UNITED STATES DEPARTMENT OF THE INTERIOR

GEOLOGICAL SURVEY

Geology and mineral deposits of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho

Abstracts and short papers selected from the symposium and poster sessions presented to the Annual Convention of the Northwest Mining Association, Spokane, Washington, December 7, 1989

edited by

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This report is preliminary and has not been reviewed for conformity with U.S. Geological Survey editorial standards and stratigraphic nomenclature.

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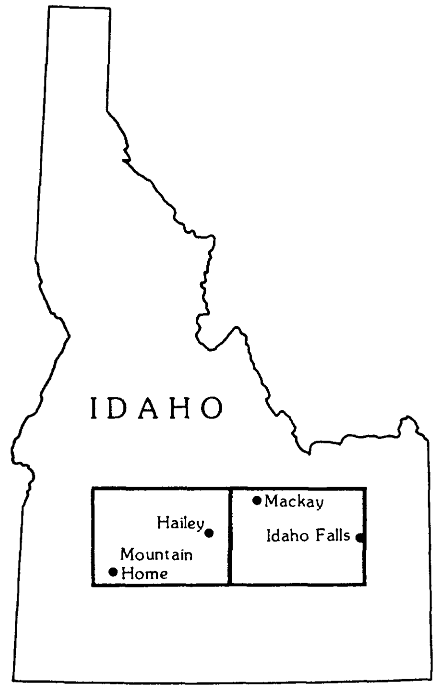
INTRODUCTION

The U.S. Geological Survey Conterminous United States Mineral Assessment Program (CUSMAP) was initiated in 1977 to provide up-to-date assessments of the mineral potential of the Nation's public lands in the lower 48 states. One major objective of CUSMAP is the development and application of new concepts for the identification of mineral resource potential in heretofore untested but possibly mineralized areas. CUSMAP is providing new information on present and potential mineral supplies and is producing important data to guide national mineral policy, land-use planning by Federal, State, and local governments, and private-sector mineral exploration.

The Hailey CUSMAP project is a cooperative venture with the Idaho Geological Survey and the Departments of Geology at Idaho State University and the University of Idaho. Project personnel have interacted closely with concurrent research investigations in the area supported by National Science Foundation Grant RUIEAR 86-18629 and an Idaho State Board Of Education Economic Incentive Grant to Idaho State University.

The Hailey project has continued to expand on previous mineral-resource assessments of nearby areas in Idaho and Montana. Investigations began in 1986 and field work was completed in 1989. Preliminary results were presentated in oral poster sessions at the Northwest Mining Association 95th Annual Convention and Trade Show, December 7, 1989, in Spokane Washington. This volume includes many of the topics presented at this convention.



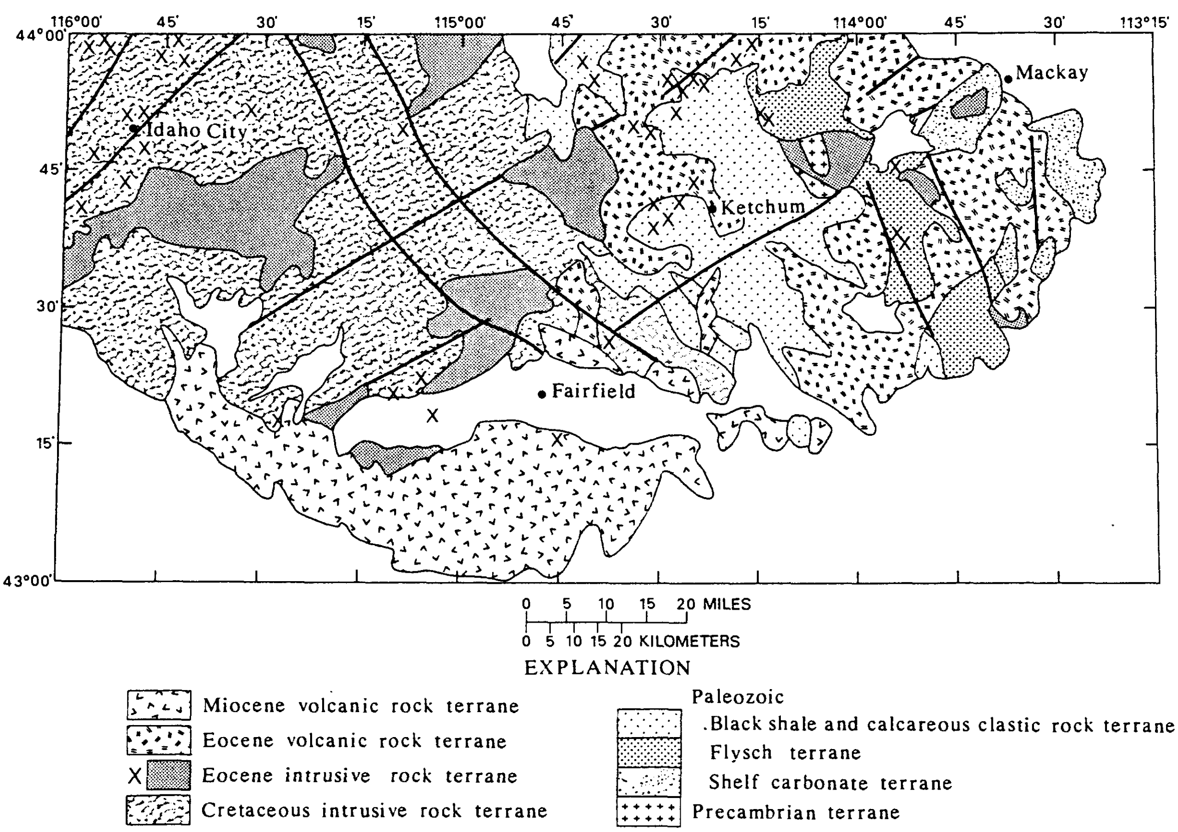


Figure 1. Maps showing location of Hailey and Idaho Falls 1°x2° quadrangles, generalized geology of the Hailey and western Idaho Falls quadrangles.

GEOLOGIC TERRANES AND MINERAL DEPOSIT TYPES IN THE HAILEY AND WESTERN IDAHO FALLS 1°x2° QUADRANGLES, IDAHO

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INTRODUCTION

Mineral resource information in this report and those that follow on the Hailey and western Idaho Falls 1°x2° quadrangles include both geologic setting and mineral deposit types. The geologic setting of the mineral deposits is discussed in terms of geologic terranes, interaction of geologic terranes and major structural features. For this project, a geologic terrane is defined as the area in which a particular assemblage of rock types crops out. The important geologic terranes and major structural features are shown on figure 1. Mineral deposit types are based on geologic characteristics of known and inferred deposits within or close to the quadrangles. The deposit types are designed to correspond to the mineral deposit models or modifications of the models presented in Cox and Singer (1986) and used in Johnson and Worl (this volume). For convenience, the models are assigned a number and a letter (or letters) that correspond to, or are modified from, their designation in Cox and Singer (1986).

This report presents the mineral-bearing geologic terranes and the known mineral deposit types present in the Hailey and western Idaho Falls 1°x2° quadrangle and briefly summarizes the metallogenic events that formed the metal deposits.

GEOLOGIC TERRANES

Quaternary rocks

Includes Quaternary sedimentary and volcanic rocks undivided.

Tertiary extrusive rock terranes

Idavada Volcanic Group: Includes all known Miocene volcanic rocks

Challis Volcanic Group: Includes all known Eocene extrusive volcanic rocks and smaller hyperbyssal bodies.

Tertiary intrusive rock terranes

Granite: Includes pink to light gray granite found in the Sawtooth, Trinity, and Soldier batholiths and many smaller plutons.

Granodiorite: Includes diorite, granodiorite, and dacite porphyry found in the border areas of the major batholiths and many smaller plutons.

Cretaceous intrusive rock terranes

Granite - granodiorite: Includes the major phases of the Idaho batholith - biotite granodiorite, muscovite-biotite granite, leucocratic granite, and aplite/pegmatite complex.

Diorite - hornblende granodiorite: Includes potassium-rich granodiorite, tonalite, and quartz diorite found mainly in the southwestern border areas of the Idaho batholith.

Paleozoic sedimentary rock terranes

Black shale: Includes dark carbonaceous, fine-grained shale, slate, argillite, impure carbonate rock, and calcareous clastic rocks. Found in the Milligen, Trail Creek, and Phi Kappa Formations,

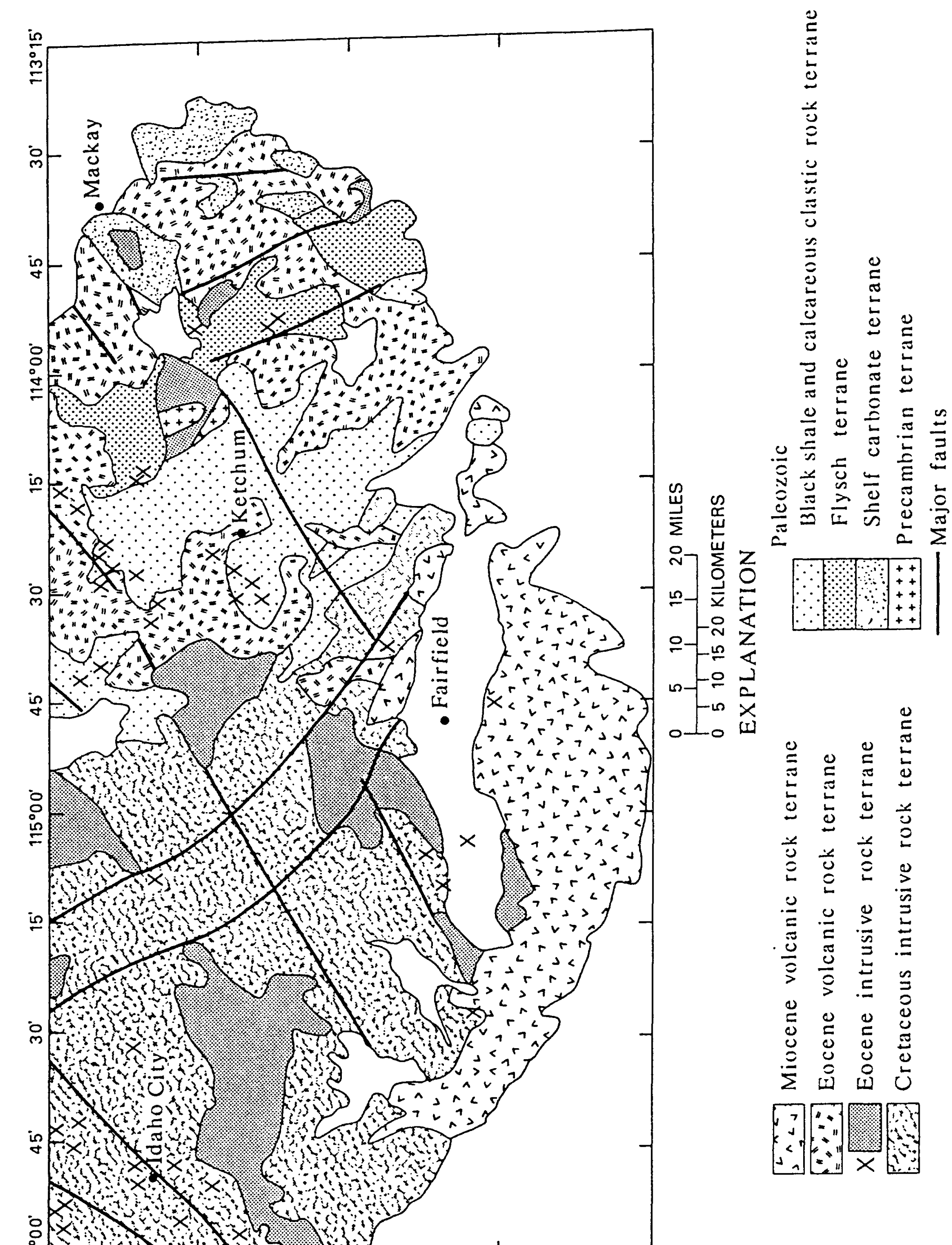

Figure 1. Map showing geologic terranes of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

undivided Devonian and Silurian rocks, and parts of the Wood River, Grand Prize, Dollarhide, and Copper Basin Formations.

Calcareous clastic rock: Includes calcareous sandstone, quartzite, metaquartzite, conglomerate, and shale and intercalated silty and sandy limestone and dolostone, but excludes carbonaceous fine-grained sandstone and shale. Found in parts of the Wood River, Grand Prize, Dollarhide, and Copper Basin Formations.

Flysch: Includes carbonate and sandstone turbidites. Found in the McGowan Creek and parts of the Copper Basin Formation.

Shelf carbonate: Includes limestone and dolostone, but excludes silty limestones in the black shale and calcareous clastic rock terranes. Found in the Carbonate Bank, White Knob, and Jefferson Formations.

Precambrian rock terrane

Includes high-grade metamorphic rocks and felsic intrusive rocks found in the Pioneer gneiss dome.

MINERAL DEPOSIT TYPES

Tungsten skarn (14A, fig. 2): Examples are the Wildhorse Creek deposits, the Ura Claim Group in the Vienna district, and the Bunker Hill mine in the Mineral Hill district.

Tungsten veins (15A, fig. 2): Examples are the Big Falls Creek prospect and the Corral Creek Tungsten deposit.

Greisen deposits (15C): Examples may be present in the Sheep Creek area located between Idaho City and Pine.

Stockwork deposits containing Mo, W, Sn, or Be (15D, fig. 2): Examples include the Walton-White Mountain prospect in the Alta district and the Rinebold prospect in the Roaring River district.

Copper skarn (18B, fig. 3): Examples are at the Empire mine and other properties in the Alder Creek district.

Iron skarn (18D, fig. 3): An example is the Saddle magnetite prospect in the Alder Creek district.

Gold skarn (18F, fig. 3): An example is at the June Day mine in the Warm Springs district.

Polymetallic skarn (18G, fig. 3): Examples are at the Eagle Bird mine in the Little Wood River District and the Champion mine in the Alder Creek District.

Polymetallic replacement deposits (19A, fig. 4): Examples are at the Triumph mine in the Warm Springs district and the Red Elephant mine in the Mineral Hill district.

Stockwork deposits containing significant gold (20D, fig. 5). An example is at the Ontario mine in the Warm Springs district.

Shear zone-hosted polymetallic veins (22CS (intrusive host), 22ES (black shale host), fig. 5): Examples are at the Alta Silver Group in the Alta district and at the El Oro mine in the Skeleton Creek district.

Shear zone-hosted precious-metal veins (22CZ (intrusive host), 22DZ (volcanic host), 22EZ (black shale host), fig. 5): Examples are at the Tip Top mine in the Skeleton Creek district, the Gambrinus mine in the Gambrinus district, and the Overlook mine in the Black Warrior district.

Complex precious-metal veins (22CY (intrusive host) and 22DY (volcanic host), fig. 5): Examples are the Pioneer ore body in the Gold Hill

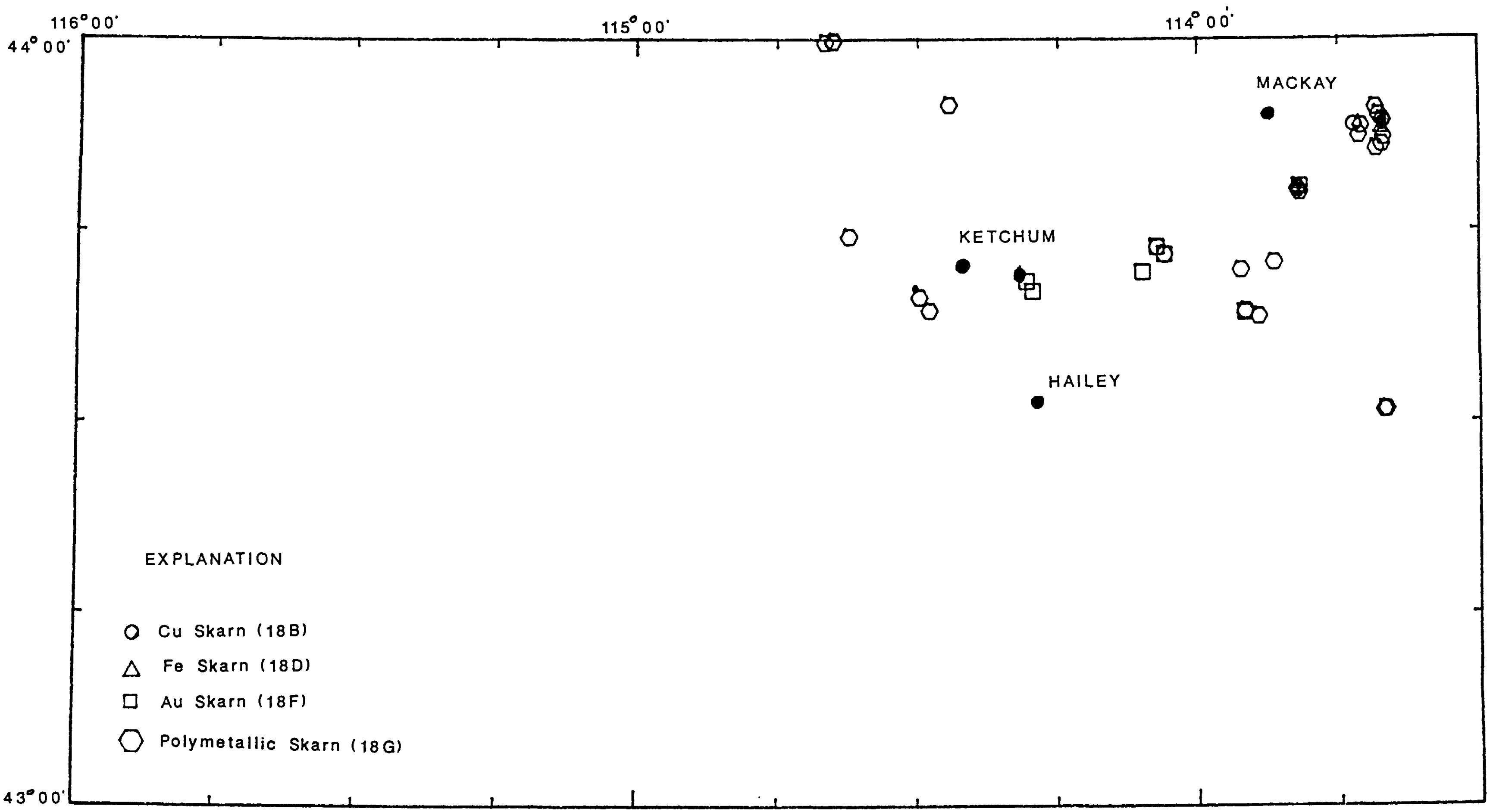

Figure 3. Map showing distribution of copper skarn (18B), iron skarn (18D), gold skarn (18F), and polymetallic skarn (18G) deposits in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

