedding is shown in Figure 2-9. A calc-silicate layer has formed boudins within the surrounding more ductile meta-argillite unit, reflecting both the extensional strain and the ductility contrast of the layers. A large pyrite nodule developed in the tensional zone between the boudins. Textures like that in Figure 2-9 record a shift from compressional tectonics and metamorphism to extensional tectonics and sulfide growth within the Stibnite area. The initial timing of that shift is unclear, but the stretching indicated in the boudinaged layer may be late in the Cretaceous deformation and formed in a relatively deep setting.

In general, where coarse-grained pyrite or magnetite is visible in hand specimen, any metamorphic calc-silicate minerals are visually unaltered, and pyroxenes and amphiboles seem to coexist with the opaque minerals. However, finer grained pyrites were typically associated with retrograde or hydrothermal alteration, as discussed in subsequent chapters.

Figure 2-8. Lower calc-silicate unit showing euhedral actinolite intergrown with pyrite (opaque, black). Sample 8EB-008 (15V024A) in plane-polarized, transmitted light at x20 magnification.

Figure 2-9. Boudinaged calc-silicate layer with coarse-grained pyrite in tensional space between the boudins, West End pit. The purple-colored rock is typical of the argillaceous impure quartzites, which locally are interbedded with impure carbonate layers that developed calc-silicate minerals. It is likely the coarse pyrite clot is unmineralized, but it was not analyzed.
Intrusive Rocks

A major objective of the study was to determine the timing and genetic relationship between the multiple intrusive phases and mineralization. One challenge for developing this framework of intrusive history is the uncertain relationship between phases found in surface mapping versus subsurface coring. Midas geologists and prior workers recognized the presence of several phases of the Cretaceous Idaho batholith in the Stibnite district, as well as a variety of younger porphyritic dikes in the district. However, nomenclature and classification of the igneous rocks has differed over the years. For this study and report, the field terms and names used by Midas Gold and most historical workers are used. No attempt was made in this study to classify the original igneous rock type beyond a field term. A summary and correlation of the intrusive rocks is given in Table 2-2.

Additional complications are the compositional changes due to hydrothermal alteration and the uncertain relationships between phases seen in the subsurface and those mapped on the surface by Stewart and others (2016). For example, the alaskites logged in core are widely present as float on the surface (geologic map unit Klg) but do not form large outcrops. This may be due to poor surface exposures or the small size of the intrusive bodies.

Table 2-2. Intrusive rocks nomenclature correlations. DH stands for drill hole.

<table>
<thead>
<tr>
<th>Rock Name (this report)</th>
<th>Midas Gold</th>
<th>Cooper (1951, USGS Bull. 969-F) and earlier workers cited therein</th>
<th>Stewart and others (2016; IGS GM-51) after Streckeisen (1976) classification</th>
<th>Lithology (GM-51 Unit Symbol)</th>
<th>Location</th>
<th>Approximate Age (this study)</th>
</tr>
</thead>
<tbody>
<tr>
<td>IDAHO BATHOLITH</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Quartz monzonite (QM or Kgd), plus earlier more mafic phases.</td>
<td></td>
<td>Quartz monzonite (QM)</td>
<td>Granodiorite (Kgd)</td>
<td>Biotite granodiorite</td>
<td>Surface, DH</td>
<td>94.21 Ma</td>
</tr>
<tr>
<td>Stibnite stock (Kgd)</td>
<td>Stibnite stock</td>
<td>NA</td>
<td>Granodiorite of Stibnite stock (Kgd)</td>
<td>Muscovite-biotite granodiorite (fine-grained)</td>
<td>Surface, DH</td>
<td>85.71 Ma</td>
</tr>
<tr>
<td>Granite (GR/Kg) - underlies Yellow Pine pit</td>
<td>Granite (GR/GRAN)</td>
<td>NA</td>
<td>NA</td>
<td>NA</td>
<td>Deep DH</td>
<td>~ 85-86 Ma</td>
</tr>
<tr>
<td>Granite (Kg) - surface</td>
<td>Granite (GR/GRAN)</td>
<td>Granite</td>
<td>Granite (Kg)</td>
<td>Biotite-muscovite granite</td>
<td>Surface</td>
<td>?</td>
</tr>
<tr>
<td>Alaskite (AK/Klg)</td>
<td>Alaskite (AK)</td>
<td>Aplite</td>
<td>Leucocratic granite (Klg)</td>
<td>Granite-mafic poor</td>
<td>Surface, DH</td>
<td>83.61 Ma</td>
</tr>
<tr>
<td>TERTIARY DIKES</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rhyolite (Tr)</td>
<td>Rhyolite (RH)</td>
<td>Rhyolite</td>
<td>Rhyolite and Rhyolite porphyry (Tr, Trp)</td>
<td>Rhyolite - quartz-bearing</td>
<td>Surface, DH</td>
<td>Eocene</td>
</tr>
<tr>
<td>Latite and quartz latite porphyry (Tlp, Tdp)</td>
<td>Latite (LA)</td>
<td>Quartz latite porphyry; dacite porphyry</td>
<td>Dacite porphyry (Td, Tdp)</td>
<td>Dacites and dacite porphyries with biotite or hornblende</td>
<td>Surface, DH</td>
<td>47.86 Ma</td>
</tr>
<tr>
<td>Lamprophyres, basalts, mafic dikes (Tb, Timi)</td>
<td>Basalt/Diabase (DB)</td>
<td>Basalt; lamprophyres; andesite</td>
<td>Intermediate and mafic intrusives (Timi)</td>
<td>Dacite to basaltic andesite</td>
<td>Surface, DH</td>
<td>Eocene</td>
</tr>
<tr>
<td></td>
<td>Trachyte</td>
<td>NA</td>
<td>Surface</td>
<td>NA</td>
<td>?</td>
<td></td>
</tr>
</tbody>
</table>

NA = not applicable
To partially mitigate this challenge, radioisotopic
geochronology utilizing both U-Pb zircon and \(^{40}\)Ar/\(^{39}\)Ar
mica and feldspar geochronology was pursued
complementary to observations of relative cross-cutting
relationships from surface and core investigations.
The results of radioisotopic geochronology using the
chemical abrasion isotope dilution thermal ionization
mass spectrometry (CA-ID-TIMS) method are
summarized in Table 2-2; details of the age systematics
are in Chapter 4. The CA-ID-TIMS method was
pursued in order to improve upon and refine preliminary
interpretations from in situ laser ablation inductive
coupled plasma mass spectrometry (LA-ICPMS)
alyses reported in Stewart and others (2016).

Major intrusive phases in the Stibnite district, as listed
from oldest to youngest, are:

**Quartz monzonite (QM)/Biotite granodiorite (Kgd)**
– This unit is the major granitic rock exposed in the
Stibnite quadrangle. Though called “QM” in this study
to correlate with historical and Midas Gold terminology
(Table 2-2), it is a granodiorite in composition according
to modern classification, as defined by Streckeisen
(1976) and mapped by Stewart and others (2016).
Biotite granodiorite contains coarse-grained quartz,
feldspar (dominantly plagioclase), and biotite, and is
the predominant Idaho batholith phase exposed at the
surface near Stibnite and in much of central Idaho
(Figures 2-1 and 2-3). Midas geologists logged most of
the granitic rocks in core as quartz monzonite or “QM.”
The QM is equivalent to the Kgd unit of Stewart and
others (2016). Locally, the biotite granodiorite is foliated
and in places porphyritic. Fresh QM/Kgd is exposed
in a large roadcut on the north side of the Yellow Pine
pit on the northwest side of the Meadow Creek fault
(Figures 2-10 and 2-11). There the rock is very fresh,
locally porphyritic, and contains approximately 15
percent biotite with \(<10\) percent potassium feldspar.
Typically, the unit contains \(>20\) percent quartz (Stewart
and others, 2016). Sparse aplites and pegmatites cut the
granodiorite, and mafic inclusions are locally present
in the granodiorite. U-Pb CA-ID-TIMS zircon analysis
returned a result of 94.21 Ma \(\pm\) 0.22 Ma. This age is
slightly older than the imprecise 89.9 \(\pm\) 1.7 Ma LA-
ICPMS zircon date of Gaschnig and others (2017).
The particular sample dated in this study could be an
older, unmapped pluton that is more mafic than typical
QM/Kgd in the region (Reed Lewis, oral commun.,
2018). Box and others (2016) obtained zircon ages on a
biotite-muscovite granite (94.9 \(\pm\) 0.8 Ma) and a biotite
granodiorite (87.2 \(\pm\) 0.7 Ma) near the Yellow Pine
deposit.

Within the Stibnite district and for this study, the host
granodiorite will be called by its historic name, quartz
monzonite (QM), and it was so labelled in drill core
logged by Midas Gold. In the Yellow Pine pit, near
faults and mineralized areas, this granitic rock has been
strongly altered by hydrothermal solutions, although
local intervals of unaltered to weakly altered quartz
monzonite are present (Figure 2-11). Figure 2-12 shows
a thin section photo of “unmineralized” (only 0.024 ppm
Au) quartz monzonite for comparison to photos of more
altered rock. This weakly altered rock was collected in the Yellow Pine pit area from drill hole 12-243 at a depth of 193 ft. Lithologically, it is very similar to the biotite granodiorite described above, and the igneous texture and biotite are preserved in the drill hole sample.

Within the mineralized part of the district, the QM is a coarse-grained quartz-potassium feldspar-plagioclase-muscovite/sericite intrusive rock with sparse amounts of mafic minerals. Locally the rock preserves the porphyritic textures of the regional Kgd, but in other places the porphyritic texture is absent. Primary feldspar compositions may be obscured by alteration. It is unknown if the quartz monzonite comprises one intrusive phase or could include several phases obscured by alteration. An example is in Figure 2-13, which also shows a contact between quartz monzonite and alaskite.

**Stibnite stock (Kgds)** – As exposed in the Stibnite pit (Figure 2-3) and intersected in MGI-10-37, from 40 to 136 ft, this is a distinct intrusion as compared to the QM. In both hand specimen and thin section, the Stibnite stock has a fine-grained, equigranular texture that is nearly sucrosic (Figure 2-14). The Kgds consists of approximately 30 percent quartz, 50 to 60 percent feldspar of subequal plagioclase and potassium feldspar composition, and about 5 percent mica, including both biotite and muscovite. Locally, magmatic microcline is visible, as is muscovite (Figure 2-14). Though recorded initially by Midas as “QM” in MGI 10-37, it was redefined by surface mapping as a distinct unit (Figure 2-3). Texturally it is like the granite (“GR” or Kg) intersected in the bottom of drill hole MGI-306 at 940 ft depth below the Yellow Pine pit. The Stibnite stock appears to have intruded partially along the West End fault zone and probably spread out in sill-like fashion along the overturned thrust fault, as shown by Stewart and others (2016). Zircon separated from the Stibnite stock contained abundant inherited cores, but magmatic overgrowths and discrete grains were dated at 85.71 Ma ± 0.10 Ma using CA-ID-TIMS methods. This age is consistent with, but more precise than the 84.4 ± 2.1 Ma LA-ICPMS date reported by Gaschnig and others (2017).
Granite (GR/Kg) – Logged by Midas as “GR/GRAN”, the granite is distinguished from quartz monzonite in several holes. Granite samples from the West End drill hole MGI-12-305, at 422 ft and 434 ft depth contain approximately 30 percent quartz, 50 to 60 percent feldspar, and as much as 15 percent muscovite. Some of the muscovite is an alteration phase and has partially replaced biotite. The samples from MGI-12-305 are medium to coarse grained with an equigranular texture and locally strong potassic alteration, which overprints sericitized plagioclase. The quartz is recrystallized and strongly sutured. The coarse-grained muscovite-bearing phases seen in drill hole may be the same rock as the coarser-grained two-mica granite (Kg) mapped in surface exposures near the Garnet pit (Figures 1-2 and 2-3). Recent logging by Midas Gold suggests that the granite intersected at depth in West End drill hole MGI-12-305 is a distinct intrusion, although it is most like the Stibnite stock. A granite sample at 422 ft depth in MGI-12-305 from the West End area was dated by U-Pb CA-ID-TIMS analysis at 85.51 Ma ± 0.06 Ma based upon the youngest single zircon fragment analysis; inherited cores are common in its zircon population.

Granite was also intersected and logged at depth on the east side of the deep Yellow Pine deposit in MGI-12-306, from 940 to 950 ft downhole. There the rock is very fine grained and equigranular, with a texture similar to the Stibnite stock. Quartz comprises about 20 percent of the rock at the bottom of MGI-12-306 with 50 percent plagioclase and about 25 percent orthoclase with lesser microcline feldspar and minor mica (Figure 2-15).
The presence of two varieties of potassium feldspar is puzzling, and though both appeared primary, more work would be needed to confirm the mineralogy and relationships. Locally the granite is cut by a fluidized breccia dike with carbonate, opaque minerals and sericite (?) alteration. An attempt to date the granite in the bottom of MGI-12-306 was made, but the zircons showed large inherited cores and very thin rims. However, two spots on the rim were analyzed with LA-ICPMS and returned ages in the mid-80s Ma range, consistent with the aforementioned CA-ID-TIMS U-Pb zircon ages of samples Kgd and Kg.

**Alaskite (AK, Klg)** – Alaskite dikes, logged as “AK” in core by Midas geologists and classified as leucocratic granite (Klg) on the surface map of Stewart and others (2016), are common in the drill core and most easily identified by their cream color and lack of mafic minerals. Small pinkish red garnets are locally present. The alaskite bodies clearly cross cut and intrude the Kgd but without a chilled margin, suggesting a continued hot thermal regime and deep emplacement. A relatively fresh sample was collected from MGI-12-243 at 170.6 ft and core diameter is 2.5 in. identified by their cream color and lack of mafic minerals. The alaskite bodies clearly cross cut and intrude the Kgd but without a chilled margin, suggesting a continued hot thermal regime and deep emplacement. A relatively fresh sample was collected from MGI-12-306 at 550 ft for U-Pb zircon dating. It consisted of a fine-grained, anhedral mosaic of potassium feldspar (75 percent) and quartz with 1 percent of remnant biotite and almost no plagioclase; some of the potassium feldspar is a result of potassic alteration. The alaskites are commonly strongly altered and mineralized and cut by dark, sulfide-bearing veinlets (Figure 2-16). The alaskite dikes appear to have been more brittle and susceptible to fracturing than the granodiorite. Alaskite dikes are the youngest of the Cretaceous batholith phases, as judged by their finer grain size and sharp contacts with the other phases. Sample 12-306-550 was dated by U-Pb CA-ID-TIMS analysis on zircon rims at 83.61 Ma ± 0.08 Ma. Inherited zircon core ages were in the 89 and 94 Ma range. Our CA-ID-TIMS crystallization age supersedes the imprecise 87.9 ± 4.9 Ma LA-ICPMS date of Stewart and others (2016) and is consistent with cross-cutting relationships between intrusive phases.

**Tertiary Porphyry** – In both surface exposures and drill core, multiple porphyry dikes intrude batholith granitic rocks and cross cut fault zones. Dikes are also cut by faults. At least one mafic dike cuts stibnite-mineralized breccia. Compositions vary from mafic (Timi) to felsic (rhyolite porphyry, Trp), with many gray-green dikes of dacite (Td) or latite composition with feldspar and biotite phenocrysts (Figure 2-17). Many of the dikes have large phenocrysts in an aphanitic to very fine-grained matrix. Macroscopically, they show weak clay or chloritic alteration and minimal fracturing in comparison to the coarse-grained intrusive and metasedimentary rocks. The dikes are not visibly mineralized in hand specimen except for sparse remobilization of sulfides; assays for the dike intervals show a sharp drop in mineralization to only weakly anomalous levels. Alteration of the dikes is noticeably more intense when seen in thin section as described later.

Zircons from a sample of latite porphyry from drill hole 13-383 at 282 ft were dated by U-Pb CA-ID-TIMS methods at 47.86 Ma ± 0.07 Ma. The same dike was also dated with $^{40}\text{Ar}/^{39}\text{Ar}$ on potassium feldspar and biotite. A flat plateau cooling age on the feldspar was
45.9 Ma ± 0.3 Ma, and the weighted average age for biotite was 46.8 Ma ± 0.3 Ma.

Figure 2-18 shows a mafic dike with a chilled margin against a stibnite-matrix black breccia from the Hangar Flat area. The dike clearly post-dates mineralization and brecciation. Earlier workers also noted that the porphyry dikes intruded along fault zones and were post-mineral in age (Cooper, 1951). The mafic dikes, sometimes referred to as “lamprophyres” but more likely of basalt to andesitic compositions, were considered to be younger in age than the lattics and rhyolites (Cooper, 1951). Very few dikes within the district have been dated by radiometric methods; thus, it is possible that some of the mafic dikes could be younger than the 46 to 48 Ma period during which most of the regional Challis-related rhyolite to latite dikes were likely emplaced.

GE/OCHEMISTRY OF INTRUSIVE ROCKS

Eight whole-rock chemical analyses of relatively unaltered intrusive rocks in or near the Stibnite district were obtained from the companion study of Stewart and others (2016, Table 2-2). The X-ray fluorescence (XRF) analytical results and location information are shown on the map of Stewart and others (2016). Six of the samples are from the Idaho batholith and two from Challis-age intrusive bodies east of the mine area. Age determinations are not available for these specific samples, all of which are from surface outcrops.

Figure 2-19 compares the Stibnite rocks to the data of Gaschnig and others (2011, 2010) who studied the chemical and isotopic characteristics of the various suites of igneous rocks that comprise the Cretaceous-age Idaho batholith and also the younger Challis intrusive province in Idaho. The batholith has been divided into two general lobes, the Atlanta lobe to the south and the Bitterroot lobe to the north. The two lobes are separated by a zone of metamorphic rocks and Tertiary intrusions near the latitude of Elk City in northern Idaho (Figure 2-1; Gaschnig and others, 2011; Lewis and others, 2012a). U-Pb geochronology based on in-situ LA-ICP-MS determinations on zircons shows that each lobe started with a belt of metaluminous plutons followed by voluminous peraluminous granitic magmatism. Early metaluminous magmas and more mafic border zone rocks of the Atlanta lobe were emplaced from about 100 to 85 Ma, followed by peraluminous magmas making up the bulk of the Atlanta lobe between about 80 and 67 Ma; the Bitterroot lobe peraluminous magmas were emplaced from about 66 to 54 Ma but further north (Gaschnig and others, 2011). The Stibnite district is situated well within the Atlanta lobe terrane at its eastern margin (Figure 2-1).

The six samples of the Cretaceous Idaho batholith rocks near Stibnite include three units: granodiorite, the Stibnite stock, and granite (Figure 2-19; Stewart and others, 2016). The samples fall dominantly within
the two fields shown for the Atlanta lobe, although one granodiorite sample (Kgd) and the Stibnite stock sample (Kgds) are slightly more alkali-rich and close to the Atlanta peraluminous field. The three granite samples (Kg) analyzed plot within the Atlanta peraluminous field of Gaschnig and others (2011). The two Tertiary samples (Timi) were described as dacite (Stewart and others, 2016) and plot within the Challis intrusive field, but on the lower silica edge. While the number of analyses is few, the general pattern supports the conclusion that, prior to hydrothermal alteration, the Cretaceous granitoids and Tertiary igneous rocks at Stibnite were compositionally typical of the Idaho batholith and Challis intrusions, respectively.

Figure 2-19. Geochemical plots of Stibnite area samples (symbols) from Stewart and others (2016) compared to compositions of the Atlanta lobe of the Idaho batholith and Challis intrusive rocks from Gaschnig and others (2011). MALI (modified alkalis) is Na₂O+K₂O-CaO, in weight percent. Fe* is FeO/(FeO+MgO) with oxides in weight percent.
CHAPTER 3

HYDROTHERMAL ALTERATION
AND GOLD-ANTIMONY-TUNGSTEN
MINERALIZATION
INTRODUCTION

Currier (1935) provides an excellent summary of the geology and major alteration and mineralization types in the “Eastern Part of the Yellow Pine District,” which has come to be known as the Stibnite district. He distinguished three types of ore deposits: “arsenical gold ores, antimony-gold-silver ores, and mercury ores” (Currier, 1935, p. 16). Low-grade Au ores are characterized by Au-bearing pyrite and arsenopyrite within silicified shear zones hosted in the “granitic country rock, chiefly the quartz monzonite facies, also to some extent in quartzite of the roof pendant” (Currier, 1935, p. 16). The later Sb deposits are partly within the Au lodes but also in narrow, higher grade veins with pyrite, quartz, and carbonate in separate fracture zones (Currier, 1935). Antimony is present almost exclusively as the mineral stibnite (Sb$_2$S$_3$). The mercury deposits are located in a distinct zone at the heads of Cinnabar and Fern Creeks (Figure 1-2) and have clear epithermal characteristics. Currier (1935) first identified potassic alteration and muscovite associated with the quartz monzonite and Au mineralization at Yellow Pine. He describes the formation of microcline and some muscovite during replacement of plagioclase and ascribes that reaction to addition of silica and potash during late-stage, high-temperature magmatic processes. He also observed that the “rocks were modified further by hydrothermal alteration, during which quartz, muscovite, sericite, titanite, apatite, micas, and pyrite were deposited” (Currier, 1935, p. 8). Currier concluded that the low-grade Au and associated potassium feldspar alteration formed from “hot solutions” originating from the magmas of the Idaho batholith, but he notes that deposition of the Au and sulfides was “from solutions at comparatively high temperature rising from depth along fault zones; wall rock alteration (silicification and slight sericitization) by these solutions” (Currier, 1935, p. 20). This period of metal deposition was followed by “Stage 6” shearing of age “time indefinite but may have been much later than stage 5 possibly attending the period of Tertiary volcanism” (Currier, 1935, p. 20). His “Stage 7” consisted of “deposition of antimony ores from hydrothermal solution rising along fractures of Stage 6 (?)” (Currier, 1935, p. 20).

Currier (1935, p. 20) inferred ages and classification of the three types of mineralization as follows:

1. Gold arsenopyrite (hypothermal), pre-Tertiary, closely following the intrusion of the Idaho batholith.
2. Antimony, (mesothermal or epithermal), with some silver and gold probably pre-Tertiary, possibly early Tertiary, later than the gold-arsenopyrite and probably earlier than the cinnabar.
3. Cinnabar (epithermal), Tertiary, succeeding the Challis volcanics.”

However, without radiometric age determinations, Currier’s prescient conclusions lacked the absolute timeline needed to more precisely correlate mineralization and alteration to other geologic events in central Idaho.

Tungsten mineralization, as the mineral scheelite (CaWO$_4$), was not recognized until 1941 by Don White of the U.S. Geological Survey while working with Bradley Mining Company as part of the strategic minerals program of the U.S. Geological Survey and the U.S. Bureau of Mines (Cooper, 1951, p. 154). Subsequent workers have provided more isotopic evidence and geochemical data, but the work to date has not fully confirmed the early observations on absolute timing of mineralization (White, 1940; Lewis, 1984; Cooper, 1951; Cooiker and others, 1987). Lewis (1984) developed four sub-stages of Au mineralization followed by Sb mineralization. Cooper (1951) provided the most complete description of the Yellow Pine district and focused on the W, an important ore during World War II. Debate over the timing of alteration and Au introduction has been ongoing since the early days and primarily focused on the Yellow Pine deposit and the Meadow Creek mine, located near the current Hangar Flats resource (Figure 1-3). Mining at the West End is more recent (1980s) and the historic accounts do not cover it in detail.

This present study was designed to use modern dating techniques in conjunction with detailed petrography to address the age and character of the ore deposits at Stibnite and to provide a new and age-constrained interpretation of the paragenesis or sequence of alteration and ore mineral deposition. This chapter describes petrographic observations, characteristics, and temporal evolution of the alteration and mineralization at Stibnite. Details of the geochronology are in the following chapter.
ORE GEOMETRY AND CONTROLS

As in many deposits, structure and stratigraphy are the two dominant ore controls and have influenced the overall geometry and localization of the ore bodies. Mineralogy is a third control on Au deposition.

Structural Controls

Both regionally and within the district, hydrothermal alteration and mineralization is primarily controlled by and localized along structures related to the Meadow Creek fault, as was recognized by early workers in the district (Figures 1-2 and 2-3; Schrader and Ross, 1926; Currier, 1935; White, 1940; Cooper, 1951; Lewis, 1984). The Yellow Pine deposit, the largest known ore concentration, is located where the Meadow Creek fault bends from a north-south to a northeast orientation. The West End deposit is localized along a major northeast-trending fault, and the Hangar Flat deposit is along the main north-south trending Meadow Creek structure south of the Yellow Pine deposit (Figure 2-3).

Figure 3-1 shows maps and diagrams of the 2014 Midas block model Au resource of the intrusive-hosted Yellow Pine deposit (Huss and others, 2014). The long section and 3D view show the historic open pit mine relative to the newly outlined resource at depth. While additional drilling has modified the details of the reserve block, the overall geometry of Au mineralization is as an irregular ovoid, with a long axis of 400 meters and a short axis of approximately 160 meters localized between the Hidden and Meadow Creek fault zones (see cross section view in Figure 3-1). The long axis parallels the local N30ºE trend of the faults. Minor mineralization is spread out as pods along the fault strands with the highest-grade ore located in the southwest part of the pit. Historic mining and more recent work by Midas have identified several cross structures with east-west and N60ºE trends linking the two major bounding faults (Cooper, 1951; Christopher Dail and Austin Zinsser, oral commun., 2017). Recent drilling results by Midas Gold and past mine maps show that the stibnite mineralization is concentrated near the southern end of the Au resource zone at Yellow Pine and that it may be localized at fracture intersections with the more east-west cross structures (Cooper, 1951; Christopher Dail, written commun., 2017).

Figure 3-2 shows maps and sections of the more complex, sediment-hosted West End deposit, which is localized along a northeast-trending fault that branches off the main Meadow Creek zone. Within the metasedimentary rocks, replacement and disseminated mineralization is present, but as the plan view and cross section 15H-15H' show, Au mineralization is spatially associated with the northeast-trending West End fault zone and subsidiary northeast-trending faults. The long section and 3-D perspective models in Figure 3-2 demonstrate that the mineralization crosses the topographic ridges, underlain by northwest-striking resistant stratigraphic units, at an oblique angle. Mineralization is localized principally along the structures with stratigraphy being a second-order control.

An outcrop-scale example of fracture-controlled veining in the Stibnite pit is shown in Figure 3-3. The Stibnite pit is grouped with the West End area. Open-space textures are visible in this structure which cross cuts the metasedimentary rocks. Brecciation is a common texture in both igneous and metasedimentary hosts. Both fault breccias and hydrothermal breccias are present.

Stratigraphic Ore Controls

The West End deposit and a number of minor prospects are hosted in the metasedimentary rocks. Although structure is the primary ore control, a secondary control is stratigraphy. Figure 3-4 summarizes the location of known sediment-hosted ores within the stratigraphic column. Mineralization is present in multiple stratigraphic units and lithologies. Favored ore hosts are the calc-silicate units and impure marble units, as well as certain contacts that were probably zones of more pronounced fluid flow. As indicated in Figure 3-4, some contacts are unconformities. Zones with local iron oxide rich laminated siltites and breccia pods are interpreted as karst horizons, and they may have also had enhanced permeability. Massive pure quartzites are less mineralized, except where strongly broken in fault zones.

Interpretation of ore-related alteration and sulfide deposition is complicated by pre-ore events, including diagenetic or regional metamorphic sulfides and sulfides related to calc-silicate rocks (Figures 2-8 and 2-9). Sulfides, principally pyrite, are locally disseminated along the relict sedimentary bedding planes. While cross-cutting fractures and veins are typically associated with mineralized intervals, mineralization is not always
Figure 3-1. Plan map and sections of mineralization block model of the Yellow Pine deposit, from the 2014 prefeasibility study (Figure 15-10 from Huss and others, 2014, p. 15-15).
Figure 3-2. Plan map and cross-sections for the West End deposit, from 2014 prefeasibility study (Figure 15-12 of Huss and others, 2014, p. 15-17). Note coincidence of reserve blocks with faults (dashed blue lines) in top left plan-view drawing.
Mineralogic Ore Controls

A third ore control is mineralogy, particularly in igneous rocks, but also in the metasedimentary rocks. In many cases, pyrite deposition during the hydrothermal phase was favored by replacement of iron-bearing silicates by pyrite and eventually arsenopyrite in a sulfidation reaction. Thus, the biotite content of the quartz monzonite made it a particularly favorable host rock. The calc-silicate horizons observed on surface exposures and in drill core from MGI 10-37, which penetrated units below the Stibnite pit (Figures 1-2, 1-3, and 2-3) were typically composed of mixed carbonate and impure silty units or argillaceous carbonates, which are typically thin-bedded (Figures 2-7 and 2-9).

In mineralized samples from MGI 10-37, and in the Stibnite and West End pits, the primary calc-silicate minerals are locally altered to carbonate, quartz, sulfides, and hydrous retrograde minerals such as amphibole, chlorite, epidote, mica, talc, serpentine, and some clay. Similar to the intrusive rocks, sulfidation of iron in the metamorphic minerals or iron-bearing carbonates may have helped precipitate the Au-bearing sulfides.

OVERVIEW OF ALTERATION AND MINERALIZATION

Introduction

Hydrothermal alteration is superimposed on and later than metamorphism and calc-silicate development. In general, alteration is strongly controlled by lithology but consists of “retrograde” assemblages within calc-silicate units. In brittle rocks, such as quartzite or marble, silification and quartz-carbonate veining is common. Potassic alteration is pervasive in the igneous rocks and it is also found locally in the metasedimentary rocks.

Figure 3-3. Quartz vein cross-cutting calc-silicate unit and metasiltstone, Stibnite pit. Location of sample 12VG148N2 with potassium feldspar envelope, dated at 51.7 Ma. No assay data is available, but relict sulfides were noted in thin section.

Figure 3-4. Stratigraphic column with zones of mineralization shown in red; Figure 7.3 from Huss and others, 2014, p. 7-14.
Gillerman and others—Chapter 3

Though chemical zoning which correlates higher Au and As with potassium enrichment has been identified in the intrusive-hosted Yellow Pine deposit (Zinsser, 2016), a simple “bullseye” pattern of the alteration and mineralization is not present in the West End area. This is due to the heterogeneous structural and lithologic patterns as well as the multiple mineralization and alteration events.

Ore-stage alteration is recognized as that alteration which has overprinted consistently across diverse lithologies in the district and cross cuts earlier alteration. The major alteration mineralogy associated with this higher-grade Au mineralization is simple. Earlier workers recognized that this alteration dominantly consists of quartz, calcite or carbonate, and potassium feldspar plus sulfide minerals (Currier, 1935; Cooper, 1951; Lewis, 1984). Figure 3-5 shows a sample of the pervasive potassic alteration from the intrusive-hosted Yellow Pine deposit.

Sulfide mineralization is typically localized in shear zones as cross-cutting veins and hydrothermal breccias, but also disseminated in intrusive rocks and favorable stratigraphic beds such as the calc-silicate units, some schists, and marbles. Gold is hosted within arsenian pyrite and to a lesser extent in arsenopyrite. Both sulfides are very fine grained in the mineralized zones and they can be difficult to see in hand sample. Disseminated pyrite is typically located near a vein or fracture or within a hydrothermal breccia matrix. Stibnite is a dark gray color and typically forms dark gray, metallic coatings on cross-cutting fractures and veins. Especially at Hangar Flats, stibnite is an abundant phase in black breccia matrix.

The Au, Sb (as the mineral stibnite), and W mineralization all show a ubiquitous and close association with quartz and carbonate veining and brecciation. Tungsten mineralization, which is late, has minimal potassium feldspar directly associated with it, but the potassium feldspar in the wall rock typically remains unaltered, suggesting the scheelite was deposited from fluids in which potassium feldspar, as well as stibnite, quartz,

Figure 3-5. Altered and mineralized quartz monzonite from the Yellow Pine pit with intense potassic alteration associated with the quartz-carbonate-sulfide assemblage and veins. Thin section 8EB-013 (15VG033) was stained yellow for potassium feldspar over the bottom ¾ of the section; unstained thin section billet on right is 2.5 cm wide.
and calcite, were stable. Table 3-1 provides a summary of the alteration and ore minerals identified in this study along with an interpreted precipitation timeline or paragenesis.

The most significant alteration types seen in this study and by Midas are: 1) alteration of biotite to sericite, and biotite to sericite plus sulfide; 2) calcite veins and calcite flooding; 3) potassium feldspar flooding and veins; and 4) silicification (silica flooding) and quartz or silica veins (Table 3-2).

Those alteration types reflect the dominance of potassic alteration (typified by introduction of potassium), the replacement of biotite by muscovite (sericite), and the ubiquitous carbonate alteration associated with quartz and sulfides. Intense alteration of feldspar directly to sericite, as would be typical of phyllic alteration in a porphyry-style deposit, is rare at Stibnite. In contrast, the volume of carbonate veining and alteration is much greater at Stibnite than in many other Au deposits visited by the senior author.