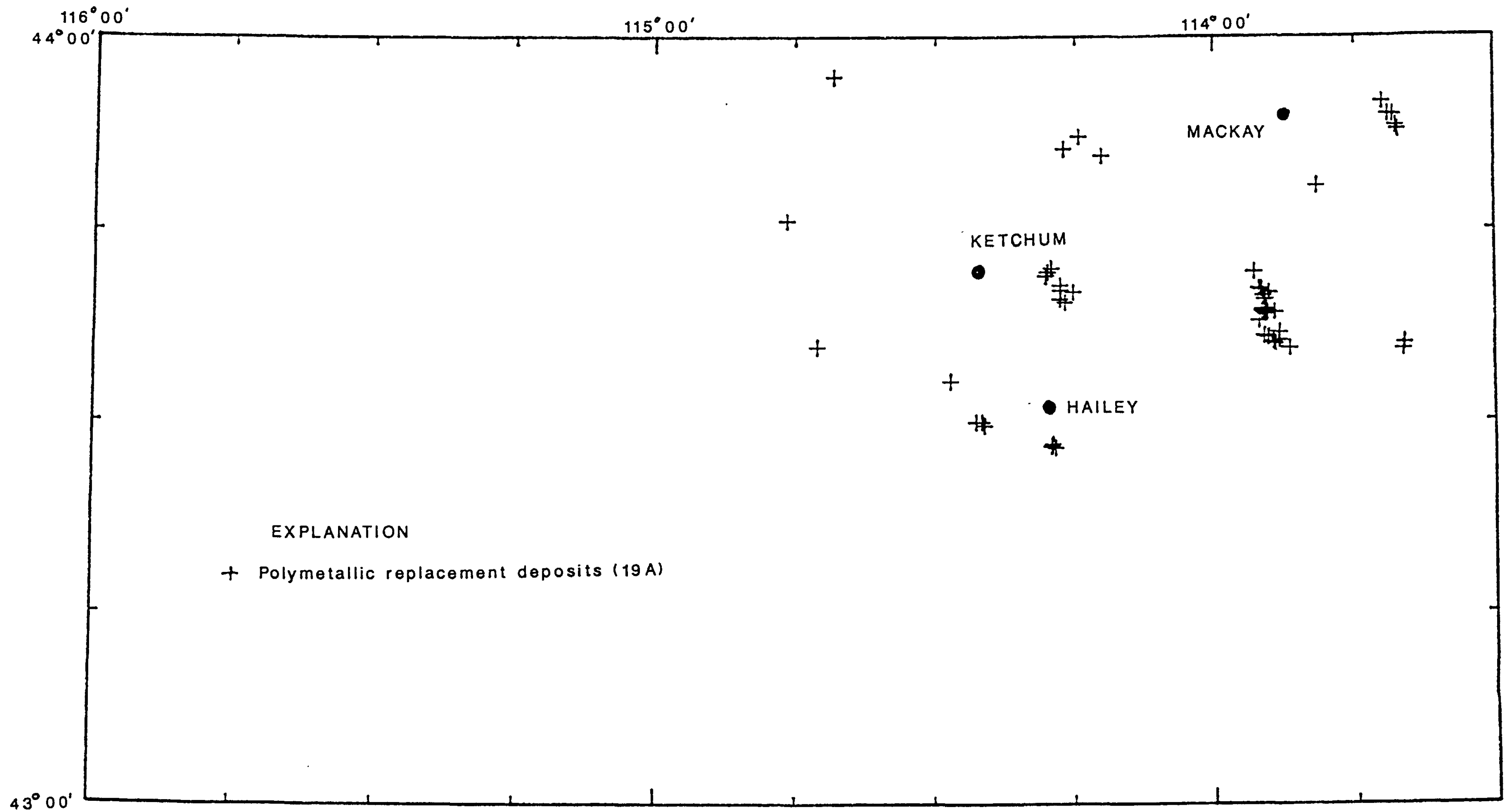

Figure 4. Map showing distribution of polymetallic replacement (19A) deposits in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

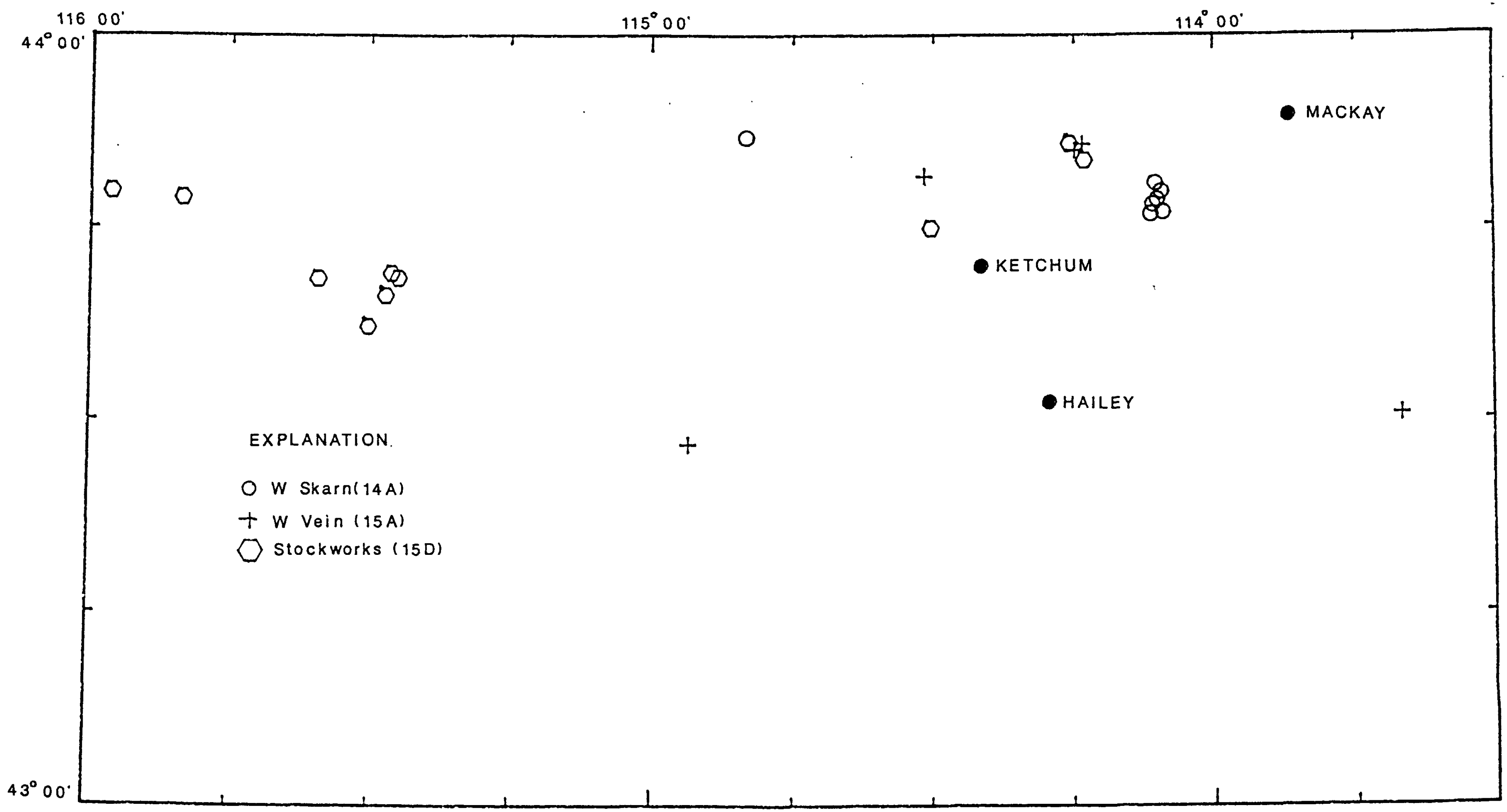

Figure 2. Map showing distribution of tungsten skarns (14A), tungsten veins (15A), and stockwork deposits containing Mo, W, Sn, or Be (15D) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

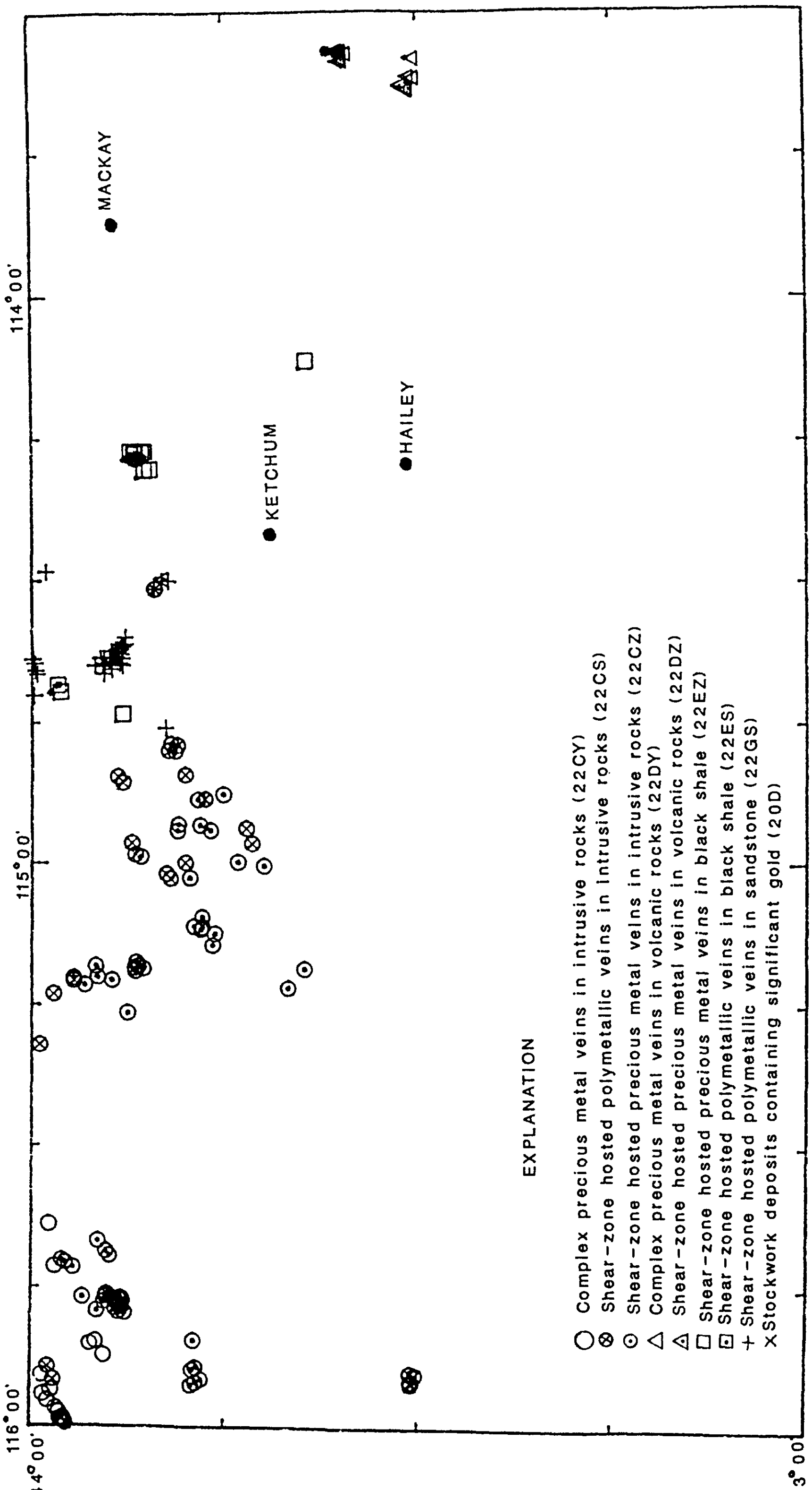

Figure 5. Map showing distribution of complex precious-metal veins (22CY and 22DY), shear zone-hosted precious-metal veins (22CZ, 22DZ, and 22EZ), shear zone-hosted polymetallic veins (22CS, 22ES, and 22GS), and stockwork deposits containing significant gold (20D) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

mine, Quartzburg district and the Horn Silver mine in the Lava Creek district.

Massive gold-quartz veins (22CX (intrusive host) and 22EX (black shale host), fig. 6): Examples are in the Hailey gold belt and Rocky Bar deposits.

Massive silver-quartz veins (22CV, fig. 6): Examples are at the Pilgrim mine, Sawtooth district, and the Webfoot mine, Vienna district.

Polymetallic veins (22C (intrusive host), 22D (volcanic host), 22E (black shale host), 22F (carbonate host), and 22G (sandstone host), fig. 7). Examples are at the Democrat and Snoose mines in the Mineral Hill district, and the Homestake mine in the Warm Springs district, Hot Springs Gold (25A, fig. 8): Examples are at the Mehie Group prospects located northeast of Fairfield, the Big John prospect east of Magic Reservoir, and in the Lava Creek district.

Volcanic-hosted epithermal precious-metal deposits (25BC): Examples may be present in the Lehman Basin area located north of Mackay.

Carbonate-hosted precious-metal deposits (26A, fig. 8): Examples may be present in the Lehman Butte area north of Mackay.

Simple antimony veins (27D, fig. 8): Examples are the Hermada antimony deposit in the Swanholm Creek area and at unnamed stibnite prospects in Navarre Creek north of Mackay.

Sedimentary stratabound lead-zinc deposits (31A, fig. 9): Examples may be present at the Triumph Mine in the Warm Springs district.

Bedded barite (31B, fig. 9): Examples are at the Sun Valley barite mine in the Warm Springs district.

METALLOGENIC EVENTS

The Hailey and western Idaho Falls quadrangles contain many types of deposits of base, precious, and ferrous metals in several geologic terranes. The deposits formed during metallogenic events including sedimentation in marine basins that were oygen-depleted during much of Devonian through Permian time, Cretaceous igneous and hydrothermal activity related to formation of the Idaho batholith, igneous and hydrothermal activity during formation of the Eocene granitic batholiths and the Challis volcanic field, hydrothermal activity related to Miocene and younger volcanic activity, and Pleistocene to Holocene weathering and erosion.

Stratabound concentrations of metals that formed during deposition of the Paleozoic black shale sequences include silver, barium, copper, molybdenum, vanadium, lead, zinc, and nickel. The metal concentrations took the form of massive sulfide lenses, stringers and disseminations of sulfide minerals, and probably metals entrapped in nonsulfide minerals. Known syngenetic deposits of precious and base metals (31A) include stratiform zinc deposits at the Hoodoo and Livingston mines in the Boulder Creek district just north of the Hailey quadrangle and some of the stratiform zinc-lead ore bodies in the Triumph Mine, Warm Springs district. Numerous other deposits are thought to be, in part, stratiform. An example of a syngenetic stratiform barite deposit (31B) is at the Sun Valley barite mine in the Warm Springs district. In addition to the stratabound deposits that may be present, the carbonaceous rocks (black shales) held an enormous reservoir of metals that could have been mobilized and concentrated by igneous and hydrothermal systems active in the Cretaceous and Tertiary.

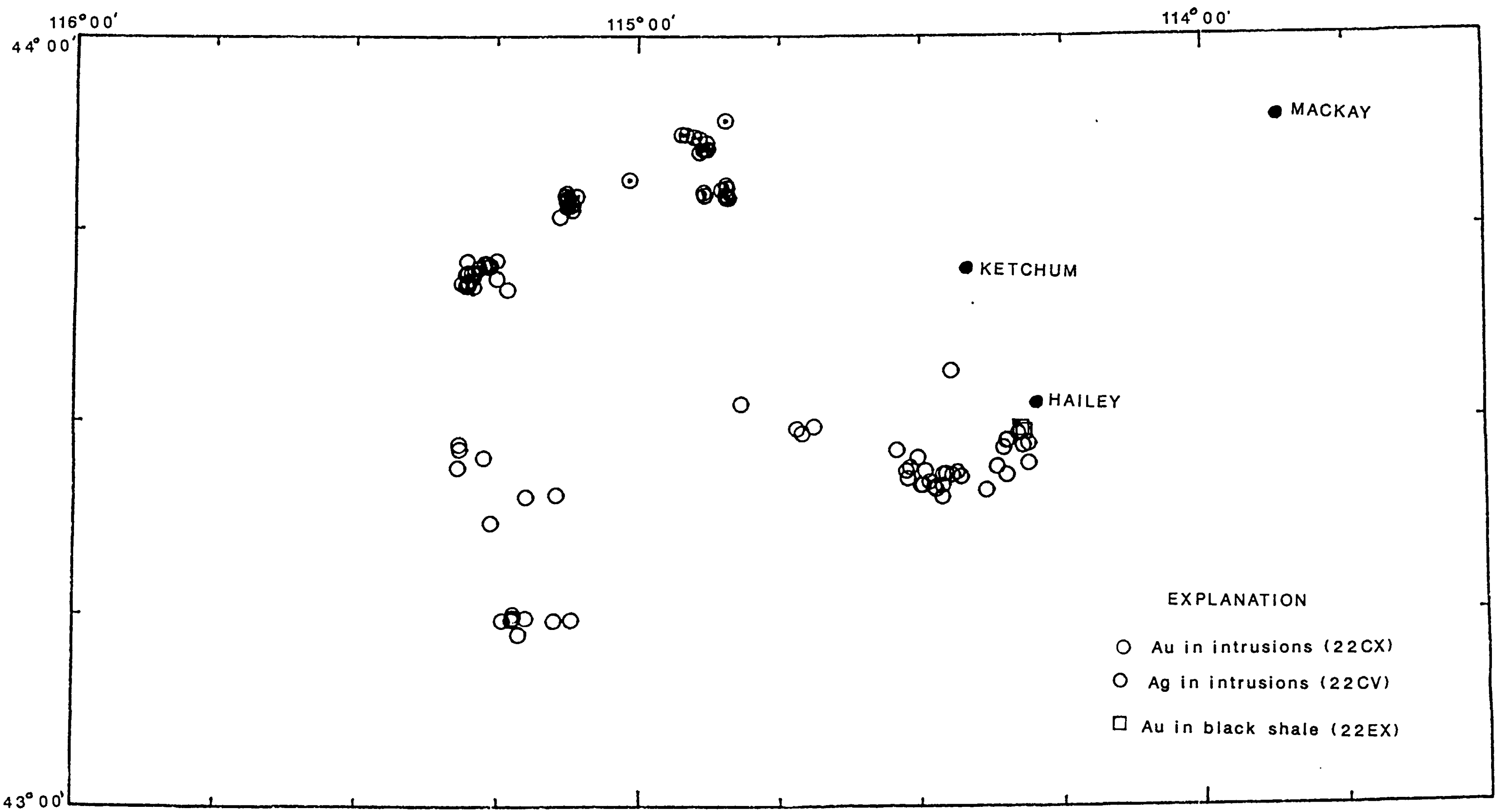

Figure 6. Map showing distribution of massive gold (22CX and 22EX) and massive silver (22CV) veins in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

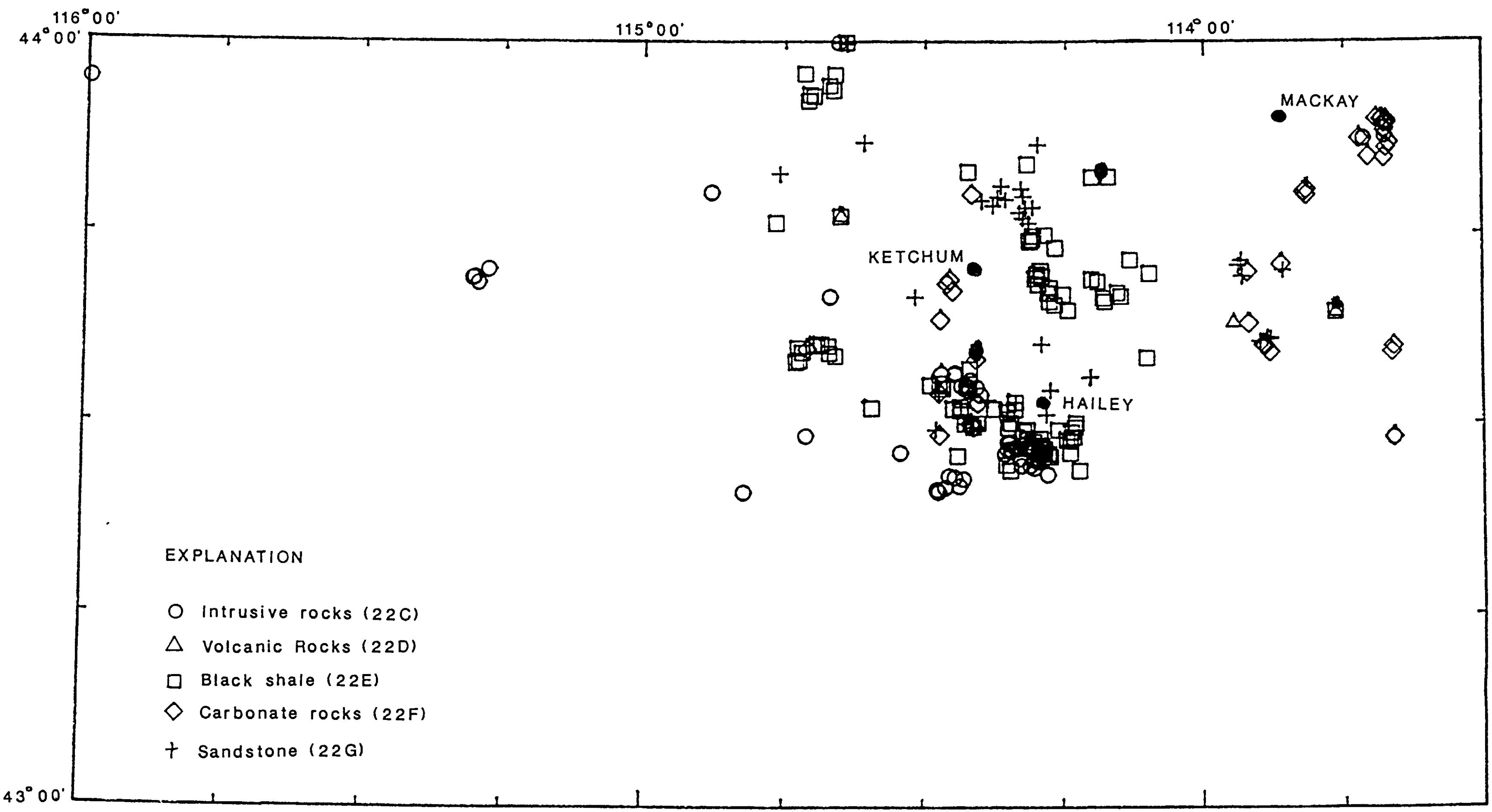

Figure 7. Map showing distribution of polymetallic veins (22C, 22D, 22E, 22F, and 22G) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