Drill Hole and Sample Locations and Methods

Core logging or inspection and sampling was conducted on ten representative diamond drill holes which are listed in Appendix A-1 along with the drill hole coordinates. The ten holes represent a small subset of the over 400 recent core holes from the property. Locations of the drill holes and geochronology samples are indicated on the district map in Figure 3-6 and on the Yellow Pine inset map A in Figure 3-7 and the West End inset map B in Figure 3-8.

Samples from both surface and drill hole locations were collected to study the alteration and ore mineralogy and to select samples for age determinations and microprobe work. Midas drill hole logs, completed by their geologists prior to this study, provided a starting framework basis for the independent petrographic work in this study. Assays and whole-rock geochemical analyses helped to correlate the rock chemistry with observations from microscope examination of over 110 thin sections (Appendices B and C). Table 3-2 lists the rock and alteration types used by Midas for core logging. Samples for age determinations and the Au content of drill hole samples used in this study are shown on left side of the Midas bar graph for MGI-12-306 shown in Figure 3-9.

The hole was drilled at a shallow angle under the center of the Yellow Pine deposit, intersecting over 900 feet of quartz monzonite cut by numerous alaskite bodies. The hole was well mineralized and as indicated in Figure 3-9, much of the core assayed >2 ppm Au with some intervals >5 ppm Au. Antimony is locally quite high (>2 percent) but much more varied in its distribution than Au. Also evident is the consistently high elemental K content (4 percent), coupled with depleted elemental Na (<0.5 percent) content compared to granitic rocks from the Stibnite area. Whole-rock analyses of five granitic rock samples from the Stibnite area indicates K2O ranges from 3.11 to 4.78 weight percent and Na2O ranges from 3.20 to 4.46 weight percent (Stewart and others, 2016). Potassium content of 4 percent in the altered rocks equals 4.82 weight percent K2O and Na content of 0.5 percent equals 0.67 weight percent Na2O, which indicates an enrichment in K and a very strong depletion of Na in the altered intrusive rocks of the Yellow Pine deposit.

Figure 3-10 shows the Midas bar graph for drill hole MGI-12-305 from the sediment-hosted West End deposit. Geochronology samples with Au contents are indicated on the left side. Mineralization in 12-305 is locally good grade (>2 ppm Au) but much less continuous than in the Yellow Pine hole. The highest grade Au is associated with a breccia zone in schist near a contact that is probably a fault. Antimony in the West End deposit is negligible, requiring a different bar scale for this hole than in Figure 3-9. The bar graphs for MGI-10-37, located near the Stibnite pit, and for MGI-12-193 from the stibnite-rich Hangar Flats deposit are in Appendix B.

SEQUENCE AND TYPES OF ALTERATION AND MINERALIZATION

The visual appearance and petrographic characteristics of the hydrothermal alteration and mineralization of the ores at Stibnite are quite varied in detail and are described and illustrated more specifically in this section. The differences reflect the varied host rock lithologies and the multiple ages of mineralization, which have had a cumulative effect on the rock characteristics. Intrusive host rock compositions are less heterogeneous than those of the metasedimentary rocks, and thus the alteration assemblages and textures
Table 3-1. List of mineral occurrences at Stibnite. Rock associations are: Metamorphic, Igneous, or Hydrothermal. Mineral abundance in each association column is indicated by: A = abundant; m = minor; r = rare. Location: Is mineral in wall rock or vein? Amount indicated by: X = abundant, x = present. Timing of mineral deposition: I = igneous, M = metamorphic, E = early hydrothermal stage, O = ore-bearing hydrothermal stage, L = late hydrothermal stage.

<table>
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<th>Hydrothermal</th>
<th>Rock</th>
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<tr>
<td>Molybdenite</td>
<td>r</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>E</td>
</tr>
<tr>
<td>Pyrrhotite</td>
<td>r</td>
<td></td>
<td></td>
<td>x</td>
<td></td>
<td>E</td>
</tr>
<tr>
<td>Pyrite (coarse)</td>
<td>m</td>
<td>m</td>
<td></td>
<td>x</td>
<td>x</td>
<td>M, E</td>
</tr>
<tr>
<td>Pyrite (fine)</td>
<td>A</td>
<td></td>
<td></td>
<td>X</td>
<td>X</td>
<td>O mostly; E-L (minor)</td>
</tr>
<tr>
<td>Arsenopyrite</td>
<td>A</td>
<td></td>
<td></td>
<td>X</td>
<td></td>
<td>O</td>
</tr>
<tr>
<td>Stibnite</td>
<td>A</td>
<td></td>
<td></td>
<td>X</td>
<td></td>
<td>O, L</td>
</tr>
<tr>
<td>Scheelite</td>
<td>m</td>
<td></td>
<td></td>
<td>x</td>
<td>x</td>
<td>L, Very L</td>
</tr>
<tr>
<td>Cinnabar</td>
<td>r</td>
<td></td>
<td></td>
<td>x</td>
<td>x</td>
<td>L (?)</td>
</tr>
</tbody>
</table>
Table 3-2. Alteration and lithology classification and abbreviations used by Midas Gold for core logging at Stibnite.

<table>
<thead>
<tr>
<th>Abbreviation (Midas Gold)</th>
<th>Alteration Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>BSSF</td>
<td>Biotite to sericite and sulfide</td>
</tr>
<tr>
<td>BSER</td>
<td>Biotite to sericite</td>
</tr>
<tr>
<td>FSER</td>
<td>Feldspar to sericite</td>
</tr>
<tr>
<td>BCHL</td>
<td>Biotite to chlorite</td>
</tr>
<tr>
<td>SILV</td>
<td>Silica veins</td>
</tr>
<tr>
<td>SILF</td>
<td>Silica flooding</td>
</tr>
<tr>
<td>CALV</td>
<td>Calcite veins</td>
</tr>
<tr>
<td>CALF</td>
<td>Calcite flooding</td>
</tr>
<tr>
<td>KSPF</td>
<td>K-feldspar flooding</td>
</tr>
<tr>
<td>DOLV</td>
<td>Dolomite veins</td>
</tr>
<tr>
<td>DOLF</td>
<td>Dolomite flooding</td>
</tr>
<tr>
<td>FEPI</td>
<td>Feldspar to epidote</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Abbreviation (Midas Gold)</th>
<th>Rock Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>OB</td>
<td>Overburden</td>
</tr>
<tr>
<td>RH</td>
<td>Rhyolite</td>
</tr>
<tr>
<td>LA</td>
<td>Latite</td>
</tr>
<tr>
<td>QM</td>
<td>Quartz monzonite</td>
</tr>
<tr>
<td>AK</td>
<td>Alaskite</td>
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<td>GRAN</td>
<td>Granite</td>
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<td>BX</td>
<td>Breccia</td>
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<tr>
<td>GO</td>
<td>Gouge</td>
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<tr>
<td>CS</td>
<td>Calc-silicate</td>
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<tr>
<td>SCH</td>
<td>Schist</td>
</tr>
<tr>
<td>QZ</td>
<td>Quartzite</td>
</tr>
<tr>
<td>CARB</td>
<td>Carbonate (marble)</td>
</tr>
<tr>
<td>QPC</td>
<td>Quartz pebble conglomerate</td>
</tr>
<tr>
<td>QZ_D</td>
<td>Quartzite (dirty)</td>
</tr>
</tbody>
</table>
Figure 3-6. Geologic map of Stibnite district with inset maps A (Figure 3-7) and B (Figure 3-8) showing location of geochronology samples and drill holes used in study. Geology from Stewart and others (2016).
Figure 3-7. Map A: Shaded relief enlargement of the Yellow Pine mine area, showing open pit (red hatched curves) and Midas resource area (blue stripes). Horizontal projections of drill holes studied are in green and Midas cross section line is in black. Age determination symbols are same as in Figure 3-6.

are simpler to work out. Hydrothermal alteration of the intrusive rocks will be described first in its interpreted sequence from early to late, followed by a description of alteration and mineralization in the metasedimentary rocks. This is followed by a description of the late Sb-W mineralization and late alteration, which affects all rock types near the Meadow Creek fault zone. Short, simplified thin section descriptions, along with selected Midas core logs, are in Appendix B, and photos of the thin section billets are in Appendix C. Geochronology and pyrite morphology and chemistry are discussed in later sections.

Early Pre-Ore Alteration

There are a few early alteration types which developed during or shortly after intrusion of the Cretaceous granitoids at Stibnite. The early alteration does not appear to be associated with significant Au mineralization. These early alteration facies include the Cretaceous quartz-molybdenite veining at West End described by Konyshev and Muntean (2016), rare garnet-pyroxene skarn which may also be Cretaceous in age, and unusual coarse muscovite bands cutting granite and metasedimentary rocks.

Calc-silicate horizons are widespread and Cretaceous in age, but true garnet-pyroxene "skarn" indicating extensive mass transfer is rare in the district and much more localized. The principal exposure of skarn was in the Garnet Creek deposit, which was mined in the 1990s and subsequently back-filled. Descriptions by
Figure 3-9. Bar graph of Midas log of DH MGI-12-306 through the core of the Yellow Pine deposit. Pink rhombs are intrusive ages. Blue triangles are alteration ages. Black dots are thin section samples; those with lime green circles are from intervals assaying >1 ppm Au. See Table 3-2 for legend. Chemical values are weight percent of the element except for Au.
Figure 3-10. Bar graph of Midas log for MGI-12-305 through the West End deposit. Black dots are thin section samples; those with green circles are from intervals assaying >1 ppm Au. Blue triangles are alteration dates; vertical green rhombs are cooling ages; horizontal pink rhombs are intrusive ages. See Table 3-2 for legend. Chemical values are weight percent of the element except for Au which is in ppm.
Cooper (1951) and other early workers indicate the Au mineralization at Garnet Creek is associated with structures rather than a carbonate/intrusive contact. Only sparse garnet was seen in the Stibnite pit; pockets of late actinolite-carbonate retrograde skarn containing remnant magnetite and two types of pyrite were collected from the West End area. A study of the skarn was not attempted. It is likely that Au in the skarn-like rocks was precipitated by later hydrothermal activity, similar to the processes described below.

The third, apparently barren, alteration type consists of bands or veins of monomineralic, coarse muscovite flakes, locally in radiating aggregates that cut granite and metasedimentary rocks. In samples from the West End and Yellow Pine areas, the mica bands are cross cut at a high angle by potassium feldspar and carbonate veins that belong to the ore-forming assemblage (Figure 3-11). Similar-looking mica was noted at the contact of strongly potassically altered quartz monzonite with a small, less potassically altered alaskite dike. A few grains of tourmaline were also seen in the alaskite in this sample (OKL-001; 12-306-382). One hypothesis is that the early mica veins are Cretaceous, greisen-like alteration features related to the alaskite intrusions. These earliest muscovites are clearly distinct from the ore-related muscovites described below. The two varieties of muscovite may lead to confusion if the sample paragenesis is not clearly understood prior to conducting age determinations.

Alteration and Mineralization in Intrusive Rocks

Early Au Ore Stage I

As the intensity of hydrothermal alteration increases, the intrusive textures become less distinct and the rock color lightens. Igneous biotite is entirely replaced by muscovite (or possibly illite) plus carbonate with titanite or rutile. However, plagioclase is only locally and partially converted to fine-grained, disseminated sericite. In mineralized zones, pyrite replaces the Fe-Ti oxides and additional pyrite and arsenopyrite are precipitated along mica cleavages and disseminated in the surrounding wall rock. Hydrothermal potassium feldspar precipitated around the earlier sericitized plagioclase. Near the ore-stage veins, igneous microcline is recrystallized to hydrothermal potassium feldspar. Carbonate veins with quartz and sulfides and envelopes of potassium feldspar locally cross cut the pervasively altered and mineralized rock. Therefore the veins must be later in the paragenesis than the sericitization of the plagioclase.

Within the intrusive rocks, the early phases of ore-related alteration were observed in samples from drill hole MGI 12-314, for which the Midas core log is shown in Appendix B. Drill hole 12-314 is on the outer fringe of the main Yellow Pine ore zone (Figure 3-7). Figure 3-12 shows hand specimen examples of wall-rock alteration surrounding a white carbonate vein from 441.7 ft in MGI 12-314. The host biotite
granodiorite is only weakly fractured and the igneous textures preserved. Within the main part of the deposit, these early reactions are largely obscured by intense alteration. The progression of alteration from the fresh rock towards the white carbonate vein was visible over a few centimeters with locations A, B, and C shown on Figure 3-12. Next to the fresh rock, an outer zone of biotite is converted to a thin zone of green chlorite plus oxides. Closer to the vein, the biotite is replaced by pseudomorphs of coarse muscovite plus carbonate and titanite, as shown in Figure 3-13. As alteration intensity increases closer to the vein, the mica sites are sulfidized and pyrite replaces the oxides and mica structure along cleavage traces. Finally, the original biotite is entirely replaced by muscovite plus sulfides with minor carbonate and local titanite (Figure 3-14). The complexities of pyrite and subsequent arsenopyrite deposition are better shown in the SEM image and compositional maps in Figures 3-15 A-D. The pyrite includes low and high As-bearing varieties and arsenopyrite typically fringes the sulfidized mica pseudomorph with remnant Ti-bearing minerals (Figure 3-15 B-D). Apatite is commonly preserved and sparse monazite grains are present. It is unclear if those were original igneous phases or if some of the phosphates may be hydrothermal. The monazite grain in Figure 3-15A is located near the late carbonate vein. Finally, a late, cross-cutting, dark gray veinlet of quartz-carbonate-sulfides with a trace of scheelite in a microbreccia texture bisects the earlier carbonate vein, as is visible in Figure 3-12. The presence of scheelite suggests that the late veinlet may belong a much later stage of mineralization.

Figure 3-12. Core from the peripheral portion of Yellow Pine deposit, drill hole MGI-12-314, at 441.7 ft showing location of three thin sections (6RJ-016, 6RJ-017, and 6RJ-018) at locations 441.7A, B, and C respectively. Freshest quartz monzonite is on far left side and biotite destruction envelope is light gray around white carbonate vein. Whitish zone between B and C shows moderate sericite alteration of plagioclase overprinted by potassium feldspar and carbonate alteration. A late gray veinlet of fine-grained quartz-carbonate microbreccia containing a tiny grain of scheelite is subparallel to but cuts the white carbonate vein. Interval assayed 0.118 ppm Au. Core is 2.5 in diameter.
Figure 3-13. Biotite replaced by muscovite, carbonate, and titanite (slightly brownish in transmitted light but hard to distinguish from carbonate) from core specimen B in Figure 3-12. Section 6RJ-017 (12-314-441.7B) at x50 magnification with yellow stain for potassium feldspar, in plane-polarized (A) and cross-polarized (B) light.

Figure 3-14. Biotite altered to muscovite plus carbonate and opaque minerals, including rutile and some sulfides. Plagioclase is almost completely replaced by sericite. Sparse euhedral apatite is gray. From section 6RJ-016 (12-314-441.7A) at x100 magnification in plane-polarized (A) and cross-polarized (B) light. From hand specimen A in Figure 3-12.
Zinsser (2016) studied the alteration zoning at the Yellow Pine deposit with visible to near infrared to short-wave infrared (VIS-NIR-SWIR) spectroscopy, and partial multi-element, whole-rock geochemistry on drill core. An assemblage of illite-chlorite was documented outside of the deposit’s bounding structures. Within the actual Au-mineralized zone, Zinsser identified a peripheral zone of illite and calcite (or iron-bearing carbonate) plus moderate potassic alteration. Overall, Zinsser (2016) describes the Au mineralization as associated with illite and ammonium-illite assemblages and extensive potassic alteration. The ferroan carbonate is probably similar to the dusty appearing carbonate seen in thin section and formed during the alteration of biotite or other iron-magnesium minerals.

Ore Stage II or Main Au Stage

Gold-related hydrothermal alteration in the quartz monzonite is dominated by extensive potassic alteration, typically related to structures and veining. Pyrite-bearing veins are typically narrow (1 to 5 mm width), with stibnite veins up to 10 mm for individual veins within thicker zones that may be 0.5 m or more across. Pyrite and arsenopyrite grains in the veins are microscopic in size; visible Au is absent.

Mineralization in the veins, crinkle veins, and breccia zones typical of the Au-bearing intervals consists of a remarkably simple mineral assemblage: quartz-carbonate-sulfides. Potassium feldspar is ubiquitous.
in the wall rock as an alteration phase and locally as a selvage adjacent to the veins. Muscovite rims sulfides or is proximal to sulfides in only a few places. Development of coarse muscovite bordering the veins may depend on the host rock composition or it may have simply formed during cooling. In one sample, coarse-grained, corroded pyrite has a muscovite rim where it adjoins microcline, suggesting conditions close to the muscovite-potassium feldspar phase equilibria boundary. In many samples, textures are very similar to classic epithermal textures such as cockscomb quartz and drusy crystals filling open spaces. Figure 3-16 shows weak peripheral style potassium feldspar veining below the Yellow Pine pit, whereas Figures 3-17 and 3-18 show photomicrographs with typical examples of strongly altered and Au-

mineralized intervals (3.7 ppm Au) in the core of the Yellow Pine deposit. The pyrites shown in Figure 3-18 carried 0.01 to 0.06 weight percent (up to 600 ppm) Au in some growth bands, and they are both within and adjacent to a quartz-carbonate vein with a potassic alteration envelope which occupies the entirety of the image width, but is difficult to see without the staining.

Previous authors described some of the alteration potassium feldspar at Stibnite to be adularia, a low-temperature potassium feldspar with a distinctive crystal habit that is frequently associated with hydrothermal veins, often in epithermal environments (White, 1940; Lewis, 1984; Cookro and others, 1987). Lewis (1984) used x-ray diffraction to identify a sample as monoclinic
potassium feldspar which he called adularia. While no attempt was made in this study to determine the precise composition or structural state of the numerous potassium feldspars observed in this study, probable adularia was noted in thin section 8EB-013 (15VG033, Yellow Pine pit) in a vein cutting igneous microcline, similar to associations noted by White (1940) and Lewis (1984).

Ore-related alteration in the Stibnite stock and granite exposed in drilling at the West End deposit is mineralogically similar to that described from the Yellow Pine deposit, but different in appearance. Distinctive pale-pink veins cut the Stibnite stock and other granitic intrusive rocks. Figure 3-19 shows a pink vein cutting granite in core from 434 ft in MGI 12-305 from the West End area (Figures 3-8 and 3-10). The pink color is likely due to Mn-bearing carbonate whereas the potassium feldspar is white. Another example of the same granite (5WR-001, 12-305-422.1) cut by a similar pink vein with potassic alteration is shown in thin section in Figure 3-20. Sample 5WR-001 returned an age of 51 Ma on potassium feldspar.

Alteration and Mineralization in Metasedimentary Rocks

The Stibnite roof pendant strata are a succession of sedimentary rock types, all metamorphosed, including marble (both dolomitic and calcareous), argillite, schist, quartzite, and interbedded impure carbonate and calcareous or dolomitic siltite (Figure 3-4 and Table 3-1). Primary sedimentary layering is still distinguishable in many cases (Figures 2-7 and 2-9). However, the original mineral assemblages have been converted to metamorphic assemblages that still preserve the compositional layering. This indicates minimal metasomatic chemical exchange across the layers during metamorphism. Where the rock has been broken, veins cut obliquely across the layers and new hydrothermal minerals have formed. Within the ore zones these typically follow the simple mineralization suite of carbonate-quartz with pyrite and potassium feldspar in suitable wall rocks.

Barren and Early Alteration

Overall, early alteration that is distinct from the metamorphism is difficult to quantify or identify. Where the anhydrous, high-temperature minerals are fresh, the rock did not appear to be mineralized. Pyroxene in the calc-silicate layers is only locally altered to amphibole with minor chlorite and epidote, but retrograde or hydrous alteration of the calc-silicate rocks and skarn is much less extensive than would typically be found in a porphyry-type setting.

Bleaching and early alteration was noted in hand specimen and thin section where thin veins or fractures cut bedding or layering at a high angle. In core from DH MGI-10-37 in the Stibnite pit (Figure 3-10, Appendix B) zones of purplish meta-argillite have retrograde or late alteration of biotite to chlorite and possible clay. In hand specimen, the original purplish color is bleached along cross-cutting fractures or converted to a pale green color, producing a “Christmas-tree”
Figure 3-19. Core showing granite cut by pink vein crosscut by sulfide-bearing pink veinlet. Sample MGI 12-305-434 from West End area (Figure 3-8 and 3-10). Core is 2.5 in diameter.

Figure 3-20. Photomicrograph of pink vein similar to that in Figure 3-19. Vein of pink carbonate-quartz with 51 Ma potassium feldspar alteration envelope is oriented horizontally. Euhedral quartz and potassium feldspar crystals line vein margin. Thicker vein is cross cut by later quartz-carbonate veinlet, oriented diagonally on left. Sample 5WR-001 (MGI-12-305-422.1) in cross-polarized light with yellow stain for potassium feldspar at x40 magnification.
appearance. However, a relation of this bleaching to mineralization has not been observed. Rather, it may be a more peripheral indicator of fracturing and fluid flow. Some zones of intense alteration are barren of Au, such as a long interval from 651 to 715 ft in DH MGI 10-37 of quartz-veined, silicified, and brecciated marble with iron oxides (Appendix B). These overturned strata likely represent a “terra rosa” unconformity or altered sedimentary karst horizon in the marble. Rare veinlets of chlorite, clay, and sparse mineralized veins of quartz-carbonate-opaques with local potassium feldspar or muscovite may be the most common peripheral indicators of mineralization. The chlorite is interpreted as forming early in the alteration sequence, although exactly when is problematic.

**Ore-Stage Alteration and Gold Mineralization**

Ore-stage alteration can be subtle and is easily overlooked in hand specimen. Figure 3-21 shows a box of unmineralized core from DH MGI-10-37 next to a box with mineralized core from the same drill hole (Figures 3-8 and 3-22; Appendix B). The lithologies in the two boxes are similar, a purplish-colored argillite with sparse calc-silicate and quartz-rich layers. However, the Au content of the unmineralized box intervals was not detectable. The mineralized box had an interval that contains over 2.2 ppm Au (Figure 3-22). Within the Au-bearing zone is a small white quartz-carbonate vein, which hosts Au-bearing pyrite. However, the amount of visible veining in the interval is minor and could easily be overlooked.

As in the igneous rocks, ore-related alteration is characterized by carbonate with lesser quartz and potassium feldspar present in veins, fracture-fillings, and breccia matrix. In a few areas of surface exposures of carbonate-rich rocks near the Fern mine and nearby prospect areas, zones of silicification traverse through sanded dolomite. Sanding is due to hydrothermal dissolution along grain boundaries, which creates a friable sand-like texture. Weathering may enhance or mimic sanding. Some jasperoids and adjacent sandy areas contain significant Au, as determined by Midas 2014 exploration work (Christopher Dail, oral commun., 2015). Similar alteration and sanding is also present in siltites from the bottom of the Stibnite pit, where the rock contains clay and secondary carbonate present in irregular veinlets and possibly replacing the matrix minerals (Figure 3-23). Sanding may be accentuated by surface weathering and disaggregation of pyrite-bearing marble in the pit.

In the West End pit thin, resistant, tan chalcedonic quartz veins and white veins of milky quartz cut the marbles at a high angle to any sedimentary structure (Figure 3-24). The chalcedony veins vary from microscopic to several centimeters thick, but are typically less than 2 mm thick. Konysh of and Muntean (2016) documented that the chalcedony veins cross cut early milky white quartz veins which must be higher temperature in origin than the chalcedony. Mineralization in the quartzite and calc-silicate units at West End is characterized by very fine grained pyrite and arsenopyrite, but the sulfides are difficult to distinguish megascopically.

Hydrothermal breccias locally cut the metasedimentary rocks as seen in drill core and in the Stibnite pit. Hydrothermal breccia was seen on the north wall of the Stibnite pit, where a 10-cm wide, dike-like vein cuts cleanly across the bedding at a high angle with a roughly north or northeast strike (Figure 3-3). Historic map records provided by Midas indicate a Au anomaly following approximately the same trend. Figure 3-25 shows some of the vein textures with quartz crystals lining clasts of siliceous and apparently brittle wall rock.

The high-angle structures and veins in the north wall of the Stibnite Pit show open-space filling, brecciation, drusy and cockscob quartz with infilling calcite, and wall-rock alteration selvages of potassium feldspar similar to features of epithermal veins from Tertiary extensional terranes common to Idaho, Nevada, and elsewhere. Vein minerals in Figures 3-3 and 3-25 are quartz, carbonate, and iron oxides, with a potassium feldspar alteration envelope adjacent to the metasedimentary wall rocks. In samples of Au-mineralized silt marble from MGI 10-37 below the Stibnite pit, calc-silicate minerals are clearly pseudomorphed by an assemblage of quartz-carbonate-muscovite or other hydrous minerals, indicative of the lower temperature of the ore-related hydrothermal event.

Figure 3-26 shows a mineralized hydrothermal breccia example from 462 ft depth in drill hole MGI 12-305 at the West End deposit (Figures 3-8 and 3-10). The mineralogy is the same as that in the thin section shown in Figure 3-27: a quartz-carbonate vein with potassium feldspar alteration and sulfides in the wall rock. At least two periods of hydrothermal veining and brecciation are visible in the sample in Figure 3-26. Many of the breccia fragments visible in Figure 3-26 are almost totally converted to white potassium feldspar. The
Figure 3-21. Unmineralized core: MGI-10-37, Box 67: 634 to 643 ft. The interval 630 to 645 ft assayed not detectable (ND) for Au. Boxes are two ft long. Drill hole location in Figure 3-8 with core log in Appendix B.

Figure 3-22. Mineralized core: MGI-10-37, Box 58: 552 to 561 ft. The interval 550 to 555 ft assayed 0.09 ppm Au, the interval 555 to 560 ft assayed 2.24 ppm Au, and the interval 560 to 565 ft assayed 2.18 ppm Au. Note possible fault gouge at 553.5 ft and sparse veins at high angle to sedimentary layering. There is a cross-cutting white vein at about 559 ft in lower left corner. Boxes are two ft long.
breccia (logged from 470 to 501 ft) has subrounded to subangular clasts of altered metasedimentary rock, and it lies between a quartzite-schist unit above and an underlying silicified and brecciated, but only weakly mineralized, carbonate unit (Figure 3-10). It is unclear whether the contact and breccia are fault related, but that is a strong possibility, although a stratigraphic contact may also have provided a preferential fluid pathway. The highest Au concentrations are from 445 to 468 ft, localized in the more brittle quartzite-schist. In thin section, strong potassic alteration is indicated by potassium feldspar that forms vein envelopes on composite quartz-carbonate veins (Figures 3-27 and 3-28). Foliated muscovite is cut off by the vein, suggesting that some of the muscovite is early and perhaps metamorphic in origin (Figure 3-28). The early muscovite is overprinted by the potassium feldspar. Approximately 0.5 percent pyrite as tiny pyritohedrons is localized in the potassic alteration envelope, but some sections show more abundant sulfides, consisting of both pyrite and arsenopyrite (Figure 3-28). Late vuggy carbonate veins cut the breccia (Figure 3-26). The interval from 453 to 460 ft assayed 6 ppm Au; the interval below, from 460.5 to 468 ft, assayed 3.4 ppm Au.

Figure 3-23. Oxidized and weakly sanded metasiltstone with pyrite in bottom of Stibnite Pit. Note the planar brown streaky appearance showing stratigraphic control of mineralization.

Figure 3-24. Quartz-calcite veins cut Fern marble in upper West End pit (15VG006). No significant sulfides are visible, and no assay was available. Compare to thin section of similar rock in Figure 3-30.

breccia (logged from 470 to 501 ft) has subrounded to subangular clasts of altered metasedimentary rock, and it lies between a quartzite-schist unit above and an underlying silicified and brecciated, but only weakly mineralized, carbonate unit (Figure 3-10). It is unclear whether the contact and breccia are fault related, but that is a strong possibility, although a stratigraphic contact may also have provided a preferential fluid pathway. The highest Au concentrations are from 445 to 468 ft, localized in the more brittle quartzite-schist. In thin section, strong potassic alteration is indicated by potassium feldspar that forms vein envelopes on composite quartz-carbonate veins (Figures 3-27 and 3-28). Foliated muscovite is cut off by the vein, suggesting that some of the muscovite is early and perhaps metamorphic in origin (Figure 3-28). The early muscovite is overprinted by the potassium feldspar. Approximately 0.5 percent pyrite as tiny pyritohedrons is localized in the potassic alteration envelope, but some sections show more abundant sulfides, consisting of both pyrite and arsenopyrite (Figure 3-28). Late vuggy carbonate veins cut the breccia (Figure 3-26). The interval from 453 to 460 ft assayed 6 ppm Au; the interval below, from 460.5 to 468 ft, assayed 3.4 ppm Au.

Figure 3-25. Quartz vein and iron-stained breccia with epithermal textures, such as open-space vugs and fillings of cockscomb quartz, Stibnite pit. Host is altered impure metasiltstone.

Figure 3-26. Mineralized hydrothermal breccia from the West End deposit. Pink carbonate vein with an open-space centerline and white potassium feldspar alteration envelope cuts earlier white potassium feldspar veins and quartz matrix breccia with clasts of altered schist. Very fine-grained sulfides are present in the schist and breccia. The interval assayed 3.44 ppm Au, and the potassium feldspar was dated at 50.8 Ma. Thin section billet 5WR-002 from MGI 12-305 at 462 ft depth. Billet is 4.5 cm x 2.5 cm.
In many samples, the sulfides are concentrated outside of the veins. Visual and assay data confirm the association of the sulfides with the veining and secondary potassium feldspar which is visible in hand specimen. Reaction of metamorphic muscovite to potassium feldspar may have assisted in precipitating the sulfides, or it may simply reflect the temperature increase and chemistry of the fluid associated with the mineralizing hydrothermal fluids. In some places, pyrite appears to have been precipitated by sulfidation of biotite or iron-titanium oxides. Typically the secondary feldspar is white in color (Figure 3-26), as opposed to the pinkish color typical of porphyry-related secondary potassium feldspar. The importance of high-angle fractures and veinlets in mineralizing the metasedimentary sequence is illustrated by the high-grade sample of mineralized siltite shown in Figure 3-29; oxidation has highlighted the thin pyrite-bearing fractures.

Surface samples from the West End pit show similar alteration to that seen in drill core. Thin section and reflected light photos in Figures 3-30, 3-31, and 3-32 illustrate a composite quartz-carbonate vein cutting marble in a sample collected from the upper West End pit. As shown in Figure 3-30, the metamorphic silicates in the impure marble are intensely retrograded to a complex mixture of fine-grained hydrous minerals (probably including talc, clays and serpentine) plus carbonate and quartz. The sulfides are concentrated in the wall rock adjacent to the quartz-carbonate veins (Figure 3-31). Textures of the sulfides are similar to those seen in intrusive rocks, including “skeletal”
Antimony and Tungsten Mineralization and Late-Stage Alteration

White (1940) discusses the Sb mineralization in the district and the views of previous workers. He notes on p. 264 that "A stibnite-silver mineralization followed an earlier high-temperature microcline (?) –albite-pyrite-arsenopyrite-gold mineralization." In terms of the absolute age, White thought that both the Au and Sb were of Tertiary age, with some of the stibnite associated closely with the albite, pyrite, and arsenopyrite, an intermediate to fairly high temperature association (White, 1940, p. 265). Given the abundance of secondary potassium feldspar and geochemical pyrite rimmed by arsenopyrite (Figures 3-31 and 3-32), suggesting a mineralogical and textural link with the geographically separate Yellow Pine deposit.

Figure 3-29. Calcareous siltite showing oxidized, brittle, fractures in high-grade (10.97 ppm Au) interval from 572 to 576 ft. Sample is core from MGI 10-37 at 572 ft. Sulfides (too fine-grained to see in photo) are concentrated in bands and altered micas. Note brown iron oxide halo around fractures, which host carbonate veinlets, as seen in thin section. Core diameter is 2.5 inches.

Figure 3-30. Altered, pyritized marble cut by near vertically-oriented carbonate-quartz vein with euhedral carbonate crystals lining vein wall. Sulfides are associated with the vein (Figure 3-31). Rock is megascopically similar to pit exposure in Figure 3-24. Section 8EB-006 (15VG017, upper West End pit) at x20 magnification in plane-polarized light.

Figure 3-31. Reflected light photo showing sulfides adjacent to quartz-carbonate vein shown in Figure 3-30. Sulfides show "skeletal" habit. Section 8EB-006 (15VG017, upper West End pit). At x20 magnification. Enlargement is in following figure.

Figure 3-32. Reflected light photo of sulfides adjacent to vein shown in Figure 3-30. Sulfides are dominantly in altered wall rock with habit of "skeletal" pyrite (yellow cores) rimmed by arsenopyrite (bright cream-colored). Skeletal habit observed here has elongate and dendritic growth pattern of pyrite. Section 8EB-006 (15VG017, upper West End pit). At x100 magnification. Image is area in mid-right of photo in Figure 3-31.
evidence from the recent exploration geochemistry, it is possible that some of his “albite” is actually hydrothermal potassium feldspar. It was shortly after White’s 1940 strategic minerals investigation, that he discovered the W in the district. Much (but probably not all) of the scheelite seems to predate or be coeval with the stibnite, as discussed below. However, the Sb will be discussed first due to its current economic importance as a critical mineral in the district.