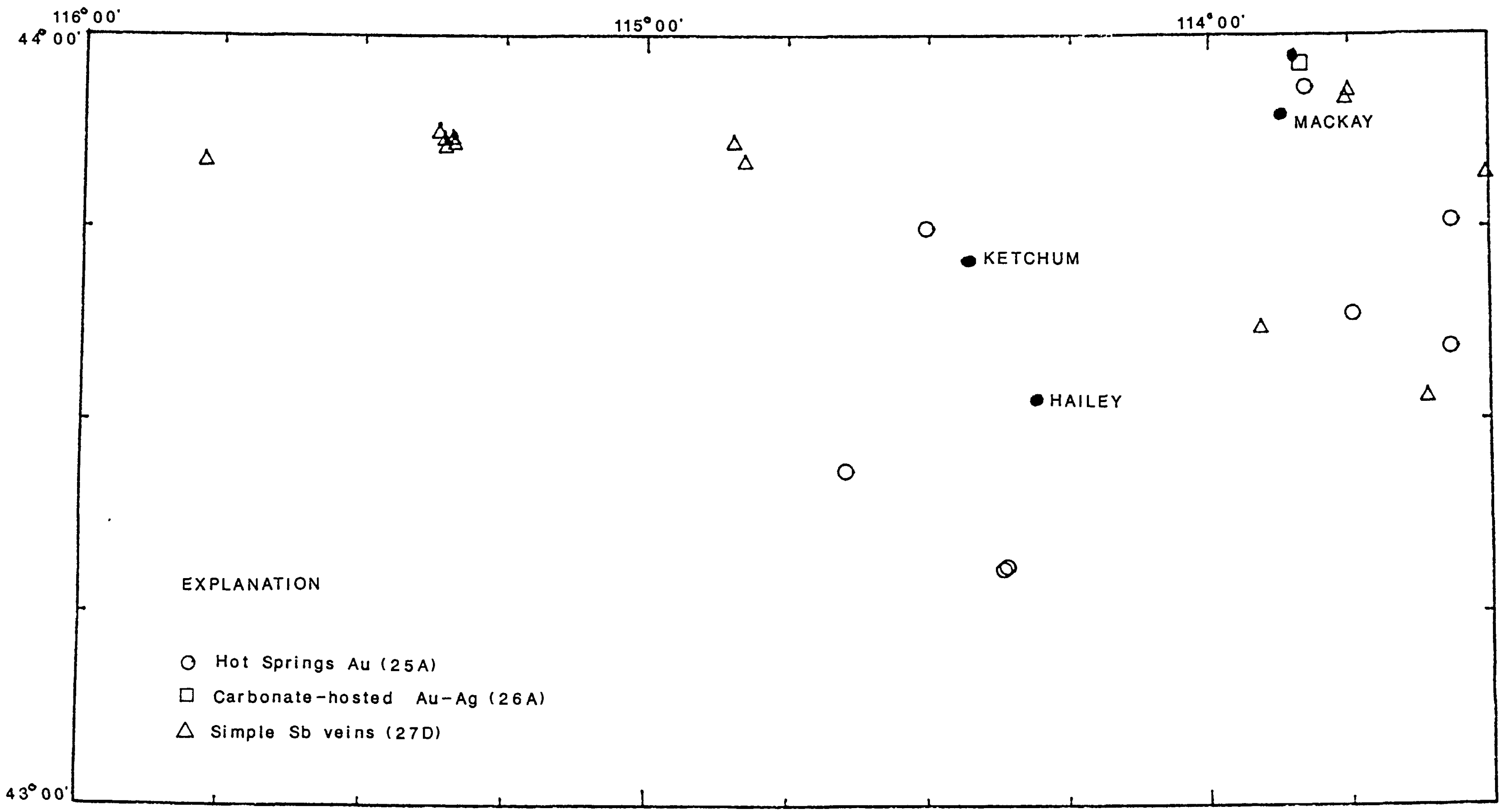

Figure 8. Map showing distribution of hot springs gold deposits (25A), sediment-hosted precious-metal deposits (26A), and simple antimony veins (27D) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

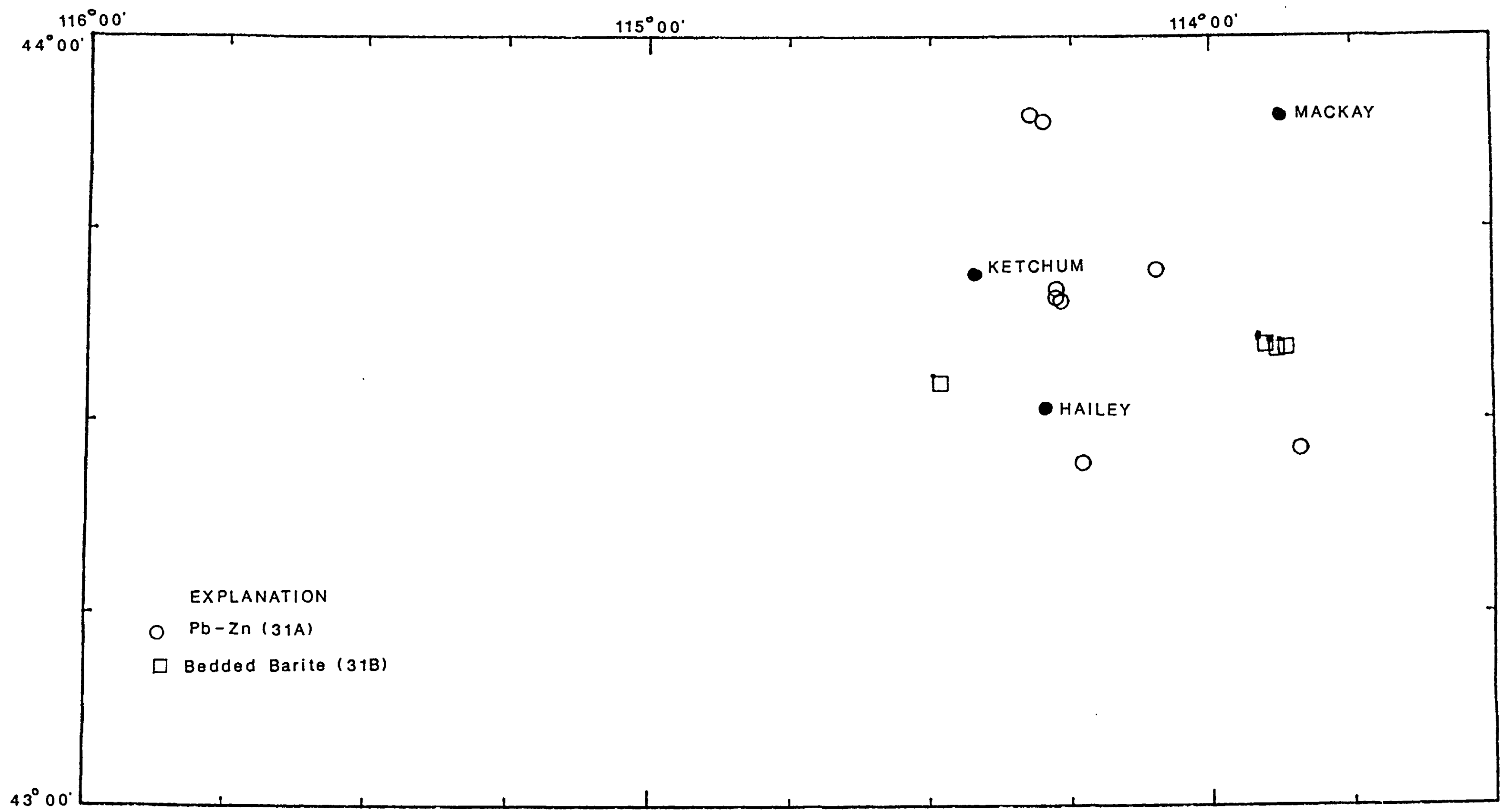

Figure 9. Map showing distribution of sedimentary stratabound lead-zinc deposits (31A) and bedded barite deposits (31B) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

Emplacement of the Idaho batholith during Cretaceous time produced several mineral deposit types. Polymetallic skarn (18G) and tungsten skarn (14A) deposits formed where plutons intruded calcareous sediments; examples are numerous in the Mineral Hill, Willow Creek, and Camas districts, but these deposits are generally small. Large polymetallic replacement deposits (19A), most notable in the Mineral Hill and Little Smoky districts, probably formed during emplacement of the batholith. These are stratabound to irregular-shaped massive lead-zinc sulfide bodies that resemble sedimentary stratabound deposits. Stockwork molybdenum and associated tungsten and base-metal vein and skarn deposits found elsewhere in Idaho in Cretaceous intrusive rocks (for example, the Thompson Creek deposit) are not known in the Hailey and Idaho Falls quadrangles.

Following emplacement of the Idaho batholith and prior to emplacement of the Eocene granitic batholiths many fissure-filling massive base- and precious-metal veins were deposited. These include the massive gold quartz (22CX) and massive silver quartz (22CV) veins hosted mainly by rocks of the Idaho batholith but locally also by rocks of the black shale terrane (22EX). Many of the massive polymetallic veins hosted by rocks of the black shale and calcareous clastic rock terranes (22E and 22G) are also thought to have formed during this metallogenic event.

Tertiary igneous and hydrothermal activity and attendant metallization took place during regional extension characterized by deep seated high-angle fractures and rift structures. Deposits formed during this event generally show some enrichment of fluorine, barium, potassium, silica, gold, silver, antimony, beryllium, iron, lithium, manganese, molybdenum, lead, strontium, tin, tellurium, tungsten, uranium, vanadium, and zinc. Where intrusions invaded calcarous rocks various skarn deposits (14A, 18B, 18D, 18F, 18G) were formed. Some stockwork deposits containing molybdenum, tungsten, beryllium, and tin? (15D) and tungsten veins (15A) formed in and along Eocene intrusive bodies. Greisens may have formed in some of the more evolved Eocene granites.

Hydrothermal activity, starting about 50 million years ago and continuing intermittently to the present, produced a variety of deposits. Many polymetallic veins in all terranes (22C, 22D, 22E, 22F, and 22G) formed in open fractures and breccia zones from hydrothermal solutions of this event. Complex precious-metal veins (22CY and 22DY) characterized by complex mineralogy and element associations including bismuth formed along major regional structures in association with emplacement of hypabyssal bodies. Many of the major regional shear zones in the area contain shear zone-hosted polymetallic (22CS, 22ES, and 22GS) and precious-metal (22CZ, 22DZ, and 22EZ) epithermal veins, veinlets, and breccias thought to have formed during the Tertiary hydrothermal events. A few small hot springs gold (25A) deposits and simple antimony veins (27D) formed during these hydrothermal events. Hydrothermal solutions coursing along major high-angle fracture systems formed large areas of jasperoid in the shelf carbonate, flysch, and black shale terranes and areas of alteration in the extrusive volcanic terranes. Some of these solutions may have been metal-bearing, in which case large low-grade sediment-hosted precious-metal deposits (26A) and volcanic hosted epithermal percious-metal deposits (25BC) may have formed.

During Pleistocene to Holocene time radioactive black-sand and gold placers formed. Metal and mineral concentrations in the late-stage crystallization products of the Idaho batholith and the large Eocene granite batholiths were the sources for the minerals in the radioactive black-sand placer deposits. Lode gold deposits were sources for the gold in the placer gold deposits.

REFERENCE

Cox, D.P. and Singer, D.A., editors, 1986, Mineral deposit models: U.S. Geological Survey Bulletin 1693, 379 p.

PRELIMINARY ASSESSMENT OF MINERAL RESOURCES OF THE HAILEY AND WESTERN IDAHO FALLS 1°x2° QUADRANGLES, IDAHO

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INTRODUCTION

The main goal of CUSMAP projects is to provide a comprehensive mineral resource analysis - determine the character and extent of known and inferred mineral deposit types, develop models to define these mineral deposit types, outline areas of mineral resource potential, and estimate the geologic character, size, grade, and number of undiscovered deposits. The Hailey CUSMAP project is in the early stages of model development.

MINERAL DEPOSIT MODELS

Mineral deposit models used in this project are based upon the guidelines provided by Cox and Singer (1986). Many of the models used in this study are taken directly from this report although several are modifications or new additions. The deposit types listed in Worl and Johnson (this volume) follow the same guidelines. Whereas the mineral deposit types represent geologic and grade-and-tonnage characteristics compiled from local analogs, the mineral deposit models are a representation of data and interpretations for a deposit type. Data for the models come from regional and international analogs. Interpretations of the formation processes, definition of diagnostic criteria, and size/grade population estimates are all part of the models.

MINERAL RESOURCE ASSESSMENT

This is a progress report and will not define models or areas of definitive resource potential. The following maps, however, do indicate areas defined as permissible for the occurrence of several of the mineral deposit types. These areas were determined by physically comparing geologic terranes, regional geochemistry, and regional geophysical data to preliminary models based only on local and regional analogs. Areas defined as permissible were drawn where one or more parameters suggest that a particular deposit type may be present. In the final analysis the number, size, and shape of these areas will probably change.

REFERENCE

Cox, D.P., and Singer, D.A., eds., 1986, Mineral deposit models: U.S. Geological Survey Bulletin 1693, 379 p.

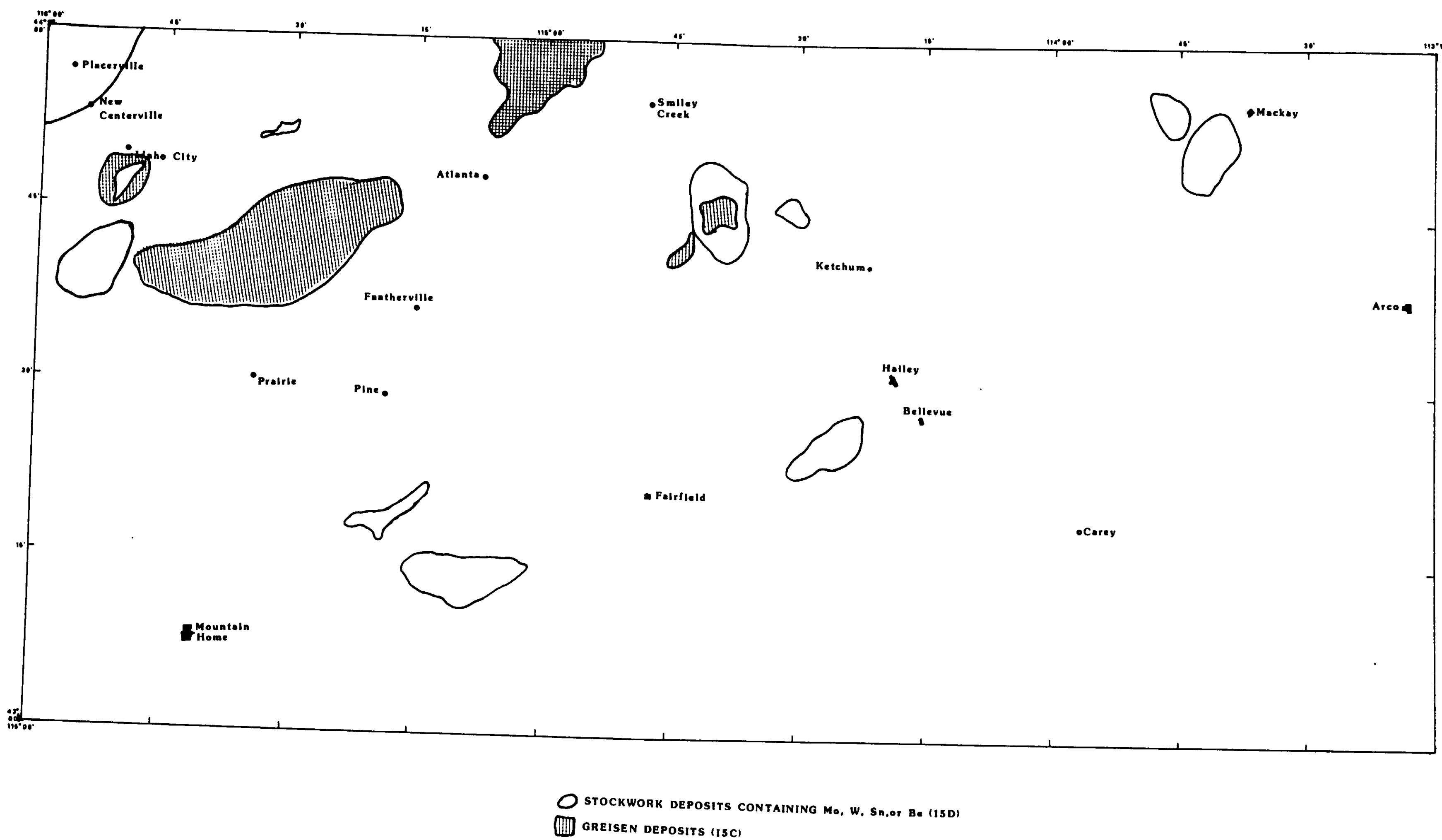

Figure 1. Map showing areas permissive for the presence of greisen deposits (15C) and stockwork deposits containing Mo, W, Sn, or Be (15D) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

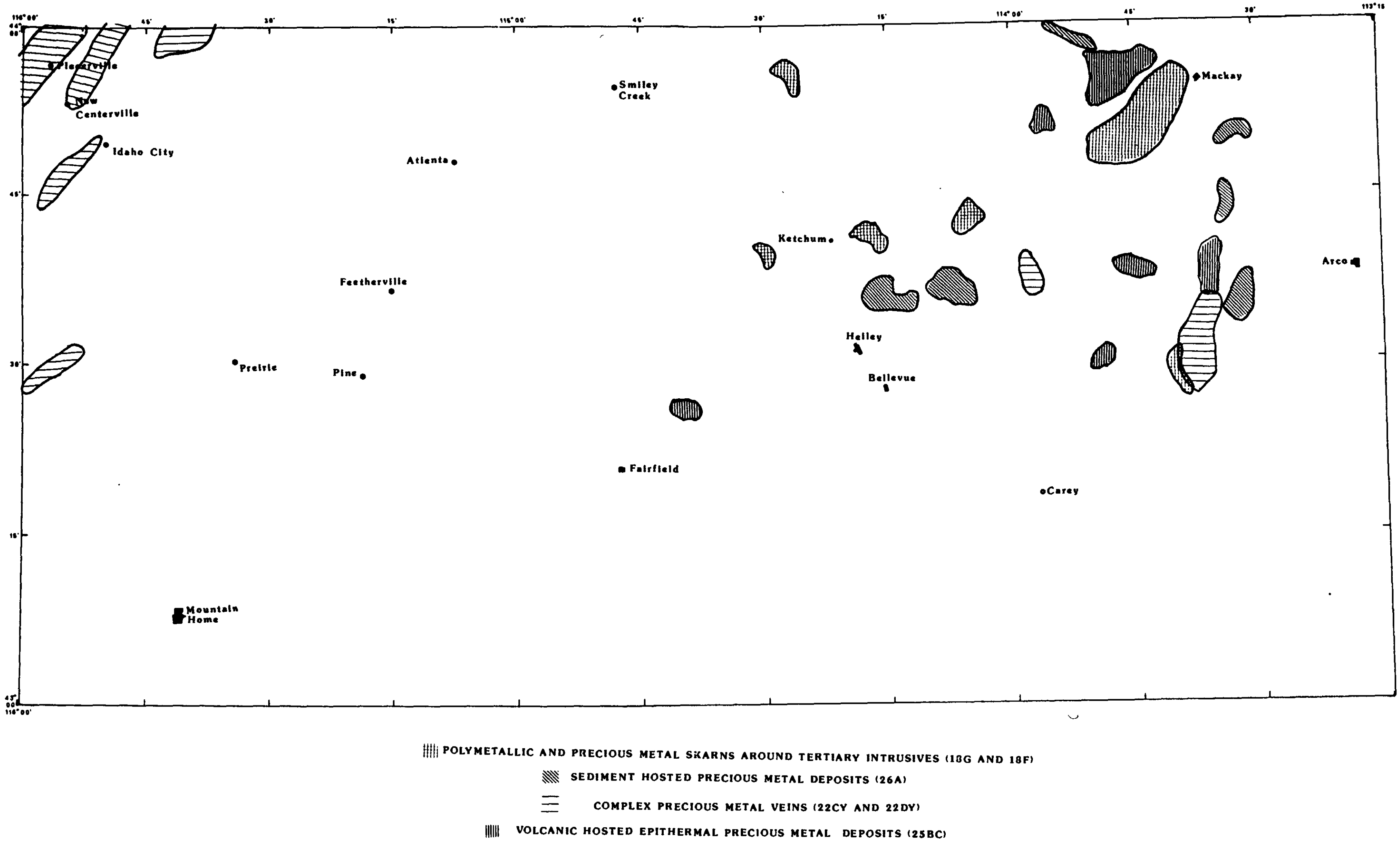

Figure 2. Map showing areas permissive for the presence of polymetallic and precious-metal skarns around Tertiary intrusives (18G and 18F), complex precious-metal veins (22CY and 22DY), volcanic-hosted epithermal precious-metal deposits (25BC), and sediment-hosted precious-metal deposits (26A) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

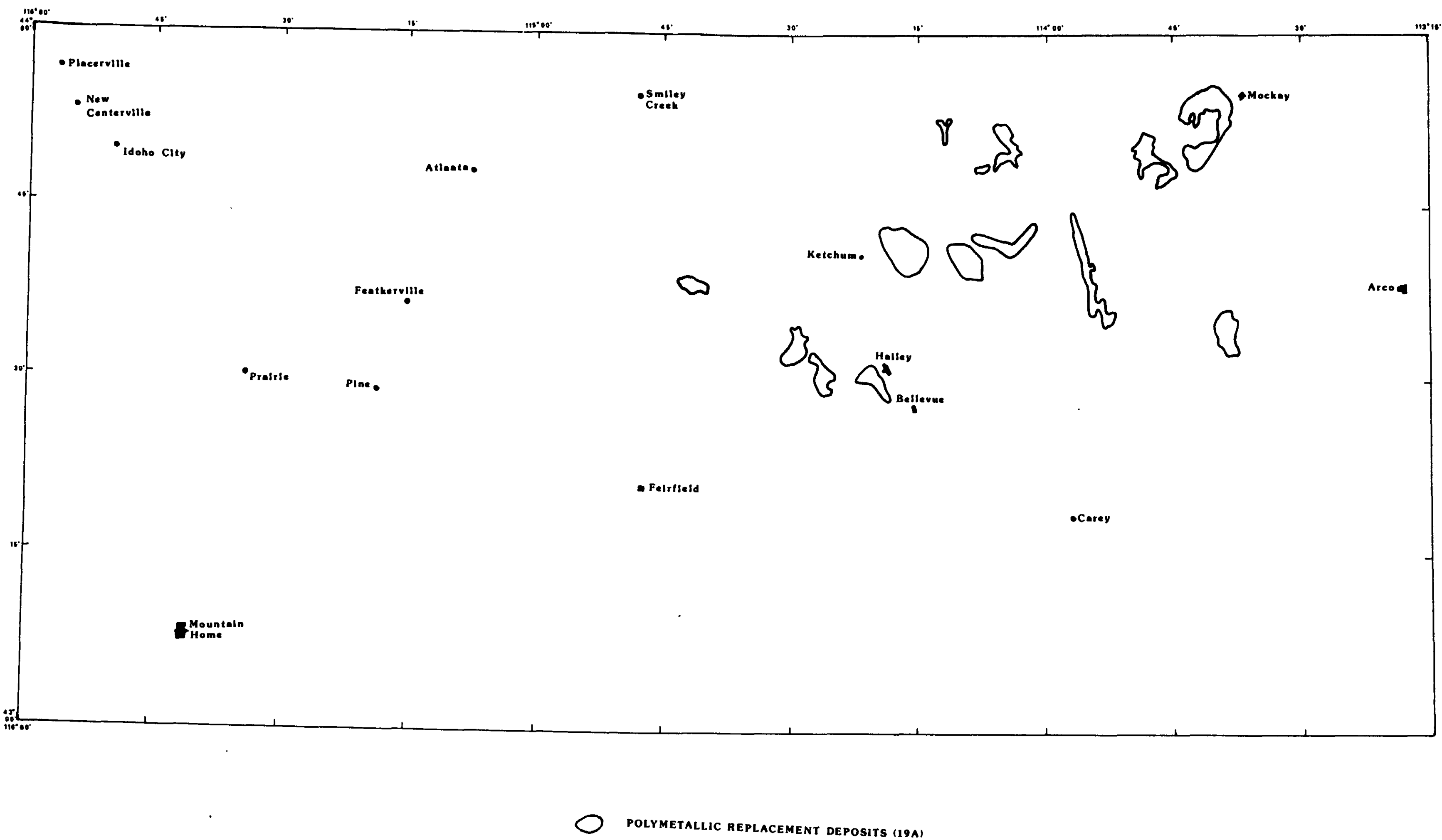

Figure 3. Map showing areas permissive for the presence of polymetallic replacement deposits (19A) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

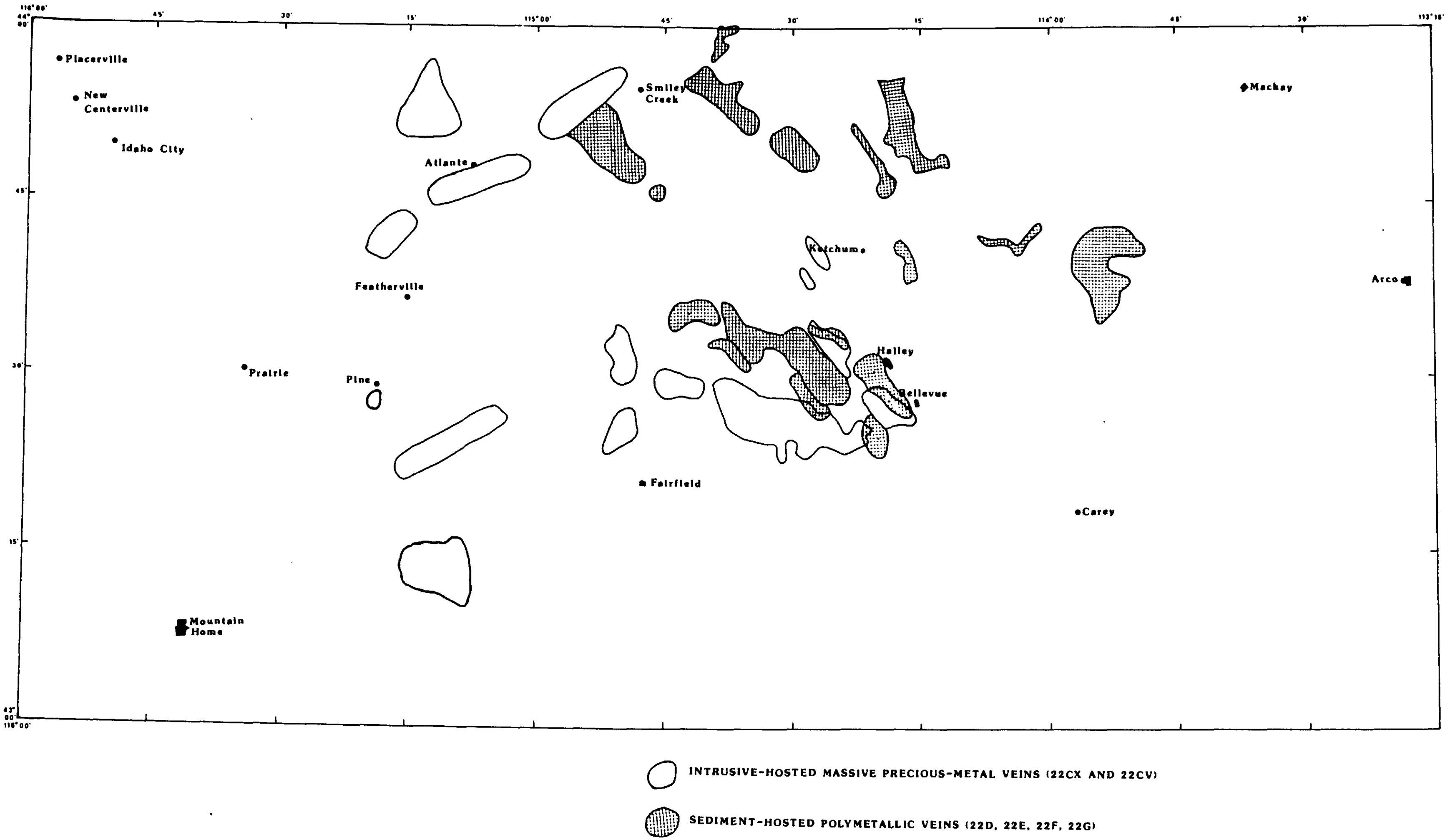

Figure 4. Map showing areas permissive for the presence of intrusive-hosted massive precious-metal veins (22CX and 22CV) and sediment-hosted polymetallic veins (22D, 22E, 22F, 22G) in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

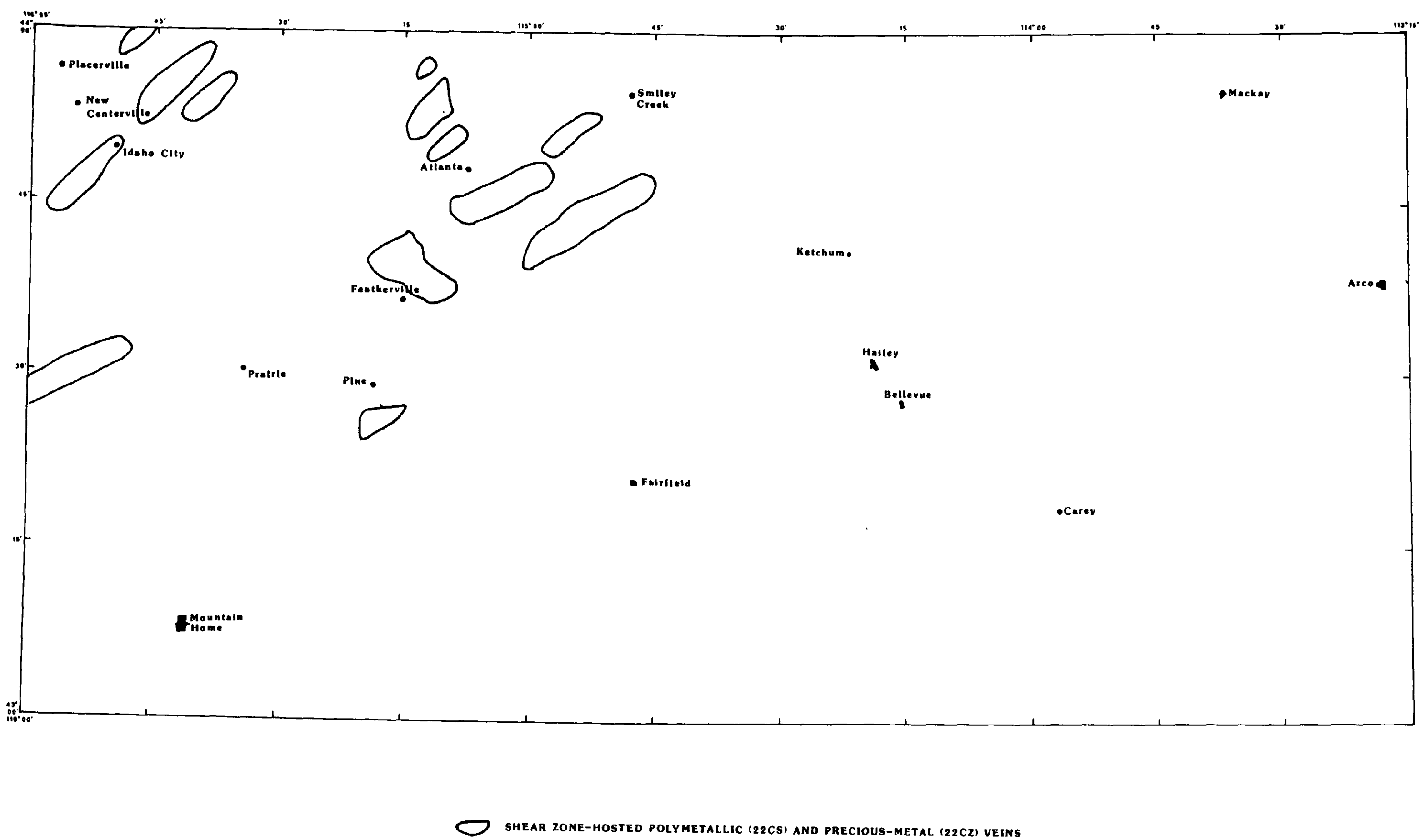

Figure 5. Map showing areas permissive for the presence of shear zone-hosted polymetallic (22CS) and precious-metal (22CZ) veins in the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

GEOCHEMICAL CHARACTERIZATION OF SOME MINERALIZED TERRANES IN THE HAILEY 1°x2° QUADRANGLE, IDAHO

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In the Hailey 1°x2° quadrangle, 1,248 geochemical samples, primarily stream sediments and soils, were collected by Savanna River Laboratory in 1980 for the National Uranium Resource Evaluation (NURE) program. In 1987 and 1988, 363 geochemical samples, primarily stream sediments, were collected by the U.S. Geological Survey as part of the Conterminous United States Mineral Assessment Program (CUSMAP). Although the combined USGS and NURE geochemical surveys have a sample density of only one sample per 4.3 square miles, most known mining districts are indicated by aomalous concentrations of one or more chemical elements.

Some anomalous sample sites, with no known mining activity in the basin drained by the sampled stream, are in areas adjacent to known mining districts. For example, a tributary of Emma Creek (fig. 1) in the north-central part of the quadrangle, has a geochemical signature (i.e. Ag, As, and Au) that is similar to that of the Vienna mining district to the north.

A few large areas that are not closely associated with known mining activity also are characterized by anomalous concentrations of some elements. Areas near Cottonwood Ranger Station, Dutch Creek Ranger Station, and Sheep Creek (all in the western part of the quadrangle) have geochemical signatures tht are distinct from those of nearby mining areas. These three areas were sampled in more detail. An additional 105 stream-sediment samples and 102 heavy-mineral-concentrate samples were collected in parts of the three anomalous areas.

The Cottonwood study area is characterized by stream sediment samples with anomalous concentrations of Ag, Bi, As, Cd, Cu, Mo, and Pb, and by samples containing anomalous concentrations of W, Sn, Mo, Th, Pb, Bi, and Ag in the nonmagnetic fraction of the heavy-mineral concentrates. This element association suggests the anomaly is caused by a shallow granitic intrusion. The presence of a Tertiary granite in the area supports this interpretation.

The Dutch Creek study area is characterized by stream sediment samples with unusually high concentrations of elements, e.g. Y, Nb, and Th, that characterize evolved granites and is adjacent to an area with stream sediment samples that contain anomalous concentrations of Sb and As. The more detailed sampling in the Dutch Creek study area found samples that contain anomalous concentrations of Au and As in the stream sediments and Au, La, Y, Pb, Th, Ag, Nb, and Bi in the nonmagnetic fraction of the heavy-mineral concentrates.

In the Sheep Creek area, stream sediment samples that are associated with a Tertiary granite are distinguished by high concentrations of Be, Y, Nb, and Th--elements that characterize evolved granites. The more detailed sampling in the Sheep Creek study area found that the nonmagnetic fraction of the heavy-mineral concentrates are characterized by high concentrations of Sn, W, Y, Nb, Bi, Be, La, and Th. The element association and rock type in the Sheep Creek area and the occurrence of topaz just east of the area, in Dismal Swamp, suggest that the granite in the Sheep Creek area is a tin granite.

The geochemical data suggest that not all Tertiary granites in the Hailey quadrangle are evolved. However, stream sediment samples associated with several Tertiary granites in the Hailey quadrangle are anomalous with respect to the ame element suite that is typical of evolved granites, and this indicates that several Tertiary granites in the Hailey quadrangle have some potential for tin enrichment.

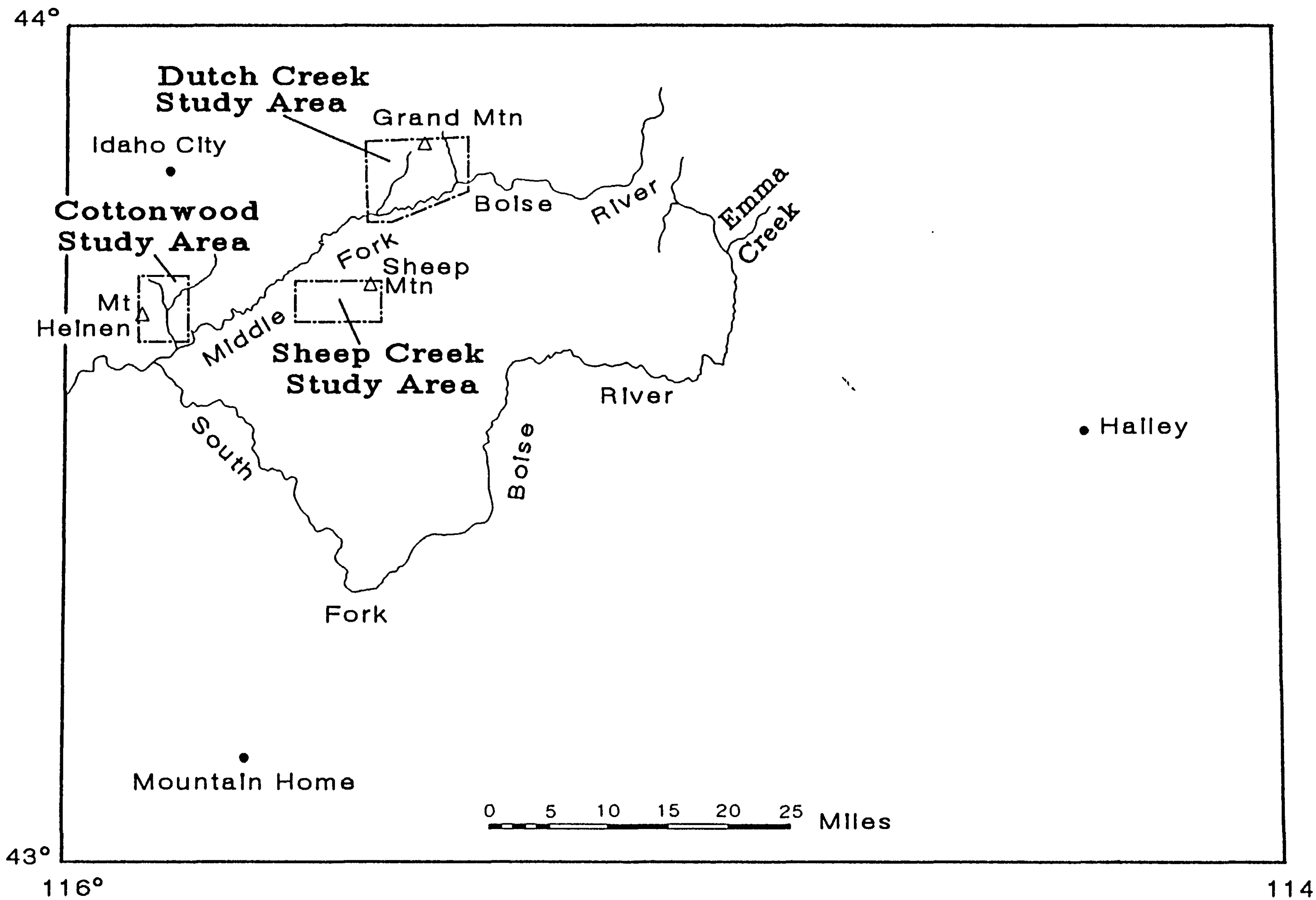

Figure 1. Map of the Hailey 1°x2° quadrangle showing locations of geochemical study areas.