**Antimony Mineralization**

In the Yellow Pine pit, stibnite-bearing veins cut intensely altered granitic rocks, as shown in Figure 3-33. The stibnite veins are localized along discrete cross-cutting vein and breccia structures which cut the earlier more disseminated alteration and Au mineralization as observed in the iron-stained pit exposures and in core (Figure 3-33; White, 1940; Cooper, 1951).

Within the late veins, euhedral and zoned quartz crystals project into a vein interior filled with carbonate, or in some cases, sulfides. This open-space filling texture is indicative of brittle fracturing of the wall in a tensional stress field. Figure 3-34 shows a sample of mineralized granite rock from the Yellow Pine pit, where stibnite (the black-appearing opaque) infills between euhedral carbonate crystals and potassium feldspar adjacent to the vein. Other samples show microcline recrystallized to hydrothermal potassium feldspar next to the vein, or even feldspar in the vein. Many samples show early sulfides predominantly localized in the wall rock rather than the veins. Some very fine grained pyrite and arsenopyrite do occur in veins and are locally enclosed by stibnite. Stibnite is clearly later than the pyrite and arsenopyrite, as determined by cross-cutting relationships like that shown in Figures 3-35 and 3-36. In many of the stibnite veins, the mineral constitutes over 50 percent of the vein.

Most commonly, stibnite fills veins and breccia matrices, whereas pyrite is localized in the wall rock (Figure 3-35). Locally stibnite fills open-space vugs at both the Yellow Pine and Hangar Flats deposits. It is typically associated with carbonate and minor quartz in the late veins and veinlets (Figures 3-34 and 3-36). Potassium feldspar appears to have been stable during at least the initial stages of stibnite deposition in the Yellow Pine and Hangar Flats deposits, but it is doubtful, though unclear, whether potassium feldspar was still precipitating during the latest phases of stibnite mineralization.

Stibnite precipitated later than, and presumably at lower temperature than, the Au-pyrite mineralization. Marsh and others (2016) concluded from fluid inclusion data that decompression and cooling caused stibnite precipitation. The strong correlation of stibnite abundance with brecciation and late fracturing is supportive of stibnite precipitation being linked to areas and times of fracture development, possibly inducing fluid release and decompression. The sample in Figure 3-37 shows crisscrossing, stibnite-filled veinlets of multiple orientations within the main vein structure, as if the area within the main vein was being hydraulically fractured and expanding. Stibnite at Yellow Pine is not always associated with scheelite, but the two are more commonly together at the Hangar Flat deposit. However, there may be a sampling bias due to the scheelite being largely mined out at Yellow Pine.

High-grade stibnite mineralization along the Meadow Creek fault has been mined historically and is present in drill core from the Hangar Flat deposit (Figure 2-3 and drill hole log for MGI-12-193 in Appendix B). Deep intercepts of stibnite-matrix breccia assay 5-20 percent Sb and variable amounts of W, as described in the next section. Stibnite is an integral part of the hydrothermal breccia matrix, but it also postdates the brecciation as late-stage veins and vug fillings.
Tungsten Mineralization

Mitchell (2000, p. 9) summarized the W association recognized by earlier workers:

“Fine-grained tungsten minerals, dominantly scheelite, formed after the gold. The scheelite occurs as disseminated grains in the gold ore, embedded in stibnite, as branching veinlets and stringers 1 to 5 millimeters thick, and as cemented breccia with scheelite in both the groundmass and the breccia clasts (Cooper, 1951). Field relationships, textural evidence, and mineral associations suggest that the deposits formed under low temperature and low pressure (epithermal) conditions (Cookro and Petersen, 1984; Cookro and Silberman, 1986).”

Cooper (1951) interpreted the scheelite, a calcium tungstate mineral, as belonging to the second of three pulses of complex, Tertiary mineralization and recognized the close spatial association with the Meadow Creek fault at both the Yellow Pine mine and the Meadow Creek mine located on the east side of the fault zone further to the south. The first of Cooper’s stages formed pyrite, arsenopyrite, and Au, along with gangue sericite, quartz, and feldspar. The second stage formed scheelite plus carbonates and quartz. The third stage formed stibnite with Ag minerals, pyrite, quartz, and carbonate minerals. Fracturing and brecciation preceded and followed each pulse, with mineralization overlapping spatially and in some cases along the same re-opened fractures. The same general sequence was noted in the petrographic work in this study and is supported by the new geochronologic data discussed later.

Little remains of the scheelite-bearing ore mined from the Yellow Pine pit during the war, but Midas drill holes in the Hangar Flats deposit intersected zones of high-grade W and Sb mineralization. Several of the W-bearing intervals were lamed and sampled for this study. The Smithsonian National Museum of Natural History (NMNH) graciously loaned a Yellow Pine scheelite sample to N. Wintzer of the U.S. Geological Survey. Textures in this sample suggest the subhedral pyrite and scheelite were deposited together or the scheelite was slightly earlier (Figure 3-38). Euhedral scheelite crystals sit in a carbonate matrix in the sample, but stibnite encloses and fills interstices of the scheelite, strongly suggesting that the Sb sulfide was precipitated later.

As seen in the Hangar Flats deposit, scheelite fluoresces pale blue under short-wave UV light. Scheelite is most abundant in the “black breccia” deep in the Hangar Flats holes (Figures 3-39 and 3-40). The black breccia is a stibnite-matrix breccia with whitish to gray-colored, altered and brecciated clasts in a very dark gray to black matrix. In some thin sections, stibnite comprised 50 percent of the matrix. It also appeared that most of the clasts were originally metasedimentary rocks, but some intrusive rock clasts are also present. The scheelite is present largely as replacements in marble clasts. The matrix contains stibnite, lesser amounts of pyrite, carbonate and quartz gangue, and sparse crystals or clasts of scheelite. Scheelite also forms veins that cross cut the breccia, and scheelite crystals precipitated with late carbonate and stibnite in vugs within the breccia.
Figure 3-35. Tiny rounded to euhedral buckshot pyrite (containing Au) in carbonate vein (recessive dark gray on left) is crosscut by stibnite breccia zone (bright pale yellow) oriented vertically. Pale peach colored opaque grain in lower right is pyrite. Section 4LO-022 (location 12VG159D) from Yellow Pine pit at x20 magnification in reflected light. Compare to transmitted light photos in following figure.

(Figure 3-41). No sign of calc-silicate or skarn minerals were seen in the thirteen thin sections examined or the corresponding hand specimens, from the three drill holes (MGI-12-192, MGI-12-193, and MGI-12-203) in the Hangar Flats deposit (Figure 3-6). The lack of calc-silicate minerals indicates scheelite deposition took place at a relatively low temperature and was not associated with the earlier skarn and calc-silicate metamorphism.

Evidence for the late deposition of scheelite in the overall paragenetic sequence and for its association with carbonate alteration was also found at the north Yellow Pine area. Trace, very fine-grained scheelite was observed in the quartz-carbonate centerline of a microbreccia vein in a sample from the north Yellow Pine area at 441.7 ft depth in drill hole MGI 12-314.

Because scheelite both replaces marble in clasts within the stibnite breccia and also occurs as crystals lining vugs filled with stibnite centers (Figure 3-41), the scheelite is interpreted as being predominantly earlier in time than the stibnite, as was concluded by Cooper (1951). However, Figure 3-42 shows rare late scheelite crystals lining the walls of late carbonate veins cutting fault breccia. The sample was taken from the waste dump at the DMEA tunnel which intersected the Meadow Creek fault. No stibnite was seen in the thin section (8EB-002; 15VG012). Thus, some scheelite was precipitated very late in the hydrothermal and deformation history of the area. Whether that W was newly introduced or simply remobilized from the early mineralization is unknown.

Post-Mineral Dikes and Late-Stage Alteration

The Tertiary dikes provide the clearest example of very late alteration in the intrusive rocks. Rhyolite to latite or dacite porphyries and basalt dikes intrude the metasedimentary and granitic intrusive rocks. Where porphyritic dikes cross cut the granites, the dikes are much less altered, and show sharp, chilled margins against the equigranular, granitic rocks. In all cases, the dikes are significantly less fractured and altered than the surrounding host rocks, and mineralization drops to near background levels (<50 ppb Au) in the dikes, as is noted by Cooper (1951) and confirmed by hundreds of analyses by Midas Gold (Christopher Dail, written commun., 2017). Macroscopically, the dikes show weak chloritic or clay alteration, but in thin section a moderate to strong low-temperature alteration assemblage is apparent. The mafic dike in the southwest part of the Yellow Pine pit is not obviously veined but a thin section reveals pervasive carbonate alteration of the mafic minerals along with carbonate in the matrix and in vesicle fillings (Figure 3-43). A spherulite of quartz in one of the vesicles is interpreted as originally having precipitated as cristobalite in a late gas cavity in the near-surface dike (Figure 3-43).

A latite porphyry dike from 646 ft in MGI 12-305 (Figures 3-8 and 3-10) in the West End area also displays intense clay-carbonate alteration of the feldspar and mafic phenocrysts and contained only 25 ppb Au. Samples collected from the DMEA tunnel dump show multiple generations of calcite veining and breccias, and are likely from where the tunnel intersected the Meadow Creek fault. Carbonate replacement-style alteration and veining must have continued along faults and permeable zones after sulfide deposition, faulting, and brecciation.

Mercury and Late Silicification

Mercury, in the form of cinnabar, was one of the first commodities mined in the Stibnite district (Mitchell, 2000). Figure 1-2 shows the peripheral position of the Fern and Hermes Hg mines relative to the main Yellow Pine and Meadow Creek mine Au-Sb deposits. The Hg mines are outside of the current Midas proposed mine area, but they are part of the complex geologic setting and hydrothermal system. Elevation at the Fern mine
Figure 3-36. Stibnite-quartz vein (black and near vertical in photo) cuts carbonate-pyrite vein (diagonal) in quartz monzonite. Section 4LO-022 at x20 magnification in plane-polarized (A) and cross-polarized (B) light. Reflected light view in previous figure. Location 12VG159D in Yellow Pine pit.

Figure 3-37. Stibnite veins filling complex fracture zone in granitic rock on northwest wall of Yellow Pine pit. Stibnite veins fill several secondary fractures, but the stibnite is largely confined within a 1-cm wide, first order fracture system oriented vertically in photo. Thin section billet 4LO-022 from location 12VG159D is 2.5 cm in height.

Workings at the Fern mine are hosted in the Fern marble, which shows significant sanding of the dolomitic marble near the workings (Figure 3-44). Trenches expose fine-grained jasperoidal veins and small irregular silicified masses that cut the altered dolomite. Cinnabar is present in the silicified zones.

At the Hermes mine, located near the town site of Cinnabar, a small collapsed stope revealed unoxidized Hg mineralization along an apparently northeast-striking, southeast-dipping fault breccia (Figure 3-45). The cinnabar grains are in a silicified breccia, and sparse, older pyrite is also present (Figure 3-46).

Another zone of late silicification is the north-northeast-trending silica rib on the east side of the Yellow Pine pit, roughly located along the trace of the Meadow Creek fault (Figure 3-47). In appearance, it differs from other quartz veins in being more intensely silicified, as well as lacking visible sulfides. The silicified rib is apparently unmineralized and was left as waste by the historic mining operations; it is not high in Hg or Au (Christopher Dail, oral commun., 2017). It may represent late, barren, silica-rich fluids moving along the Meadow Creek fault in the vicinity of the modern Yellow Pine deposit. The relationship of the silicified rib to the late carbonate veining on the Meadow Creek fault further south or to the silicified cinnabar-bearing zones is unknown.
Figure 3-38. Scheelite crystals with adjacent subhedral pyrite with corroded core; both minerals are surrounded by interstitial stibnite and minor carbonate. Large crystal in lower half is scheelite. In cross-polarized (A) light and reflected light (B) at x100 magnification. Section of Yellow Pine sample NMNH 107667 from N. Wintzer.

Figure 3-39. Black breccia and crackle breccia from the Hangar Flats area with stibnite and scheelite. Carbonate is also in the matrix. Scheelite is blue fluorescing mineral. Core is 2 in diameter and from MGI-12-193, 936 ft under normal light (A) and a closer view under a short-wave ultraviolet (SW UV) lamp (B).
Figure 3-40. High-grade scheelite mineralization in breccia and contact (arrow) with possible alaskite from Hangar Flats area. Core is 2.5 in diameter. From MGI-12-203 at 1055 ft under normal light (A) and SW UV lamp (B).

Figure 3-41. Vug-filling stibnite in W-mineralized black breccia from Hangar Flats area. Stibnite (black) is present in the matrix, vug, and vein cutting the breccia. Scheelite grains show orange-colored birefringence and carbonate rhombs are white of the highest order. Section 6GE-005 (12-203-1054) at x100 magnification in plane-polarized (A) and cross-polarized (B) light.
Figure 3-42. Very late scheelite crystals in calcite-veined fault breccia. Tiny orange birefringent crystals on vein walls are scheelite. Section 8EB-002 (15VG012) from DMEA adit dump at x20 in plane-polarized (A) and cross-polarized (B) light.

Figure 3-43. Mafic dike from Yellow Pine pit showing vesicle filled with carbonate and radiating cristobalite, probably inverted to other form of quartz. Clay alteration is likely present in groundmass and lining the vesicle; no x-ray work was done to positively identify the clay. Section 8EB-017 from 15VG036d at x20 magnification in cross-polarized light.
Figure 3-44. Outcrop of silicified veins cutting Fern marble at Fern mine trench. The trench wall is approximately 7 m (21 ft) in height.

Figure 3-45. Late cinnabar (red color) is in silicified gray fault breccia (sample15VG028m) from Hermes mine. Field of view is approximately 10 cm across.
Figure 3-46. Cinnabar (red in A) with earlier pyrite in silicified breccia from Hermes mine. Pyrite is opaque. In reflected light (B), the pyrite is a bright cream color and surrounded by cinnabar. Section 8EB-010 (12VG028) at x20 magnification in cross-polarized light (A) using condenser lens to show red color.

Figure 3-47. Outcrop of the silicified rib of Meadow Creek fault breccia from the east side of Yellow Pine pit (15VG015). Horizontal late slickensides are visible on the north-south-striking fault plane surface to the lower left of the hammer.
DISCUSSION OF HYDROTHERMAL ALTERATION AND MINERALIZATION

Observations made in this study confirm and amplify conclusions of earlier workers, some of whom could only access one deposit in the Stibnite district. As noted by earlier workers:

- The stratigraphic succession in the Stibnite roof pendant consists of intensely folded Neoproterozoic to Paleozoic metasedimentary rocks with mixed siliciclastic units and dolomitic to calcareous carbonate lithologies (Smitherman, 1985; Stewart and others, 2016).

- Multiple distinct intrusive phases of the Idaho batholith are exposed within the district at the surface and in the subsurface. All of the granitic rocks host mineralization and alteration to some degree. Characteristics, composition, and Cretaceous ages of the granitic rocks at Stibnite are very similar to other rocks of the Atlanta lobe of the batholith.

- Tertiary dikes in the district are typical of the suite of Eocene Challis magmatic event dikes and show significantly less fracturing and shearing than adjacent batholith rocks. Dikes are locally altered with carbonate and clays, but they do not host ore and typically contain only background to trace values of ore metals. Sparse mineralization has been reported locally in rhyolite and fractured mafic dikes (Figure 2-18; Cooper, 1951).

- As well-documented by many earlier workers, ore is structurally localized along the north-south-trending Meadow Creek fault zone and northeast-trending structures that intersect or splay off the Meadow Creek fault. Three main resource areas of Au mineralization are outlined by current Midas exploration work and coincide with historic mining at the Yellow Pine pit, the West End area, and Stibnite pit. During the potassic alteration, untwinned potassium feldspar replaced and veined plagioclase or igneous microcline or orthoclase, and overprinted fine-grained sericite in plagioclase. Secondary potassium feldspar typically coexists with carbonate and quartz, as well as with pyrite and arsenopyrite.

- An earlier and peripheral alteration is the replacement of igneous or metamorphic biotite by muscovite plus carbonate and rutile or titanite, followed by sulfidation and pyrite precipitation. This pyrite, which partly replaced and then nucleated upon the altered mafic minerals, is equivalent to the “disseminated” sulfide noted by earlier workers. The disseminated pyrite was locally surrounded by arsenopyrite. In some samples, secondary potassium feldspar appears to be stable with the muscovite, but elsewhere the feldspar veined and overprinted the micas. Illite, a very fine grained sheet silicate mineral related to micas and clays and noted in spectroscopic work by Midas, could be associated with some of the coarser grained muscovitic mica visible under the microscope, but the exact location of the illite is unknown.

- The muscovite replacing biotite alteration at Stibnite is very different in appearance and association from the phyllic alteration (quartz-sericite-pyrite) typical of a porphyry system.
• Carbonates are ubiquitous and abundant in veins and within pseudomorphic replacements of primary minerals. Gold-bearing pyrite is present within carbonate veins cutting altered and highly fractured granitic rocks.

• Gold-related alteration in the metasedimentary rocks is the same as in the granitic rocks: carbonate-quartz-potassium feldspar plus sulfides.

• Sulfide precipitation shows a pattern of pyrite rimmed by arsenopyrite crystals. In places the sulfides form skeletal-shaped, branching growths suggestive of either rapid cooling or replacement and growth along cleavage traces in a prior silicate mineral. In general, sulfides are most abundant in the wall rocks adjacent to quartz-carbonate veins, and many of the veins show multiple generations of fracturing and vein filling. Some veins, particularly in the Stibnite pit near the West End area, show classic epithermal textures such as cockscomb quartz, brecciation, and open-space filling.

• As noted by earlier workers, Au-bearing disseminated mineralization and potassic alteration at the Yellow Pine deposit is cross cut by fracture-controlled stibnite mineralization. At Hangar Flats, stibnite forms a breccia matrix and is present as late veins and vug fillings. The West End deposit is distinct and contains very little Sb mineralization.

• Known W mineralization, as scheelite, is confined to the Yellow Pine and Hangar Flats deposits and is closely associated with, but at least locally predates, stibnite precipitation. Scheelite grains, or grain fragments, in late quartz-carbonate veins from the Yellow Pine deposit periphery, and the descriptions of Cooper (1951) also indicate that the W and Sb zones are distinct.
CHAPTER 4

GEOCHRONOLOGY
INTRODUCTION

A major goal of this project was to determine the age of the Au-Sb-W mineralization at Stibnite and to place the mineralization into a district and regional chronologic framework. The study also provided the opportunity to acquire new age constraints on the host rocks. Stratigraphic maximum depositional ages based on detrital zircon analysis and three intrusive ages, all based on U-Pb LA-ICPMS methods, were reported by Stewart and others (2016) on the Stibnite 7.5' quadrangle map. The detrital zircon data confirmed that the upper stratigraphic units exposed at Stibnite are Ordovician in age (Table 2-1, Figures 2-3 and 2-4; Stewart and others, 2016). The intrusive age reported by Stewart and others (2016) of 91.2 ± 2.2 Ma on the granodiorite (Kgd, quartz monzonite sample 10RL889), was revised slightly to 89.9 ± 1.7 Ma by Gaschnig and others (2017). An age of 87.9 ± 4.9 Ma was reported by Stewart and others for the leucocratic granite (Klg, alaskite sample MGI-10-20). The 84.9 ± 2.0 Ma on the Stibnite stock (Kgds, sample 10RL887) was revised slightly to 84.4 ± 2.1 Ma by Gaschnig and others (2017). The Kgd and Klg samples were from drill core in the Hangar Flats area. Those ages were obtained from zircon rims, because many zircons exhibit large Mesoproterozoic to Neoproterozoic cores (Stewart and others, 2016; Gaschnig and others, 2017).

Gillerman and others (2014) reported initial results for new U-Pb zircon ages on intrusive rocks of the Idaho batholith and 40Ar/39Ar ages for potassium feldspar associated with West End mineralization. In total, seven new U-Pb zircon ages on intrusive rocks, three cooling ages based on U-Pb methods on rutile and titanite minerals, and eight new 40Ar/39Ar ages on alteration feldspars and mica were obtained for this study (Tables 4-1 and 4-2). In addition, ages from LA-ICPMS U-Pb measurements of scheelite were determined in conjunction with Wintzer and others (2016, 2017). Figures 3-6 through 3-10 plus Appendix A locate the drill holes sampled relative to the district geology and ore zones.

Samples were collected by V.S. Gillerman from drill core provided by Midas Gold or from surface exposures. For U-Pb dating, a 5 to 15-pound bag of core was collected over a 1-to-30 ft intercept of the least-altered zone of the desired rock type (Table 4-1). Heavy minerals were separated using conventional density and magnetic methods in the Isotope Geology Laboratory at Boise State University. For 40Ar/39Ar analysis on potassic alteration minerals, a selected piece of core or hand sample was sawed to separate and isolate the alteration mineral and sent to the Geochronology Laboratory at the University of Alaska, Fairbanks, for mineral separation and dating. Appendix E contains the reports including analytical methodology from the University of Alaska 40Ar/39Ar work.

GEOCHRONOLOGY OF INTRUSIVE ROCKS AND COOLING AGES

Crystallization and cooling ages for intrusive rocks were determined using 40Ar/39Ar analysis on biotite and potassium feldspar, and via the U-Pb decay scheme in zircon, titanite, and rutile using both the isotope dilution thermal ionization mass spectrometry (ID-TIMS) and laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) methods of isotopic analysis. 40Ar/39Ar methodologies are those summarized in Benowitz and others (2011). Analytical methods for U-Pb geochronology are those summarized in Rivera and others (2013) and Macdonald and others (2018). LA-ICPMS is a high throughputs, in situ technique that can rapidly acquire isotope ratios on a spatially restricted domain of an individual crystal (25 to 40-micron spot analyses), but it is subject to higher analytical uncertainties due to the small sampling volume and nature of mass spectrometry. As such it is well-suited to detrital zircon analysis and rapid screening of igneous samples for the presence of inherited cores versus magmatic overgrowth. When measured with a quadrupole mass spectrometer, the ablation materials also yield their trace element compositions in addition to U-Pb isotope ratios and dates, which can be useful for crystallization thermometry and petrogenetic interpretation (Ferry and Watson, 2007; Hayden and others, 2008). ID-TIMS uses an isotopic spike mixed with a dissolved crystal to quantify isotope ratios and can achieve a factor of 20-100 times greater temporal resolution, although at the expense of spatial resolution. By applying chemical abrasion (CA) in concentrated hydrofluoric acid at high temperatures, the CA-ID-TIMS method can eliminate Pb-loss in zircon crystals that results from radiation damage (Mattinson, 2005), and thus mitigate against anomalously young apparent
Table 4-1. New geochronology of intrusive rocks at Stibnite. Deposit or pit name closest to drill hole is listed.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Unit</th>
<th>Mineral/Lithology</th>
<th>Location</th>
<th>Lab</th>
<th>Method</th>
<th>Results</th>
</tr>
</thead>
<tbody>
<tr>
<td>13-383-285</td>
<td>Dacite porphyry (Tdp)</td>
<td>Potassium feldspar; fresh biotite-feldspar porphyry</td>
<td>Yellow Pine pit; MGI-13-383, 285 ft</td>
<td>UAF</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Plateau age: 45.9 ± 0.3 Ma; very flat age spectrum</td>
</tr>
<tr>
<td>13-383-285</td>
<td>Dacite porphyry (Tdp)</td>
<td>Biotite</td>
<td>Yellow Pine pit; MGI-13-383, 285 ft</td>
<td>UAF</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>46.8 ± 0.3 Ma; weighted average</td>
</tr>
<tr>
<td>13-383-282</td>
<td>Dacite porphyry (Tdp)</td>
<td>Zircon; fresh feldspar-biotite porphyry</td>
<td>Yellow Pine pit; MGI-13-383, 282 ft</td>
<td>BSU</td>
<td>U-Pb ID-TIMS</td>
<td>47.86 ± 0.07 Ma; weighted mean of 4 crystals</td>
</tr>
<tr>
<td>MG-306-550</td>
<td>Alaskite (AK/Klg)</td>
<td>Zircon; alaskite of potassium feldspar, quartz, and trace mica</td>
<td>Yellow Pine pit; MGI-12-306, 543-564 ft</td>
<td>BSU</td>
<td>U-Pb ID-TIMS</td>
<td>83.61 ± 0.08 Ma; weighted mean of 4 crystals; common inherited cores</td>
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<td>MG-243-175</td>
<td>Alaskite (AK/Klg)</td>
<td>Zircon; alaskite cutting Kgd and Kg</td>
<td>Yellow Pine pit; MGI-12-243, 175 ft</td>
<td>BSU</td>
<td>U-Pb LA-ICPMS</td>
<td>84.8 ± 1.3 Ma; weighted mean of 17 youngest crystals; common inherited cores</td>
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<td>MG-306-940</td>
<td>Granite (GR/Kg) (unmineralized)</td>
<td>Zircon rims; fine-grained granite; plagioclase-orthoclase-quartz-mica rock</td>
<td>Yellow Pine pit; MGI-12-306, 929-956 ft</td>
<td>BSU</td>
<td>U-Pb LA-ICPMS</td>
<td>86.0 ± 5.4 Ma; weighted mean of 4 zircon rims; dominantly inherited cores with thin rims</td>
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<tr>
<td>12-305-422</td>
<td>Granite (GR/Kg)</td>
<td>Zircon; medium-grained two mica granite, no veins, altered</td>
<td>West End pit; MGI-12-305, 422 ft</td>
<td>BSU</td>
<td>U-Pb ID-TIMS</td>
<td>85.51 ± 0.06 Ma on youngest zircon</td>
</tr>
<tr>
<td>10-37/50</td>
<td>Stibnite stock (Kgds)</td>
<td>Zircon; fine-grained granite of quartz-microcline-plagioclase and mica</td>
<td>Stibnite pit; MGI-10-37, 40-55 ft</td>
<td>BSU</td>
<td>U-Pb ID-TIMS</td>
<td>85.71 ± 0.10 Ma weighted mean of 4 crystals; common inherited cores</td>
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<td>14VG8-4B</td>
<td>Biotite granodiorite (QM/Kgd)</td>
<td>Zircon; fresh biotite-rich granodiorite</td>
<td>Roadcut N of Yellow Pine pit; 44°55.97' N 115°20.08'W</td>
<td>BSU</td>
<td>U-Pb ID-TIMS</td>
<td>94.21 ± 0.22 Ma on four zircon crystals</td>
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Table 4-2. Geochronology results for metamorphic mineral cooling ages in metasedimentary and igneous rocks.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Unit</th>
<th>Mineral/Lithology</th>
<th>Location</th>
<th>Lab</th>
<th>Method</th>
<th>Results</th>
</tr>
</thead>
<tbody>
<tr>
<td>4LO-008</td>
<td>Calc-silicate schist</td>
<td>Metamorphic titanite in layered calc-silicate</td>
<td>Stibnite pit; surface sample 12VG147C-2</td>
<td>BSU</td>
<td>U-Pb ID-TIMS</td>
<td>85.31 ± 0.05 Ma on titanite grains</td>
</tr>
<tr>
<td>4LO-013</td>
<td>Quartzfeldspathic schist</td>
<td>Rutile in schist near Stibnite stock</td>
<td>Stibnite pit; surface sample 12VG149-1</td>
<td>BSU</td>
<td>U-Pb LA-ICPMS; in-situ</td>
<td>78.5 ± 4.4 Ma; Zr-inrutile temp ~650°C</td>
</tr>
<tr>
<td>5XG-006</td>
<td>Granite (Kg) with strong potassic alteration.</td>
<td>Rutile in granite; overgrown with pyrite</td>
<td>MGI 12-305-422.1 in West End area</td>
<td>BSU</td>
<td>U-Pb LA-ICPMS; in-situ</td>
<td>79.5 ± 2.6 Ma; Zr-inrutile temp 700-740°C</td>
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</tbody>
</table>
dates due to open-system behavior in zircon. Both LA-ICPMS and CA-ID-TIMS methods were pursued on crystals characterized by cathodoluminescence (CL) imaging, in order to avoid as much as possible inherited cores that could bias apparent dates too old compared to the true magmatic crystallization age. Analytical data for U-Pb determinations are compiled in Appendix D. The interpreted zircon crystallization ages are listed in Table 4-1, and the three metamorphic or cooling ages are listed in Table 4-2. Zircon and titanite results are illustrated in the U-Pb concordia diagram of Figure 4-1.

Results for Intrusive Rocks

Crystallization ages for the Cretaceous intrusive rocks at Stibnite range from 94.2 to 83.6 Ma, thus corresponding to ages of the early metaluminous suite and Atlanta peraluminous suite of the Idaho batholith, as distinguished by Gaschnig and others (2011). All of the samples had common inherited cores to the zircon crystals, with overgrowths of a magmatic phase of zircon. Details of the analytical results for each sample follow:

**Biotite granodiorite (Kgd/QM; sample 14VG8-4B; YP)**

This unaltered sample correlates to the altered quartz monzonite (QM) which hosts the Yellow Pine deposit. CL imaging of equant to dominantly elongate prismatic zircon crystals revealed higher U (non-luminescent) rims and overgrowths on commonly brighter (low U) resorbed cores. Six higher aspect ratio grains were chosen for CA-ID-TIMS analysis and yielded a range of 206Pb/238U dates from 94.48 to 93.77 Ma. Given the evidence for inherited cores the oldest date was ignored in the weighted mean age; similarly, the pervasive

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**Figure 4-1.** 206Pb/238U vs. 207Pb/235U concordia diagram of intrusive rock and metamorphic ages at Stibnite. Error ellipses for individual zircon or titanite crystals are plotted at the 2-sigma confidence level; the concordia curve is labeled in millions of years. Colors represent analyses from same sample; crystallization ages are interpreted from the reproducible 206Pb/238U dates of clusters of analyses (filled ellipses). Pale grey polygon highlights zircon crystals (open ellipses) from the Stibnite stock and alaskite intrusion with older ages (ca. 87, 89, and 90 Ma) that likely represent inherited cores from earlier intrusions.
intrusion of younger magmas may produce cryptic overgrowth on pre-existing crystals leading to the youngest apparent date. The remaining four grains also span nearly 250 ka, which may represent the intrusion and solidification interval of the pluton; the weighted mean of these four analyses, including apparent geological scatter to encompass the solidification time, is 94.21 ± 0.22 Ma.

**Stibnite stock (Kgds; sample MG 10-37/50; WE)**

CL imaging of the sparse population of equant to slightly elongate zircon crystals from this granodiorite revealed a diverse population of moderately to brightly luminescent, weakly zoned crystals with non-luminescent overgrowths. Only 18 of 38 LA-ICPMS spot analyses produced Cretaceous dates, with the rest recording Precambrian inherited core ages of ca. 2.6, 1.7, 1.4, and 1.1 Ga. Ten zircon grains selected for CA-ID-TIMS analysis yielded a range of 40Ar/39Ar dates of 1370 to 85.5 Ma. The youngest four grains are concordant and equivalent at 85.71 ± 0.10 Ma (mean squared weighted deviation, MSWD = 2.07), which we interpret as dating the intrusion and crystallization age of the Stibnite stock.

**Granite (Kg or “GR”; sample 12-305-422; WE)**

CL imaging of a moderately abundant population of equant to slightly elongate zircon crystals from this granite revealed a consistent population of moderately to brightly luminescent, weakly zoned crystals with non-luminescent overgrowths. A minority of crystals were more strongly oscillatory zoned. LA-ICPMS analysis produced Precambrian inherited core ages of ca. 2.7, 2.3, 1.7, 1.4, 1.1, 0.7, and 0.5 Ga, in addition to >60 spots with Cretaceous dates. Four zircon grains selected for CA-ID-TIMS analysis yielded a range of 83.61 ± 0.08 Ma (MSWD = 3.56), which is interpreted as dating the intrusion of the alaskite.

**Alaskite (Klg or “AK”; sample MG 243-175; YP)**

CL-imaging of a moderately abundant population of equant to slightly elongate zircon crystals from this alaskite revealed a consistent population of moderately to brightly luminescent, weakly zoned crystals with non-luminescent overgrowths. LA-ICPMS analysis produced Precambrian inherited core ages of ca. 2.5, 1.7, 1.4, 1.1, and 0.6 Ga, with >30 Cretaceous dates on rims. The 17 youngest spot analyses produce a weighted mean 206Pb/238U date of 84.8 ± 1.3 Ma. Again, we correlate the intrusion to others dated at ca. 83.6 Ma.

**Alaskite (Klg or “AK”, sample MG 306-550; YP)**

CL-imaging of the sparse zircon crystals from this alaskite sample revealed a heterogeneous population of variably luminescent zircon crystals with common signs of zoning truncations and inherited cores. Laser ablation screening of these crystals confirmed the diversity of the crystal load, with only 17 of 44 spot analyses yielding Cretaceous (ca. 90 Ma) dates. Inherited cores yielded dates of ca. 1.7, 1.3, 1.0 and 0.7 Ga. Twelve grains were selected for microsampling and CA-TIMS analysis based on the LA-ICPMS spot ages; crystal rims were fragmented from these grains and subjected to chemical abrasion. Eight crystal fragments yielded older, mixed-age results variably biased by inheritance. The remaining four of the twelve analyses are concordant and equivalent, with a weighted mean 206Pb/238U date of 83.61 ± 0.08 Ma (MSWD = 3.56), which is interpreted as dating the intrusion of the alaskite.