GRAVITY AND MAGNETIC ANOMALY PATTERNS APPLIED TO MINERAL RESOURCE EXPLORATION, HAILEY 1°x2° QUADRANGLE, IDAHO

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INTRODUCTION

Mines and prospects containing precious metals are widely scattered in the Hailey 1°x2° quadrangle and the western part of the Idaho Falls quadrangle. Many of the deposits are associated with granitic intrusions and fault zones of the trans-Challis fault system. Earlier studies of the geophysical expressions of trans-Challis fault system for the mineral-resource assessment of the Challis quadrangle are published by Mabey and Webring, 1985. The Hailey mineral belt is of particular interest to an assessment of the Hailey quadrangle inasmuch as it occupies a position south of, but approximately parallel to, the trans-Challis system.

Interpretation of aeromagnetic and gravity anomaly data are an integral part of the assessment. The anomaly data are applied in conjunction with geological investigations to map buried intrusive complexes, delineate major fault and shear zones, and identify localities that might be hydrothermally altered or mineralized. Comparison of the magnetic and gravity anomaly data with the generalized geology illustrates the geophysical signatures of the various structural and lithologic units and suggests projections of these features in the subsurface (Kleinkopf and others, 1988a, b). The aeromagnetic anomaly data are applicable to direct detection of mineralized features, and the gravity anomaly data define structural settings and distinguish between low density Tertiary granitic intrusions and Cretaceous granitic intrusions of the Idaho batholith.

MAGNETIC DATA

The residual total intensity magnetic map (fig. 1) was compiled using data from ten aerial surveys shown on the index map (fig. 3). The surveys were flown at line spacings ranging from 0.5 to 3 miles either at a constant barometric elevation or a constant altitude above terrain. Data from these surveys were adjusted and analytically continued to a datum of 1,000 feet above terrain. The Definitive International Geomagnetic Reference Field (DIGRF), updated to the date each survey was flown, was removed before merging the surveys to obtain residual magnetic anomaly data.

In this region of Idaho, a number of significant magnetic and gravity surveys for mineral-resource exploration and regional framework studies of the region have been conducted over the past 30 years. Published state magnetic and gravity anomaly maps, at scales of 1:500,000 and 1:1,000,000 have been particularly useful for describing the general geophysical setting of the region (Bonini, 1963; Bankey and others, 1985; Zietz and others, 1978).

GRAVITY DATA

The Bouguer gravity anomaly map (fig. 2) is controlled by 2,843 unevenly spaced gravity stations. Most of this control is from the data set compiled for the recently published Bouguer gravity anomaly map of Idaho (Bankey and others, 1985). The principal facts for these data are available on magnetic

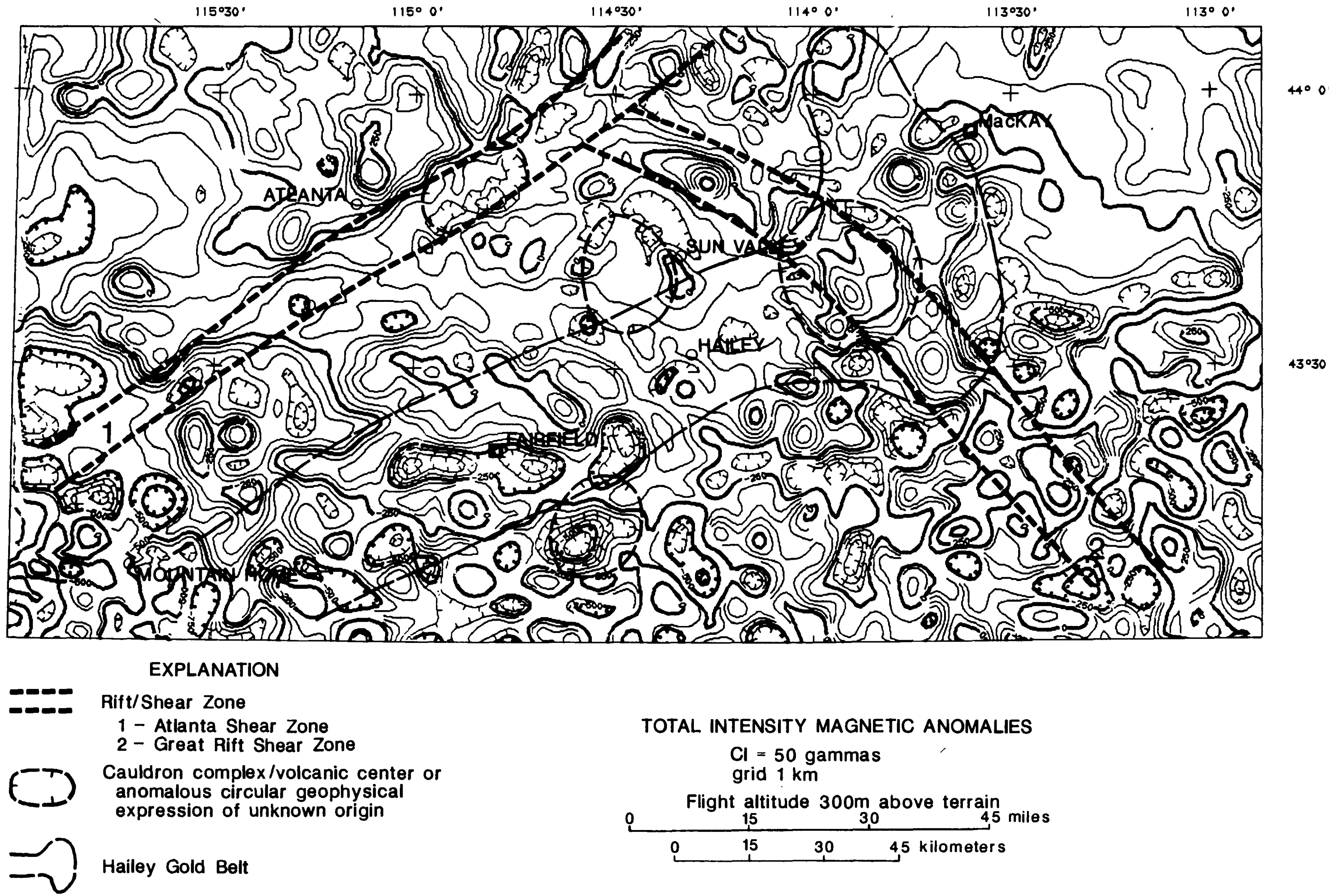


Figure 1. Residual total intensity magnetic map of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.

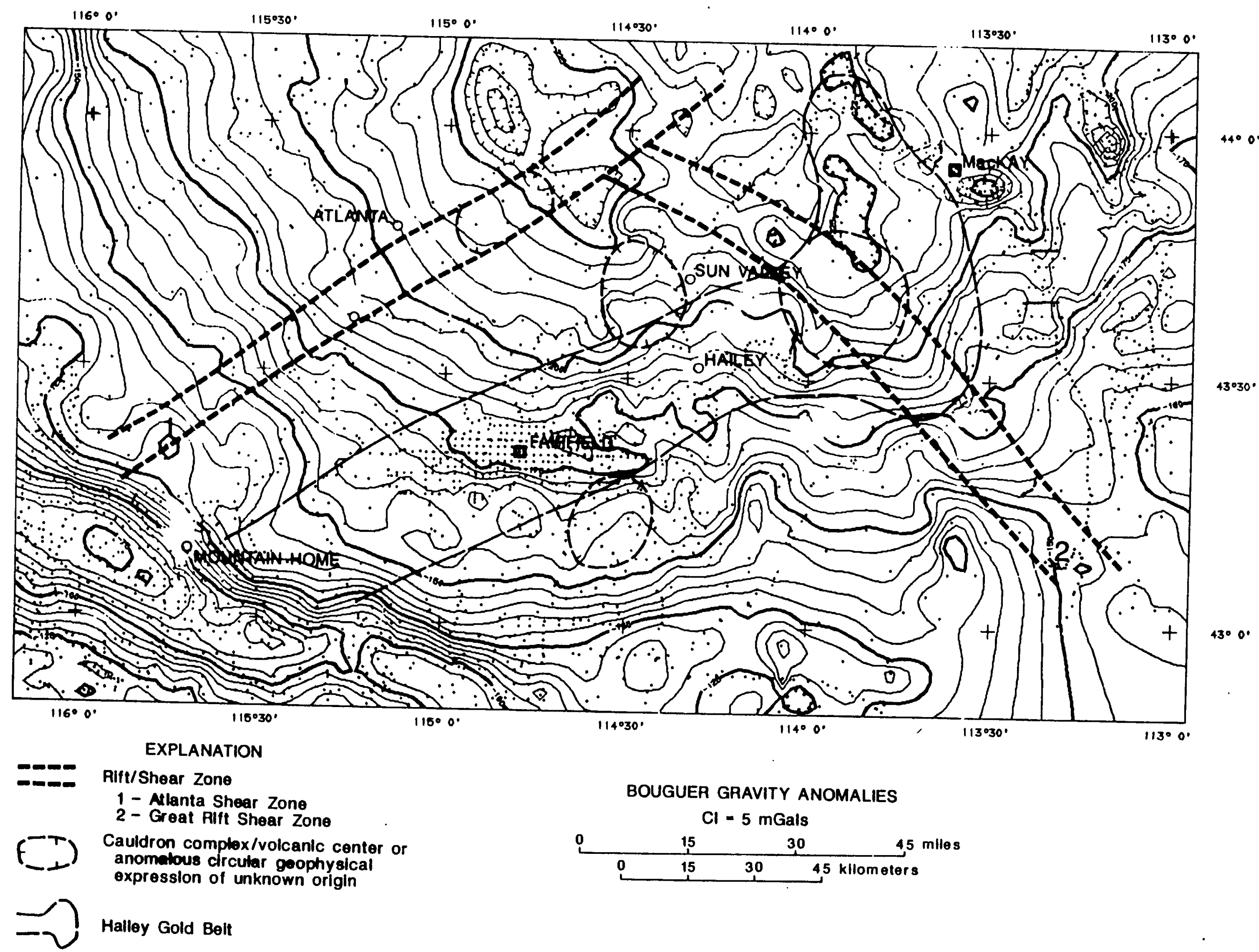
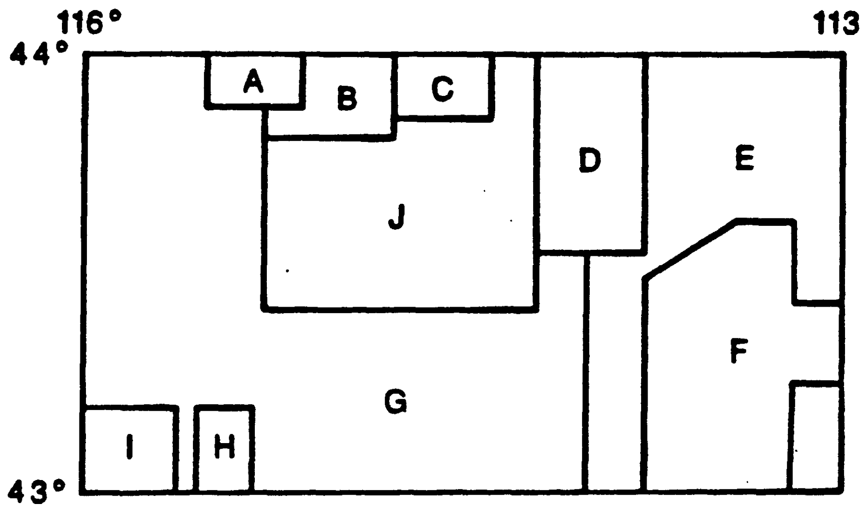


Figure 2. Bouguer gravity anomaly map of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho.



Index map showing sources of magnetic data. The direction, altitude, and spacing of flight lines are listed below

A. North-South, 1,000 feet terrane clearance, 0.5 mile (Mabey, 1986).
B. North-South, 12,000 feet barometric, 1 mile (Kiilsgaard and others, 1970).
C. North-South, 12,000 feet barometric, 1 mile (Mabey and Tschanz, 1986).
D. North-South, 12,000 feet barometric, 1 mile (USGS 1979a).
E. East-West, 400 feet terrane clearance, 3 miles (GeoMetrics, Inc. 1979a).
F. East-West, 9,000 feet barometric, 1 mile (U.S. Geological Survey, 1979b).
G. East-West, 400 feet terrane clearance, 3 miles (GeoMetrics, Inc. 1979b).
H. East-West, 5,500 feet barometric, 1 mile (U.S. Geological Survey, 1982).
I. North-South, 8,000 feet barometric, 2 miles (U.S. Geological Survey, 1974).
J. North-South, 12,000 feet barometric, 1 mile (Unpublished data).

Figure 3. Index map of individual aeromagnetic surveys compiled and merged for the Hailey quadrangle.

tape from the Data Service Officer, EROS Data Center, U.S. Geological Survey, Sioux Falls, SD 57198. These data were compiled from surveys conducted by U.S. Geological Survey, Department of Defense, and universities. About 120 stations were recently collected by the U.S. Geological Survey and are unpublished. Also included are data for nearly 300 stations collected as part of a M.S. thesis for an area of Camas Prairie (Cluer and others, 1988).

The gravity data are referenced to the IGSN'71 datum, and data are reduced to the Bouguer anomaly using the 1967 gravity formula (International Association of Geodesy, 1971) with an assumed crustal density of 2.67 g/cm^3. The recently collected gravity data were tied to local gravity bases which had been established at the towns of Bellevue, Ketchum, Fairfield, Featherville, Atlanta, and Idaho City. These bases were referenced to Department of Defense bases located at Pocatello and Boise. Terrain values were computer-calculated radially from each station to a distance of 167 km using the method of Plouff (1977). The Bouguer gravity anomaly values for the stations are estimated to be accurate to one mGal (milligal).

CRUSTAL PROVINCES

Three crustal provinces can be distinguished by their contrasting aeromagnetic and gravity signatures and provide a framework for the mineral resource evaluation: (1) In the northwest, Cretaceous granitic terranes of the Idaho batholith are weakly magnetic and have slowly varying, highly negative Bouguer anomaly values. (2) In the northeast, heterogeneous terranes of thrust and block-faulted Paleozoic sedimentary rocks, Precambrian metamorphic rocks of the Pioneer Mountains core complex, Tertiary granitic rocks, and deposits of Eocene Challis and Miocene volcanic rocks produce a variety of weak to strong magnetic anomalies. This assemblage gives an overall negative mass effect on the gravity field similar to that of the Idaho batholith. (3) In the south, the Snake River Plain province, composed predominantly of Cenozoic basalt, is much more magnetic and is denser than the terranes to the north. Boundaries between the three provinces are irregular. Granitic blocks probably locally underlie basalt near the northern margin of the Snake River Plain.

IDAHO BATHOLITH

The highest magnetic values correlate spatially with high elevation Tertiary epizonal granitic intrusions. Examples are the Sawtooth batholith, north of Atlanta; the stock at Steel Mountain, just north of Rocky Bar; and the major intrusion just east of Twin Springs. These plutons lie along the northwest side of a prominent northeast-trending steep magnetic gradient zone that suggests a buried shear zone perhaps related to northeast-structures of the trans-Challis fault system. Regional faults, such as the Deer Park, Montezuma, and Sawtooth faults, and correlative magnetic trends systematically change strike from south-southeast north of the postulated shear zone to southeast on the south side of this steep gradient zone. Along the magnetic gradient zone the gravity map exhibits low amplitude anomalies and subtle changes of gradient.

There is a change in character of the magnetic and gravity anomaly patterns across the boundaries of the Idaho batholith, which suggest that the border zones of the Idaho batholith are quite irregular with exotic blocks of granite in volcanic terranes of the Snake River Plain. In fact, the Idaho batholith probably extends beneath Paleozoic allochthons east of the Wood River graben and beyond as far as the Mackay stock.

MINERAL RESOURCES

In mineral-resource studies, magnetic anomaly data generally provide more definitive information than gravity anomaly data for identifying and mapping igneous plutons and faults that may have a potential for mineralization. This relates to nearly continuous data that are collected along the flight lines in comparison with isolated data points obtained from randomly spaced gravity stations. The gravity anomaly data provide constraints on the lithology and thickness of granitic units and on the continuity of major structural features.

The Hailey mineral belt comprises variably mineralized ground and precious- and base-metal occurrences that extend from near Mountain Home to the vicinity of Mackay, and includes parts of the White Knob and Pioneer Mountains. The belt extends some 150 km to the southwest across the southeastern part of the Idaho batholith as far as the Snake River Plain. There are a number of areas of potential mineral resources along the Hailey mineral belt. For example, an area of possible mineralization interest is beyond the Lucky Boy Mine near the intersection of strong east-west and northeast-trending structures of the trans-Challis fault system. In the eastern part of the Hailey mineral belt, the magnetic anomaly data suggest that the Mackay and Lake Creek stocks are more extensive in the subsurface, which may have implications for exploration for additional mineralized areas. These stocks are located about 10 and 5 km northeast and south-southeast respectively from Copper Basin.

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PLUTONIC ROCKS IN THE SOUTHEASTERN PART OF THE IDAHO BATHOLITH AND THEIR RELATIONSHIP TO MINERALIZATION

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Geologic mapping and geochemical sampling by the U.S. Geological Survey in the southeastern part of the Idaho batholith have revealed three plutonic rock groups of probable Late Cretaceous age. The oldest of these, the Croesus quartz diorite, is exposed south of Hailey, Idaho (fig. 1). It contains abundant pyroxene, which is absent elsewhere in the batholith, and is locally cross-cut by lamprophyre dikes. The second group, exposed primarily in the Hailey Gold belt area, is a K-rich suite, consisting of hornblende-biotite granodiorite and granite. This suite, which includes the Hailey granodiorite unit of Schmidt (1962), intrudes the Croesus quartz diorite . The third group is composed of Na-rich biotite granodiorite and Na-rich hornblende-biotite granodiorite. The Na-rich suite is widely exposed in the south-central part of the batholith. The biotite in these rocks is in thin, anhedral books, and contrasts with thick, subhedral books of biotite in the K-rich suite.

For a given SiO_2 content, the K-rich suite is higher in Rb, Cs, U, B, Mg, Sc, Cr, and Sb than the Na-rich suite. The Na-rich suite is high in Al, Ba, Sr, Zr, and light rare earth elements, and probably originated in a source area more depleted in large-ion lithophile elements than that of the K-rich suite. These suites are analogous to the sodic series and the main (K-rich) series of the Boulder batholith of Montana described by Tilling (1973).