**Tertiary porphyry (Tdp, sample 13-383-282; YP)**

CL-imaging revealed a consistent population of moderately to brightly luminescent zircon crystals. Most grains contain one or more of the following zones: a luminescent, homogenous core, a surrounding oscillatory-zoned mantle, and a luminescent rim overgrowth. Some crystals capture the full tripartite zoning. LA-ICPMS analysis confirmed a predominant Eocene population of crystals with sparse Cretaceous cores. Four zircon grains selected for CA-TIMS analysis based on CL pattern yielded concordant analyses with a small range in 206Pb/238U dates, and together give a weighted mean 206Pb/238U date of 47.86 ± 0.07 Ma (MSWD = 8.2), which is interpreted as dating the intrusion of this porphyry dikes.

**Tertiary porphyry (Tdp, sample 13-383-285; YP)**

Biotite and potassium feldspar separates from a sample of the same porphyry dike discussed above were also analyzed by 40Ar/39Ar geochronology at the University
of Alaska, Fairbanks (Table 4-1). The biotite separate returned a weighted average age of 46.8 ± 0.3 Ma, and the potassium feldspar separate had a very flat spectrum with a plateau age of 45.9 ± 0.3 Ma (Figure 4-2). The ages of the biotite and feldspar are appropriately lower than the U-Pb intrusive age (47.86 Ma) because they document cooling of the magmatic minerals through their ‘closure temperature’ (Dodson, 1973). For biotite, that closure temperature is approximately 300°C (Harrison and others, 1985). For potassium feldspar, it is variable over a range from 150°C to approximately 300°C, depending on domains of crystallinity, grain size and mineral structure (Henry and others, 1997; Love and others, 1998). Thus, the dike cooled to below about 150°C within approximately a million years of its emplacement in the Eocene.

Cooling Ages

Metamorphic mineral (titanite and rutile) ages were obtained from a variety of metasedimentary and altered granitoid lithologies and are reported in Table 4-2. Titanite and rutile are common metamorphic minerals in calc-silicate and low-Fe, quartzo-feldspathic clastic metasediments, respectively. Rutile can also crystallize in low-Fe granitoids, such as the West End granite in sample 5XG-006 discussed below. Pb diffusion in titanite is relatively slow and can be quantified through closure temperatures (Dodson, 1973) in the range of 700 to 750°C; U-Pb ages for titanite are thus interpreted as dating metamorphic mineral growth during magmatic intrusion, heating, and hydrothermal circulation. Zirconium substitution into rutile has been calibrated as a thermometer (Ferry and Watson, 2007), thus giving the temperature of rutile growth due to the slow diffusion of Zr in rutile. By comparison, Pb diffusion in rutile is relatively fast, such that nominal closure temperatures of 400°C demand that rutile U-Pb dates likely represent cooling ages. The concordia diagram of rutile cooling age determinations is in Figure 4-3.

Titanite in calc-silicate schist (sample 4LO-008; WE)

Abundant titanite in a calc-silicate schist sampled at the surface in the Stibnite pit exhibits the characteristic lentil-shaped metamorphic habit, and most crystals contain abundant mineral inclusions. Five grains of variable size (t1 to t5 arranged smallest to largest) were selected for dissolution and ID-TIMS analysis.

![Figure 4-2](image_url)

Figure 4-2. Age spectra from magmatic Tertiary porphyry sample 13-383-285 (BI is biotite in blue and FS is potassium feldspar in gray). Sample of alteration feldspar 12-305-462 is also shown in red.
The four coeval titanite crystals also form isochronous arrays in total Pb isotope (radiogenic + initial Pb) space; they yielded estimates of the Pb isotope composition of the impure carbonate at the time of metamorphism: $^{206}\text{Pb}/^{204}\text{Pb} = 21.03 \pm 0.11$ and $^{207}\text{Pb}/^{204}\text{Pb} = 16.047 \pm 0.040$, which may be compared to the common Pb isotope data reported below.

**Rutile in quartzofeldspathic schist (sample 4LO-013; WE)**

A psammitic metasedimentary quartzofeldspathic schist was sampled in the Stibnite pit within a few meters of the calc-silicate schist from which titanite was analyzed. Rutile crystals were analyzed by LA-ICPMS in situ within thin sections. Rutile crystals are moderately U-rich and Th-poor, with variable amounts of initial Pb. Zirconium concentrations were converted to temperature using the calibration of Ferry and Watson (2007) with activities for titania and silica of unity. Resulting temperatures are tightly clustered around 650°C, consistent with textural evidence for growth during amphibolite-facies metamorphism of the schist. Using the common Pb composition of the nearby Stibnite stock, rutile spot analyses were corrected using the $^{207}\text{Pb}/^{206}\text{Pb}$-$^{206}\text{Pb}/^{238}\text{U}$ concordance method, which is equivalent to the common Pb anchored, semi-total Pb isochron method of Ludwig (1998). Seven rutile spots regressed on a Tera-Wasserburg concordia diagram yield a lower intercept age of 78.5 ± 4.4 Ma, which is interpreted as a cooling age following regional magmatism and metamorphism (Figure 4-3).

**Rutile in granite with potassic alteration (sample 5XG-006; WE)**

Figure 4-4 shows a magmatic rutile grain in 5XG-006, a section from granite in DDH 12-305 in the West End area. The coarse, euhedral rutile is interpreted as a primary igneous phase. Twenty-five spot analyses on these rutile crystals yielded relatively high-U, relatively radiogenic Pb/U ratios. Zirconium concentrations indicate temperatures of 740 to 700°C, consistent with a eutectic granitoid composition. Using the common Pb composition of the nearby Stibnite stock, rutile spot analyses were regressed using the $^{207}\text{Pb}/^{206}\text{Pb}$-$^{206}\text{Pb}/^{238}\text{U}$ concordance method, which is equivalent to the common Pb anchored, semi-total Pb isochron method of Ludwig (1998). The resulting lower intercept yields a date of 79.5 ± 2.6 Ma, which is interpreted as a cooling age following regional magmatism and metamorphism (Figure 4-3). The corroded rim must have developed after the 79 Ma cooling age analyzed for the magmatic rutile. Most likely the corrosion records uplift and hydrothermal activity before and during the start of pyrite deposition, which was likely in the Eocene, as described later.

The largest crystal yielded a resolvable older $^{206}\text{Pb}/^{238}\text{U}$ date of 88.87 Ma, which is interpreted to contain either a small nucleus of an earlier phase of metamorphic titanite or perhaps zircon growth. The remaining four analyses are concordant and equivalent, with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ date of 85.31 ± 0.05 Ma (MSWD = 0.31), which is interpreted as dating the solid-state recrystallization of the impure carbonate protolith. This age is just slightly younger than the intrusion and crystallization of the Stibnite stock (Table 4-1), thus we correlate the metamorphic event with hydrothermal circulation associated with this magmatic intrusion.
DISCUSSION OF INTRUSIVE AND COOLING AGES

• Based on the metamorphic and cooling ages recorded by titanite and rutile, the 94 to 83 Ma deep-seated intrusions of the Idaho batholith at Stibnite started cooling shortly after emplacement, reaching temperatures of <400°C (the closure temperature of rutile) by ca. 78 Ma. There is no record of magmatism in the region from the ca. 83 Ma alaskite intrusions until the subsequent appearance of the Challis magmatic rocks at around 51 Ma. By this Eocene time, the region had apparently experienced significant uplift, based on the abundant porphyry textures suggestive of more rapid cooling in a hypabyssal (subvolcanic) setting.

• The magmatic gap at Stibnite from about 80 to 50 million years is common in southern and central Idaho. But in northern Idaho, that time interval (80 to 50 million years) is partly occupied by rocks of the Bitterroot peraluminous suite, for which Gaschnig and others (2011) determined ages from 66 to 53 Ma. They also note that a few younger ages for the Atlanta lobe peraluminous rocks exist.

• One latite porphyry dike from drill hole MGI 13-383 (285 ft) was analysed by three techniques. The U-Pb age of 47.86 ± 0.07 Ma constrains the crystallization age. The $^{40}$Ar/$^{39}$Ar ages of 46.8 Ma and 45.9 Ma on biotite and potassium feldspar, respectively, from the same rock represent cooling ages. The relatively short time (1-2 million years) between crystallization and cooling below biotite and feldspar closure temperature (ca. 150°C) confirm the shallow emplacement of dikes during the Eocene magmatic event. However, given the many distinct dikes observed and mapped regionally, it is likely that volcanism and dike emplacement spanned several million years.

• A study by the U.S. Geological Survey to sample and date volcanic rocks and dikes in the region is ongoing. Preliminary results of five new ages on dikes within the district and region were graciously provided by Steve Box and Dave John of the USGS (written commun., 2017). The five ages, which include two U-Pb and three $^{40}$Ar/$^{39}$Ar dates on biotite, range from 49.6 to 47.0 Ma, confirming the Eocene age for dike intrusion at Stibnite.

• Ages of the multiple inherited zircon cores in the Cretaceous and Eocene intrusive rocks provide a

Figure 4-4. Magmatic rutile grains (medium gray with laser holes) surrounded by corroded rim of rutile with carbonate and other minerals and late overgrowth of pyritohedral pyrite (white color). Backscatter image of 5XG-006 (12-305-422.1)
glimpse into the age and availability of crust below
the Stibnite district at the time of magma generation.
For the Cretaceous granitic rocks, the core ages are
older batholith phases plus Precambrian inherited
cores of ages ca. 2.5, 1.7, 1.4, 1.1, and 0.6 Ga. This
suggests a mixed, deep-crustal architecture at 94 to
83 Ma with Archean components, Mesoproterozoic
components similar to the well-exposed Belt-
age rocks, plus some “Grenville-age” crust, and
Neoproterozoic rocks. In contrast, Eocene zircons
record a more juvenile magmatic path with much
less inheritance, as is also evident in the lead
isotope signatures.

GEOCHRONOLOGY
OF ALTERATION AND
MINERALIZATION

Introduction and Previous Geochronology

Uncertainty over the timing of mineralization in
the Stibnite district began with early workers who
lacked the advantage of modern radioisotopic dating
tools. However, all the early geologists and miners
recognized that there are two suites of intrusive
rocks in the Stibnite district, the Cretaceous batholith
granitoids and the Tertiary porphyritic dikes, and that
regional metamorphic mineral assemblages predate the
intrusions and Au mineralization.

An important but difficult question is whether there
was significant hydrothermal alteration and Au or other
mineralization associated with the Cretaceous intrusions
at Stibnite. Gammons (1988) reported a suite of Late
Cretaceous ⁴⁰Ar/³⁹Ar ages (approximately 78 to 70 Ma)
on gangue muscovite in the Edwardsburg area about 15
miles to the north of Stibnite (Figure 2-2). Gammons
also reported a 77.9 Ma age on “coarse muscovite (after
biotite) in sheared and silicified wallrock” at the “Yellow
Pine mine”, and a 57 Ma age on sericitized wall rock
next to “scheelite-rich quartz-carbonate veins” at Quartz
Creek, which is located closer to the town of Yellow
Pine. Lewis (1984) did an extensive study of the Stibnite
District and obtained K-Ar ages on purified adularia
from pervasive potassium metasomatism associated
with Au mineralization at the Yellow Pine mine. Two
samples produced three K-Ar ages of 56.2 ±1.3 Ma,
58.9± 1.4 Ma, and 57.1 ± 1.3 Ma which averaged to
an approximate mineralization age of 57 Ma. He noted
that the age of the ore-related potassium feldspar was
significantly older than ages of the Challis volcanic
rocks and nearby Thunder Mountain caldera. Table 4-3
summarizes the published and newly determined ages
at Stibnite.

Recently, Konyshev and Muntean (2016) reported
an early assemblage of milky white quartz veins with
minor molybdenite cutting metasedimentary rocks
from the West End pit. They obtained a Re-Os age of
86.3 Ma ± 0.4 Ma on the molybdenite, indicating a
Cretaceous mineralization event. The molybdenite and
possibly other sparse ore minerals, such as chalcopyrite,
deposit.
feldspar have preferred ages of 51.7 Ma, 51.0 Ma, and 50.8 Ma, with margins of error of 0.4 Ma or less. The argon plateaus do not contain any hint of older ages or evidence of reheating complications (Figure 4-8). The age of this alteration and related sulfides is clearly Tertiary, with mineralization occurring at about 51 Ma. It is likely veining and mineralization took place over a period of several thousands of years. The porphyry dike sample, also dated by the same lab and method, returned a cooling age date of 46 Ma, as compared to a U-Pb zircon crystallization age of 47.86 Ma. The dates agree well with field and core observations that the rhyolite and latite or dacite porphyry dikes are consistently post-Au mineralization and similar in appearance and age to dikes of the Challis magmatic group.

Ages of Alteration Minerals at Yellow Pine Deposit

A sample of altered quartz monzonite from 382 ft depth in MGI 12-306 below the Yellow Pine pit was submitted for $^{40}$Ar/$^{39}$Ar dating (Figures 3-9 and 4-9). The interval assayed 1 ppm Au or less. In thin section (OKL-001), the quartz monzonite shows strong potassic alteration along with replacement of igneous biotite by muscovite plus titanite and opaque minerals. The gray-green color is due to the muscovite. Plagioclase in the section retains some sericitized cores, but they are surrounded and veined by potassium feldspar.

Three distinct and homogenous mineral separates were prepared with the goal to distinguish the age of any early sericite or mica alteration. Previous work by Gammons (1988) and Lewis (1984) in the 1980s indicated the possibility of Yellow Pine mineralization being significantly older than the ages obtained from the West End veins.

The coarse-grained muscovite separate returned the oldest age, 61.5 Ma ± 0.3 Ma; a potassium feldspar separate returned an age of 59.6 Ma ± 0.4 Ma; and the fine-grained sericite had an age of 53.9 Ma ± 0.3 Ma. Step-heating release spectra of the muscovite and feldspar analyses are in Figures 4-10 and 4-11. The fine-
Table 4-3. Summary of available geochronology for Stibnite region.

<table>
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<th>Sample Number</th>
<th>Mineral/Rock Unit</th>
<th>Age (Ma)</th>
<th>Error Range (Ma)</th>
<th>Method</th>
<th>Mineral</th>
<th>Deposit/Area</th>
<th>Comments</th>
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<td>outside altered</td>
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<td>This study</td>
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<td>Gaschnig and others (2017)</td>
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<td><strong>Cooling Ages</strong></td>
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<td>0.05</td>
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<td>Stibnite Pit</td>
<td>This study</td>
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<td>U-Pb, LA-ICPMS</td>
<td>rutile</td>
<td>West End</td>
<td>Stibnite Pit</td>
<td>This study</td>
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<td>muscovite</td>
<td>Yellow Pine</td>
<td>This study</td>
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<td>K-feldspar</td>
<td>59.6</td>
<td>0.4</td>
<td>^40Ar/^39Ar</td>
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<td>This study</td>
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<td>0.3</td>
<td>^40Ar/^39Ar</td>
<td>sericite</td>
<td>Yellow Pine</td>
<td>This study</td>
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<td>^40Ar/^39Ar</td>
<td>K-feldspar</td>
<td>West End</td>
<td>Stibnite Pit</td>
<td>This study</td>
</tr>
<tr>
<td>12-305-462</td>
<td>K-feldspar/breccia</td>
<td>50.8</td>
<td>0.3</td>
<td>^40Ar/^39Ar</td>
<td>K-feldspar</td>
<td>West End</td>
<td>This study</td>
<td></td>
</tr>
<tr>
<td>YP-203</td>
<td>K-feldspar (adularia)</td>
<td>56.2</td>
<td>1.3</td>
<td>K-Ar</td>
<td>K-feldspar</td>
<td>Yellow Pine</td>
<td>Lewis (1984)</td>
<td></td>
</tr>
<tr>
<td>YP-211</td>
<td>K-feldspar</td>
<td>58.9</td>
<td>1.4</td>
<td>K-Ar</td>
<td>K-feldspar</td>
<td>Yellow Pine</td>
<td>Lewis (1984)</td>
<td></td>
</tr>
<tr>
<td>YP-211 (2)</td>
<td>K-feldspar</td>
<td>57.1</td>
<td>1.3</td>
<td>K-Ar</td>
<td>K-feldspar</td>
<td>Yellow Pine</td>
<td>Lewis (1984)</td>
<td></td>
</tr>
<tr>
<td>1B-27/Gammons</td>
<td>Muscovite</td>
<td>77.9</td>
<td>0.3</td>
<td>^40Ar/^39Ar</td>
<td>muscovite</td>
<td>Yellow Pine</td>
<td>plateau age</td>
<td>Gammons (1988)</td>
</tr>
<tr>
<td>83-60/Gammons</td>
<td>Sericite</td>
<td>57.0</td>
<td>0.2</td>
<td>^40Ar/^39Ar</td>
<td>muscovite</td>
<td>Quartz Creek</td>
<td>maximum age</td>
<td>Gammons (1988)</td>
</tr>
<tr>
<td><strong>Tungsten Mineralization Ages</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6RU-013</td>
<td>Scheelite - Type 1/2</td>
<td>45.6</td>
<td>1.6</td>
<td>U-Pb, LA-ICPMS</td>
<td>scheelite</td>
<td>Hangar Flats</td>
<td>12-192-1061</td>
<td>This study</td>
</tr>
<tr>
<td>6GE-005</td>
<td>Scheelite - Type 1</td>
<td>45.4</td>
<td>0.9</td>
<td>U-Pb, LA-ICPMS</td>
<td>scheelite</td>
<td>Hangar Flats</td>
<td>12-203-1054</td>
<td>This study</td>
</tr>
<tr>
<td>6EW-002 (1)</td>
<td>Scheelite - Type 1</td>
<td>43.5</td>
<td>0.8</td>
<td>U-Pb, LA-ICPMS</td>
<td>scheelite</td>
<td>Hangar Flats</td>
<td>12-193-970A</td>
<td>This study</td>
</tr>
<tr>
<td>6EW-002 (2)</td>
<td>Scheelite - Type 2/3</td>
<td>35.8</td>
<td>0.9</td>
<td>U-Pb, LA-ICPMS</td>
<td>scheelite</td>
<td>Hangar Flats</td>
<td>12-193-970A</td>
<td>This study</td>
</tr>
<tr>
<td>12234-314</td>
<td>Scheelite - Type 2/3</td>
<td>42.8</td>
<td>1.4</td>
<td>U-Pb, LA-ICPMS</td>
<td>scheelite</td>
<td>Hangar Flats</td>
<td>12-234-314</td>
<td>Wintzer and others (2016, 2017)</td>
</tr>
<tr>
<td>12234-316</td>
<td>Scheelite - Type 1</td>
<td>44.7</td>
<td>1.3</td>
<td>U-Pb, LA-ICPMS</td>
<td>scheelite</td>
<td>Yellow Pine</td>
<td>pit</td>
<td>Wintzer and others (2016, 2017)</td>
</tr>
<tr>
<td>NMNH 107667-00</td>
<td>Scheelite - Type 1</td>
<td>44.33</td>
<td>0.4</td>
<td>U-Pb, LA-ICPMS</td>
<td>scheelite</td>
<td>Yellow Pine</td>
<td>pit?</td>
<td>Wintzer and others (2016, 2017)</td>
</tr>
<tr>
<td><strong>Other Mineralization</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Konyshev-Mo</td>
<td>Molybdenite</td>
<td>86.3</td>
<td>0.4</td>
<td>Re/Os</td>
<td>molybdenite</td>
<td>West End</td>
<td>milky quartz veins</td>
<td>Konyshev and Muntean (2016)</td>
</tr>
</tbody>
</table>
Table 4-4. Alteration and mineralization ages for the Stibnite/Yellow Pine district. Quartz is abbreviated “qtz.” Samples labelled “NW” are from Wintzer and others (2016 and 2017).

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Unit</th>
<th>Mineral/Lithology</th>
<th>Location</th>
<th>Lab</th>
<th>Method</th>
<th>Results</th>
</tr>
</thead>
<tbody>
<tr>
<td>12-305-422.1</td>
<td>Vein envelope</td>
<td>K-feldspar/with pink carbonate-qtz-vein cutting granite.</td>
<td>Below West End pit; MGI-12-305, 422 ft</td>
<td>UAF</td>
<td>⁴⁰Ar/³⁹Ar weighted average age</td>
<td>51.0 Ma ± 0.4 Ma</td>
</tr>
<tr>
<td>12-305-462</td>
<td>Vein/breccia – hydrothermal</td>
<td>K-feldspar/qtz-carbonate vein with pyrite</td>
<td>Below West End pit; MGI-12-305, 462 ft</td>
<td>UAF</td>
<td>⁴⁰Ar/³⁹Ar plateau age</td>
<td>50.8 Ma ± 0.3 Ma</td>
</tr>
<tr>
<td>12VG148N2</td>
<td>Vein envelope</td>
<td>K-feldspar/qtz-carbonate-limonite vein</td>
<td>Stibnite pit, north wall</td>
<td>UAF</td>
<td>⁴⁰Ar/³⁹Ar plateau age</td>
<td>51.7 Ma ± 0.3 Ma</td>
</tr>
<tr>
<td>12-306-382 (FS)</td>
<td>Altered quartz monzonite</td>
<td>K-feldspar; potassic alteration</td>
<td>Below Yellow Pine pit; MGI 12-306, 382 ft</td>
<td>UAF</td>
<td>⁴⁰Ar/³⁹Ar weighted average age</td>
<td>59.6 Ma ± 0.4 Ma; K-feldspar separate.</td>
</tr>
<tr>
<td>12-306-382 (MU)</td>
<td>Altered quartz monzonite</td>
<td>Muscovite; coarse-grained; alteration of biotite</td>
<td>Below Yellow Pine pit; MGI 12-306, 382 ft</td>
<td>UAF</td>
<td>⁴⁰Ar/³⁹Ar plateau age</td>
<td>61.5 Ma ± 0.3 Ma on coarse mica separate.</td>
</tr>
<tr>
<td>12-306-382 (SER)</td>
<td>Altered quartz monzonite</td>
<td>Muscovite; coarse-grained; alteration of biotite or feldspar</td>
<td>Below Yellow Pine pit; MGI 12-306, 382 ft</td>
<td>UAF</td>
<td>⁴⁰Ar/³⁹Ar weighted average age</td>
<td>53.9 Ma ± 0.3 Ma on sericite separate.</td>
</tr>
<tr>
<td><strong>SCHEELITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6RJ-013</td>
<td>Black stibnite breccia</td>
<td>Scheelite – type 2; veins in granite</td>
<td>Hangar Flats; MGI 12-192, 1061 ft</td>
<td>BSU</td>
<td>U-Pb; LA-ICPMS</td>
<td>45.6 Ma ± 1.6 Ma</td>
</tr>
<tr>
<td>6GE-005</td>
<td>Stibnite veins, breccia</td>
<td>Scheelite - type1/2; in clasts of breccia, veins in marble and granite</td>
<td>Hangar Flats; MGI 12-203, 1054 ft</td>
<td>BSU</td>
<td>U-Pb; LA-ICPMS</td>
<td>45.4 Ma ± 0.9 Ma</td>
</tr>
<tr>
<td>6EW-002 (1)</td>
<td>Black stibnite breccia</td>
<td>Scheelite – type 1; brecciated quartzite, granite</td>
<td>Hangar Flats; MGI 12-193, 970 ft</td>
<td>BSU</td>
<td>U-Pb; LA-ICPMS</td>
<td>43.5 Ma ± 0.8 Ma</td>
</tr>
<tr>
<td>6EW-002 (2)</td>
<td>Vugs in stibnite breccia</td>
<td>Scheelite – type 2/3; vug linings with stibnite</td>
<td>Hangar Flats; MGI 12-193, 970 ft</td>
<td>BSU</td>
<td>U-Pb; LA-ICPMS</td>
<td>35.8 Ma ± 0.9 Ma</td>
</tr>
<tr>
<td>12234-314 (NW)</td>
<td>Breccia</td>
<td>Scheelite – type 2/3</td>
<td>Hangar Flats; MGI 12-234, 314 ft</td>
<td>BSU</td>
<td>U-Pb; LA-ICPMS</td>
<td>42.8 Ma ± 1.4 Ma</td>
</tr>
<tr>
<td>12145-436 (NW)</td>
<td>Type 1</td>
<td>Scheelite – type 1</td>
<td>Yellow Pine pit</td>
<td>BSU</td>
<td>U-Pb; LA-ICPMS</td>
<td>44.7 Ma ± 1.3 Ma</td>
</tr>
<tr>
<td>NMNH 107667 (NW)</td>
<td>Breccia</td>
<td>Scheelite – type 1; clasts</td>
<td>Yellow Pine pit</td>
<td>BSU</td>
<td>U-Pb; LA-ICPMS</td>
<td>44.3 Ma ± 0.4 Ma</td>
</tr>
</tbody>
</table>
grained sericite had a more complex spectrum with a weighted average age of 53.9 Ma ± 0.3 Ma. The younger age of the fine-grained sericite may reflect alteration of feldspar or muscovite on cooling (Appendix E).

Figure 4-9. Yellow Pine sample 12-306-382 submitted for $^{40}$Ar/$^{39}$Ar dating. Rock is altered quartz monzonite. Igneous biotite is altered to coarse-grained, greenish-gray muscovite. Longest length of core shown in photo is approximately 4 in.

Figure 4-8. Step-heating release spectra and related graphs for alteration feldspar sample 12-305-462 (shown in Figure 4-7) from the West End deposit with a plateau age of 50.8 Ma ± 0.3 Ma. Plots for additional samples are in Appendix E.

Scheelite Ages

Wintzer and others (2016, 2017) presented the preliminary results of an innovative effort to date scheelite by U-Pb methods using LA-ICPMS. Three textural types of scheelite were defined: 1) Type 1 scheelite was in brecciated clasts, 2) Type 2 scheelite was in veins or disseminated grains, and 3) Type 3 scheelite formed in veins and vug linings, some of which are infilled with stibnite. The oldest Type 1 scheelite co-exists with pyrite but is cross cut or surrounded by stibnite. Type 2 scheelite is locally intergrown with stibnite. Results from six samples, which include four from the Hangar Flats deposit along the Meadow Creek fault zone and two from the Yellow Pine deposit, are shown in Table 4-4 and Figure 4-12. The scheelite ages are distinctly younger than the Au mineralization with the scheelite ages in the 46 to 43 Ma range. One sample of vug-filling scheelite crystals was dated at about 36 Ma. Error ranges on the scheelite ages are approximately ± 1 Ma.

DISCUSSION OF ALTERATION AND MINERALIZATION AGES

- Table 4-3 summarizes the ages for magmatic, metamorphic and hydrothermal rocks at Stibnite. Cretaceous alteration seems to have been minimal and unrelated to Au mineralization; however, it is possible that some early sericite or potassic alteration accompanied the weak Cretaceous skarn development, possibly related to the 77.9 Ma $^{40}$Ar/$^{39}$Ar mica age obtained by Gammons (1988) and/or the molybdenite age of 86 Ma determined by Konyshev and Muntean (2016). The 77.9 Ma age on Yellow Pine muscovite determined
Figure 4-10. Step-heating release spectra for 12-306-382 MU muscovite separate with plateau age of 61.5 Ma ± 0.3 Ma from the Yellow Pine deposit.

Figure 4-11. Step-heating release spectra for 12-306-382 FS K-feldspar separate with weighted average maximum age of 59.6 Ma ± 0.4 Ma. A minimum age of 58.5 ± 0.4 Ma was also calculated. Sample is from Yellow Pine deposit.
by Gammons is nearly identical to cooling ages determined in this study, and a preferred hypothesis is that the 78 Ma ages record the time of rapid uplift and exhumation of the deeply emplaced plutons and their metamorphic wall-rocks.

- The new $^{40}$Ar/$^{39}$Ar dates for Yellow Pine alteration minerals at 61 to 60 Ma are similar to, but more precise than, the approximately 57 Ma K-Ar ages on potassium feldspar alteration determined by Lewis (1984) at Yellow Pine. Unpublished data by the USGS returned slightly older ages (approximately 65 Ma) on alteration feldspar at Yellow Pine (Al Hofstra, written and oral commun., 2017), supporting the Paleocene age of Yellow Pine potassic alteration and the disseminated Au mineralization associated with it.

- Potassium feldspar vein envelopes and alteration potassium feldspar returned dates of 51.7 to 50.8 Ma from three localities in the West End area. One of those samples, 12-305-462, contained coarse, foliated muscovite veins (pre-ore to early stage) cut by a quartz-carbonate-sulfide vein and well-mineralized breccia with potassically altered clasts. This higher grade, epithermal-like mineralization is interpreted to have taken place approximately 51 Ma ago and to overprint the “early” more replacement-style mineralizing event at 60 Ma. The 51 Ma mineralizing event probably redistributed and added Au to the system, thereby enhancing the Au resource.

- Mineralization at the West End deposit is slightly different in character than that at the Yellow Pine or Hangar Flats deposits and frequently exhibits textures of classic epithermal ores, such as open vugs in banded chalcedonic to drusy veins of quartz.

Figure 4-12. Summary compilation of ages for Stibnite district from this study plus others (Konyshev and Muntean, 2016; Stewart and others, 2016; Gammons, 1988; Lewis, 1984, and Wintzer and others, 2016, 2017. Error bars on the newer analyses are hidden by symbol size. YP is Yellow Pine, WE is West End, HF is Hangar Flats, ID BATH is Idaho batholith, and Tfp is Tertiary feldspar porphyry. QC is Quartz Creek W deposit about 12 miles west of Stibnite. Refer to Tables 4-1, 4-2, 4-3, and 4-4 for sample details.
and carbonate, locally with sulfides. Cookro and others, 1987, identified these textures at the West End area, and earlier workers also noted epithermal features locally in the district. The textural difference between the Yellow Pine disseminated Au-As mineralization and the more epithermal-style mineralization in some West End ores is compatible with the different ages determined for the two deposits.

- The U-Pb scheelite ages (mostly 46 to 43 Ma) at Stibnite suggest late fluid flow along the Meadow Creek fault and possibly a third pulse of even younger mineralization. The close association of the stibnite and scheelite also suggest that at least some of the Sb is coeval or younger than the W. It is possible that the hydrothermal fluids may have recrystallized or remobilized older tungsten, as suggested by Cookro and others (1987), though Tertiary W mineralization has been documented in the region (Table 4-3). However, the U-Pb scheelite ages are younger than the few known ages (48 to 46 Ma range) on igneous dikes in the district, and more work is needed to decipher this conundrum. At Stibnite, at least one mafic dike clearly cuts the stibnite breccia which carries scheelite in the Hangar Flats area (Figure 2-18). It is possible that some of the mafic dikes are younger than 44 Ma.

- Additional sampling is needed to fully document the timing of the complex and overprinted polymetallic mineralization, particularly at the Yellow Pine deposit. Gold mineralization in the district is Paleogene in age and took place in at least two pulses prior to initiation of the Eocene Challis magmatic episode in central Idaho (Figure 4-12). Tungsten-antimony mineralization is Eocene in age. The overall span of alteration and mineralization ages from at least 61 to 44 Ma demonstrates that the mineral deposits at Stibnite were produced through the coalescence and interaction between multiple, Paleogene and Eocene hydrothermal systems.
CHAPTER 5

LEAD ISOTOPE STUDY
INTRODUCTION

A common lead isotope study was initiated in order to establish the diversity and provenance of the ore-forming fluids in the Stibnite mining district. Lead (Pb) is a trace metal of the ore assemblage, and due to its combined lithophile and chalcophile geochemical affinity, is incorporated into a variety of ore (pyrite, arsenopyrite, stibnite) and gangue (potassium feldspar, calcite) minerals. Lead isotopic variation occurs due to the time-dependent ingrowth of radiogenic Pb from radioactive U and Th. Igneous and metasedimentary rocks of differing age thus have widely varying Pb isotopic compositions, expressed as the isotopic ratios, $^{208}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{206}\text{Pb}$, and $^{208}\text{Pb}/^{206}\text{Pb}$.

Stacey and Kramers (1975) defined a two-stage (two stages of evolution with differing $^{238}\text{U}/^{204}\text{Pb}$ or $\mu$ values) Pb isotope evolution model for the bulk upper mantle and crust, on the basis of stratiform ores and mantle-derived igneous rocks of known age. This model evolution curve provides a reference frame for juvenile magmas, fluids, and rocks extracted from the mantle at a given age. Many juvenile arc magmas reflect these model compositions, however continental arc magmas like those of the Idaho batholith can be displaced from the Stacey and Kramers model through assimilation and/or anatexis of older continental crust. Upper crustal rocks in particular evolve to higher, more ‘radiogenic’ values of $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ due to their time-integrated high U and Th concentrations. For example, Archean crust of the Wyoming craton is known to have elevated $^{207}\text{Pb}/^{204}\text{Pb}$ relative to $^{206}\text{Pb}/^{204}\text{Pb}$ due to the evolution of high U/Pb crustal domains from the Paleoproterozoic (Wooden and Mueller, 1988). High U/Pb and Th/Pb source rocks can produce extremely radiogenic Pb isotope ratios (high $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, $^{208}\text{Pb}/^{204}\text{Pb}$, and low $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{208}\text{Pb}/^{206}\text{Pb}$) beyond the isotopic composition of ‘modern Pb’ defined by Stacey and Kramers (1975). These highly radiogenic Pb compositions are a hallmark of mineral deposits sourced from crustal fluids scavenging metals from older sedimentary and plutonic continental crustal rocks (Goldhaber and others, 1995; Panneerselvam and others, 2006).

Lead isotope ratios can be measured by a variety of mass spectrometric techniques. We utilized two methods of analysis: 1) dissolution of ‘bulk’ pyrite, stibnite, carbonate, and leached potassium feldspar mineral separates, separation of Pb by ion chromatography, and high-precision analysis by TIMS; 2) in situ sampling of sulfide minerals via laser ablation and low-precision isotope ratio analysis by quadrupole ICPMS. These two techniques have complementary strengths and weaknesses. TIMS analysis produces intense ion beams and high-precision isotope ratios ($\leq 0.01$ percent relative standard deviation) at the expense of spatial resolution and grain-scale Petrographic context. TIMS analysis also includes accurate measurement of the minor $^{204}\text{Pb}$ isotope, thus allowing calculation and visualization of the $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{206}\text{Pb}$, $^{208}\text{Pb}/^{206}\text{Pb}$ ratios. LA-ICPMS analysis can sample at the scale of 30 microns within a thin section, but suffers from much lower precision ($\leq 0.5$ percent relative standard deviation), which when combined with a persistent mercury interference on the minor $^{204}\text{Pb}$ isotope limits the measurement and interpretation to $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{208}\text{Pb}/^{206}\text{Pb}$ ratios. Thin section petrography guided our sampling strategies for relatively simple (TIMS) versus complex (LA-ICPMS) mineral assemblages.

METHODS

Thermal Ionization Mass Spectrometry

Mineral samples were separated, hand picked, cleaned via ultrasonication, chemically prepared, and analyzed by TIMS. Samples included eight feldspar separates, three stibnite separates, and six pyrite separates. Feldspar samples included potassium feldspar from altered intrusive host rocks, particularly the same intervals as were dated by U-Pb zircon analysis, as well as two samples of potassium feldspar from wall rock alteration envelopes. Stibnite samples included two from the Hangar Flats core and a single sample from the Yellow Pine core. Pyrite proved to be the most difficult mineral to separate, due to its fine grain size and multiple occurrence modes. All of the pyrites separated were from the West End area (drill core or surface samples from the Stibnite Pit). Some samples contained two distinct shapes of pyrite, as described in the sample list in Appendix F-1.

For Pb isotopic analysis, approximately 50 mg aliquots of sulfide minerals were completely dissolved in concentrated nitric acid. For feldspar analysis, approximately 200 mg of separated crystals from each
sample were sequentially leached using modifications to the method of Housh and Bowring (1991). Feldspar separates were transferred to 15 mL Savillex PFA Teflon beakers in ethanol and rinsed three times with 1 mL H₂O. Each separate was first fluxed in 1 mL 3.5M HNO₃ for 30 minutes on a 120°C hot plate, and then rinsed three times with 0.5 mL H₂O. This step was followed by fluxing in 1 mL 6M HCl for 30 minutes at the same temperature and rinsing with three aliquots of 0.5 mL H₂O. Both nitric and hydrochloric acid leachates were generally discarded; an exception was sample 12VG148J, an intimate mixture of carbonate and potassium feldspar. In this case the nitric acid leachate was retained as representative of the carbonate mineral isotopic composition, and the insoluble residue was processed as the silicate feldspar component.

Residual feldspar separates were then sequentially dissolved in 1 mL 1M HF + 0.1 mL concentrated HNO₃ in 30 minute intervals on a 120°C hotplate. After each sequential dissolution step, acid was pipetted away from the crystal residue and transferred to a clean 5 mL Savillex beaker and rinsed three times with 0.5 mL H₂O. Each rinse was combined with the retrieved leachate. Sequential dissolution was repeated up to five or six times depending on the size of the remaining sample after each leaching step.

Lead was separated from each dissolved mineral fraction by anion exchange in dilute HBr medium using AG1-X8, 200-400 mesh Eichron resin in 100 μL columns. Purified Pb samples were loaded on outgassed Re center filaments in 2 μL of a silica-gel/phosphoric acid mixture (Gerstenberger and Haase, 1997) and isotopic measurements made on an IsotopX Isoprobe-T multicollector thermal ionization mass spectrometer equipped with nine Faraday cups. Isotope ratios were measured in static Faraday mode; external fractionation correction was made on the basis of repeated measurements of the NBS-981 standard during the course of analysis.

Laser Ablation Inductively Coupled Plasma Mass Spectrometry

Polished thin sections were mapped with reflected light microscopy to identify and target sulfide crystals in petrographically characterized domains for LA-ICPMS. The use of quadrupole mass spectrometry allowed the simultaneous measurement of Pb isotope ratio measurements and trace metal concentrations, including Ag and Au. Sulfide grains were targeted using a New Wave Research UP-213 Nd:YAG UV (213 nm) laser ablation system (30 μm aperture, 5 J/cm² fluence, 5 Hz firing repetition, 15 sec gas blank, 30 sec sample ablation) connected to a ThermoElectron X-Series II quadrupole ICPMS. Ablated material was carried by a ~1 L/min He gas stream to the plasma. Quadrupole mass dwell times were 2 milliseconds (ms) for ²⁹Si, ³¹S; 5 ms for ⁴⁴Ca, ⁴⁷Ti, ⁵¹V, ⁵²Cr, ⁵⁵Mn, ⁵⁷Fe, ⁶⁰Co, ⁶⁰Ni, ⁶⁵Cu, ⁶⁶Zn, ⁷⁵As, ¹²¹Sb; 20 ms for ¹⁰⁷Ag, ¹⁹⁷Au; and 100 ms for ²⁰⁸Hg, ²⁰⁴Pb, ²⁰⁶Pb, ²⁰⁷Pb, ²⁰⁸Pb. Background count rates for each analyte are obtained prior to each spot analysis and subtracted from the raw count rate for each analyte. For concentration calculations, background-subtracted count rates for each analyte were internally normalized to ³⁵Fe (pyrite) or ¹²⁵Sb (stibnite), and calibrated with respect to stoichiometric pyrite, arsenopyrite, and stibnite (Fe, S, As, and Sb) or NIST 610 and 612 glasses (remaining trace metals) as primary standards. BIR-1G, BCR-2G, GSE-1G, and GSD-1G glasses were measured in the same experiments to assess accuracy within and reproducibility across experiments. Accuracy and reproducibility of the suite of minor and trace elements in these glasses were within 5-10 percent for elements at concentrations >50 ppm, and ≤20 percent for concentrations >1 ppm. While absolute bias due to the differential ablation of sulfide unknowns versus silicate standards is likely, this should be mitigated by internal standardization, and relative variations in abundances will be robust. LA-ICPMS determined As and Au contents in many pyrites are comparable to concentrations measured by electron microprobe (Chapter 6; Appendix G). Pb isotope ratio fractionation was monitored using NIST 610 glass and was indistinguishable from zero in all experiments. Recovered ²⁰⁶Pb/²⁰⁴Pb and ²⁰⁷Pb/²⁰⁶Pb ratios for NIST 610, NIST 612, BCR-2G, BIR-1G, GSE-1G and GSD-1G glass standards were within 0.5% (1σ) or better of their true values (Appendix F-4).

RESULTS

Thermal Ionization Mass Spectrometry

Results of TIMS Pb isotope ratio analyses are shown in Figure 5-1 as plots of ²⁰⁶Pb/²⁰⁴Pb versus ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁸Pb/²⁰⁴Pb versus ²⁰⁶Pb/²⁰⁴Pb. Samples are denoted by mineralogy as well as their location in three resource
Figure 5-1. Plot of common Pb isotopes for Stibnite district ore and gangue minerals analyzed by TIMS. Sample points labeled ‘Au’ contained >2 ppm Au in bulk rock assays from the same core interval. A) $^{207}\text{Pb}/^{204}\text{Pb}-^{206}\text{Pb}/^{204}\text{Pb}$ isotope correlation diagram; B) $^{208}\text{Pb}/^{204}\text{Pb}-^{206}\text{Pb}/^{204}\text{Pb}$ isotope correlation diagram. The Stacey and Kramers (1975) 2-stage isotope evolution model is shown for reference, marked in increments of 100 Ma. Data for the Idaho batholith, reported by Gashnig and others (2011), are illustrated for comparison.
areas: Yellow Pine (YP), West End (WE), and Hanger Flats (HF). Also shown on these figures are the whole-rock compositions of a variety of plutonic rocks of the Idaho batholith from Gaschnig and others (2011), and the Stacey and Kramers (1975) Pb isotope evolution model curve (labeled in increments of 100 Ma, and ending at modern Pb). Numerical results are also in Appendix F-2.

With the exception of the single Eocene dike sample, which plots near the modern Pb of Stacey and Kramers (1975) and within the field of other Tertiary Challis intrusive rocks, the Pb isotopic values for feldspars, carbonate, and sulfide ore minerals of the Stibnite district are highly radiogenic, with elevated $^{208}\text{Pb}/^{204}\text{Pb}$ (40.21 to 41.92), $^{207}\text{Pb}/^{204}\text{Pb}$ (15.82 to 16.12), and $^{206}\text{Pb}/^{204}\text{Pb}$ (20.29 to 22.13). These ratios are significantly offset to more radiogenic values from the majority of Idaho batholith granitoid compositions, including metaluminous and peraluminous compositions of the Cretaceous intrusive rocks, and the later Eocene Challis-event intrusions. Elevated $^{206}\text{Pb}/^{204}\text{Pb}$ is also matched by radiogenic $^{207}\text{Pb}/^{204}\text{Pb}$, necessitating an ancient high-$U$/Pb source of Pb, and elevated $^{208}\text{Pb}/^{204}\text{Pb}$, requiring a high-Th/Pb source compared to those of the Idaho batholith magmas.

When differentiated by ore body, the potassium feldspar compositions of the Yellow Pine and Hanger Flats altered alaskites and pegmatites are the most radiogenic and tightly clustered. By contrast, the potassium feldspars and associated carbonate of both altered granitoids and discrete vein envelopes of the West End deposit, including the Stibnite pit, exhibit a much wider range of compositions defining a correlated array in $^{206}\text{Pb}/^{204}\text{Pb}-^{207}\text{Pb}/^{204}\text{Pb}$ space, from values comparable to the Yellow Pine samples to much less radiogenic compositions. Pyrite separates from the West End deposit overlap with the less radiogenic feldspar compositions, as do stibnite compositions from both the Yellow Pine and Hanger Flats deposits. Two samples of pyrite and potassium feldspar hosted in metasedimentary rocks from the West End deposit have anomalously elevated $^{207}\text{Pb}/^{204}\text{Pb}$ compared to the broadly linear array; this Pb signature is likely due to the greater incorporation of significantly older sedimentary-hosted Pb that evolved in a high-$\mu$ environment. In $^{208}\text{Pb}/^{204}\text{Pb}-^{206}\text{Pb}/^{204}\text{Pb}$ space, the Yellow Pine/Hanger Flats samples are clearly offset to higher $^{208}\text{Pb}/^{204}\text{Pb}$ compared to the West End compositions, necessitating their evolution from a higher Th/Pb crustal reservoir.

Laser Ablation Inductively Coupled Plasma Mass Spectrometry

In order to extend the scope of the Pb isotope analysis to additional sulfide minerals, particularly more sulfide-poor samples of the Yellow Pine deposit, and to examine the Pb isotopic variability within individual crystals and across domains at the scale of a thin section, we applied the technique of Pb isotope ratio analysis using LA-ICPMS. Figure 5-2 presents a series of $^{208}\text{Pb}/^{206}\text{Pb}$ vs. $^{207}\text{Pb}/^{206}\text{Pb}$ isotope ratio plots. Each of the plots shown in Figure 5-2 B to 5-2 G presents data for sulfide grains from six individual thin sections. Different domains (indicated by capital letters) within a thin section are illustrated in different colors according to the legend of each panel. A domain is defined as an area within the section that contains specific petrographic relationships and which was analyzed; analyses are of arsenian pyrite unless otherwise noted.

For comparison with the in-situ LA-ICPMS data, Figure 5-2A illustrates the previously described TIMS data along with the whole-rock Idaho batholith samples of Gaschnig and others (2011) on a graph of $^{208}\text{Pb}/^{206}\text{Pb}-^{207}\text{Pb}/^{206}\text{Pb}$. Consistent with Figure 5-1, all Stibnite district gangue and ore minerals are significantly offset from most Idaho batholith magmatic compositions. Yellow Pine and Hanger Flats feldspars are clustered at the lowest $^{207}\text{Pb}/^{206}\text{Pb}$ isotope ratios, with feldspars and pyrites from the West End forming a correlated array to higher $^{207}\text{Pb}/^{206}\text{Pb}$ isotope ratios. As in Figure 5-1B, the West End array is offset from the more thorianic, higher $^{208}\text{Pb}/^{206}\text{Pb}$ Yellow Pine and Hanger Flats feldspars and stibnites.

LA-ICPMS analyses of arsenian pyrite and arsenopyrite in thin sections of four of the six samples are tightly clustered in Pb isotope ratios. These include a veined quartz monzonite in core from beneath the Yellow Pine pit (6RJ-019, Figure 5-2B), fault breccia in the DMEA adit of the Hanger Flats area (8EB-002, Figure 5-2C), and granite with potassic alteration in core from beneath the West End (5XG-006, Figure 5-2D) that exhibit progressively higher $^{208}\text{Pb}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ isotope ratios from Figures 5-2B to 5-2C and then 5-2D). These trends closely mimic those illustrated by the sparser TIMS data (Figure 5-2A). Importantly, there is no isotopic overlap between the Yellow Pine and West End samples, while the Hanger Flats sample contains a bimodal assemblage intermediate between the Yellow Pine and West End samples. Pyrite in a second sample of
Figure 5-2. Pb isotope correlation diagrams for Stibnite district ore and gangue minerals analyzed by TIMS (5-2A) and LA-ICPMS (5-2B-G). Panel A illustrates TIMS measurements, which are reproduced in gray outlines in subsequent panels. Symbols follow those in Figure 5-1. In panels B-G, different spatial domains (also indicated by capital letters) within a thin section are illustrated in different colors according to the legend of each panel; analyses are of arsenian pyrite unless otherwise noted. Pyrite analyses with high Au contents are illustrated with thicker symbol outlines. Rock types are quartz monzonite (QM) and Cretaceous granite (Kg). The Stacey and Kramers (1975) 2-stage isotope evolution model is shown for reference, marked in increments of 100 Ma.
Figure 5-2C and D.
Figure 5-2E and F.
altered quartz monzonite (5XG-011, Figure 5-2E) from the Yellow Pine deposit is very similar to the Hanger Flats fault breccia pyrites; stibnite from this sample is consistently offset to higher $^{208}\text{Pb}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ isotope ratios compared to pyrite.

Two thin sections of altered quartz monzonite from the Yellow Pine deposit (4LO-022 and 4LO-023) preserve more complex relationships between petrographically distinct pyrite hosted in the wall rock, pyrite in cross-cutting veins, and pyrite and stibnite in late veins cutting the aforementioned earlier pyrite-bearing veins (Figures 3-34 through 3-37). Each petrographic domain preserves unique Pb isotopic compositions in their sulfide minerals. The domain is indicated by the capital letter after the sample number and mineral name in Appendix F-3. In section 4LO-022 (Figure 5-2F) stibnite and gold-enriched arsenian pyrite grains within a younger potassium feldspar-carbonate-stibnite breccia vein (stibnite in domain D and pyrite in domain E) span a wide array of isotopic compositions but are systematically offset to lower $^{208}\text{Pb}/^{206}\text{Pb}$ at a given $^{207}\text{Pb}/^{206}\text{Pb}$, compared to gold-bearing and gold-poor pyrites in older thin carbonate veins (domains B and C). In section 4LO-023 (Figure 5-2G) skeletal arsenian pyrite (domain A) and arsenopyrite have the lowest $^{208}\text{Pb}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of any measured sulfides. The skeletal forms are distinct from pyrite and stibnite associated in a cross-cutting potassium feldspar-carbonate-stibnite breccia vein (domain C) that have Pb isotope values ranging up toward the highest $^{208}\text{Pb}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of any measured sulfides.

Figure 5-3 illustrates probability density functions that aggregate the $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of sulfide analyses from each thin section. This alternative representation emphasizes the relative homogeneity or heterogeneity of each sample, and it displays the differences between samples taking into account analytical errors. Modes of isotopic compositions are more readily identified compared with bivariate diagrams. Multiple modes of isotopic composition are illustrated by most samples, both across pyrites of different petrographic domains, and between pyrite and stibnite. It is immediately evident that early pyrites of the Yellow Pine samples (6RJ-019, 4LO-023 A, 4LO-022) preserve a distinctly lower $^{207}\text{Pb}/^{206}\text{Pb}$ (and $^{208}\text{Pb}/^{206}\text{Pb}$) isotopic composition, compared with sulfides within late stibnite plus pyrite-bearing breccia veins or the altered West End granite.

An added benefit of our analytical approach was the simultaneous measurement of Au and other elemental...
Pyrite analyses with high Au contents are illustrated with thicker symbol outlines. Gold is most closely correlated with Cu contents and shows a weak negative correlation with Pb concentrations. Lead is most closely correlated with Sb (Appendix F-3). These trace metal contents demonstrate that the isotopic compositions of Pb in pyrite are directly fingerprinting the provenance of the Au, Sb, and Pb incorporated in the same mineral phase at the time of precipitation, and thus from the same ore-forming fluids.

DISCUSSION

On the \(^{207}\text{Pb}/^{206}\text{Pb}\) vs. \(^{206}\text{Pb}/^{204}\text{Pb}\) and \(^{208}\text{Pb}/^{204}\text{Pb}\) vs. \(^{206}\text{Pb}/^{204}\text{Pb}\) plots in Figure 5-1, the sulfides and alteration feldspars generally plot in a broad linear array above and to the right of the Stacey and Kramers evolution model curve. Such radiogenic isotopic compositions require the ingrowth of radiogenic Pb within a very high U/Pb and Th/Pb crustal setting. The lack of overlap with most whole rock plutonic Pb isotope compositions of the Idaho batholith reinforces the conclusion that the ore minerals have distinctive crustal source(s) that differ in character and/or proportion compared to the magmatic compositions of the Cretaceous or Eocene granitoids. The ores are not simply magmatic derivatives. Conversely, the tail to more radiogenic Pb isotopic compositions exhibited by a handful of early metaluminous suite samples, and the consistently elevated \(^{207}\text{Pb}/^{206}\text{Pb}\) relative to \(^{206}\text{Pb}/^{204}\text{Pb}\) for the Atlanta lobe peraluminous suite plutons suggest that these magmas variably interacted with and assimilated a high U/Pb crustal component or that they were altered by the same characteristic fluids that resulted in ore mineralization at Stibnite.

In order to better constrain the provenance of the metals in these ore-forming fluids, we forward modeled an ensemble of multi-stage Pb-isotope evolution curves and compared the results to the higher resolution TIMS data (Figure 5-4). All models utilized the first and second stage \(^{238}\text{U}/^{206}\text{Pb}\) (\(\mu\)) and \(^{232}\text{Th}/^{208}\text{U}\) (K) of Stacey and Kramers (1975), representing the ancient evolution of the upper mantle through time until the episode of crustal extraction. They differ in the timing of initiation and variation in \(\mu\) and K for the third stage of Pb-isotope evolution.
evolution, which is the effective crustal residence time prior to extraction of Pb (and other metals) into the ore-forming fluid. We first focus on the $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ systematics given their simpler modeling parameters and greater temporal significance (due to the time-sensitive ingrowth of $^{207}\text{Pb}$ from $^{235}\text{U}$ decay). Presuming that the correlated Pb isotope array defined by most samples represents mixing, our first goal was to model possible end members of the array. Our secondary goal was to constrain the nature of the crustal reservoirs that could form the highly elevated $^{207}\text{Pb}/^{204}\text{Pb}$ values of the two anomalous West End samples (pyrite and feldspar).

Our first model ensemble assumes third stage crustal extraction in the late Archean, at 2.55 Ga. Examining a late Archean third stage extraction is warranted by the age of basement rocks underlying southern Idaho, whether as a westward extension of the southern province of the Wyoming craton (Shirley, 2013) or as a distinctive Grouse Creek block (Mueller and others, 2011). A series of evolution curves corresponding to $\mu$ values from 12 to 16 are illustrated in Figure 5-4A, with 100 Ma increments ending at the present day. These hyperbolic growth curves express a fanning pattern extending to a wide range of elevated $^{207}\text{Pb}/^{204}\text{Pb}$ values, reflecting early ingrowth of $^{207}\text{Pb}$ from $^{235}\text{U}$ decay. Present-day Pb isotopic compositions for higher $\mu$ growth curves are generally similar to the compositions observed in the two anomalously high $^{207}\text{Pb}/^{204}\text{Pb}$ ore samples derived from metasedimentary rocks. The growth curve with a moderately high $\mu$ of 12 results in a present-day isotopic composition just to the left of the end of the Stibnite ore array (Figure 5-4A). We interpret the results of this modeling as successfully demonstrating that late Archean high-$\mu$ crust is an appropriate source of some components of the ore-forming fluids, particularly those producing the West End arsenian pyrite and the Hanger Flats stibnite ores.

Our second model ensemble assumes third stage crustal extraction in the late Neoproterozoic, at 0.6 Ga (Figure 5-4A). This is a reasonable age to explore as it corresponds to Windermere Supergroup rocks exposed along the North American Cordillera, including a variety of igneous volcanic and plutonic rocks (Lund and others, 2003; 2010). However it is immediately apparent that even with extreme $\mu$ values, the subhorizontal growth curves do not reach sufficiently high $^{207}\text{Pb}/^{204}\text{Pb}$ or $^{206}\text{Pb}/^{204}\text{Pb}$ values to explain the Stibnite ore compositions. The same would be true for younger third stage separation in the Phanerozoic. We conclude that Neoproterozoic and younger crustal reservoirs are not important sources or sites of equilibration for the ore-forming fluids.

Our third model ensemble assumes third stage crustal extraction in the Mesoproterozoic, at 1.4 Ga (Figure 5-4A). This is an important period of ‘anorogenic granite’ emplacement through Laurentia, including the local intrusion of plutons in northern and eastern Idaho (Evans and Zartman, 1990; Aleinikoff and others, 2012). This is also the time of subsidence of the Belt basin and its infill with clastic sedimentary rocks and associated extension-related volcanic and plutonic rocks (Evans and others, 2000). A suite of models with $\mu$ values reaching 24 produce present-day Pb isotopic compositions proximal to the more radiogenic end of the Stibnite ore array. While such $\mu$ values are extreme, they are not unheard of in upper crustal sedimentary compositions. We interpret these models as indicative of Mesoproterozoic sedimentary rock units as important sources or sites of equilibration for the ore-forming fluids.

In summary, there is clear evidence from Pb-isotope modeling for predominantly crustal provenance of the ore-forming fluids and their metals budget. These two crustal sources include: rocks of at least Late Archean age or the eroded and redeposited equivalents; and metasedimentary rock packages of Mesoproterozoic age. Younger Neoproterozoic to Phanerozoic rock packages appear to contribute little to the Pb-isotopic composition, and by inference the metals budget, of the ore-forming fluids.

We now address the apparent dichotomy between Pb-isotope compositions of the Yellow Pine and West End deposits. Potassic alteration of rocks of the Yellow Pine deposit are tightly clustered at the more radiogenic end of the $^{207}\text{Pb}/^{204}\text{Pb}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ array, and thus consistent with a predominantly Mesoproterozoic metasedimentary rock provenance for the ore-forming fluids. By contrast in the West End deposit, Pb-isotopic compositions scatter across the $^{207}\text{Pb}/^{204}\text{Pb}$ or $^{206}\text{Pb}/^{204}\text{Pb}$ array, suggesting the addition of an Archean crustal component in the ore-forming fluids and metals budget, which mixed with the predominantly Mesoproterozoic crustal provenance exhibited at Yellow Pine. The influence of Archean crustal Pb is also consistent with the high $^{207}\text{Pb}/^{204}\text{Pb}$ samples at West End. Stibnite compositions similarly range across the $^{207}\text{Pb}/^{204}\text{Pb}$ -
Figure 5-4. Pb isotope evolution models illustrated on $^{207}\text{Pb}/^{204}\text{Pb} - ^{206}\text{Pb}/^{204}\text{Pb}$ (A) and $^{208}\text{Pb}/^{206}\text{Pb} - ^{207}\text{Pb}/^{206}\text{Pb}$ (B) correlation diagrams, superposed onto Stibnite district ore and gangue minerals from this study and Idaho batholith whole-rock analyses from Gaschnig and others (2011). Symbols follow those in Figure 5-1.
\(^{206}\text{Pb}/^{204}\text{Pb}\) isotope array, suggesting mixing between Archean and Proterozoic crustal metals sources.

In order to further explore the mixing scenario, we modeled the coupled thorogenic and uranogenic Pb isotope signals. Figure 5-4B illustrates these models on a \(^{206}\text{Pb}/^{204}\text{Pb} - ^{207}\text{Pb}/^{206}\text{Pb}\) isotope correlation diagram. On this diagram, the isotopic compositions of West End are clearly offset to lower \(^{206}\text{Pb}/^{204}\text{Pb}\) for a given \(^{207}\text{Pb}/^{206}\text{Pb}\) compared to Yellow Pine and Hanger Flats samples. These two groups define separate arrays in Figure 5-4B that on Figure 5-4A overlapped in isotope space. We selected \(\mu\) values of 12 and 24 for the Archean and Mesoproterozoic third stage models, respectively, and varied \(^{232}\text{Th}/^{238}\text{U}\) (K) values. Both Archean and Mesoproterozoic models required distinct K values to model the ends of the two (West End and Yellow Pine-Hanger Flats) sample arrays. K values were consistently lower for the West End models, compared to the Yellow Pine-Hanger Flats models. We fit mixing lines (red lines on Figure 5-4B) between the Archean and Proterozoic third stage model present-day isotopic compositions to illustrate the linearity of these data arrays and their conformity to relatively simple two-component mixing. The thorogenic Pb system complements the conclusions from \(^{206}\text{Pb}/^{204}\text{Pb} - ^{207}\text{Pb}/^{204}\text{Pb}\) modeling in requiring both Archean and Proterozoic crustal metals sources, but also further differentiates the sources for the three resource areas in terms of their time-integrated \(^{232}\text{Th}/^{238}\text{U}\).

In summary, the changes in Pb-isotope compositions and interpreted Precambrian crustal metals provenance between the Yellow Pine, West End, and Hanger Flats (stibnite) deposits are consistent with the time-transgressive nature of emplacement of these ore bodies:

- The earliest (ca. 61 Ma) Yellow Pine mineralization is more clustered isotopically and arguably sourced from a more monotonous metals source from relatively high U/Pb, high Th/U Mesoproterozoic crustal protoliths.

- Subsequent West End mineralization (ca. 51 Ma) saw the introduction of a new predominantly Archean metals source that mixed with the earlier Proterozoic-sourced metals of the Yellow Pine deposit, potentially through reworking of the earlier deposits, but also with new contributions from lower Th/U Mesoproterozoic crust.

- The latest stibnite mineralization (ca. 45 Ma), which demonstrably cross cuts and reworks the older phases of mineralization at Yellow Pine and Hanger Flats, also shows a mixed isotopic provenance, consistent with ore remobilization as well as new metals contributions biased toward higher Th/U Archean crustal sources.

These conclusions from modeling of the TIMS Pb-isotope measurements are also consistent with the isotopic variations correlated to cross-cutting relationships observed at the thin-section scale in lower resolution LA-ICPMS Pb isotope measurements.
CHAPTER 6

GOLD-BEARING PYRITE

AND MINERAL CHEMISTRY
INTRODUCTION AND METHODS

Pyrite is one of the most common minerals in hydrothermal ore deposits, including at Stibnite. Numerous journal articles describe the diverse characteristics of pyrite in different types of deposits. This chapter records our research describing the appearance and chemistry, including the distribution of As and Au, of pyrite in the Stibnite district. The main focus was to determine the distribution of As in the arsenian pyrite and also Au, and to examine any multi-element correlations or evolutionary trends in the pyrite. Other ore minerals were analyzed to determine composition and trace-element compositions, but in fewer samples. As expected, pyrite from Stibnite exhibits significant physical and chemical diversity at multiple scales.

Twenty polished thin sections were examined by electron microprobe and scanning electron microscope (SEM) to analyze mineral compositions, look for compositional zoning, and determine Au and As contents in pyrite (Table 6-1; Appendix G, particularly Appendices G-1 and G-2). Samples were selected to include a variety of mineralized and non-mineralized rocks. Analyses were completed at Washington State University on a JEOL JXA-8500F field emission electron microprobe/SEM during two multi-day sessions in 2015 and 2016. Qualitative identification of phases was done by SEM and energy dispersive X-ray spectroscopy (EDS) prior to analysis by the electron microprobe. For sulfides (pyrite, arsenopyrite, stibnite), all samples were analyzed for Fe, S, As, Au, Co, Cu, Hg, Pb, Sn, M, and Ni, and some for Ag, Te, and Se (Appendix G-1, G-2, G-3, and G-4). Scheelite was analyzed in three samples (Appendix G-4). Silicates (feldspars and micas) were analyzed quantitatively in five sections and identified qualitatively by EDS in a few other samples (Table 6-1, Appendix G-5). Feldspar and muscovite were determined to be K-rich, though the muscovites showed interesting Fe-rich variations worthy of additional follow up.

After detecting Au in a few analyses of pyrite, instrument settings and count times were adjusted to maximize detectability and precision of the Au measurements. Count times were lengthened to better analyze for Au and other trace elements in the pyrites. Nine of the polished sections contained multiple analyzed pyrite points with detectable Au in concentrations from trace to 0.1 weight percent (1000 ppm) Au. For the remainder of this discussion, “percent” will refer to weight percent, unless specified otherwise. Two additional sections had pyrite with only one or two points containing Au. Detection limit for Au was approximately 0.015 percent for the initial 2015 pyrite runs, but it dropped to 0.006 percent (60 ppm) Au for the optimized subsequent “redo” runs on 4LO-023 and 5XG-011 (Appendix G-1). The detection limit for Au was also about 0.006 percent for the microprobe analyses done in 2016 (Appendix G-2). Compositional maps were made of selected areas of interest in a few pyrite samples. Count times varied by element and sample; the ZAF intensity matrix correction was used. For the 2015 analytical runs, an accelerating voltage of 25 kV was used with a 50 nA beam current and a beam size of 0-1 μm, providing excellent spatial resolution. In 2016, the accelerating voltage was 20 kV and for the pyrites, a 200 nA beam current was used. Appendix G contains the microprobe analytical results for pyrite, arsenopyrite, stibnite, scheelite, and silicates in the 2015 and 2016 sessions. Averages of multiple points taken in specific domains of the grains and at specific sites are also provided in Appendix G.

Prior to the electron microprobe sessions, transmitted and reflected light petrography was conducted on each polished thin section to identify minerals and textures of interest. Specific sites (labelled A, A-2, B, etc.) were marked in ink on the section, which was then scanned to provide an image and map for the microprobe work. Under the much greater magnification of the microprobe and SEM, multiple grains of pyrite (for example) could be seen. Grains picked for analysis were designated by number (1, 2, 3, etc.). Where intra-mineral zoning or distinct domains could be seen, those were described and descriptively named. For example, pyrites often displayed core, middle, and rim zones, light and dark bands evident in backscatter mode (BSE) image or a skeletal habit. Thus, in the analytical results pyrite spreadsheets in Appendix G, the spot labelled “4LO-023_Site A-2 Light” refers to a light-colored pyrite at site A-2 which is near Site A in polished section 4LO-023. The label “4VF-006_Pyrite B_Corroded” would refer to a pyrite with corroded texture at site B in section 4VF-006. To statistically better characterize a particular zone or grain, several spot analyses were done; averages are also given in Appendix G. Some grains were analyzed using an automated linear traverse
Table 6-1. List of electron microprobe samples, Stibnite district, Idaho. Py is pyrite; Stbn is stibnite; Apy is arsenopyrite; Sch is scheelite; Kfs is potassium feldspar; Pl is plagioclase; Ms is muscovite; Sulf is sulfide; Unk is unknown. NA is not assayed (surface sample) or not analyzed, with an estimate of likely or unlikely gold mineralization. BDL is below detection limit in assay. Yellow Pine DH refers to a drill hole sample from below the Yellow Pine pit, while Yellow Pine pit refers to a surface sample. Two sections (5XG-011 and 4LO-023) were rerun in 2015 due to equipment issues and those analyses are listed as “redos” in 2015 in Appendix G-1.