Most of the production from the southeastern part of the Idaho batholith has been from gold-bearing quartz veins. This mineralization is probably Late Cretaceous in age. In places the veins are found alongside lamprophyre dikes, but this association is not typical. Pyrite, chalcopyrite, and arsenopyrite are the most common sulfide minerals, and the predominant alteration type is sericitic. Numerous vein deposits containing Pb-Ag-Zn minerals are present within Paleozoic metasedimentary rocks near the batholith. They probably reflect remobilization of stratiform sulfides within the metasedimentary rocks and, like the gold deposits, may be Late Cretaceous in age.

The distribution of mines and prospects in and adjacent to the southeastern part of the Idaho batholith is depicted in figure 1. Note that the Croesus quartz diorite and the K-rich suite are more highly mineralized than the Na-rich suite. The scarcity of mineralization in the Na-rich suite is perhaps a reflection of the relatively depleted source area.

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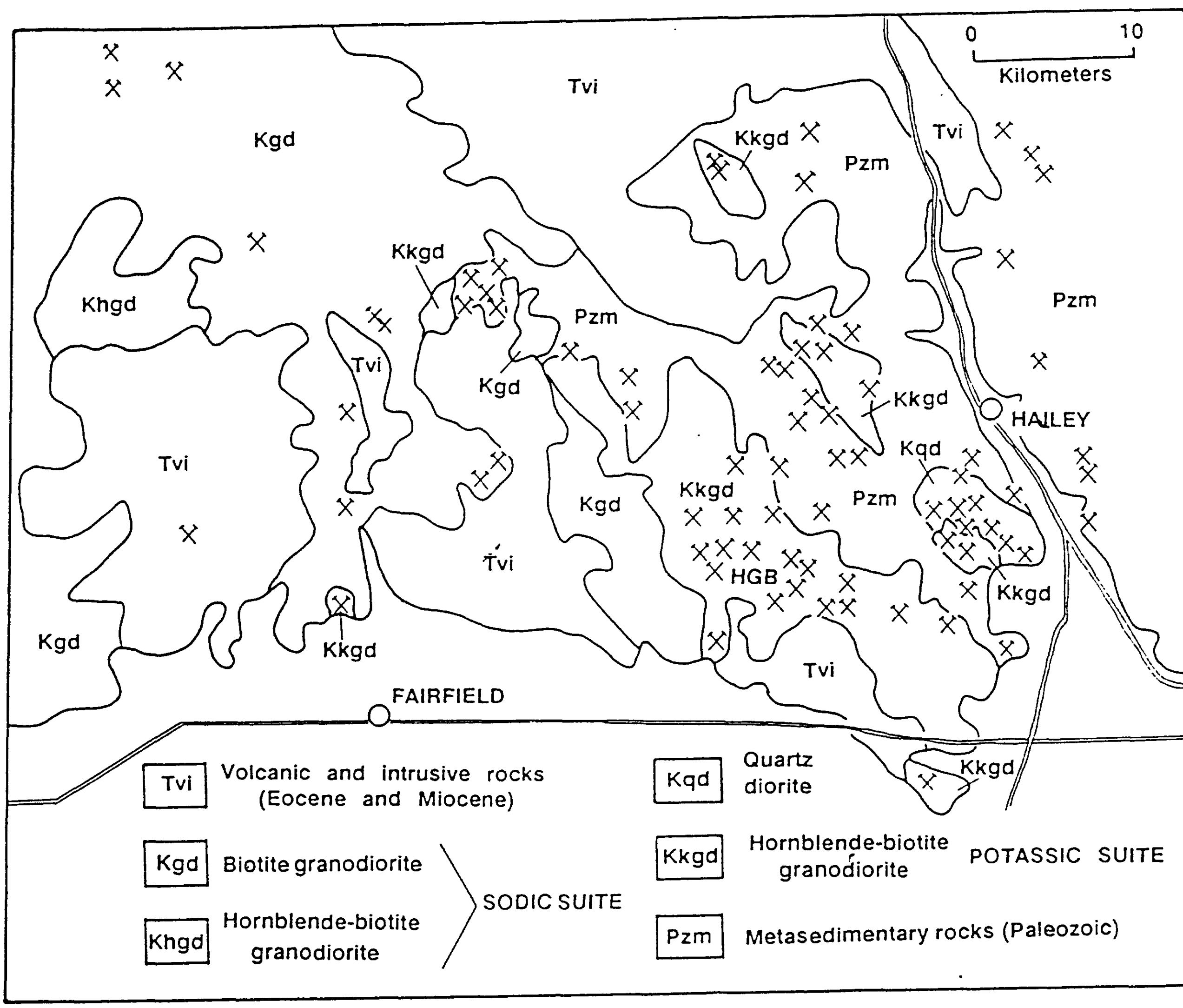

Figure 1. Simplified geologic map of the southeastern part of the Idaho batholith. HGB = Hailey Gold belt.

REGIONAL GEOLOGY OF THE NORTHWESTERN PART OF THE HAILEY 1°x2° QUADRANGLE AND ORE DEPOSITS OF THE ATLANTA DISTRICT, IDAHO

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The northwestern part of the Hailey 1°x2° quadrangle is underlain chiefly by biotite granodiorite and two-mica granite of the Idaho batholith, of Cretaceous age. These rocks are intruded by Eocene plutonic rocks that range from diorite to granite in composition. Both the Cretaceous and Eocene plutonic rocks are cut by swarms of dikes that range from lamprophyre to rhyolite in composition. Most of the dikes are porphyritic and all of them, with the possible exception of some lamprophyres, appear to be related to the Eocene plutonic rocks. The dacitic dikes are compositionally similar to the dioritic intrusives and the rhyolites are compositionally equivalent to the granite. Rhyolitic dikes, some of which cut dacitic dikes, appear to be the youngest of the porphyritic dikes. Basaltic dikes that may be related to the Columbia Plateau basalts, or even to younger basalts, are the youngest dikes.

The Cretaceous rocks and some of the Eocene plutonic rocks are cut by regional faults, the principal sets of which strike northeast or northwest. Large exposures of Eocene plutonic rocks are elongated along northeast trends, as though they were emplaced along northeast-trending faults. Most of the porphyritic dikes strike northeast, and commonly follow along northeast-striking faults.

Precious metal deposits in the Boise Basin, the Neal District, and at Atlanta are associated with northeast-trending faults. The mineralized deposits either are along the northeast-trending faults or in tensional fractures that extend obliquely away from the shear zones. Deposits in the western part of the Boise Basin are associated with faults of the trans-Challis fault system (Kiilsgaard and Lewis, 1985; Bennett and Knowles, 1985; Kiilsgaard and Bennett, 1985; Kiilsgaard, Fisher, and Bennett, 1986). Veins in the Quartzburg District strike northeast, subparallel to faults of the trans-Challis fault system. Veins in the Gambrinus District, however, strike northwest, a feature, which along with other geologic criteria led Anderson (1947) to conclude that the veins were older than those in the Quartzburg District. Our studies suggest that the northwest-striking veins are along tensional fractures formed from lateral displacement along the trans-Challis fault system and are of the same age as those in the Quartzburg District. Veins near Quartzburg cut rhyolite dikes, which, in turn, intrude Eocene granodiorite, thus the mineralized veins can be no older than Eocene.

More gold has been produced from the Boise basin than any other area in Idaho. Gold production, through 1982, is estimated at about 2.8 million ounces.

The Neal District is about 15 miles east of Boise, and has produced about 35,000 ounces of gold. The principal part of the district was being intensively explored by drilling in 1989. Gold has been mined chiefly from northeast-trending veins that cut Cretaceous biotite granodiorite. Subparallel to and near the veins are pre-mineral lamprophyric dikes. A number of rhyolite dikes crop out near and in the mineralized area, which is bounded by northeast-trending regional faults. The Sunshine vein strikes northwest and appears to be a gash vein related to tensional adjustment along the northeast-trending faults.

The Atlanta lode was discovered in 1864 and has been mined sporadically since then, principally through three underground mines. Production since 1900 amounts to about 260,000 ounces of gold and 990,000 ounces of silver. Total production of gold is not known, but based on incomplete figures probably exceeds 400,000 ounces. At times, the district has been the largest gold producer in Idaho. The lode is about 2 miles long and ranges from 40 to 120 feet in thickness. It consists of argillized biotite granodiorite of the Idaho batholith in which are silicified and mineralized zones. The lode strikes N. 50 to 75 E., and dips steeply. Ore shoots within it have been mined to a depth of 1,100 feet. The lode is terminated at the northeast end by the Montezuma fault.

The Atlanta lode is a strong northeast-trending fault zone that cuts biotite granodiorite of the Idaho batholith. The granitic rocks in and along the zone are argillically altered and, locally, were subsequently silicified. Ore minerals were deposited concurrently with silicification (Anderson, 1939). Perhaps the most striking features of the lode are the extensive gash veins that intersect the lode obliquely and which extend northwest and southeast away from the lode. The gash veins are tensional fractures that have formed from structural adjustment along the lode, and, together with structural openings within the lode made the area permeable to mineralizing solutions from which were deposited the ore minerals.

Exploration of the Atlanta lode was resumed in 1985 by the Atlanta Gold Corporation, who, through 1988, drilled 452 reverse circulation and diamond drill holes. This drilling program has outlined two blocks of gold-bearing material that may be mined by open-pit methods. Reserves of the East block are calculated at 8.2 million tons that average 0.087 oz gold per ton, whereas reserves of the West block are calculated at 6.2 million tons that average 0.049 oz gold per ton. Much of the gold is contained in sulfide minerals, for which concentration by flotation is planned. The current objective is to begin mining operations in 1990.

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$^{40}Ar/^{39}Ar$ THERMOCHRONOLOGY OF MINERAL DEPOSITS IN THE SOUTHERN PART OF THE IDAHO BATHOLITH

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Age and origin of mineral deposits developed in roof pendants and Cretaceous-age plutons of the Idaho batholith have been debated since Anderson (1951) attributed the origin of many of the deposits to Eocene hydrothermal activity. $^{40}Ar/^{39}Ar$ thermochronologic studies of the Marshall Mountain (M), Warren (W), Atlanta (A), and Rocky Bar (R) gold mining districts, which produced nearly 25 percent of Idaho's total 8.3 million ounces of gold between 1862 and 1962, and the Hermada antimony mine (northwest of Atlanta), which was active during World War II, are underway. Argon isotopic dates indicate that these deposits are Cretaceous and Paleocene in age. Muscovite from quartz veins at the Lucky Strike mine (M) is 75.6±0.4 Ma (millions of years old). In the Warren district, quartz vein muscovite at the War Eagle mine is 74.5±0.4 Ma and nearby wall rock muscovite is 74.6±0.4 Ma. Coarse- and fine-grained muscovite from quartz veins near the glory hole at the Atlanta mine (A) give plateau dates of 69.1±0.4 and 68.9±0.4 Ma, respectively. Muscovite from altered granitic wall rock near the North Vein (Atlanta district) shows argon loss from 69 to 57 Ma and muscovite from the vein yielded an identical age spectrum; K-feldspar cogenetic with North Vein quartz and molybdenite has argon loss from 74 to 58 Ma. Hydrothermal muscovite formed in fractures in altered Cretaceous granitic rock along the Yuba River, west of Atlanta, has a plateau date of 66.8±0.4 Ma with minor argon loss from 74.5 to 66 Ma. Three muscovite separates from the Ada-Elmore and Ida-Elmore mines at Rocky Bar have plateau dates of 57.5±0.3, 57.3±0.3, and 58.3±0.3 Ma. Muscovite from the Hermada mine shows severe argon loss from the Cretaceous, but has a plateau date of 61.2±0.3 Ma, which is thought to be the age of alteration associated with stibnite deposition.

These new data combined with previously reported results from the Edwardsburg, Profile Gap, Buffalo Hump, Dixie, and Center Star districts (north and east of districts included in this study) provide information about age, duration, number of thermal pulses, and temperature of mineralization for deposits throughout much of the southern part of the Idaho batholith. Based on over 50 age spectra, mineralization took place regionally between 78 and 57 million years ago. Within that time-range, most thermal and alteration activity occurred during four periods, namely 78-76, 74-70, 68-66, and 61-57 Ma; a geographic zonation in the ages of mineralization is apparent in the data. However, most of the larger deposits were affected during more than one of these episodes and the effects of the younger episodes are commonly exhibited by minerals first-formed during older events. Evidence for Tertiary thermal activity at temperatures of greater than 250 °C is generally lacking but low-temperature (150 °C or less) activity is recorded in three samples at 52 Ma. Thus, regional-scale propylitic alteration apparently caused by meteoric-hydrothermal activity spatially related to Eocene (about 48-46 Ma) plutonic rocks had no thermal effect on the muscovite or K-feldspar from any of these deposits.

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ORIGIN OF ORE DEPOSITS IN THE CENTRAL IDAHO BLACK SHALE BELT--IMPLICATIONS FROM LEAD ISOTOPES

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INTRODUCTION

The black shale belt of central Idaho has been a significant area of silver-lead-zinc production since the late 1800's (Hall and others, 1978; Hall, 1985). This area was one of the first mining districts to be studied using lead isotopes. Lead isotopes in galena from the Minnie Moore mine in the Big Wood River valley, central Idaho, were noted as unusually radiogenic for deposits of this size in the western U.S. (Doe, 1978). Metals were thought to be derived from the enclosing host rocks by lateral secretion (Hall and others, 1978). Regression of $^{207}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ data yielded isochron ages of 2500 and 1400-1600 Ma (million years) for the primary source material (Small, 1968; Doe and Delevaux, 1985). Sulfur isotopic data suggest that sulfur was derived from crustal sources such as syngenetic sulfide minerals in sedimentary rocks (Howe and Hall, 1985). Hydrogen and oxygen isotopic studies suggest that the hydrothermal fluid that transported the metals was meteoric water driven by heat from igneous intrusions (Criss and others, 1984, 1985; Howe and Hall, 1985). The present study aims to extend the earlier isotopic studies to cover much of central Idaho and to investigate in more detail the sources of lead and sulfur as a function of deposit type and host rock lithology.

METHODS

This study synthesizes lead isotopic data from current studies, unpublished work of Doe and Delevaux (1965), and previously published studies (Small, 1968; Davis, 1978; Hall and others, 1978). Sample material includes galena from veins and from epigenetic concentrations comformable to bedding (including both syngenetic ore and bedding-parallel epigenetic ore), whole rock pyrite-rich argillites, whole rock granitoids and volcanic rocks, and feldspar separates from granitoids. This data set includes results for 63 samples obtained by Doe and analyzed under the supervision of M.H. Delevaux (Doe and Delevaux, 1985) by Stephen Kish, Ake Johansson, and Holly Stein during 1982-1984, 59 samples analyzed by Wooden in 1989, 19 samples by Small (1968), 9 by Davis (1978), and 6 reported by Hall and others (1978).

Samples of Doe and Delevaux (1985) were analyzed as follows. Lead separated from whole-rocks and minerals were analyzed by the surface-emission (silica gel) ionization-technique of solid-source mass spectrometry. The mass-analysis methods for all samples and lead-separation techniques used for silicates are those of Doe and Delevaux (1980). For lead-rich samples, e.g. galena, jamesonite, and boulangerite, an HCl dissolution was substituted for the HF-HCl_4 decomposition step on silicates and thence the dissolved material was transferred directly to anodic electrodeposition. The rest of the procedures were the same except when bismuth or thallium signals were observed. Although isotopes of these elements do not directly interfere with lead isotopes, they are at adjacent masses and add to uncertainties in the background. These samples were reprocessed on chloride-anion resin columns eluted with water to separate bismuth and thallium from lead.

A slightly different variation of the silica-gel technique was used by Wooden. Sulfide samples were prepared by hand picking a few milligrams of ore and dissolving it in 1-2 ml of concentrated HNO_3. After 12 hours, this solution was diluted to about 15 ml total volume with distilled water and heated at 50-80°C for another 12 hours. Approximately 10 microliters of this final solution were loaded directly on a bed of silica gel-H_3PO_4 that had been partly dried on a single Re filament. Whole-rock shale samples were treated with concentrated HNO_3 to dissolve sulfides and then given a normal HF dissolution treatment for the silicate residue. The total sample was centrifuged to remove graphite and then lead was separated by the standard HBr-HCl technique on anion exchange columns. Loading of samples for mass spectrometry followed standard procedures. All samples were analyzed for lead isotopic composition by simultaneous collection of all four isotopes on multicollector Finnigan-MAT mass spectrometers. Isotopic ratios are corrected for thermal fractionation by 0.11 percent per mass unit based on average values of numerous analyses of NBS-981 and 982 determined at the same operating conditions as samples.

$^{208}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ RESULTS

Results are presented by lithostratigraphic unit. The Milligen Formation in the Hailey 1°x2° quadrangle is the major Devonian unit. The undated Triumph block may be correlated with the Milligen (Otto and Turner, 1988) or the Permian Dollarhide (Wavra, 1988). Mississippian units include the Salmon River Group in the Challis quadrangle and Copper Basin Formation, mostly in the Idaho Falls quadrangle. Recent work suggests that the Salmon River Group may be Mississippian. In this paper, the Triumph data are presented with the Milligen Formation data and the Salmon River Formation data are presented with the Mississippian data simply for convenience. We do not necessarily endorse either assignment, and our conclusions based on lead isotopes are believed to be valid regardless of assignment. The Pennsylvanian-Permian Wood River Formation was sampled in the Challis and Hailey quadrangles, and the Permian Dollarhide and Grand Prize Formations were sampled in the Hailey quadrangle.

The variation in $^{208}Pb/^{204}Pb$ is moderately independent of $^{206}Pb/^{204}Pb$ and has been used successfully in discriminating between source terranes (e.g. Zartman, 1974). The $^{207}Pb/^{204}Pb$ ratio is more correlated with $^{206}Pb/^{204}Pb$ but can also be useful.

Samples from the Milligen Formation and Triumph block include galena from epigenetic veins in the Milligen Formation such as at the Silver Star Queen mine, galena from conformable ore of the Triumph deposit, which is thought to be syngenetic, galena from epigenetic veins in the Triumph block, and pyrite-rich argillite whole rocks (fig. 1).

The $^{208}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ diagram shows two distinct groups, one with $^{206}Pb/^{204}Pb$=19.2-20.2 and $^{208}Pb/^{204}Pb$=39.2-40.0 (type B) and the other with $^{206}Pb/^{204}Pb$=20.2-20.6 and $^{208}Pb/^{204}Pb$=40.8-41.4 (Type A2). The first group comprises all the whole rocks, all the syngenetic Triumph mine ore, and all the samples of vein galena from the Triumph block except for one from the Old Triumph mine. The second group comprises all the galena samples from the Silver Star Queen and the one sample from the Old Triumph mine.

Samples from Mississippian lithologies include galena from bedded replacement ore of the Phi Kappa Mine, galena from veins in the Copper Basin and Salmon River Formations, and galena from syngenetic ore in the Livingston and Hoodoo mines in the Salmon River Formation (fig. 2).

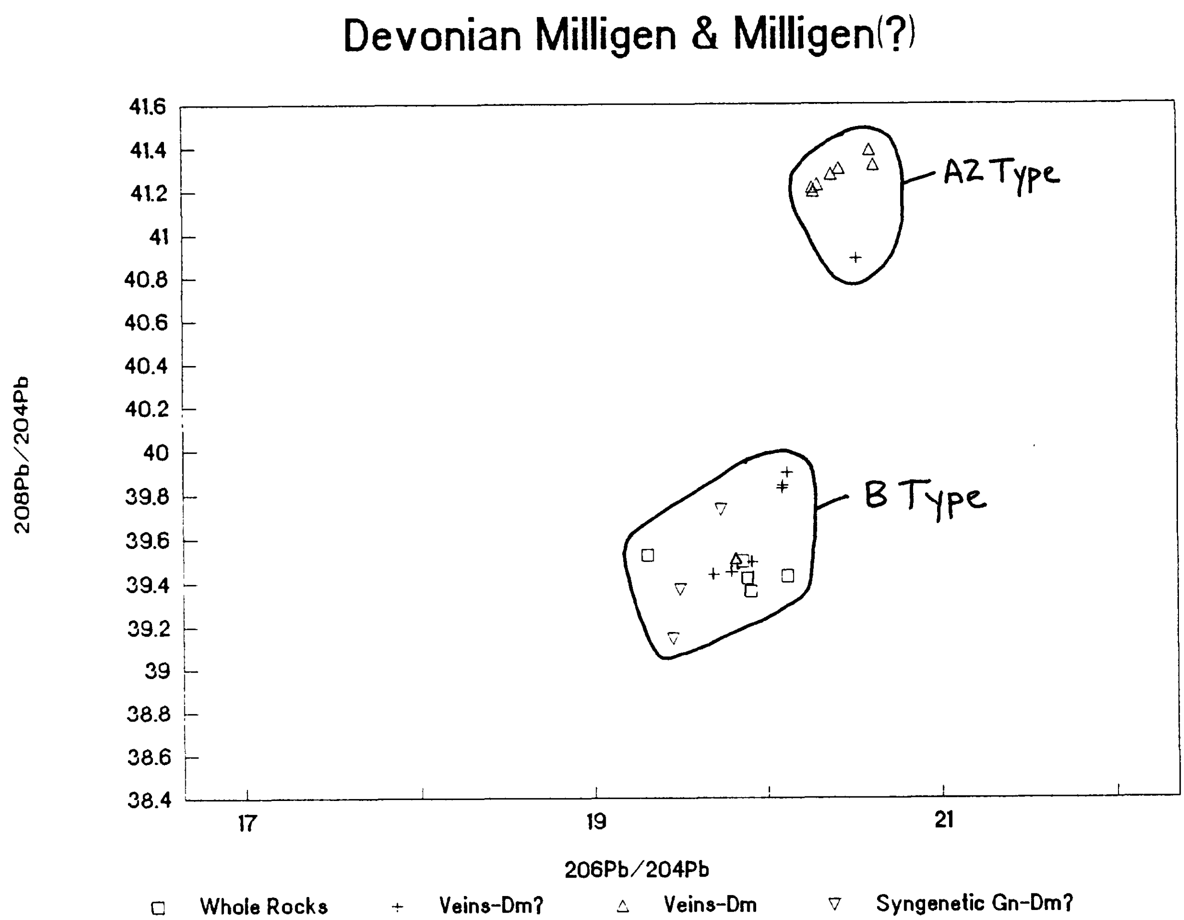

Figure 1. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram for vein and syngenetic galena and whole-rock samples from the Milligen Formation (Dm) and Triumph block (Dm?). Squares, pyrite-bearing argillite whole rocks; crosses, vein galena from the Triumph block and Snoose mine; upward-pointing triangles, vein galena from the Milligen Formation; downward-pointing triangles, conformable-syngenetic galena (Gn) from the Triumph mine.

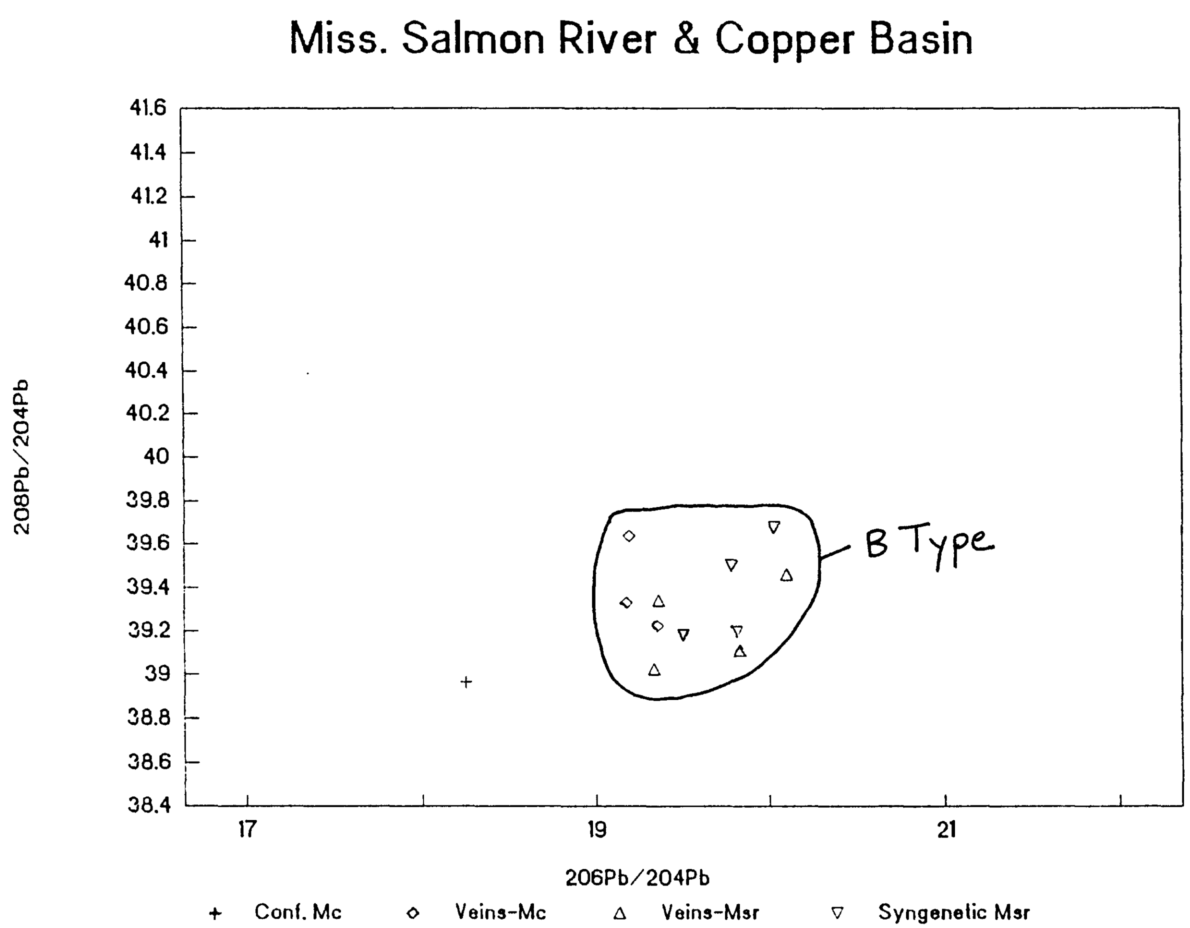

Figure 2. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram for vein and syngenetic galena from the Mississippian Salmon River Group (Msr) and Copper Basin Formation (Mc). Cross, replacement galena from the Phi Kappa mine, Copper Basin Formation; diamonds, vein galena from the Copper Basin Formation; upward-pointing triangles, vein galena from the Salmon River Group; downward-pointing triangles, conformable-syngenetic galena from the Triumph mine.

All but one of the samples falls in a group like the lower $^{208}Pb/^{204}Pb$ group in figure 1. Only the galena from the Phi Kappa mine is anomalous.

Samples from Pennsylvanian and Permian rocks include galena from epigenetic veins in the Wood River, Dollarhide, and Grand Prize Formations, galena from bedding-conformable concentrations in the Dollarhide Formation, and pyrite-rich argillites from the Dollarhide Formation (fig. 3).

Isotopic ratios define three, perhaps four, groups of samples. One group, having $^{206}Pb/^{204}Pb$=18.0-19.2 and $^{208}Pb/^{204}Pb$=39.5-40.0, comprises some Dollarhide-hosted vein and bedding-conformable galena. Deposits in this group include the Buttercup mine and all those sampled from the Carrietown mining district A second group, having $^{206}Pb/^{204}Pb$=19.9-20.4 and $^{208}Pb/^{204}Pb$=40.5-41.1, comprises the remainder of the Dollarhide-hosted vein and bedding-conformable galena. This group includes the Eureka, Liberty Gem, Jay Gould, and Idahoan mines, for example. A third group, having $^{206}Pb/^{204}Pb$=19.2-20.4, comprises four whole rocks from the Dollarhide and the one vein galena from the Grand Prize Formation in a subgroup of lower $^{208}Pb/^{204}Pb$ and all the vein samples of galena in the Wood River Formation in a subgroup having higher $^{208}Pb/^{204}Pb$.

Lead isotopes were analyzed for epigenetic vein galena in Cretaceous and Tertiary granitoids and Eocene Challis volcanic rocks (fig. 4). The entire set of igneous-rock hosted galena samples covers a wide range of $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ values, but all have $^{206}Pb/^{204}Pb$ less than 19.2. The veins from the Challis quadrangle have a wider range than those from the Hailey quadrangle. Those from the Hailey quadrangle can be divided into three moderately overlapping groups corresponding to geographical areas within the Idaho batholith. One group, having $^{206}Pb/^{204}Pb$=17.0-18.0, includes the three samples from the south-central part of the Idaho batholith, representing the Featherville, Pine, and Volcano mining districts. A second group, having $^{206}Pb/^{204}Pb$=17.8-18.4, comprises all the samples from the Vienna District in the east-central part of the Idaho batholith. A third group, having $^{206}Pb/^{204}Pb$=18.0-19.2, comprises veins in the southeast part of the Idaho batholith, composed of the Big Smokey, Camas, Mineral Hill, and Soldier mining districts.

In summary, the $^{208}Pb/^{204}Pb$ results indicate three clearly distinct groups of lead isotope ratios, low $^{206}Pb/^{204}Pb$-low $^{208}Pb/^{204}Pb$, high $^{206}Pb/^{204}Pb$-high $^{208}Pb/^{204}Pb$, and intermediate $^{206}Pb/^{204}Pb$-low $^{208}Pb/^{204}Pb$. The first two groups correspond to Doe's "A" type (Doe and Delevaux, 1985) and are designated A1 and A2, respectively for purposes of this paper. They also correspond to type Ia of Zartman (1974). The third group corresponds to Doe's group "B" and Zartman's type II, and it will be referred to as type B.

$^{207}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ RESULTS

Based on the above groupings, $^{207}Pb/^{204}Pb$ data are presented separately for types A and B lead (figs. 5 and 6). The type A lead (fig. 5) displays a fairly linear trend that yields a Pb-Pb isochron age of 2.2 $\pm$ 0.2 Ga. Four samples are clearly anomalous. Visually, the A1 and A2 groups appear to have slightly different slopes.

The $^{207}Pb/^{204}Pb$ data for type B samples reveal two subgroups (fig. 6). Syngenetic galena from the Triumph, Hoodoo, and Livingston mines as a group (B1) have higher $^{207}Pb/^{204}Pb$ than whole rock samples from the Milligen and Dollarhide Formations as a group (B2) for the same $^{206}Pb/^{204}Pb$ values. Vein

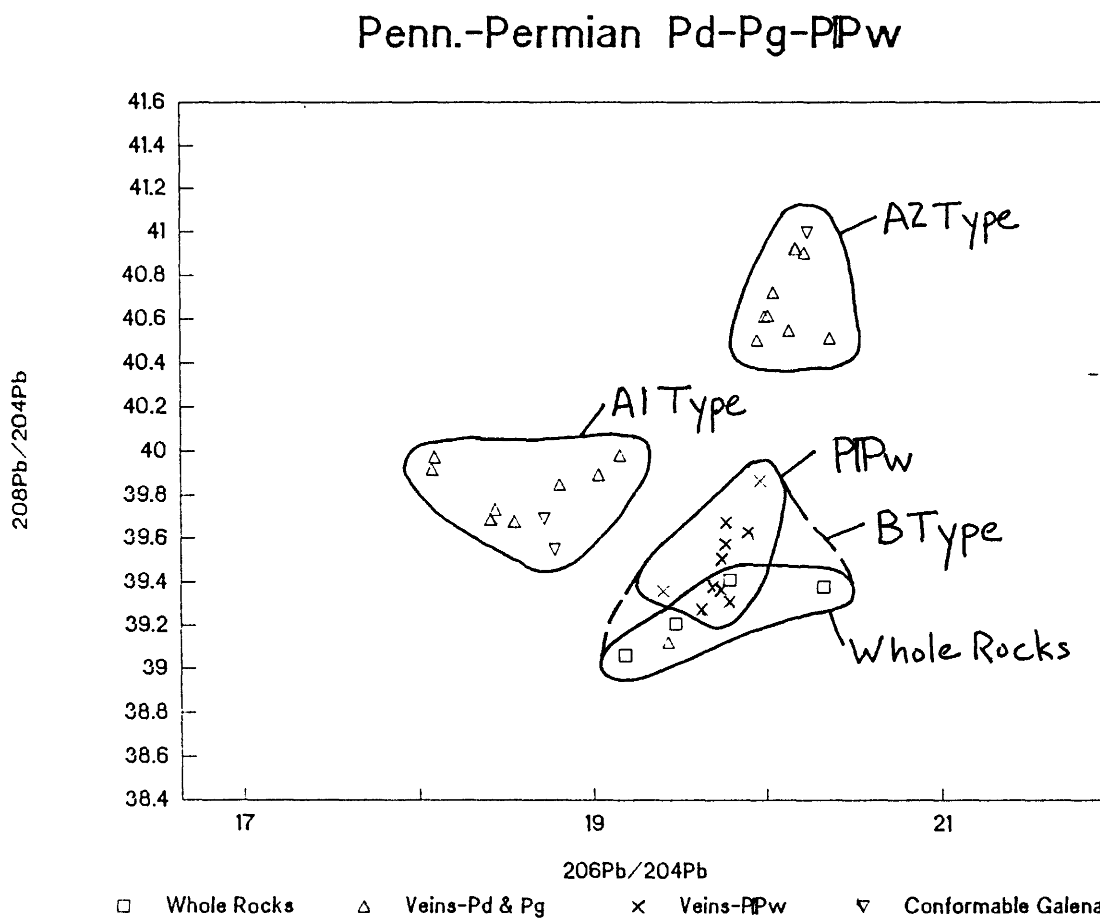

Figure 3. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram for vein and conformable-epigenetic galena and whole rocks from the Wood River (PPw), Dollarhide (Pd), and Grand Prize (Pg) Formations. Squares, pyrite-bearing argillite whole rocks from the Dollarhide Formation; upward-pointing triangles, vein galena from the Dollarhide and Grand Prize Formations; X's, vein galena from the Wood River Formation; downward-pointing triangles, conformable-epigenetic galena from the Dollarhide Formation.

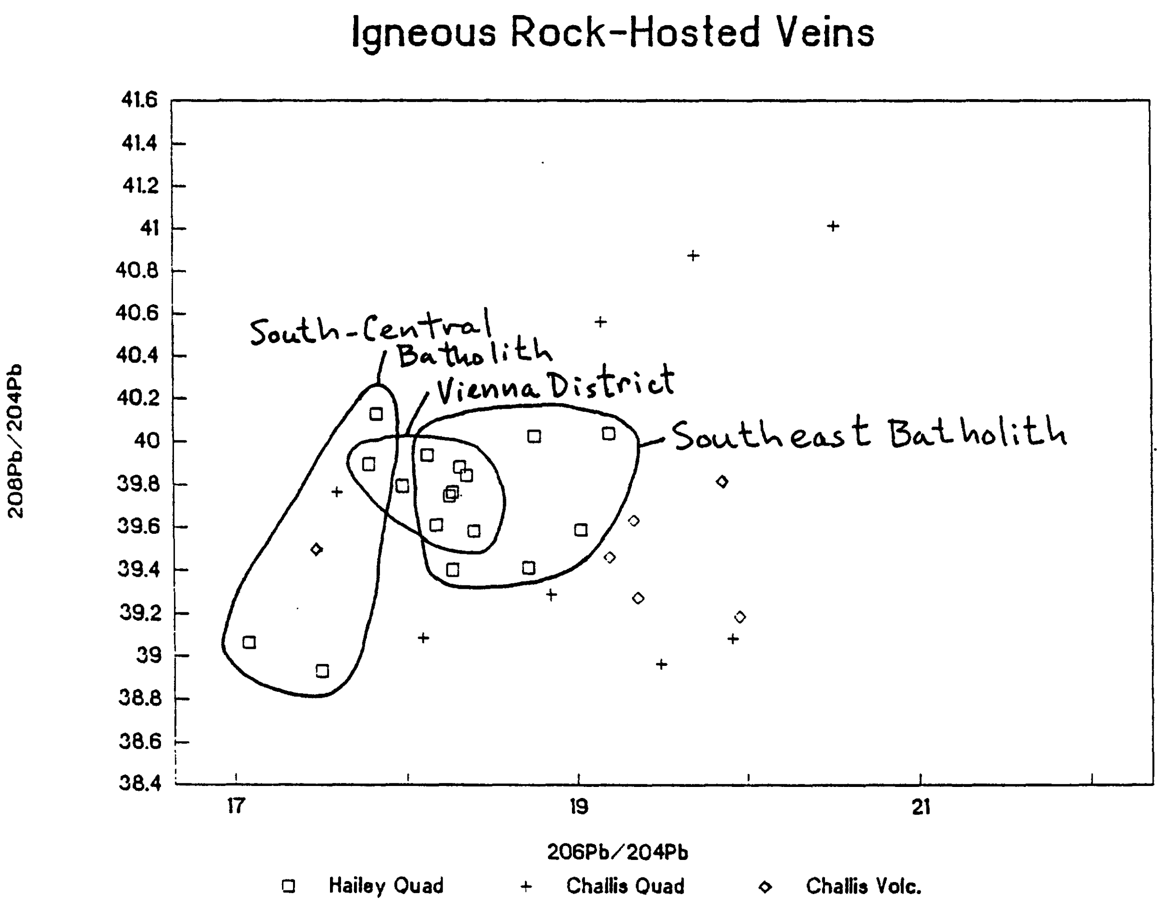

Figure 4. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram for vein galena from igneous rocks. Squares, galena from granitoid-hosted veins in the Hailey 2 degree quadrangle; crosses, galena from granitoid-hosted veins in the Challis 2 degree quadrangle; diamonds, galena from the Challis volcanics. Lines outline samples from south-central, east-central, and southeast Idaho batholith, in order of increasing $^{206}Pb/^{204}Pb$.

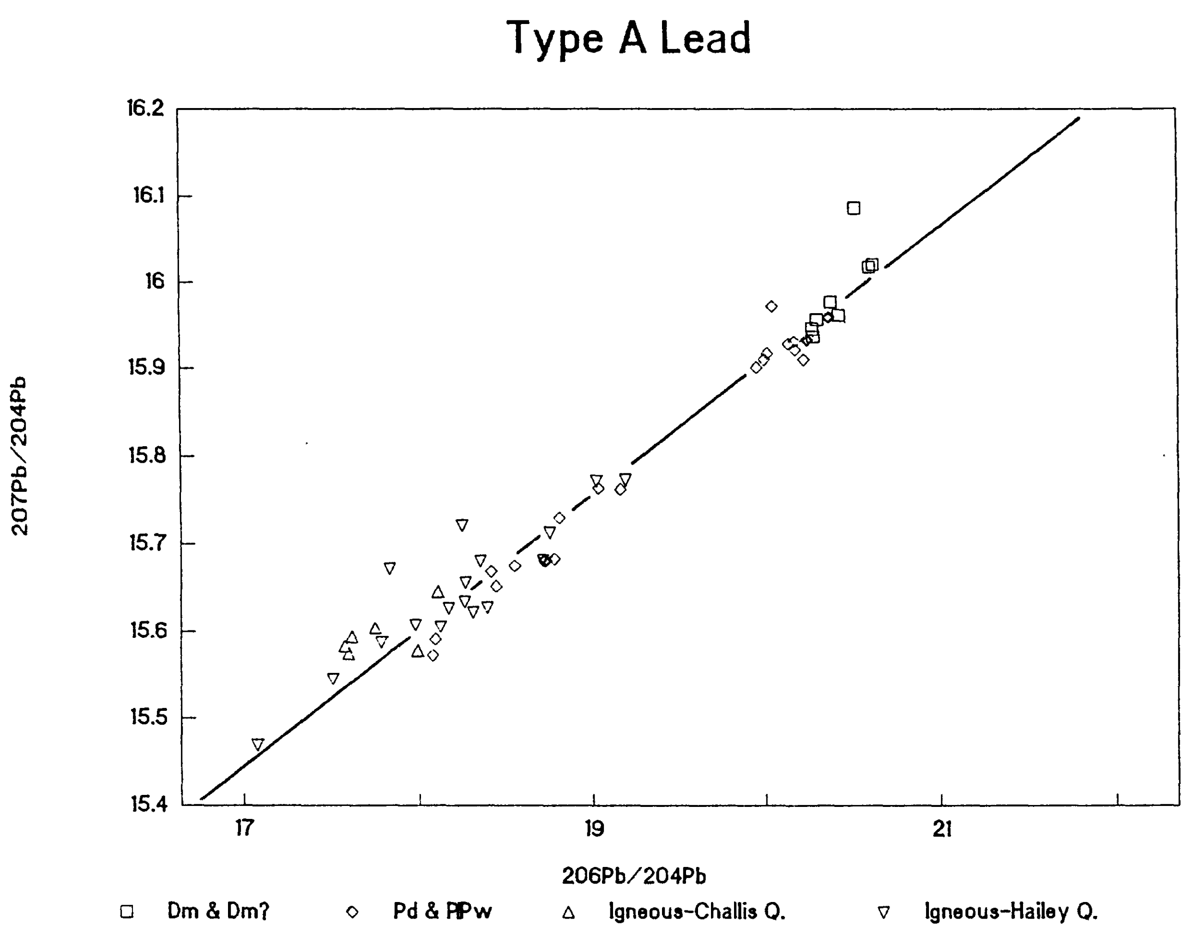

Figure 5. $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram for type A lead. Dm, Milligen Fm.; Dm?, Triumph block and Snoose area, possible Milligen Fm.; Pd, Dollarhide Fm.; PIPw, Wood River Fm.