<table>
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<tr>
<th>TS #</th>
<th>WAYPOINT / DH</th>
<th>LOCATION</th>
<th>LITHOGRAPHY</th>
<th>GOLD ASSAY FOR INTERVAL (ppm)</th>
<th>GOLD DETECTED IN PYRITE (Probe)</th>
<th>MINERALS PROBED (DATE)</th>
</tr>
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<td><strong>Outside of Resource Area</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>8EB-008</td>
<td>15VG024A</td>
<td>Ridge N of Fern mine</td>
<td>Calc-silicate w/Py</td>
<td>NA, but unlikely</td>
<td>No</td>
<td>Py (2016)</td>
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<td><strong>West End Area/Stibnite Pit</strong></td>
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<td>4LO-013</td>
<td>12VG149-1</td>
<td>Stibnite pit</td>
<td>Meta-siltstone</td>
<td>NA</td>
<td>No</td>
<td>Py (2015)</td>
</tr>
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<td>10-37-120.8</td>
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<td>1.69 ppm</td>
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<td>Py (2016)</td>
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</tr>
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<td>NA</td>
<td>No</td>
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<td>Granite w/ pink vein</td>
<td>0.206 ppm</td>
<td>Yes</td>
<td>Py, Kfs, Ms (2016)</td>
</tr>
<tr>
<td>6RJ-008</td>
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<td>Granite</td>
<td>0.209 ppm</td>
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<td>Py (2015)</td>
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<td>West End DH</td>
<td>Quartzite w/breccia</td>
<td>6.79 ppm</td>
<td>Yes</td>
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<td>12-305-462</td>
<td>West End DH</td>
<td>Metasediment w/breccia</td>
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<td><strong>Yellow Pine Deposit</strong></td>
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<td></td>
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<td>4LO-023</td>
<td>12VG159F</td>
<td>Yellow Pine pit</td>
<td>Quartz monzonite</td>
<td>NA, but likely</td>
<td>Yes</td>
<td>Py, Apy, As-Fe-S Unk, Kfs, Ms (2015)</td>
</tr>
<tr>
<td>4LO-022</td>
<td>12VG159D</td>
<td>Yellow Pine pit</td>
<td>Quartz monzonite</td>
<td>NA, but likely</td>
<td>Yes</td>
<td>Py (2016)</td>
</tr>
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<td>Yellow Pine pit</td>
<td>Quartz monzonite</td>
<td>NA, but likely</td>
<td>Yes</td>
<td>Py (2016)</td>
</tr>
<tr>
<td>5XG-011</td>
<td>12-306-425p</td>
<td>Yellow Pine DH</td>
<td>Quartz monzonite</td>
<td>3.73 ppm</td>
<td>Yes</td>
<td>Py, Kfs, Ms (2015)</td>
</tr>
<tr>
<td>5XG-013</td>
<td>12-306-550p</td>
<td>Yellow Pine DH</td>
<td>Alaskite w/ breccia</td>
<td>0.523 ppm</td>
<td>Yes</td>
<td>Py (2016)</td>
</tr>
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<td>6RJ-016</td>
<td>12-314-441.7a</td>
<td>N. Yellow Pine DH</td>
<td>Quartz monzonite</td>
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<td>Py, Apy, As-Fe-S Unk, Kfs, Sch (2015)</td>
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<td>12-314-446.9</td>
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<td>Kfs, Pl, Ms (2016)</td>
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from core (grain’s center) to rim of the grain to look for compositional zoning within the pyrites.

Pyrite exhibits considerable compositional variation and complex zoning textures. Each sample is described separately to document the heterogeneity; a summary discussion follows the descriptions. Stibnite and scheelite are close to stoichiometric in composition. Arsenopyrite is nearly stoichiometric as well, though the few samples that were analyzed contained slightly greater S, less As, and trace Sb, in comparison to ideal arsenopyrite. Metallurgical work commissioned by Midas Gold also found that the arsenopyrite is deficient in As (Christopher Dail, oral commun., 2017). Three samples contain an unknown Fe-As sulfide. Analyses of the unknown were reproducible, suggesting that the analyses are not a mixture or analytical error. The unknown phase appears to be an early metastable mineral with a slightly variable, non-stoichiometric composition of approximately 35 percent Fe, 24 percent S, and 41 percent As, thus intermediate in composition between arsenopyrite and arsenian pyrite. The analysis of a typical unknown point in section 4LO-023 returned a calculated formula of $\text{Fe}_{0.72} \text{As}_{0.4} \text{S}$, somewhat analogous to an arsenian pyrrhotite. The mode of occurrence of the unknown Fe-As sulfide is compatible with the unknown mineral being precipitated early, possibly pre-ore, and at a higher temperature than the more abundant As-rich pyrite. Alternatively, the unknown phase could represent an early (Cretaceous?) pyrrhotite that underwent later sulfidation and reaction. The zone or subzone visible in BSE mode, and photographed in the images reproduced herein, is labelled for most points analyzed by the microprobe, and it was used to define the pyrite paragenesis as the grains grew from the core out to later rim deposition in an ideal case.

Spongy or moth-eaten textures are defined as pyrite with multiple, irregularly shaped small holes that have other minerals filling the holes. Sulfide intergrowths in several samples show elongate branching forms in thin section; these elongate growths were originally called “skeletal” by the senior author because they resemble a backbone with vertebrae protruding. However, a more common and appropriate mineralogic term would be dendritic, as the crystal intergrowths also resemble branching, treelike forms. True “skeletal” crystal habits show hollow depressions where the crystal faces are, due to more rapid growth of ions along the edges and corners of growing crystals. In thin section, the two dimensional view can show only the dendritic form. Dendritic crystals are thought to develop during conditions of high supersaturation and rapid growth. Common examples are frost crystals and Mn oxides on fractures (Hibbard, 2002). Botryoidal pyrite is rare and forms rounded layers draped over earlier pyrite.

The following description of pyrite samples from the Stibnite district is arranged from least mineralized zoning in some crystals. However, limitations in time and probe resolution did not allow determination of the compositional variation of these micron-scale lamellae. Pyrite also exhibits variations in crystal habit, including cubic, pyritohedral, dendritic or skeletal, and botryoidal forms. Also evident are poikilitic intergrowths of early pyrite with metamorphic minerals, and in the ore zones pyrite displays multiple resorption (also known as dissolution) and regrowth textures. Pyrite from the Yellow Pine deposit is the most complex.
samples to the most mineralized and it is subdivided into the three main deposits. Discussion of the overall observations is at the end of the chapter. Appendices G-1 and G-2 contains analytical results for pyrite analyzed by electron microprobe. For purposes of discussion, the samples are separated according to their intensity of Au mineralization, whether identified from Midas assays or inferred from location, appearance, and sulfide content. “Barren” samples are those with non-detectable to background (<50 ppb) Au contents. Weakly mineralized rocks would be those with Au contents under approximately 0.3 ppm. Mineralized rocks at Stibnite are those with assays typically of 1 ppm Au or greater, i.e. “ore grade.” No samples of pyrite in un-mineralized igneous rocks were analyzed. Mineral compositions are reported in weight percent of the element or oxide, and “percent” in the text will refer to weight percent.

Pyrite from Barren and Very Weakly Mineralized Metasedimentary Rocks

Some calc-silicate metasedimentary rocks contain coarse-grained, bright pyrite visible in hand specimen. In polished section, those pyrites are unzoned or simply zoned and homogeneous in composition. Two samples (8EB-008, 8EB-009) of pyrite-bearing metasedimentary rocks from outside the resource area were probed (Table 6-1). The pyrite is intergrown with fresh amphibole as shown in Figure 2-8. These pyrites contained neither detectable As nor detectable Au, Cu, Sb, or Ag. They did contain as much as 2 percent Co and trace Ni (Appendix G-2). The Co and Ni appear to be indicators of “sedimentary” or “metamorphic” pyrite.

Sample 4LO-013 of siltstone from the bottom of the Stibnite pit contained euhedral to anhedral pyrite with trace Co and Ni, but also trace As, reflecting its location near mineralization.

Pyrite and pyrrhotite were both present in section 5SZ-003 of unmineralized siltstone from 631.6 ft depth in drill hole MGI 10-37 below the Stibnite pit in the West End area (Figure 3-8; Appendix B). The pyrrhotite shows a pitted texture in contrast to the bright, solid pyrite that contains trace Co (Figures 6-1 and 6-2). Neither sulfide mineral contains As (Appendix G-1).

WEST END DEPOSIT

Pyrite from Mineralized Intrusive Rocks

Two sections (5XG-006 and 6RJ-008) from core in MGI 12-305 (Figures 3-10 and 3-12) of weakly mineralized, but strongly potassically altered, granite from the West End deposit were analyzed by microprobe. Both sample intervals assayed only 0.2 ppm Au; the granite is cut by pale pink carbonate-quartz veins with local pyrite. Figure 6-3 shows an image of Pyrite A in section 6RJ-008 (12-305-433.7). Cores contain 4 to 5 percent As with rims of less than 1 percent (Appendix G-1). No other trace elements were detected other than a single outer rim point with detectable Au but low totals in 6RJ-008. Thin laminae are present in the rims of some pyrites in the sample.

Sample 5XG-006 (12-305-422.1) is from an intensely potassium feldspar flooded granite cut by a quartz-carbonate-potassium feldspar-pyrite vein. The vein potassium feldspar was dated at 51.0 Ma. Some pyrites sit adjacent to and apparently growing on corroded rutile grains with cooling ages of 78 Ma on the rutile cores (Figure 4-3). Figure 6-4 shows a photo of a wall-rock pyrite from this sample. This pyrite has detectable Au in the middle portion which was also the zone highest in As (3-5 percent). Three tiny homogeneous-appearing, pyritohedral pyrites in the vein contain <1 percent As and consistent trace Cu without any detectable Au (Appendix G-2).

Sample 4VF-002 (MGI 10-37 at 120.8 ft) is from the mineralized Stibnite stock in an interval that assayed 1.6 ppm Au. Chalcedonic quartz-carbonate-pyrite veins cut the granite, which contains both potassium feldspar alteration and local muscovite. The sample contains several well-zoned, very fine grained pyrites, one of which contains detectable gold and is shown in Figure 6-5. A graph of chemical zoning along a transect is shown in Figure 6-6. The As content peaks in the middle zone which is characterized by prominent lamellar banding. Gold does not correlate exactly with As content and appears to be highest both before and after the highest As peak. Trace Cu (<0.1 percent) is ubiquitous in the pyrites in this sample (Appendix G-2).
Figure 6-1. Pyrrhotite grain with pitted texture from section 5SZ-003 (MGI 10-37-631.6). No trace elements were detected.

Figure 6-2. Pyrite C with light and dark zones that do not contain As or Au and are nearly stoichiometric. From section 5SZ-003 (MGI 10-37-631.6). Trace (0.1 to 0.4 percent Co) is present, although the probe totals were low (about 97 percent) in the darker pyrite C.
Figure 6-3. BSE image of Pyrite A, section 6RJ-008 from MGI 12-305-433.7. Lighter cores contain 4 to 5 percent As but no other trace elements. The darker thick rim zone with the probed points in red contains less than 1 percent As.

Figure 6-4. Pyrite 1 growing on corroded rutile rind in section 5XG-006 at Site 1 (12-305-422.1). Core points 300 and 301 had less than 0.5 percent As with trace Co and Ni. Points 303 and 304 had detectable Au and 3 to 5 percent As. Points 305 and 306 in the solid rim are less than 0.5 percent As.
Figure 6-5. Pyrite 1 at Site C in and near a vein in the Stibnite stock (section 4VF-002 at 10-37-120.8). The granitic rock is cut by quartz-carbonate veins containing pyrite, one of which is shown here. Arsenic is highest in the middle zone of transect as shown in Figure 6-6 below.

Figure 6-6. Compositional zoning along core-rim transect in Pyrite 1, Site C, 4VF-002, shown above.
Pyrite in Mineralized Metasedimentary Rocks

Sample 8EB-004 (15VG016) is a veined, sandy carbonate unit from the West End pit (Table 6-1). The section contains larger pyrites which are 200 to 400 microns across with complex core-middle-rim geometries as shown in Figure 6-7. The larger cores are irregular and contain trace Co but are barren of As (Appendix G-2). These large and complex pyrites show a resorption or solution boundary from the core to the middle zone. Smaller pyrites 10-20 microns in size with discernible cores and rims are also present. Arsenic contents of the smaller pyrites are in the 1.5 to 3.5 percent range with only weak zoning from higher in the cores to lower in the rims. Smaller pyrites appear to lack the early resorbed cores that carry the “sedimentary” signature of trace Co and Ni without As.

Sample 4VF-006 (10-37-575.6) is of fractured sandy calcareous marble to calcareous siltite from a well-mineralized (10 ppm Au and 8600 ppm As) interval in drill core underneath the Stibnite pit (Figure 3-8; Appendix B-2). Figure 3-29 shows a hand sample of the high grade core interval just uphole from the section analyzed. Calc-silicate metamorphic minerals are retrograded and the rock is cut by carbonate veins. A number of pyrites exhibit complex textures and zoning in As content, but Au was not detected in the pyrites from this section (Appendix G-1). Figure 6-8 shows an example. In general, the cores contain arsenian pyrite with 7 to 10 percent As, and in one case, the unknown corroded phase intermediate between Fe sulfide and arsenopyrite. The unknown mineral is somewhat similar to non-stoichiometric, As-rich pyrrhotite and consistently is early in the paragenesis. Cubic rims typically contain <1 percent As.

Figure 6-7. Pyrite 1, Transect 1, Site A, Section 8EB-004 from West End pit (15VG016). This pyrite shows complex growth history with irregular zoning including a corroded core, laminated middle zone, and outer euhedral rim. Points 326-328 in core have trace Co up to 2.5 percent and no As. Points 329-333 contains 1.7 to 2.5 percent As but no Co. Bright diamond-shaped grains at top of photo are arsenopyrite.
Sample 4VF-005 (10-37-559.5) is from a quartz-carbonate veined calc-silicate metasedimentary horizon above 4VF-006 (Appendix B-2). The interval contained 2.24 ppm Au. Pyrite C, one of the coarser pyrites analyzed, contained up to 0.03 to 0.04 percent Au and trace Cu localized in the intermediate banded zone (Figures 6-9 and 6-10). Cores of pyrite C were stoichiometric pyrite with trace Co. The boundary of the core is extremely irregular in shape and resorbed with numerous pits marking the dissolution boundary to the core. The core is mantled by the lighter banded zone of arsenian pyrite with 5 to 6 percent As and minor Cu and Au. The uniform, subhedral rim has 1.2 to 2.7 percent As and trace Cu with one of four points containing detectable Au (Appendix G-1). The section also contains many small, disseminated skeletal to dendritic pyrite grains surrounded by stoichiometric arsenopyrite, as seen in Figure 6-10. The skeletal pyrite cores contain trace Sb but no detectable Au.

Two samples of mineralized, brecciated metasedimentary rocks in drill hole MGI 12-305 (Figures 3-8 and 3-10) from the West End deposit were probed. Section 6RJ-010 from 456 ft in 12-305 is in an interval (453 to 460.5 ft) that assayed 6.8 ppm Au with 5700 ppm As. The rock is a silicified breccia containing abundant pyrite and arsenopyrite. Section 5XG-008 is from 462 ft in 12-305, and the interval 460.5 to 468 ft assayed 3.4 ppm Au with strong potassic alteration in the breccia clasts (Figure 3-28). Potassium feldspar from this zone was dated at 50.8 Ma.

Figure 6-11 shows pyrite A in 6RJ-010. Pyrite B exhibits a similar texture. Darker cores are stoichiometric pyrite with nil to very low trace elements. The lighter zones in the BSE images are arsenian pyrite with 1 to 4 percent As and 0.03 to 0.08 percent Cu (Appendix G-1). The lighter material mantles the resorbed cores and appears to alter the cores along healed cracks. A single point in the light pyrite had detectable Au and all had trace Cu. There is a sharp outer discontinuity with low As (1.3 percent) subhedral pyrite in the rim zone. The rims are unconformable on the middle zones. Based on a semi-quantitative SEM-EDS analysis, the bright sulfide that veins and fills interstices within the broken pyrite is tetrahedrite-tennantite. Two bright grains outside of the pyrite are arsenopyrite and chalcopyrite, also based on SEM analysis. Midas also noted fracture-fillings of late...
Figure 6-9. Pyrite C in Section 4VF-005 (10-37-559.5) with core, intermediate light-banded zone, and subhedral rims. Bright diamond-shaped crystals on the perimeter are arsenopyrite. Gold is detectable in the As-rich, light-banded zone (LB points).

Figure 6-10. Larger view of section 4VF-005 from below Stibnite pit (10-37-559.5). Pyrite C is surrounded by multiple, fine-grained skeletal pyrite rimmed with arsenopyrite.
Three of the four pyrites probed in section 5XG-008 (12-305-462) have detectable Au with pyrite A being particularly enriched. Figure 6-12 shows pyrite D, a large complex grain with three clearly identifiable zones. The pockmarked, dark cores are nearly stoichiometric pyrite with about 0.5 percent As and some Cu (Appendix G-1). The cores are resorbed and cracked with a lighter middle band of high As (4 to 5.5 percent) pyrite and outer darker rims of low As (1 to 2 percent) pyrite. Only one point, B4, in the weakly laminated middle bands of pyrite D has detectable Au. Pyrite C (not pictured), another large grain in section 5XG-008, is about 400 microns across with a large homogeneous core of stoichiometric pyrite containing 1 percent Co locally. A thin light-colored and weakly laminated middle zone of high As (2 to 5 percent) pyrite is rimmed by euhedral low As pyrite. A small pyrite (Pyrite A, 5XG-008) with three zones also has detectable Au in the middle zone.

YELLOW PINE DEPOSIT

Yellow Pine Pyrite from Weakly Mineralized Rocks

The least mineralized intrusive rock probed was section 6RJ-016 from MGI 12-314 (Figures 3-7 and Figure 3-12, Appendix B) at 441.7 ft depth in an interval assaying 0.118 ppm Au and 1890 ppm As. The rock is altered quartz monzonite cut by a thin composite carbonate veinlet. The younger veinlet is quartz-carbonate with trace scheelite. As illustrated in the chemical maps of the sulfidized mica in Figures 3-14 and 3-15, the pyrite partially replaced igneous biotite in textures that initially mimic the mica structure. However, pyrite also nucleated outside of the mica with complex zoning revealed by the chemical maps in Figures 6-13 and 6-14. These images of a cluster of intergrown, subhedral, pyritohedral pyrites show complexly zoned grains.
The As map in Figure 6-14 reveals stoichiometric pyrite cores containing almost no As (<1 percent As) and displaying multiple dissolution textures and healed fractures. Filling in the cracks and deposited on the core is a middle zone of arsenian pyrite with 2 to 8 percent As plus trace Cu or Sb. The outermost rim pyrite zone contains less As (2 to 6 percent As), possibly reflecting depletion of arsenic following the precipitation of the high-As band (yellow and orange color) at the middle-rim zone boundary (Figures 6-13 and 6-14). In this and other samples, arsenopyrite in tiny, euhedral, diamond-shaped crystals typically rims the pyrites, particularly the skeletal pyrites. Typically the pyrite rims contain lower As than the pyrite middle zones. This could indicate saturation and precipitation of arsenopyrite, which depleted the fluid of As. Arsenopyrite crystals typically abut the earlier pyrite crystals, providing textural evidence of the paragenetic sequence.

Yellow Pine Pyrite from Mineralized Rock and Ore

Three surface samples (4LO-023, 4LO-022, and 8EB-014) from the western bench of the Yellow Pine pit were analyzed with the electron microprobe (Table 6-1). All samples are strongly altered and Fe-stained quartz monzonite. The Yellow Pine samples are strongly mineralized with disseminated sulfides and display potassic alteration, local muscovite, and cross-cutting stibnite veins visible in hand sample (Figures 3-33 and 3-37). All three samples contained pyrite with detectable Au in the microprobe analyses and displayed complex textures of sulfide growth.

Section 4LO-023 (12VG159F) revealed spectacular elongate and branching, platy to dendritic or skeletal sulfide growths with cores of the unknown Fe-As-S sulfide phase containing approximately 39 percent Fe, 31 percent S, and 30.5 percent As (Figure 6-15). This unique habit may be due to the sulfide initially being precipitated along cleavages in altered biotite sites. Alternatively, the branching habit resembles dendritic crystals in slags and spinifex texture in olivines. By analogy, the dendritic sulfide growths could be due to rapid crystallization and cooling of the hydrothermal fluid. One way to achieve such conditions would be faulting, depressurization, and influx of cooler water into the system. Euhedral, terminated arsenopyrite
Figure 6-13. BSE image of pyrite in section 6RJ-016 from MGI 12-314 at 441.7 ft, north Yellow Pine area.

Figure 6-14. Arsenic map of pyrite in section 6RJ-016. Warm colors indicate higher As contents. Pyrite is from MGI-12-314 at 441.7 ft in the north Yellow Pine area.
Figure 6-15. Branching dendritic to skeletal sulfide intergrowths at Site A in section 4LO-023 from the Yellow Pine pit. Elongate cores of light gray color are unknown Fe-As-S phase, rimmed by pointy crystals of arsenopyrite (whitish gray) surrounded by dark botryoidal-like bands of Au-bearing pyrite. Very bright, younger sulfide is stibnite. Figure 6-16 shows an enlarged view of a similar intergrowth.
Table 6-2. Summary of 4LO-023 sulfide analyses in weight percent. Full analyses are in Appendix G-1. Values are averages for the phase: A-1 Dark pyrite is points D-1 to D-4. A-1 Light Unknown is points L-1 to L-6 as marked in Figure 6-16. BDL is below detection limit. The detection limit for Au in this run was about 0.005 percent. Cobalt is sporadic, but averaged at or close to detection limit (DL) of 0.02 percent.

<table>
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<th>Site/Phase</th>
<th>Fe</th>
<th>S</th>
<th>As</th>
<th>Au</th>
<th>Co</th>
<th>Cu</th>
<th>Sb</th>
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<td>&lt;DL</td>
<td>0.02</td>
<td>0.04</td>
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<td>0.03</td>
<td>0.30</td>
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crystals have nucleated on the outside of the unknown sulfide phase. Locally, very dark, high-As, Au-bearing pyrite precipitated in botryoidal bands or layers, only a few microns thick, that overlay the unknown mineral (Figures 6-15 and 6-16). The Au-bearing, arsenian pyrite bands appear to be roughly contemporaneous with the arsenopyrite crystals that occupy a similar paragenetic location. Figure 6-16 shows an enlargement of Site A-1 in 4LO-023 with probed points marked. Table 6-2 summarizes microprobe analyses from an analytical run on sulfides in section 4LO-023 (Appendix G-1). Site A-2 also has late stibnite that precipitated around the spherical, dark gray arsenian pyrite which contains 0.04 percent Au (Table 6-2). Other enigmatic textures in sample 4LO-023 include a round corroded pyrite-arsenopyrite composite grain (Site B, not pictured) coated by muscovite flakes intergrown with rimming arsenopyrite that contains sporadic trace Au and Te (averaging 0.22 percent; Appendix G-1, G-3, and G-5), according to the microprobe analysis (4LO-023 Site B pyrites in Table 6-2; Appendix G-1). One hypothesis is that the muscovite plates grew during cooling, which may also have led to the complex corrosion in the pyrite.

Section 4LO-022 (12VG159D) is altered quartz monzonite cut by stibnite veins (Figure 3-37). The carbonate-pyrite vein is cross cut by a later stibnite breccia vein (Figures 3-35 and 3-36). BSE images reveal complex textures and multiple zones in the tiny “buckshot” pyrites in the vein (Figures 6-17 and 6-18). These pyrites contain detectable Au. A transect across Pyrite 1 at Site B in 4LO-022 displays a regularly and sharply laminated core with 2 to 4 percent As, 0.039 to 0.063 percent Au, and minor Cu (Figures 6-18 and 6-19). The analytical resolution is insufficient to fully determine the compositions of the very fine light and dark laminae, which are only about a micron thick in Pyrite 1. One interpretation is the fluids initially precipitated light-colored pyrite laminae, perhaps due to higher As contents in the fluids, and as that fluid was depleted of As, the dark pyrite (lower As) laminae precipitated with a sharp boundary against the next outer lamina. The proposed diffusion-related process is somewhat similar to liesengang Fe banding but with slow diffusion in the hydrothermal fluid. Alternatively, the laminae could reflect periodic influxes of new fluid or changes in the physical conditions. The Pyrite 1 core is surrounded by a thicker, pyritohedral rim zone with diffuse and poorly defined bands exhibiting local dissolution and incorporation of earlier pyrite pieces. The inner rim is lighter colored and highly enriched in As (to >9 percent), while the outermost rim is darker. The outer rim does not contain Au and has less Cu than the core, but the rim is slightly enriched in Mn relative to the rest of the crystal (Figure 6-19; Appendix G). Pyrite 2 at site B also had the highest Au and As in the core and significant amounts in the middle zone (Appendix G-2). Pyrite at sites C and D in the section also showed strongly laminated cores of Au-bearing arsenian pyrite and less-developed rims chemically and visually (Appendix G-2).

A third sample, 8EB-014 from the west wall of the Yellow Pine pit, was also analyzed with the electron microprobe. Sample 8EB-014 is strongly potassically altered quartz monzonite cut by a quartz-carbonate-stibnite vein (Appendices B and C). Figure 6-20 shows an image of Pyrite 1 at Site C in the vein. This pyrite...
Figure 6-16. Site A-1 in section 4LO-023 from the Yellow Pine pit (12VG159F). Points L-1 to L-6 are light gray unknown phase (Appendix G-1); points R-1 to R-6 are arsenopyrite rim analyses (Appendix G-1); points D-1 to D-4 are dark As-rich pyrite with Au and trace Cu and Ag (Appendix G-1); Sb points are stibnite.

Figure 6-17. Zoned pyrites in carbonate vein at Site B, section 4LO-022 from Yellow Pine pit.
Figure 6-18. Section 4LO-022, Site B, Pyrite 1 with marked core to rim transect.

Figure 6-19. Chemical zoning of 4LO-022 Site B Pyrite 1 along core to rim transect.
Figure 6-20. BSE image of Pyrite 1, Site C in section 8EB-014, with core to rim traverse marked.

Figure 6-21. Zoning of Pyrite 1 at Site C, section 8EB-014, Yellow Pine pit.
is one of a cluster of well-laminated pyrites which are disseminated in a quartz-carbonate vein. Here, Au is not present in the high As core, but peaks mid-way to 0.097 percent, coincident with As peaking at >10 percent (Figure 6-21). As evident in the analytical data in Appendix G and Figure 6-21, the two highest Au values are coincident with the two highest As values, though there is not a perfect numerical correlation of Au and As in this sample or in general. Trace Cu and Sb are highest in the outer half of the grain. Pyrite 2 in the cluster at Site C shows a similar relationship with laminated cores of 6 to 10 percent As, but no Au, and thick banded rims containing 2 to 5 percent As with Au contents of 0.02 to 0.06 percent and trace Cu and Sb (Appendix G-2). Other vein pyrites at sites D and E showed coarse zoning of lighter As-rich cores with sporadic Sb out to darker rims with trace Cu, but none of the points had detectable Au. A large, “dog-shaped” pyrite in the wall rock at Site A contained no detectable Au but did contain As (Figure 6-22). This pyrite shows a pitted core without laminations and a cubic pyrite inclusion, evidence of a complex precipitation history.

Two samples from drill hole MGI 12-306 (Figures 3-7 and 3-9) under the Yellow Pine pit were probed (Table 6-1). Sample 5XG-013 at 550 ft is brecciated alaskite (Klg) in an interval with 0.523 ppm Au (Appendix B-5). The breccia matrix is composed of quartz-carbonate-sulfides, and the wall rock contains sericitized plagioclase that has been partially altered to potassium feldspar. Two types of pyrites were noted in reflected light.

At Site A in 5XG-013, an aggregate of finely intergrown arsenopyrite and rounded pyrites contained Au (0.01 to 0.03 percent) and trace Cu in the pyrites (Figure 6-23). Figure 6-24 shows a photo of an individual pyrite (Site A, Pyrite 1) with an ovoid-shaped, banded core surrounded by a dark distorted cubic rim. Though no

![Figure 6-22. “Running dog” pyrite from 8EB-014 Site A in wall rock. Dark pitted core contains about 1 percent As and 0.03 percent Co, while thin light band has 3 to 5 percent As and 0.04 percent Co. The brighter cubic pyrite inclusion contains 4 to 7 percent As and trace Sb.](image-url)
Au was detected in Pyrite 1 at Site A, the light bands contain up to 7 percent As and 0.1 percent Cu. The dark core has <1 percent As and the very dark rims have no detectable As and trace Cu. Pyrites from Site C in 5XG-013 show botryoidal light and dark bands surrounding clusters of pyrite with irregular pitted cores. Arsenic is higher in the light-colored bands, but Cu, Sb, and Ag are consistently higher in the dark bands (Appendix G-2).

Sample 5XG-011 from MGI 12-306 (Figures 3-7, 3-9 and 3-18) at 425 ft is of altered quartz monzonite from an interval assaying 3.7 ppm Au with 0.12 percent Sb. Potassic alteration is intense next to the quartz-sulfide-carbonate vein. Coarse-grained mica is more abundant than in many samples and the mica is cut by the mineralized vein. Several of the pyrites probed have detectable Au. The Au-bearing pyrites at Site A are tiny with diameters of 20 to 50 microns (Figure 6-25). Rounded outlines of the grains exhibit resorption and pitting on the exterior and particularly along an interior boundary from lighter colored cores to dark mid-zones. The outermost, lighter rim locally shows pyritohedral outlines. Three transects from core to rim show Au contents of 0.02 to 0.06 percent across the width of the grains. Arsenic varies by approximately a factor of two with lower amounts in the core rising to 8 to 9 percent in the lighter middle band, then falling in the rim (Figures 6-25 and 6-26; Appendix G-1). Trace amounts of Cu, Co, Mn, and Ag are also present. These auriferous pyrites are characterized by well-defined narrow laminae throughout the grains but especially in the middle zones. Pyrite at Site A, Transect 3a, in 5XG-011 also contains Au-bearing, light-colored, As-enriched bands, but the grains are broken, possibly due to late hydrothermal activity.

**HANGAR FLATS DEPOSIT**

Only one pyrite sample (6RJ-006, MGI 12-193-930.3) from the Hangar Flats deposit (Table 6-1, Figure 3-6, Appendix B) was analyzed with the electron microprobe. Sparse disseminated pyrite is present in brecciated and strongly mineralized (Au, W, Sb) host rock largely comprised of granitic rocks but with local clasts of marble. As no metasedimentary rocks have been described in the Hangar Flats area, the marble could be pieces offset by faulting and transported laterally or vertically from the roof pendant. Zoning is similar to pyrites from the other deposits with a stoichiometric pyrite core, surrounded by and veined by an As-rich middle zone with laminae and local dissolution pockets (Figure 6-27). The euhedral rim contains As but in lower concentrations than the middle zone (Figure 6-28). Gold was not detected in the analyses, but trace Cu (<0.11 percent) is present in the light-colored bands and rim zones of the pyrite (Appendix G-1). Stibnite surrounds and is clearly later than the pyrite (Figure 6-27).
Figure 6-25. Pyrite at Site A in section 5XG-011 (12-306-425) from Yellow Pine deposit, with transects T1 and T2 marked. Note irregular shaped to corroded fine grains with prominent lamellae.

Figure 6-26. Section 5XG-011 Site A Pyrite, Transect 2 with As, Au and Cu contents from core to rim. Transect location is shown in Figure 6-25.
Figure 6-27. BSE image of pyrite in section 6RJ-006 (12-193-930.3) from the Hangar Flats deposit. White mineral surrounding the pyrite is stibnite. Non-sulfides appear black.

Figure 6-28. Arsenic map of pyrite in section 6RJ-006; warmer colors indicate higher As contents. Core area (black in image) had no detectable As.
DISCUSSION OF PYRITE CHEMISTRY AND TEXTURES

Based on the new age determinations of alteration minerals and scheelite (summarized in Table 4-3 and Figure 4-10), the three different Au deposits (Yellow Pine, West End, and Hangar Flats) are different in their timing and character of mineralization. In spite of the three geographically and temporally distinct Au deposits in the Stibnite district, as well as the heterogeneous host rocks, there is a consistent pattern in appearance and composition of pyrite in the deposits. An idealized growth sequence through time is proposed in Table 6-3. Figure 6-29 shows a sketch of a simplified, idealized pyrite grain found in the deposits at Stibnite.

Table 6-3. Sulfide paragenesis model for the Stibnite district, Idaho.

| Early Fe sulfide, oxide, or Fe-bearing silicate. Trace Co, Ni in sulfide. | Igneous oxides (i.e. magnetite, rutile). Igneous biotite and muscovite. |
| Metamorphism: Growth of coarse pyrite with trace Co and Ni, and local magnetite, pyrrhotite especially with iron-bearing calc-silicates or micas. No As or Au. Sparse quartz-Mo veins and fracture coats in skarn. | Cretaceous alteration: Minimal or weak; possible disseminated pyrite. |
| Uplift ~~~~~~~~~~~~~~~~~~~~~~~~~~~~ | Uplift ~~~~~~~~~~~~~~~~~~~~~~~~~~~~ |
| Resorption of metamorphic pyrite cores. | Corrosion of rutile and any early sulfides. |
| Early stage/pre-ore: New pyrite precipitates on earlier resorbed sulfide grains. Cements earlier pyrites. | Early stage/pre-ore: Disseminated, stoichiometric pyrite; sulfidation of biotite to form dendritic to skeletal sulfides; unknown Fe-As-S phase. New pyrite nucleates on corroded rutile rims and Fe-bearing micas. |
| Main As-Au ore stage: Subhedral, As-bearing pyrite deposited in bands around old cores and locally nucleating new pyrite grains, especially in veins and disseminations in permeable sediments. Pyrite continues to replace earlier Fe-bearing silicate phases. | Main As-Au ore stage: Arsenic-bearing pyrite deposited in bands around old dendritic to skeletal cores and locally nucleates new pyrite grains, especially in carbonate veins and adjacent wall rocks. Arsenopyrite deposition in crystals that form overgrowths. |
| Dissolution and pitting. | Local grain dissolution and pitting. |
| ~~~~~~~~~~~~~~~~~~~~~~~~~~~~ | ~~~~~~~~~~~~~~~~~~~~~~~~~~~~ |
| Main/late stage: Deposition of outer, uniform rim pyrite in cubic or pyritohedral crystal habit. Decreased concentration of As in crystal as arsenopyrite precipitates. Outermost rims are low in As. Pyrites incorporate trace Cu, Sb, and other elements. Late tetrahedrite veinlets. | Main/late stage: Deposition of outer, rim pyrite zone in cubic or pyritohedral crystal habit. Decreased concentration of As in crystal as arsenopyrite precipitates. Outermost rims are low in As. Pyrites incorporate trace Cu, Sb, and other elements. |
| Tungsten and antimony stage: Scheelite deposition. Stibnite infills around earlier sulfides (pyrite, arsenopyrite). | Tungsten and antimony stage: Scheelite veins? Stibnite crosscuts and infills around earlier sulfides (pyrite, arsenopyrite). |

Despite the variations of sulfide textures and compositions observed at Stibnite, there are a few commonalities:

- Pre-ore and barren pyrite, whether of sedimentary or metamorphic origin, does not contain As but may contain trace amounts of Co and Ni. Barren pyrite is typically coarser grained and may nucleate later ore-bearing pyrite.
- Trace elements in pyrite associated with the Au-Sb-W mineralization include Sb, Cu, Ag, and Mn and major-element amounts of As. Ore-stage pyrite is very fine grained, typically less than 50 microns in diameter, and locally displays oscillatory zoning (i.e. the alternating light and dark laminae).
Gold mineralization is clearly associated with As-bearing pyrite, and bulk geochemical analyses district-wide show the excellent correlation of As and Au. However, traverses across single, zoned pyrite grains suggest slight differences in timing. The highest As-in-pyrite zones (5-10 percent As) do not always contain the greatest amount of Au in the pyrite. An interpretation from this study is that As was deposited first in arsenian pyrite. But as time went on, arsenopyrite became saturated and precipitated, with the available As partitioning into arsenopyrite over pyrite. That shift to lower As contents in the hydrothermal fluid changed the pyrite composition to a lower As content (2-5 percent As) and may have promoted saturation and precipitation of gold in the arsenian pyrite lattice. Temperature and kinetics would also have influenced precipitation and localization of Au in this complex and changing environment.

The early-stage mineralization that formed the As-rich, disseminated Au is related to sulfidation of biotite. A S-bearing (probably H₂S) hydrothermal fluid altered the biotite and released Fe which helped precipitate Fe-bearing sulfides. Some of the dendritic to elongate skeletal crystal intergrowths may have started as replacements of the igneous biotite, but the dendritic and branching habit may also reflect rapid crystal growth and rapid cooling, particularly where evidence for silicate replacement is lacking.

The ubiquitous presence of zones of corrosion, resorption, and discontinuities in growth zones of mineralized pyrite requires a dynamic fluid environment that changed composition over time. The major pyrite zones illustrated in Figure 6-29 are interpreted to broadly correlate with the pre-ore or early stage, main stage and late stage of Au mineralization, all of which may have taken place in pulses. The zones of etching correspond to the millions of years between the mineralized pulses. The late Sb mineralization is largely post-pyrite at any one location, at least in the samples studied.
CHAPTER 7

DISCUSSION AND CONCLUSIONS
INTRODUCTION

Results of this investigation on geology and Au-Sb-W mineralization in the Stibnite district, Idaho, answer some questions but raise many others. The discussion is organized by topics and it summarizes the main conclusions and questions of this study. Additional information from the literature is presented briefly to fill in aspects of the geology at Stibnite that were not addressed in the current investigation but bear on the conclusions presented herein.

CRETACEOUS INTRUSIVE ROCKS AND TECTONICS

Dating of alteration and mineralization at Stibnite has clearly established the post-Cretaceous age of the mineralization. Although Au mineralization is not related to Cretaceous magmatism, Mesozoic tectonics and the characteristics of the Cretaceous granitoids had important roles in localizing mineralization, particularly in the Yellow Pine deposit.

In summary, Mesozoic events influenced the Au-Sb-W mineralization at Stibnite by generating:

• Chemically reactive, biotite-bearing, granodiorite host rock.

• Iron-oxide and Fe-sulfide minerals in calc-silicate metasedimentary rocks and small, localized skarn pods, that provided chemically reactive nuclei for later ore-bearing pyrite.

• Multiple intrusions with contrasting physical and chemical characteristics, including brittle alaskite bodies that fractured easily.

• Preservation of a roof pendant of deformed and faulted impure metasedimentary rocks.

• Crustal thickening that positioned Precambrian crust into the deep thermal regime conducive to magma genesis and later hydrothermal fluid flow.

• Incipient structural weaknesses that resulted in both northeast-trending faults from Mesozoic compression and north-south faults which are subparallel to the old rifted margin and Salmon River suture zone (Figures 2-1 and 2-2; Tikoff and others, 2017; Lewis and others, 2012a).

Reactivation of those north-south structures in the late Tertiary to Quaternary developed linear north-south oriented valleys that are important transportation routes to Stibnite and other parts of west-central Idaho.

• Protracted crustal stress and strain over tens of millions of years.

Gaschnig and others (2010, 2011, 2013, and 2017) provide comprehensive descriptions of the Idaho batholith, its ages, and its isotopic signatures, as well as data on the spatially overlapping, but younger Challis magmatic event. The town of Yellow Pine and the Stibnite district lie within the Atlanta lobe, the larger and southern of two lobes of the relatively homogeneous batholith that occupies a very large area in central Idaho (Lewis and others, 2012a). The intrusive rock descriptions and ages presented here and in Stewart and others (2016) contribute to the overall characterization of the batholith. The analyzed intrusive rocks at Stibnite fall generally within the compositional field of the Idaho batholith and are not unique in chemistry (Figure 2-19).

Ages of the multiple phases of the Idaho batholith at Stibnite range from approximately 94 Ma for the porphyritic, biotite-rich granodiorite to 83.6 Ma for the alaskite, with at least one other phase, granite, in between at about 85.5 Ma. A cryptic 89 Ma phase is indicated by ages of inherited zircon cores in the younger plutons. The presence of two-mica granite and the garnets in the alaskite suggest that those units are peraluminous, although the earlier biotite granodiorite is metaluminous. All zircons dated have large Precambrian cores typical of the Idaho batholith and provide evidence of incorporation of metasedimentary rocks from the continental substrate into the batholith magmas. Petrographic characteristics and the few geochemical analyses available suggest that the granodiorite overlaps with the early metaluminous suite, and the granite and alaskite are part of the Atlanta peraluminous suite (Figure 2-19). Gaschnig and others (2011) interpreted the more biotite-rich metaluminous suite as a hybrid mix of crust and mantle melts, whereas the later peraluminous rocks are derived from crustal melting. Both suites post-date the convergence and accretion of the Blue Mountains and other terranes onto western North America along the Salmon River suture. Different tectonic models, including a subduction model, a terrane collision model, and a terrane accretion model, have been proposed for the formation of the batholith (Gaschnig and others, 2011; Tikoff and others, 2017). Zircon core ages and
Pb isotopic data support the interpretation of significant crustal melting involved in magma genesis from 83 to 94 Ma in central Idaho, though the exact tectonic model involved is speculative. Ironically, the age of mineralization at Stibnite most closely overlaps the age of the Bitterroot lobe in northern Idaho, not the age of the Atlanta lobe where the district is situated. Gaschnig and others (2011) determined a 66 to 53 Ma age range of the Bitterroot lobe and suggested it was a crustal melt derived by partial melting of largely Paleoproterozoic basement gneisses but with some Archean and Belt sedimentary components.

Mineralogically, the early biotite-rich Cretaceous granodiorite provided an excellent host rock with mineral and chemical reactivity that facilitated sulfide precipitation. The number of intrusions and the complex interweaving with deformed roof pendant rocks provided variations in chemistry and in physical properties that facilitated subsequent fracturing due to tectonic strain or changes in fluid pressure.

At Stibnite, the batholith phases clearly postdate folding and thrust faulting, as shown in Stewart and others (2016). The importance of the northeast-trending faults (e.g. West End and Garnet Creek) was recognized by early workers in the district (Cooper, 1951). A reasonable hypothesis from the map pattern is that some of the northeast structures are likely to have originated as tear faults which offset bedding-parallel thrusts and the folded strata (Stewart and others, 2016). The tear faults were likely reactivated in the Tertiary and provided fluid conduits for mineralization.

TERTIARY IGNEOUS ROCKS AND TECTONICS

Field evidence alone brackets the Au-Sb-W mineralization at Stibnite between the ages of the Cretaceous host rocks and the age of the porphyritic dikes that are typically unmineralized. Our 47.86 Ma U-Pb zircon date from a typical latite porphyry dike and the similar, but appropriately younger cooling ages on the latite dike, confirm the prior assignment of the dikes in the district to the Eocene Challis magmatic event (Table 4-3).

Products of Challis magmatism are well exposed in the voluminous volcanic sequence of nested calderas exposed a few miles east of Stibnite in the Thunder Mountain area (Figure 2-2). Latite to dacite porphyries, rhyolites, and andesites are all common dike lithologies in the Eocene-age Challis Volcanic Group at Stibnite and across the state. A few of the mafic dikes at Stibnite may be younger than Eocene, but there are currently no radiometric ages available for these rocks.

The 47.86 Ma crystallization age for a typical latite porphyry dike at Yellow Pine is within the middle of the Challis episode (Table 4-3). Combined with other unpublished age determinations from recent work by the U.S. Geological Survey, our age determinations suggest an interval from about 48 to 46 Ma for dike intrusion in the district and region, within the overall time of Challis volcanism from 51 to 43 Ma (Gaschnig and others, 2009; Steve Box, written commun., 2017). The 48 to 46 Ma time was also the peak of volcanic activity in the Thunder Mountain area (Leonard and Marvin, 1982), although lack of detailed mapping and less-precise K-Ar ages hinder detailed comparison.

Many epithermal Au-Ag prospects and deposits, including the historic Dewey Au mine and the Thunder Mountain and Lightning Ridge mines from the 1980s, are hosted in the Thunder Mountain volcanic terrane approximately 15 kilometers to the east of Stibnite (Gillerman and Mitchell, 2005; Tate and others, 2016; Figure 2-1). The northeast-trending Custer graben further south also hosts numerous precious metal deposits (Moye and others, 1988; Fisher and others, 1992). The Custer graben, located west of Challis, is part of the northeast-trending trans-Challis fault system with its extensional structures, Eocene volcanic rocks, and ore deposits (Kilsgaard and others, 1986). The N35ºE-trending Big Creek graben is also of Eocene age, and subparallel to but outside of the trans-Challis fault system (Figures 2-1 and 2-2).

As noted by many workers, the Challis magmatic event records a shift to extensional tectonics in the northern Rockies, after the earlier compressional environment. Gaschnig and others (2011, p. 2397) ascribe the shift to “extensional collapse of an over-thickened crust.” However, many other tectonic models have been proposed to explain this Paleocene to Oligocene shift from compression to extension with magmatism progressing from British Columbia southwards to Nevada. These tectonic models include changes in the rate of plate convergence, intracratonic rifting, a slab window, ridge subduction, flat-slab subduction and/or slab rollback (Moye and others, 1988: Humphreys, 2009; Gaschnig, and others, 2011). A common driver to those hypotheses is melting of either lithospheric
mantle or asthenosphere. The proposed mantle melts may have introduced heat to melt the overlying crust. The heat also drove large-scale circulation of water, creating regional-scale hydrothermal alteration. Whether the magmas introduced metals or metals were leached from pre-existing rock is problematic from the field and tectonic evidence alone. However, the flare-up of Tertiary magmatism in a complex extensional setting is clearly associated with numerous ore deposits in the Rocky Mountain and Great Basin provinces.

At Stibnite, there is evidence for multiple ages and types of faulting with reactivation of earlier structures. The Meadow Creek fault and other north-south structures parallel the Salmon River suture and Neoproterozoic rifted margin (Figure 2-1; Lewis and others, 2012a) and the north-south faults, such as the Johnson Creek shear zone near the town of Yellow Pine (Figure 2-2) and the Meadow Creek fault, may represent old, deep-seated fractures that channeled fluid circulation deep into and through the old metasedimentary piles and shallow roof pendants. Typical Eocene features in Idaho commonly have northeast-trending orientations, and it is likely the northeast-trending faults in the Stibnite district were active in creating structural loci for ore deposition, although some of the northeast orientations may be inherited from Mesozoic tear faults or older structures. At least some of the right-lateral, strike-slip movement on the Meadow Creek was likely coincident with timing of the Yellow Pine deposit mineralization, and in general, dilatational zones at fault jogs and fault intersections have localized ore deposition in the district (Cooper, 1951). Ore textures suggest that precious metal mineralization in the West End deposit took place in an extensional environment, and the fracture-hosted and breccia-hosted Sb mineralization may have been emplaced in an extensional or trans-tensional setting as well.

MINERALIZATION AND ALTERATION AT STIBNITE

Mineralization at Stibnite is not tied directly to any igneous rocks exposed in the district or region. Instead, the surface and subsurface geology, the petrographic relationships, the isotopic characteristics, and age determinations obtained in this study indicate the more important roles of structure, crustal-derived fluids and metals, and regional heating, although the cause is unknown at present. The complex mineralizing process was repeated in the Paleocene, Eocene, and possibly later, though with different metals and characteristics.

Timing

Results of the geochronological study confirm the multiple Tertiary ages for economic Au-Sb-W mineralization at Stibnite. Table 4-3 and Figure 4-10 summarize the ages determined at Stibnite. Figure 7-1 summarizes the age progression of measured and interpreted tectonic, magmatic, and mineralization events in the Stibnite district. Cretaceous granitic rocks of the Idaho batholith probably generated weak quartz-molybdenite veining and sparse base-metal sulfides localized in weakly developed pods of skarn. Petrographic relationships and radiometric ages of alteration feldspar, mica, and scheelite clearly tie Au deposition to Paleocene and Eocene times, well after intrusion of the Atlanta lobe of the Idaho batholith, though partially overlapping the age of the Bitterroot lobe.

Gold-arsenic mineralization at the Yellow Pine deposit occurred during the Paleocene at approximately 62 to 61 Ma, while a second period of epithermal-like Au mineralization has been dated in the West End deposit almost 10 million years later, just prior to and at the earliest stages of eruption and emplacement of the Challis Volcanic Group. Ages for the secondary potassium feldspar at the Yellow Pine pit (60 Ma) and
the potassium feldspar at the West End deposit (51 Ma) are analytically distinct, and they must reflect two different pulses of Au-As mineralization in the Stibnite district. As discussed in Chapter 5, the Pb isotope ratios of the two deposits are also slightly different (Figure 5-1). The West End deposit contains noticeably less Sb and W than the Yellow Pine system, which is an independent indication of two chemically and temporally distinct systems. It is unclear whether all Sb and W mineralization was confined to the third late stage of mineralization, localized along the Meadow Creek fault, or whether scheelite and stibnite deposition also followed and was associated with the early event at Yellow Pine. Mining of the W ore during World War II has eliminated much of the evidence of this event.

Hydrothermal Alteration and Ore Paragenesis

Previous workers distinguished two main stages of hydrothermal alteration and mineralization at Stibnite. In broad terms, this study supports the observations and conclusions of White (1940) and Cooper (1951) that there are two main stages to the economic Au-Sb-W mineralization: 1) an early stage of disseminated Au-As mineralization, possibly with late scheelite; and 2) a subsequent fracture-controlled, more epithermal-style mineralization event with stibnite and trace Ag. Lewis (1984) provided a more comprehensive petrographic and geochemical study of the Yellow Pine mine. His observations are generally consistent with
the paragenesis and geochronology determined in this study. The West End mine and other small pits mined for oxide Au in the 1980s and 1990s were not described until the work of Cookro and others (1987), which emphasized the epithermal nature of those ores.

The paragenesis determined by Lewis (1984) is summarized in Table 7-1. In addition to his own work, Lewis (1984) used observations of earlier workers, particularly Cooper (1951) and White (1940), who mapped in the pit or underground when the mines were operating. Observations from our study support Lewis’ general model, although the absolute timing of the initial scheelite and stibnite mineralization at the Yellow Pine deposit is still uncertain. The proposed sequence of mineralization and alteration phases determined from our work is in Table 7-2. The major sulfide etching event between the earliest large pyrites (“Gold-1”) and the “Gold-2” stage of Lewis (1984) fits polished section observations documented in the current study and is correlated with the core to middle zone transition, as summarized in Chapter 6 and Table 6-3. The elongate blebs of sulfides, which formed along replaced mica cleavage planes, as noted by Lewis (1984), are the skeletal sulfide intergrowths illustrated in Chapter 6 of the current study.

The three statistically different radiometric ages determined for the three major deposits create challenges in correlating multiple absolute ages to the petrographic mineralization stages. Results of the ages and petrographic evidence suggest three distinct mineralizing stages (Early, Main, and Late W-Sb), as shown in Table 7-2. The 17 million years between the oldest alteration muscovite and potassium feldspar age (approximately 61 Ma in this study) and the most common young age from scheelite (close to 44 Ma) is an indication of the lengthy duration of a robust

<table>
<thead>
<tr>
<th>Mineralization Phase at Yellow Pine mine (Lewis, 1984)</th>
<th>Alteration</th>
<th>Mineralization</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gold - 1</td>
<td>K-metasomatism; plagioclase and microcline replaced by adularia; biotite replaced by sericite; quartz added.</td>
<td>Pyrite as large, euhedral, hydrothermal pyrite (disseminated).</td>
<td>Disseminated.</td>
</tr>
<tr>
<td>Gold – 2 (most abundant volumetrically)</td>
<td>Plagioclase and igneous microcline replaced by adularia; biotite replaced by more quartz and sericite.</td>
<td>Pyrite, arsenopyrite, (pyrrhotite).</td>
<td>Sulfides form elongate blebs along mica cleavage planes. Sulfide precipitation by iron released from biotite.</td>
</tr>
<tr>
<td>Gold – 2 late</td>
<td>Sericitization is abundant; dolomite and calcite + adularia first introduced.</td>
<td>Disseminated sulfides; pyrite in altered biotite.</td>
<td>Alteration is texturally destructive.</td>
</tr>
<tr>
<td>Gold – 3</td>
<td>Adularia replaced by coarse-grained, optically distinct sericite.</td>
<td>No sulfides.</td>
<td>Due to cooling temperatures?</td>
</tr>
<tr>
<td>Non-ore phase</td>
<td>Quartz-ankerite-adularia.</td>
<td></td>
<td>Fill open fissures.</td>
</tr>
<tr>
<td>Tungsten phase</td>
<td>Quartz-adularia-dolomite-calcite.</td>
<td>Scheelite, pyrite.</td>
<td>Infill open fissures.</td>
</tr>
<tr>
<td>Silicification and minor antimony</td>
<td>Intense silicification to fine-grained quartz, minor sericite.</td>
<td>Stibnite-(pyrite).</td>
<td>Southeast side of Yellow Pine mine.</td>
</tr>
</tbody>
</table>
hydrothermal system. The complex sulfide textures and pyrite zoning, combined with the ages, suggest repeated and additive episodes of Au mineralization. We hypothesize that addition of Au from multiple, vein- and fracture-dominant stages of mineralization increased average Au concentrations from one ppm Au to several ppm Au locally, greatly improving the ore grades. However, the mineralization stages varied in intensity in separate geographic areas of the district.

Petrographic observations and microprobe work reveal a variety of complex textures and compositional details of the pyrites in the Stibnite district. Features which have enhanced the Au grades and contributed to the major mineral deposit at Stibnite include:

- Presence of locally abundant Fe sulfides or oxides of sedimentary, regional metamorphic, or skarn origin helped nucleate ore-stage pyrite.
- Abundant igneous biotite in the Cretaceous quartz monzonite to provide Fe which helped precipitate pyrite when S-bearing hydrothermal fluids were introduced.
- An unknown As-Fe-S phase that precipitated early and perhaps quickly, possibly as an As-rich pyrrhotite. This confirms the As-rich nature of the early stage fluids. The metastable phase helped nucleate later sulfides.

Table 7-2. Proposed mineralization sequence in Stibnite district. A stage may predominate in one area but be absent in another. A stage can be repeated with pulses of similar mineralization and discontinuities within it.

<table>
<thead>
<tr>
<th>Hydrothermal sequence in Stibnite district (this study)</th>
<th>Alteration</th>
<th>Mineralization</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pre-ore (metamorphic, deuteritic or magmatic-hydrothermal)</td>
<td>Calc-silicate minerals, metamorphic micas, minor garnet-bearing or amphibole skarn. Sericite? Quartz veins.</td>
<td>Magnetite, pyrrhotite, pyrite (coarse-grained). Rutile (magmatic); weak quartz veining with molybdenite.</td>
<td>Pyrite is nearly stoichiometric.</td>
</tr>
<tr>
<td>Uplift</td>
<td>Discontinuity</td>
<td>Corrosion of rutile, pyrite etching.</td>
<td>Oxidation? Major dissolution of pyrite cores?</td>
</tr>
<tr>
<td>Pre-ore/Early stage - 1 (Au?)</td>
<td>Biotite to chlorite and then muscovite (coarse grained); Weak sericite in plagioclase.</td>
<td>Apatite and monazite, unknown As-Fe-S phase.</td>
<td>Magmatic or alteration phases? High-temperature, As-Fe-sulfide unknown at Yellow Pine.</td>
</tr>
<tr>
<td>Early stage - 2 (Au, As)</td>
<td>Biotite to muscovite-(carbonate-quartz-rutile/titanite); hydrothermal potassium feldspar replaces igneous plagioclase and microcline.</td>
<td>Pyrite (low and high As) followed by arsenopyrite. Some Au-bearing pyrite.</td>
<td>Dendritic to elongate sulfide intergrowths, locally along mica cleavages. Sulfides disseminated in QM host, favorable metasedimentary rocks.</td>
</tr>
<tr>
<td>Discontinuity</td>
<td>Additional resorption to form unconformity between pyrite cores and middle zones?</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Main stage (Au, As)</td>
<td>Intense potassic alteration with local sericite. Metamorphic minerals altered to carbonate-quartz-hydrous minerals with sulfides. Carbonate-quartz-potassium feldspar, locally in banded veins.</td>
<td>Pyrite (fine-grained, banded, As-rich and locally Au-rich) and arsenopyrite. Sulfides are disseminated and in veins.</td>
<td>Transition to more vein and fracture-controlled mineralization with breccia zones. Epithermal textures at West End.</td>
</tr>
<tr>
<td>Late tungsten-antimony stage (W, Sb)</td>
<td>Carbonate-quartz potassium feldspar stable. In Yellow Pine and Hangar Flats areas.</td>
<td>Scheelite, pyrite, stibnite; late Ag minerals. Stibnite veins cross-cut calcite-pyrite veins. Stibnite surrounds pyrite.</td>
<td>Stibnite is clearly later than most pyrite and scheelite. Sparse very late scheelite.</td>
</tr>
<tr>
<td>Mercury and late silicification</td>
<td>Fine-grained quartz.</td>
<td>Cinnabar, minor pyrite.</td>
<td>Shallow outer zone of system? Late along major structure.</td>
</tr>
<tr>
<td>Late stage</td>
<td>Carbonate, clays, chalcedony.</td>
<td>Sparse remobilized stibnite, scheelite. Alters dikes. Thin veins and vug-fillings along fault zones.</td>
<td></td>
</tr>
</tbody>
</table>
• Locally abundant Fe-bearing silicates or carbonates that served as chemical traps within permeable metasedimentary rocks, particularly in the calc-silicate units.

Other characteristics illustrate components of the paragenesis and mineralization:

• Multiple and consistent unconformities within the pyrite crystals that indicate consistent and laterally traceable breaks in sulfide deposition, resulting from dissolution or etching as fluid compositions and physical conditions changed (Figure 7-2).

• Pyrite grains that are texturally zoned with three major zones: cores, middle zones, and rims (Figure 7-2). The middle zone seems to correlate best with the main stage of ore deposition. Gold-bearing pyrite is typically very fine grained and contains significant As.

• Pyrite, particularly in Au-bearing intervals, that typically has multiple bands and repeated lamellae indicating cyclic deposition, possibly from fault-induced pressure release, seasonal fluid variations, multiple magma pulses or other dynamic and repetitive causes.

• Specific areas and pyrite bands that have very high Au contents, as measured by electron microprobe, in the tenth of a weight percent range. The pyrite bands are similar in appearance to Au-rich bands found in epithermal veins. The maximum Au content analyzed was about 0.1 percent or 1,000 ppm Au in the pyrites. Ore grades for the bulk-mineral resource are in the 1 to 3 ppm Au range, so a small amount of the very high grade, Au-bearing pyrite is significant.

• A rough correlation of pyrite and Au with the onset of arsenopyrite deposition. Later pyrite may be lower in As due to precipitation of arsenopyrite.

Figure 7-2. Cluster of zoned pyrite grains from the Yellow Pine pit, showing a well-defined etching or unconformity between the core and middle zones and a prominent, well-formed rim. Lighter middle bands are higher in As than the darker zones. Bright grain is stibnite. Section 8EB-014, Site B, from location 15VG033.
Metal Sources and Fluid Chemistry

In order to formulate a more complete picture of the origin of mineralization at the Stibnite district, the following geochemical information is summarized from the literature as no stable isotopic or fluid inclusion studies were conducted in this current study.

Hydrothermal Fluid Characteristics

Lewis (1984) studied mineral compositions and oxygen, carbon, and sulfur isotopes of the Yellow Pine ores. His work provides partial constraints on temperature and fluid chemistry and sources, particularly in the context of evaluating the influence of meteoric water hydrothermal systems surrounding Tertiary plutons in Idaho, as described by Criss and Taylor (1983). The extensive fluid circulation system indicated by the isotopic depletions could promote interactions between the deeply circulating hydrothermal fluids and deep basement rocks containing metals.

Using an As-geothermometer for co-existing pyrite, arsenopyrite, and pyrrhotite, Lewis (1984) suggested temperatures for the Au phase from less than 100 to over 400°C, and he settled on a 260°C average. Lewis (1984) also calculated temperatures from δ¹⁸O values of coexisting calcite and dolomite, obtaining a range of 180 to 400°C for the Au phase and less than 100 to 180°C for his Sb-Ag stage (Table 7-1). More recently, Marsh and others (2016) examined fluid inclusions in quartz and stibnite from the district. Inclusions in quartz associated with the metalliferous hydrothermal fluids yielded homogenization temperatures (T_h) of 175 to 233°C, at pressures of 9 to 28 bars, salinities of 6 to 9 weight percent NaCl, and low contents of CO₂ or CH₄. From primary inclusions in unfractured stibnite, the microthermometry indicated T_h of 160 to 189°C and P_h of 6 to 11 bars with salinities of 4 to 11 percent and no gas component. In summary, the temperatures for ore deposition fall into the moderate and sub-magmatic category for early Au phases, extending down to much cooler temperatures for the later Sb mineralization. The pressures are also low, supporting observations of the earlier workers that the Yellow Pine ore zone was funnel-shaped and flared upwards as if it was affected by the paleosurface. Marsh and others (2016) concluded the ores were epizonal.

Lewis (1984) analyzed carbon and oxygen isotopes in samples from the Yellow Pine ore body and compared the results to isotopic depletions identified by Criss and Taylor (1983) around Tertiary plutons intruding the Idaho batholith. Carbon isotopes in coexisting calcite-dolomite pairs were different in the early Au phase and the lower temperature Sb-Ag phase (Lewis, 1984). Oxygen isotope measurements on specific minerals (feldspar, sericite, and quartz) of the different mineralizing stages are slightly lower than values for magmatic minerals in unaltered igneous rocks, but Lewis (1984) noted different isotopic values for the Au-related phases and the Sb and Hg stages of mineralization (Lewis, 1984; Guilbert and Park, 1986). Lewis (1984) suggests the isotopic variation is partly due to temperature dependence of mineral-fluid isotopic equilibration and his determinations, cited above, of a significant temperature decrease from the earlier Yellow Pine Au phase to the lower temperature Sb and Hg mineralization events. Although there is overlap in his δ¹⁸O values for minerals at the various stages, Lewis (1984) used his temperature measurements to calculate the δ¹⁸O value of the fluid which precipitated the alteration mineral, based on theoretical fractionation curves, and determined that the carbonate-bearing Au phase fluid had a δ¹⁸O value of 0 to -6 ‰, while the later and cooler Sb-Ag hydrothermal fluid was more depleted with δ¹⁸O values of -6 to -14 ‰. The more depleted oxygen isotope values generally matched the depletions observed and calculated by Criss and Taylor (1983) for Eocene meteoric water in central Idaho at similar latitudes. Thus, Lewis (1984) concluded that not only did the later mineralizing events form at lower temperatures, but they involved a source fluid of Eocene meteoric water. Sulfur isotope determinations on a small suite of samples show a shift between the Au pyrite stage and the stibnite or Sb stage, with average values compatible with a magmatic H₂S input (Lewis, 1984).

The earlier Paleocene age obtained here for the Yellow Pine mineralization may slightly change the accuracy of
Lewis’ comparison to Eocene meteoric waters, but the multiple mineralization ages and stages determined in our study are compatible with evolving isotopic shifts and temperatures. His isotopic data also confirm the importance of regional hydrothermal circulation in ore deposition at Stibnite.

**Metal Sources**

Lead isotope signatures of the ores, discussed in Chapter 5, are highly radiogenic and indicate that the hydrothermal fluids had significant interaction with and derived much of the Pb in the ore minerals from older sedimentary or metasedimentary units, rather than magmas of either the Idaho batholith or Challis magmatic rocks. Common Pb in the alteration feldspars and sulfide minerals at Stibnite was derived from highly radiogenic crustal sources, which modeling suggests included predominantly Mesoproterozoic Pb (1.4 Ga) with some late Archean Pb sources (Figures 5-1 and 5-4). Zircons in the Idaho batholith granitoids contain inherited cores of similar age, supporting the presence of Precambrian rocks of those ages below central Idaho.

Earlier Yellow Pine mineralization (ca. 61 Ma) obtained its Pb from a homogeneous Mesoproterozoic crustal protolith with high U/Pb and Th/U ratios. Later West End mineralization (ca. 51 Ma) shows more isotopically diverse Pb values, suggesting more heterogeneous metal sources including highly radiogenic Archean material mixed with the Proterozoic Pb seen in the Yellow Pine deposit. Stibnite Pb is far more radiogenic than any known magmatic rocks, and the primary provenance of the Pb in the fluids and ores was from ancient crustal rocks or reworked crust. This does not eliminate introduction of some components (S, As?) from deep magmas, but it does suggest that remobilization of Pb and perhaps other components (As, Sb or even Au?) by deeply circulating hydrothermal fluids was a critical process in formation of the Stibnite ores.

**STRUCTURAL AND TECTONIC SETTING OF MINERALIZATION**

**Regional Tectonic Setting**

Age determinations from this study bracket Au-W-Sb mineralization at the Stibnite district to a 20 million-year-long period from approximately 62 to 44 Ma during Paleocene and Eocene times. In Idaho and adjoining states, magmatic events within that time frame include the peraluminous Bitterroot lobe (66 to 54 Ma) of the Idaho batholith in northern Idaho, the intrusive and volcanic rocks of the Challis magmatic event (51 to 43 Ma) just to the east of Stibnite, the Challis-age volcanic rocks of the Republic graben area of Washington, and the middle Cenozoic ignimbrite flareup recognized from the northern and southern Rockies, Nevada, and south into Mexico (Lipman and others, 1972; Henry and others, 2010; Gaschnig and others, 2011; Henry and others, 2012; Henry and John, 2013). As noted by a number of workers, the ages of the volcanic rocks are older (i.e. Eocene) to the north but become younger to the south with caldera-forming ignimbrites in southern Nevada having Oligocene to Miocene ages (Henry and others, 2012; Henry and John, 2013). In their detailed study of a number of Nevada ignimbrite fields, Henry and John (p. 985, 2013) noted that the “plutons associated with significant mineral deposits in the western Nevada volcanic field are older than nearby calderas.” Nor is there always a clear relationship between the plutons and mineralization in Nevada or in Idaho’s Challis magmatic group. Our study shows the age of Au-related alteration at Stibnite pre-dates the known range of volcanic rock ages from the Thunder Mountain field and the presumably related dikes.

The southwestern sweep of mid-Tertiary magmatism across Nevada is well documented, and many of Nevada’s largest Au, Au-Cu, and Mo deposits show a similar southward progression of ages. Gold deposits in the Carlin Trend in northeastern Nevada are 40 to 36 Ma, while at the Cortez district, further south, the sediment-hosted Au deposits are 36 Ma in age (Henry and John, 2015). Henry and John (2015) suggested the Nevada magmatic event followed a 20 million period of flat-slab subduction and was accompanied by a change from contraction to extension, possibly due to slab collapse. The same shift from a compressional environment to extension is evident in the rocks and structures at Stibnite. In Idaho, magmatic activity and metallogenesis started earlier in Eocene and even Paleocene times.

The metal suite of the Carlin-type ores is Au-As-Sb-Hg, similar to the geochemical association at Stibnite, though with significant differences. Both deposit types lack a clear connection to igneous rocks, have similar silty carbonate to granitoid host rocks, similar metal
suites, and contain Au- and As-enriched pyrite. However, arsenopyrite and stibnite are common minerals at Stibnite, while realgar and orpiment are the As minerals in Carlin-type deposits. Another major difference is that decalcification or leaching of carbonate to create porosity is a distinctive feature of Carlin-type deposits (Muntean and others, 2004), but at Stibnite, carbonate is precipitated within the ore zones and throughout the district.

Ore Deposition and Localization

At the district scale, the dominant ore control at Stibnite is structural. Observations of historical workers in the district, plus those of Midas geologists, are well summarized by Huss and others (2014). Au-Sb-W mineralization at Stibnite is clearly associated spatially with the north-south trending Meadow Creek fault system, including its various northeast-trending splays and subsidiary structures. Most noticeable at the Yellow Pine deposit, the intense pull-apart fracturing has shattered the quartz monzonite between complex, en echelon and bending faults to create dilational fault jogs, breccias, and a variety of complex dilatant zones. The West End and Garnet Creek deposits are located along northeast-trending faults that splay off or intersect the Meadow Creek structure.

Antimony mineralization, in the form of stibnite veins and breccia matrix fillings, cross cuts earlier pyrite-bearing veins and suggests a more complex sequence with evolving open structures through time. Stibnite veins in the Yellow Pine deposit typically occupy northwest-trending fractures, in contrast to earlier disseminated Au mineralization found at Yellow Pine in dispersed pyrite grains and small pyrite-bearing veinlets (Figures 3-33 and 3-35; Huss and others, 2014; Austin Zinsser, written commun., 2017).

Any structural model needs to incorporate the new timing, mineralogy, petrographic observations, and information on metal sources. Complex structural patterns over the millions of years of magmatism and hydrothermal activity at Stibnite undoubtedly reflect changing regional stresses which are evident in detailed structural analysis (Austin Zinsser, written commun., 2017).

Textures and mineralogy of the multiple ore stages suggest additional, finer controls on ore deposition. Initial, disseminated replacement sulfides were overprinted and added to by repeated fracture-controlled mineralization as temperatures lowered in the system. Pyrites with multiple growth bands and lamellae and veins with local crustiform banding are interpreted as precipitations of pyrite, quartz, and carbonate due to a pressure-release, crack-and-seal mechanism. The large volume of carbonate gangue at Stibnite, from sedimentary dolomite to ferroan or Mn-bearing carbonate and pure calcite, is quite notable and distinct from many hydrothermal ore deposits.

The structural controls evident at Stibnite are similar to those in numerous mining districts. McKinstry (1948) provides numerous illustrations of vein and fault patterns, including mineralized dilational zones at fault bends. Sibson (1987) summarized the geometry of ore veins and ore shoots in epithermal ore-forming systems, and he proposed a causal connection between faulting, earthquakes, and the timing and cause of ore deposition. Simplistically, fault movement (which likely would generate an earthquake) may be enhanced or catalyzed by fluid flow. However, the same hydrothermal fluid may precipitate minerals due to pressure changes created by the fault movement. As at Stibnite, the polymetallic, siderite-rich Illinois Creek deposit in Alaska is located in a terrane of Cretaceous granodiorite that intruded a folded sequence of Paleozoic sedimentary rocks with diverse chemical and physical properties. The Illinois Creek district is also near large, regional-scale strike-slip faults. Gold at Illinois Creek is hosted in a polymetallic, carbonate-bearing breccia (Gillerman and Sibson, 1988). The pattern of dilatant zones at bends between two strike-slip faults is similar to the geometry of the Yellow Pine deposit, which is located between strands of the Meadow Creek fault (Figures 2-1 and 2-3). At Stibnite, the eastward bend in the right-lateral Meadow Creek fault creates an obvious dilational zone (Cooper, 1951). Cooper (1951) also notes that the older West End fault may be partly responsible for the bend in the Meadow Creek fault, and thus for the ore there.

Sibson’s (1987) fault-rupture model proposes that episodic mineral deposition is generated by the pressure change from rupturing on flanking faults at relatively shallow crustal levels. The hydraulic fracturing and brecciation is a consequence of fluid overpressure and pressure release creating open space, and possible boiling or chemical changes, that allow fluids to ingress repeatedly into the system. If there is abundant CO$_2$ in the environment, the expansive and explosive effects of such fault-induced pressure changes would be
magnified. Carbonate is an abundant gangue mineral in virtually all mineralized rocks at Stibnite, and it is the main gangue for veins containing tiny, lamellar-zoned, Au-bearing pyrite at Yellow Pine. Linking fault ruptures and ore deposition at Stibnite seems very plausible.

ORE DEPOSIT MODEL AND TEMPORAL EVOLUTION

The prefeasibility study for the Midas Stibnite Au project correctly concluded that no single deposit model is applicable to deposits in the Stibnite district (Huss and others, 2014). In terms of conventional ore deposit models and classification schemes, Stibnite incorporates features of orogenic gold systems, epithermal systems, Carlin-type deposits, and hydrothermal veins in general. The Stibnite district displays multiple ore depositional settings and controls, including structural, petrologic, and stratigraphic loci of high permeability and chemical reactivity along with multiple times of metal deposition, as supported in the complex geochronology detailed in this study. A composite deposit evolution of the Stibnite district through geologic time is illustrated in Figure 7-3. While many uncertainties exist, it incorporates the best evidence to date for the timing and paragenesis of mineralization.

COMPARISON TO OTHER GOLD-ANTIMONY-TUNGSTEN DEPOSITS

The significant Sb and W production makes the Stibnite district more complex than just a Au deposit. In pressure-temperature space, Stibnite district mineralization transitions over time from a deeper orogenic setting to a shallower epithermal setting. In terms of type and abundance of ore and trace metals (Au, As, Sb, Hg), sedimentary or metasedimentary host rocks, arsenian pyrite, tectonic setting, and Tertiary age, Stibnite shares many similarities with Nevada’s famous Carlin-type Au deposits. There are notable differences. Whereas Carlin-type deposits are characterized by decalcification, i.e. leaching of carbonate wall rock, and silicification, the veins at Stibnite precipitated abundant carbonate along with the ore pyrite. The association of scheelite with Au is not found with Carlin-type deposits. However, W is also found in Au-Sb vein deposits in orogenic and epithermal settings (Table 7-3).

The association of Au-Sb-W has been noted in other deposits around the world, and a few of those deposits are summarized in Table 7-3 for comparison to the Stibnite district. In addition, Sb vein deposits are sometimes noted as containing Au (Lindgren, 1933; Guilbert and Park, 1986). Global Sb production and supply has long been dominated by China, where rich deposits in Hunan province have been mined for decades (Guilbert and Park, 1986). In 2013 the U.S. imported 85 percent of the country’s consumption of Sb, most of that from China (Wintzer and Guberman, 2015). Idaho is one of the few states with historic Sb production. In addition to the Stibnite district, Sb has been produced from tetrahedrite ores of the Sunshine mine and at other properties in the Coeur d’Alene district (White, 1940; Bennett and others, 1989; Leach and others, 1998; Bennett and Gillerman, 2002; Wintzer and Guberman, 2015). Table 7-3 summarizes some of the Chinese deposits and a few other Sb-Au-W occurrences.

Table 7-3 lists only a few of the deposits worldwide with partly similar metal suites to Stibnite. Other Sb deposits include veins in Spain, Portugal, Bolivia, and others in China, and the few in Idaho. A few generalizations about other Sb deposits are relevant to the Stibnite district:

• Most of the Sb and Sb-Au deposits are hosted in Paleozoic terrain with weakly to moderately metamorphosed, mixed clastic and carbonate strata.

• The deposits are younger than, but localized within, orogenic belts and hosted by or near major shear zones.

• Some authors report Sb and other metals to be enriched in carbonaceous limestones or clastic rocks of the sedimentary section.

• Intrusive rocks are present regionally, but not necessarily close to or related to the ore deposits.

• Many of the Sb veins are polymetallic and contain Au; in places, W and base metals are also reported.

• Quartz is the major gangue but carbonate is very common, particularly where host rocks are calcareous. Wall-rock alteration is not prominent.

• Temperatures from fluid-inclusion work are relatively low (<250°C).
Figure 7-3A. Sedimentation during Neoproterozoic through mid-Paleozoic, with explanation for Figures 7-3A through 7-3G. Geology modified from Stewart and others (2016). Age constraints are from this study or literature (Table 4-3).
Figure 7-3B. Contraction during Jurassic to Early Cretaceous.

Figure 7-3C. Magmatism and intrusion of Idaho batholith during Cretaceous with minor molybdenum mineralization (100 to 80 Ma).
Figure 7-3D. Uplift and erosion during Late Cretaceous (80 to 66 Ma).

Figure 7-3E. Early-stage Au-As mineralization during Paleocene (62 to 60 Ma).
Figure 7-3F. Extension and main-stage Au-As-Ag mineralization during Eocene at approximately 50 to 52 Ma.

Figure 7-3G. Late-stage W-Sb mineralization during Eocene at approximately 44 to 46 Ma, based on U-Pb dates on scheelite (Table 4-3).
Where radiometric ages on mineralization have been determined, the deposits post-date a major deformational and/or intrusive event.

Mineralization took place in an extensional environment, which occurred in some cases after a major compressive orogeny.

Many deposits have a complex paragenesis with multiple stages of mineralization and several authors invoke remobilization or overprinting of earlier mineralization to explain the deposit genesis.

In contrast to traditional lode Au deposits classified as belonging to the orogenic model, the Sb or mixed Sb-Au deposits seem to be more epizonal in character and post-metamorphic in timing. It is difficult to fully ascribe the origin of the Stibnite district hydrothermal fluids to devolatilization or other metamorphic processes that are part of the orogenic Au model (Groves and others, 1998). However, features such as fluid overpressures and their influence on fault controls may apply in more epizonal settings.

In comparing Stibnite to other districts of Sb-Au mineralization, the south China deposits, the Hillgrove deposit in southeast Australia, and the Lake George deposit in New Brunswick seem most analogous. At Hillgrove in the Permo-Carboniferous New England orogen, Ashley and Caw (2004) noted the association of auriferous arsenopyrite-pyrite-quartz-carbonate veins, and local scheelite with the mesothermal deposits formed under brittle-ductile conditions. In contrast, the shallower quartz-stibnite veins and Hg mineralization are associated with brittle structures. The Stibnite district is similar with the deeper early stage Au-As mineralization at Yellow Pine followed by a younger epithermal system and late stibnite. As in Idaho, the quartz-stibnite veins at Hillgrove are associated with breccia and open-space fillings. Late carbonate-stibnite veins are present at Hillgrove and along the Meadow Creek fault at Stibnite. Ashley and Caw (2004) propose that the overprinting of ductile features by brittle vein systems at Hillgrove records rapid exhumation and uplift of as much as 12 kilometers during a 10-million-year span.

The Lake George deposit in New Brunswick is more closely associated with Late Silurian granitic rocks, and Mo and W are more prevalent in the veins, which also include a late stibnite-quartz set. Seal and others (1988) documented an evolution from high-temperature W-Mo stockwork with quartz and alkali feldspar alteration to later and much cooler stibnite-bearing veins along fractures that cut the cupola and surrounding sediments. The intrusions postdate deformed and greenschist-metamorphosed clastic sediments.

Devonian strata, including shales and black carbonaceous limestone, host a number of polymetallic, stibnite-bearing deposits in Guangxi Autonomous Region in southern China. The southern China strata are of similar age and lithology to those in the more famous Hunan Sb belt to the north, which hosts the giant Xikuangshan Sb deposit (dated at approximately 156 Ma; Table 7-3) and the Wuxi Au-Sb-W district (Table 7-3; Guilbert and Park, 1986). The southern Chinese deposits include the Dachang district with several mines exploiting Sn-Zn-Cu-Sb ores in a variety of fold and fault structures along the Danchi ore belt (Table 7-3; Minghai and others, 2007). Fan and others (2004b) make a case for Sb and other trace-element enrichments in the original sediments, which were deposited in a northwest-southeast trough surrounded by a carbonate platform in the Devonian. However, the ore deposits are associated with a belt of folded sedimentary rocks that include anticlines intruded by granites and porphyries of Yanshanian age. The Jurassic to Early Cretaceous Yanshanian event (180 to 90 Ma) affected much of eastern and southeastern China and with the earlier Indosinian compression records the subduction and accretion of the Pacific plate under the Eurasian plate (Pirajno and Bagas, 2002; Chen and others, 2017). The Yanshanian is associated with numerous ore deposits, and the late Yanshanian is associated with the conversion from compression to extension at Dachang (Cheng and Peng, 2014). The Dachang deposits are epigenetic quartz veins and replacements in dark argillite and limestone. Ore samples at Dachang and elsewhere in the belt contain visible stibnite in massive sulfide lenses and quartz veins with base metal and sulfosalts mineralization. The Dachang ores are post folding and exhibit controls by brittle fracturing. Sparse, undeformed porphyritic intrusions seen in a field visit by the senior author in 2016 appeared identical to Tertiary rocks in Idaho and Nevada, and indicative of a shallower subvolcanic setting, typical of emplacement in post-collisional extensional settings. Lead isotopes of the ores at Dachang show both upper crust and mantle signatures (Cheng and Peng, 2014).
Table 7-3. Selected Sb deposits for comparison to the Stibnite district.

<table>
<thead>
<tr>
<th>Deposit</th>
<th>Metals</th>
<th>Resource/Grades</th>
<th>Host rocks</th>
<th>Deformation</th>
<th>Intrusions</th>
<th>Deposit age</th>
<th>Mineralization</th>
<th>Alteration</th>
<th>Structure/Origin</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Xikuangshan, Hunan, China</td>
<td>Sb</td>
<td>Reserve of 2.11 million metric tons Sb metal. Grade 4 wt. % Sb. World's largest Sb deposit; mined for over 100 years.</td>
<td>Upper Devonian and Lower Carboniferous carbonatic, local siltstone, black shale.</td>
<td>Gently plunging anticlines; regional fault intersections; strathform but locally discordant. Passive continental margin with fault basins.</td>
<td>Pre-ore lamprophyre; Indosinian (Triassic), Yanshanian (Jurassic-Cretaceous) granites in region.</td>
<td>Sm/Nd ages on calcites: early syn-sulfide calcite is 155.5 Ma; late syn-sulfide calcite is 124.1 Ma.</td>
<td>Massive stibnite with only trace other minerals. Some open-space textures. Mesothermal to Epithermal.</td>
<td>Silicification and quartz-calcite. Minor barite, fluorite. Quartz-stibnite is early, with calcite-stibnite ores late.</td>
<td>4 en echelon anticlines control ore bodies, cut by faults. Hydrothermal replacements; Role of magmas or orogeny (213-140 Ma) is questionable. Controversial.</td>
<td>Hu and Peng, 2018; Liu and others, 2010; Peng and others, 2003; Hu and others, 1996b, 1996c; Fan and others, 2004a; Laznicka, 1999; Ottens, 2007; Wu, 1993; Gilbert and Park, 1986.</td>
</tr>
</tbody>
</table>
Table 7-3. Selected Sb deposits for comparison to the Stibnite district—continued.

<table>
<thead>
<tr>
<th>Deposit</th>
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<th>Structure/Origin</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Dachang, Danchi mineralized belt, Guangxi province, China</strong></td>
<td>Sn-Zn-Cu-Sb (Au, W)</td>
<td>Large Sn-producing district and current producer; multiple mines in operation.</td>
<td>Devonian black shale and bioclastic limestone, chert in basin surrounded by platform carbonates.</td>
<td>Anticlines with extensional faults.</td>
<td>Sparse porphyry dikes; biorite granite of Late Yanshanian age (93 Ma).</td>
<td>Post-magmatic; U-Pb on cassiterite is 90 to 95 Ma.</td>
<td>Varies: cassiterite, stibnite, sphalerite, jamesonite, arsenopyrite and over 100 minerals. Large district (100 x 30 km).</td>
<td>Quartz veins and replacements with minimal wall rock alteration; not well known.</td>
<td>Vein stockworks, stratified bodies, replacements. Controversial origin: magmatic-hydrothermal? Some sedimentary sources?</td>
<td>Guo and others, 2018; Minghai and others, 2007; Fan and others, 2004b; Cheng and Peng, 2014; Ottens, 2007; Lattanzi and others, 1989.</td>
</tr>
<tr>
<td><strong>Carlin Trend, Nevada</strong></td>
<td>Au</td>
<td>Associated As-Sb-Hg geochemistry.</td>
<td>Paleozoic organic-rich silty limestone, dolomitic siltstone; debris flows, calcareous sandstone.</td>
<td>Gentle folds cut by numerous high-angle faults. NW-trending belt on major anticlinal hinge and culminations of thrust plate. Tertiary extensional tectonics overprint.</td>
<td>Sparse porphyry dikes. Many Nevada deposits have no known intrusions.</td>
<td>Late Eocene, especially around 39 Ma. Interval is 42 - 36 Ma. Post-folding and major magmatism.</td>
<td>Au-bearing arsenian pyrite, diagenetic pyrite, realgar, orpiment, and local barite. Reduced ore fluids, breccias, replacement. Sb is trace element.</td>
<td>Decalcification (carbonate dissolution), argillic alteration, silicification, organic C. Calcite veins peripheral.</td>
<td>NNE- and NNW-trending high-angle structures were fluid conduits. Multiple deposits in 60 km belt.</td>
<td>Teal and Jackson, 1997; Muntean and others, 2004; and many others.</td>
</tr>
<tr>
<td><strong>Stibnite District, Idaho</strong></td>
<td>Au-Sb-W (Hg)</td>
<td>Historic production: Approx. 1 million troy ounces (31 metric tons) Au; 40,000 metric tons Sb; 6100 metric tons W. Midas Gold current measured and indicated resource: 5.6 million troy ounces (174.5 metric tons) Au; 203,838,000 pounds (92,456 metric tons) Sb.</td>
<td>Paleozoic to Neoproterozoic metacarbonates and mixed silicilastics; Cretaceous Idaho batholith granitoids.</td>
<td>Overturned folds and Eocene dikes in district. Challis volcanic rocks in region.</td>
<td>Mesozoic plutons and Eocene dikes in district.</td>
<td>Tertiary with ages of approximately 61 Ma for Au-As disseminated Yellow Pine deposit, 51 Ma on epithermal veins. Ages on scheelite about 45 Ma.</td>
<td>Au-bearing arsenian pyrite, arsenopyrite, stibnite, scheelite. Cinnabar in outer or higher elevation areas of district.</td>
<td>Complex. Dominantly potassic alteration with quartz-carbonate-sulfide veins. Early sericitic alteration of biotite. Abundant calcite, dolomite, and mixed carbonates.</td>
<td>Structurally controlled by dilational jogs on strike-slip faults and intersections with splays, plus favorable stratigraphy in metasedimentary rocks. Late Sb is in breccias and stockwork. Overprinting of deposits.</td>
<td>Midas Gold Press Release of Feb. 15, 2018; this study and references therein.</td>
</tr>
</tbody>
</table>
Similarities of the Stibnite district, Idaho, to other Au-Sb-W districts include a complex geologic history, often with multiple pulses of mineralization, a post-convergence setting, a transition to extensional environments, polymetallic mineralization, Precambrian to Paleozoic host rocks, regional structures, few igneous rocks in district, no direct connection to coeval magmatism, and orogenic uplift most likely in an extensional setting. Mineralization at Stibnite is not orogenic in the classic sense of fluids being derived from deep metamorphic devolatilization, as in Archean greenstones. Rather, the Stibnite ores combine magmatic components and metals derived from the host rocks in hydrothermal fluids circulating deeply and widely in a syn-orogenic, post-convergent setting that evolved from deeper ductile to shallower brittle conditions. Ore deposition in multiple pulses and individual events is hypothesized to be closely related to faulting along the long-lived Meadow Creek and northeast-trending structures in the district.

CONCLUSIONS

The new petrographic and microprobe work presented in this study supports many of the general conclusions of prior studies in the Stibnite Au-Sb-W district, Idaho, but adds crucial new geochronological data for both the intrusive rocks and the ore-forming systems. Transmitted and reflected-light microscopy of samples from all three deposits at Stibnite characterized the major hydrothermal alteration associated with the multiple pulses of mineralization, although much more work is needed to fully map the zoning and details of the large district. Micro-analytical work, including SEM imaging and element mapping of Au-bearing pyrite, reveals chemical and spatial characteristics of Au deposition. The new results allow comparison of Stibnite to other districts within western North America, and worldwide, and properly positions Au-Sb-W mineralization at Stibnite in the chronologically correct tectonic setting.

Major results from this study include:

Plutonic rocks at Stibnite range from 94 to 83 Ma and are compositionally similar to other rocks of the Idaho batholith. Cretaceous intrusive rocks are the host for the Yellow Pine deposit and portions of mineralization elsewhere in the district, but they are not genetically related to the Au mineralization. Calc-silicate horizons, weak skarn, and sparse quartz-molybdenite veining are of Cretaceous age and origin. Biotite in the granitoids and metasedimentary rocks was an important source of Fe that assisted in mineralization by reacting with the introduced S and precipitating pyrite which largely hosts the Au.

Gold mineralization in the Stibnite district took place in at least two pulses in the early Tertiary. Intrusive-hosted Au-As mineralization at Yellow Pine has been dated by \(^{40}\)Ar/\(^{39}\)Ar on alteration mica and potassium feldspar at 62 to 60 Ma (Paleocene). Gold mineralization is early Tertiary, not Cretaceous in age, and is over 20 million years younger than the youngest phase of the Idaho batholith in the district, namely the alaskite/leucogranite at 85 to 83 Ma (Table 4-3). Potassium feldspar from epigenetic veins and breccia zones with ore-grade Au values at the West End area return an age of 52 to 50 Ma (Eocene). The Eocene age closely matches the known onset of Challis volcanism and magmatism in the region. The 52 to 51 Ma event overprinted the earlier, more disseminated style of Au-As mineralization. The presence of adularia, a low-temperature hydrothermal potassium feldspar, and estimates of temperatures under 300°C from fluid inclusions and isotopic work (Lewis, 1984; Marsh and others, 2016) are compatible with an Eocene, subvolcanic setting for mineralization of the second and third stage of Au and Sb-W mineralization. The 50 Ma main-stage event is hypothesized to have added Au (and Ag) to the district by locally precipitating bonanza-grade Au laminae on earlier ore-bearing pyrite and precipitating very fine grained, Au-rich pyrite in carbonate veins similar to those dated as Eocene. However, some of the high-grade pyrite may represent the late phases of the early (61 Ma) event. Textures observed in pyrite include at least two major discontinuities and are compatible with pulses of mineralization interrupted by dissolution and resorption of the pyrite in a hydrothermal system. Lead isotope values of alteration feldspars and sulfide minerals differ for the three deposits, Yellow Pine, West End, and Hangar Flats, suggesting not only an age difference, but an evolving mixture of different sources for the common Pb in the ores of the three deposits.

Early-stage fluids sulfidized igneous biotite and converted it to an aggregate of muscovite, sulfides, and other phases. Potassium was introduced in both early and main stages of mineralization and resulted in pervasive replacement of plagioclase and igneous microcline by hydrothermal potassium feldspar. In the Yellow Pine
deposit, where mineralization is highest grade, potassic alteration is most intense. Peripheral zones of alteration were not the focus of this study but work by Midas Gold has identified Fe-bearing carbonate in the quartz monzonite and chlorite from spectroscopic studies of core of altered granitic rocks. Thin sections of peripheral Yellow Pine samples also show local chlorite alteration of biotite and abundant carbonate associated with the muscovite alteration of igneous biotite. Carbonate is a ubiquitous and common phase in alteration assemblages in the district.

Antimony mineralization is late and cross cuts the earlier ores along fractures and in late breccias along the Meadow Creek fault. Tungsten mineralization, as scheelite, postdates the Au and predates or is synchronous with and closely associated with the Sb mineralization. Scheelite precipitation was undoubtedly localized in large part by areas of calcareous rocks as sediments or as hydrothermal calcite in the Yellow Pine deposit. Scheelite U-Pb ages of approximately 46 to 43 Ma, and locally younger at 36 Ma (Table 4-3; Wintzer and others, 2016, 2017), plus petrographic evidence of multiple generations of scheelite, indicate that the W-bearing portion of the hydrothermal system may have been active for 10 million years along the fault. Major district mineralization took place in three discrete pulses which are documented from about 62 to 44 Ma, an interval of over 15 million years. Some hydrothermal activity may be younger.

Multiple generations of carbonate exist in the district, from that associated with early Au mineralization to late carbonate veins with scheelite, suggesting high CO₂ pressures were common, if not ubiquitous during mineralization. In contrast, massive silicification appears to be a localized, late stage, and poorly understood feature of the district. Carbonate and quartz coexist together in many veins. Many veins, particularly in the metasedimentary rocks, are composite veins, suggesting multiple periods of vein opening and filling, indicating possible crack and seal mechanisms of mineral precipitation.

Textures of intrusive rocks, mineralization, and structures record two and perhaps three major episodes of uplift in the geologic history of the Stibnite area. Following the deep emplacement of the Idaho batholith, a period of major uplift and erosion must have taken place between about 75 and 51 Ma, prior to initiation of the Challis magmatic event in a shallower crustal setting, as supported by porphyritic Eocene dikes and open-space vein textures in the district. Latest phases of scheelite and stibnite mineralization, and the Hg mineralization, may reflect additional uplift during younger Eocene times, as well as continued hydrothermal flow along the Meadow Creek fault and connecting structures.

Mineralization of all three stages is spatially associated with the north-south-striking Meadow Creek fault. Ore bodies have developed in dilatant zones at bends in the strike-slip faults and at intersections with northeast-striking splays and other structural loci. As noted by other workers (Cooper, 1951; Huss and others, 2014), it is likely that the Meadow Creek structure has had a complex history. Rock textures suggest a period of more ductile shear, possibly reflecting an early origin as a tear fault or regional shear. The origin of the northeast-trending structures is also complex. Favorable stratigraphic units may host disseminated mineralization where intersected by structure. The long history of deformation and intrusion also provided good ground preparation prior to the ore-forming event.

Early sedimentary or metamorphic sulfides or Fe-bearing silicates served as nuclei to precipitate later ore-bearing pyrite. Regionally, W and Sb appear to be distributed along northerly-striking structural zones that transect the region of mixed batholith and roof pendant rocks (Figures 2-1 and 2-2). Gold deposits are more abundant in the northeast-trending extensional features that cut the Challis volcanic rocks, but also along the large north-trending structures in Idaho. The older Au-As mineralization at the Yellow Pine deposit is close in age to the inferred 61 Ma (bracketed between 64 and 50 Ma) age of the main stage, As-rich base metal veins at Butte, Montana (Lund and others, 2002), and the fault-hosted approximately 68 Ma Au-As mineralization at the Beartrack mine in Lemhi County, Idaho (Hawksworth and others, 2003). The association of Au with large regional faults or shear zones is not a coincidence, but an indicator of the importance of these structures. An interpretation from our work is that the fault processes were triggers for mineral deposition, but in addition, the large amounts of hydrothermal fluid flow along the regional structures may have promoted movement along those faults.

Geology of the Stibnite region records a shift from convergent tectonics related to Mesozoic compression, folding, and batholith generation to Tertiary extensional or trans-tensional tectonics with volcanic activity,
shallow-level magmatism, and Au mineralization. In Idaho this transition took place in the Paleocene to Eocene time, while similar events were slightly younger in Nevada. Eocene magmatism and development of extensional features and uplift have been ascribed to a variety of slab-removal models but they do not fully explain generation of the specific types of metal concentrations. Though sometimes compared to Carlin-type deposits, Stibnite is different in character from Au deposits of Tertiary age in Nevada, such as porphyry-related or Carlin-type Au deposits. Rather, the Stibnite district ores combine characteristics of several ore deposit models, including orogenic Au, reduced intrusion-related Au, epithermal precious metal veins, Carlin-type Au, and polymetallic vein models.

The Stibnite district is unique in Idaho for its metal suite, size, and diversity of both host rocks and ores. Critical features to formation of the Au-W-Sb-(As) ores at Stibnite are postulated to include:

- Regional heat flow and juvenile magma generation during the early Tertiary as a consequence of far-field regional tectonic events such as slab rollback or delamination.
- A regional to local trans-tensional to extensional stress regime that allowed fluid circulation to deep crustal levels over millions of years along major faults.
- Repeated cycles of deposition of Au in arsenian pyrite, as well as quartz and carbonate, within the same vein or small volume of rock, probably as a result of fluid pressure release during faulting.
- Thick sequences of clastic sedimentary or low- to medium-grade metasedimentary Proterozoic to Paleozoic rocks at depth that could serve as sources for metals such as As, Sb, and perhaps Au. At Stibnite, some Archean input is also evident from the Pb isotopes.
- An alkaline fluid, possibly buffered by the voluminous amounts of granodioritic rocks of the Idaho batholith at depth, along with carbonate sediments or metasedimentary rocks to supply CO$_2$ to the hydrothermal fluid.
- Gold, S, and perhaps W supplied by Tertiary magmas at depth, and probable leaching of Au, plus Sb and As, from the deep metasedimentary pile or regionally.
- Chemical traps provided by mafic minerals of the Cretaceous plutons or metamorphosed sedimentary rocks.
- Structural weaknesses inherited from Mesozoic fold and thrust deformation, or other early tectonic fabrics that could be reactivated as later Tertiary structures.
- Multiple, overprinted hydrothermal and ore-forming systems active over a longtime frame.

Gold deposits of Eocene age are common in Idaho, but rarely as large as at Stibnite. The Boise Basin in southern Idaho has a recorded production of approximately 2.5 million troy ounces of Au from placers and veins associated with Eocene porphyry dikes (Anderson, 1947; Kiilsgaard and others, 1997). However, the host rock in the Boise Basin is exclusively plutonic, and Stibnite's complex structural and sedimentary architecture and Mesoproterozoic to Paleozoic metasedimentary sequence, is lacking there.

**SUGGESTIONS FOR FUTURE WORK**

The Stibnite district hosts Idaho’s largest known Au resource, approximately 6 million troy ounces, in addition to significant Sb and historically important W and Hg production. Whether the objective is exploration and mine development or simply to better understand the geologic evolution of central Idaho, additional studies are recommended. Specific topics and questions for other studies include:

- Additional age determinations on alteration minerals at the Yellow Pine deposit, along with fluid inclusion or isotopic work. Monazite and apatite are locally present in the Yellow Pine rocks. Are they relict and show ages reset by the ore fluids or even related to the hydrothermal fluids?
- Additional study of scheelite, including more regional samples and perhaps Nd-Sm dating or U-Pb ID-TIMS dating and Pb isotopic work.
- Detailed study of the multiple carbonate phases involved in the alteration and mineralization, including ages, isotopic signatures, and compositions in a 4-D (3-D space plus time) framework.
• More petrographic and geochemical work on dike compositions, age determinations, and isotopic signatures.

• New geochronological work and detailed mapping of the volcanic rocks at Thunder Mountain east of Stibnite.

• Structural studies, including reconstructions of post-ore offsets along the Meadow Creek fault system.

• Additional geochemical studies including measurements on temperature and composition of fluid inclusions and review of As and Sb sulfide phase equilibria.

• Additional Pb isotope work to expand the distribution of the sampling in this study.

• Regional tectonic studies of uplift, structures, and mineralization.

The complexities seen in the geologic history at the Stibnite or Yellow Pine district in central Idaho are typical of large mineralized systems. The Pb isotopic differences determined in this study confirm the division into three deposits distinguished by location, age, and metal content and structure, but all within a few miles. In addition, multiple sources of metals are indicated – both plutonic-hosted and epithermal Au, W potentially scavenged from older rocks, and Sb from deeply buried basinal sediments or basement terranes. The multiple factors of host rock composition, structures, tectonic position, fluid chemistry, and critical timing have intersected to contribute to the large mineral endowment localized in the Stibnite district of central Idaho.
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