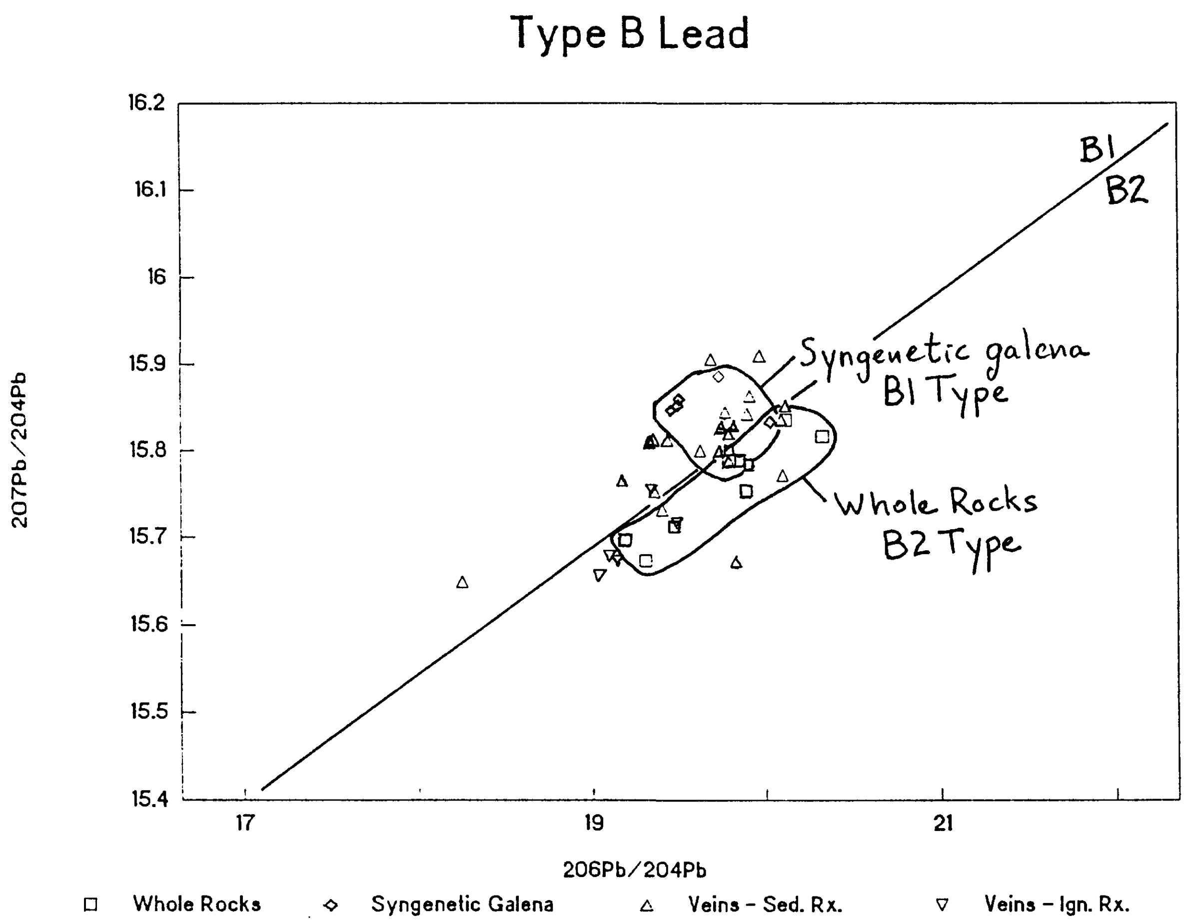

Figure 6. $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram for type B lead.

galena from sediment-hosted deposits have $^{207}Pb/^{204}Pb$ values that span the range of subgroups B1 and B2, but most correlate isotopically with group B1, the syngenetic galenas, rather than B2, the whole-rock argillites.

DISCUSSION

All whole-rock argillite samples in this study have type B2 lead. We therefore cannot use the lead isotopes to resolve stratigraphic problems such as whether the Triumph block argillite correlates with the Milligen or Dollarhide Formation. However, this fact makes it easier to recognize non-argillite sources of lead in ore deposits. Most significantly, lead in galena from nearly all sampled syngenetic and vein deposits probably did not come from the host or underlying argillites as commonly thought. For vein deposits having type A lead, the $^{208}Pb/^{204}Pb$ ratio is consistently too high, and the range in $^{208}Pb/^{204}Pb$ is too great, compared to the type B2 lead in the whole-rock argillites (fig. 7). For vein and syngenetic deposits having type B1 lead, the $^{207}Pb/^{204}Pb$ ratio is typically too high compared to the argillites.

Compared to the field of whole rock $^{207}Pb/^{204}Pb$ ratios (fig. 6), the Triumph lead is clearly distinct, and the Hoodoo and Livingston lead show minimal overlap. Some of the lead in these three deposits may have come from the enclosing sediments, but there is clearly an additional component having higher $^{207}Pb/^{204}Pb$. If these syngenetic deposits formed from "black smokers" in a submarine rift basin, as is currently thought, then the hydrothermal fluid appears to have tapped a deeper, perhaps older source of lead than that contained in detrital material being eroded from the craton.

Comparison of veins having type B lead with syngenetic deposits and whole rock argillites shows that nearly all B type veins correlate isotopically with syngenetic deposits and not with whole rocks (figs. 6 and 7). Only three of the 23 sampled B type deposits have B2 type $^{207}Pb/^{204}Pb$ values, i.e., in the field of whole rock $^{207}Pb/^{204}Pb$, the Snoose, Emperium, and White Cloud. Apparently, either the type B1 lead in the remaining 20 veins came from remobilized deposits such as the Triumph, or it came from a source at depth that is isotopically similar to the source of the syngenetic deposits (fig. 8).

The isotopic characteristics of type A lead rule out both the argillites and remobilized syngenetic deposits as possible sources. The Silver Star Queen, having type A2 lead, is hosted by the Milligen or Dollarhide Formation, but the deposit must have derived lead from a non-argillite source (fig. 8). The one sample from the Old Triumph Mine also suggests lead derived from a non-argillite source. All of the sampled deposits in the Dollarhide Formation exhibit lead from a non-argillite source.

Even the bedding-conformable galena in the Dollarhide Formation must have been introduced from a non-argillite source. Thus the conformable ore in the Dollarhide Formation is simply vein filling that happened to be deposited in cavities parallel to bedding, showing that textural evidence is not always a reliable guide to syngenetic ore.

Galena in Dollarhide-hosted deposits formed from two distinct sources of lead (fig. 8). Type A1 lead in the Dollarhide Formation (fig. 3) corresponds almost exactly to lead from the adjacent southeastern Idaho batholith deposits (fig. 4) and suggests a similar source. We do not have lead-isotope data on whole rocks or feldspar separates from intrusions in this area of the Idaho batholith, but judging from data for the Challis quadrangle, a Cretaceous or Tertiary igneous source for the type A1 lead is highly likely.

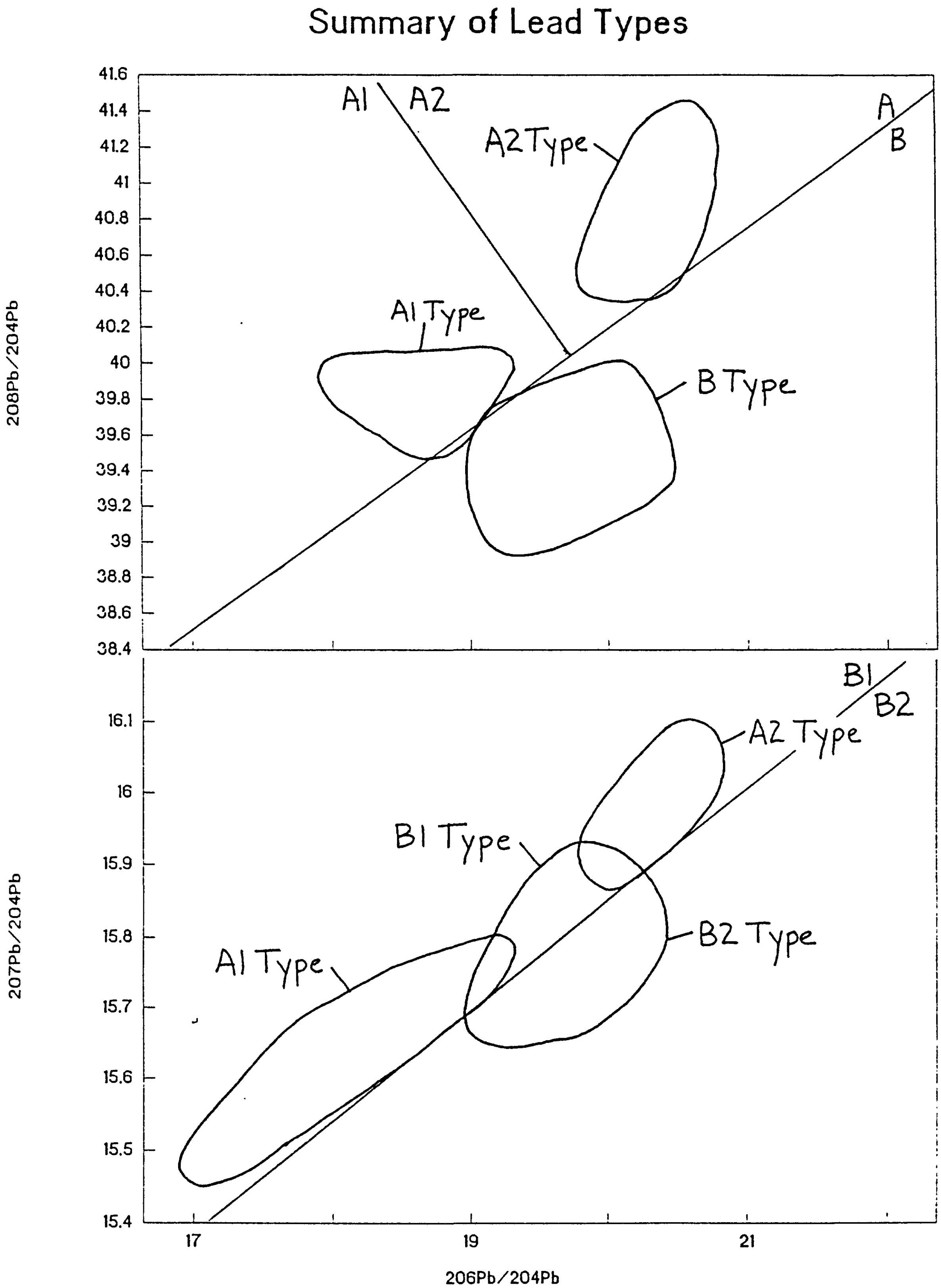

Figure 7. $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagrams summarizing lead isotope types.

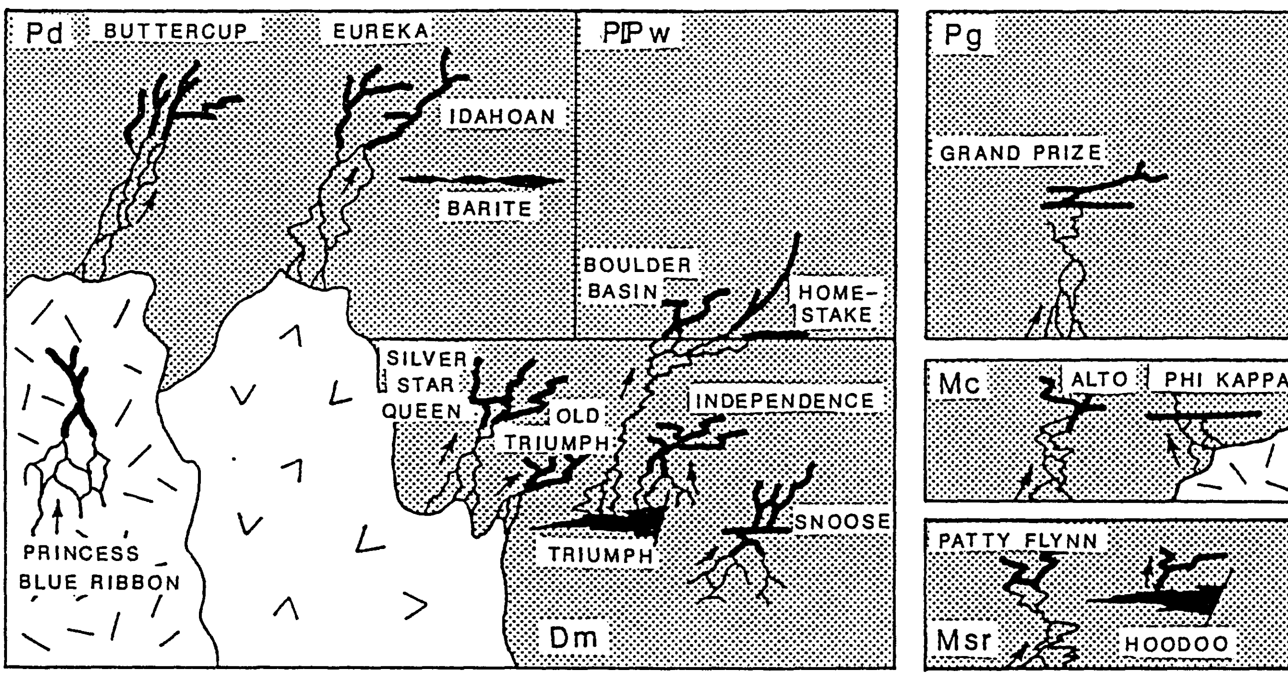

Figure 8. Schematic diagram showing sources of lead for representative examples of deposit types in each lithostratigraphic unit. Bold lines indicate mineralization; light lines indicate unmineralized fractures. See discussion section of text for explanation. Pd, Dollarhide Fm.; PIPw, Wood River Fm.; Pg, Grand Prize Fm.; Dm, Milligen Fm.; Mc, Copper Basin Fm.; Msr, Salmon River Group.

The type A2 lead in the Dollarhide Formation coincides in value with lead from the Silver Star Queen and Old Triumph. The source of this lead is unknown, but the argillites cannot be a source, as discussed above, and the Idaho batholith cannot be a source because the type A2 $^{206}Pb/^{204}Pb$ ratio too high. The problem of the highly radiogenic vein lead compared to known lead in igneous rock and sediments has been discussed previously elsewhere (e.g. Doe and others, 1979), and the most likely source appears to be upper crustal Precambrian material. A component of Archean crustal material is also required to account for the highly radiogenic A2 lead.

Deposits sampled from the Wood River Formation cover a large area but consistently exhibit type B1 lead suggesting a source in syngenetic ore like that in the Triumph deposit but not an argillite source such as the Milligen or Salmon River Formations. Figure 8 is meant to show that Triumph-like deposits may be a source, but is not meant to imply that the Triumph deposit itself was the source for all the lead concentrations in the Wood River Formation.

In summary, lead isotopes in central Idaho can reveal whether the galena in a deposit is syngenetic or epigenetic and can indicate the type of source rock, whether Paleozoic argillaceous sediments, remobilized syngenetic ore, Cretaceous-Tertiary intrusions, or a forth, as yet unidentified, upper crustal source (figs. 7 and 8). Future isotopic work will attempt to characterize better these source rocks and will incorporate more data from other isotopes such as sulfur.

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We thank Reed Lewis, Paul Link, Falma Moye, Sandy Soulliere, Craig Wavra, and Anna Wilson for collecting some of the samples used in this study. Discussions with Gary Winkler were very productive. Larry Dee, U.S. Bureau of Land Management, provided all sorts of assistance. Larry Snee carefully reviewed the manuscript.

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STRATIGRAPHIC SETTING OF SEDIMENT-HOSTED MINERALIZATION IN THE EASTERN HAILEY 1°x2° QUADRANGLE, BLAINE, CUSTER, AND CAMAS COUNTIES, SOUTH-CENTRAL IDAHO

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ABSTRACT

A variety of base- and precious-metal mineral deposit types are hosted by Paleozoic sedimentary rocks in the eastern half of the Hailey 2-degree quadrangle, south-central Idaho. Deposits occur in dark-colored, carbonaceous shales, siltstones and limestones of Ordovician, Silurian, Devonian, Mississippian, Pennsylvanian, and Permian ages; anoxic environments of deposition apparently existed through much of Paleozoic time. Although initial syngenetic stratiform mineralization is probable, most mineral deposits are the products of hydrothermal cells driven by Cretaceous and Eocene magmatism. Mineralized structures include unconformities, probable Mesozoic thrust faults, northwest-striking Cretaceous faults, Paleogene detachment faults, and high-angle northeast-striking Eocene shear zones. Replacement deposits occur in calcareous rocks adjacent to mineralized structures and along silicification fronts produced during fluid migration. Skarn deposits are locally present in calcareous host rocks adjacent to plutons.

INTRODUCTION

This short review provides a preliminary synopsis of the stratigraphic and structural setting of sediment-hosted mineral deposits in the eastern half of the Hailey 1°x2° quadrangle (fig. 1). The area includes the "central Idaho black-shale mineral belt" of Hall (1985). Mineral deposits contain primarily silver-lead-zinc. They are hosted by fine-grained carbonaceous Paleozoic strata, including early Paleozoic deep-water pelitic rocks (Ordovician and Silurian Phi Kappa and Trail Creek Formations, Devonian Milligen Formation, and unnamed correlative units), the Mississippian Copper Basin Formation, and the Pennsylvanian-Permian Wood River, Dollarhide, and Grand Prize Formations. The deposits are included within mineral deposit areas (fig. 2), which have been defined for use in all reports stemming from the Hailey CUSMAP project.

PREVIOUS WORK

The silver-lead-zinc mineralization of the "Wood River area" (near Bellevue and Ketchum) has been extensively studied by W.E. Hall and his colleagues over the last 20 years (Hall and Czamanske, 1972; Hall and others, 1978; Hall, 1985; Howe and Hall, 1985; Hall, 1987a; 1987b). Hall concluded that the deposits were formed by hydrothermal circulation systems developed during both Cretaceous and Tertiary magmatism. These hydrothermal systems derived metals from a country rock source (the early Paleozoic "black shales"). In the Wood River area, metal deposition was localized below the Wood River thrust, which acted as a permeability barrier to mineralizing solutions.

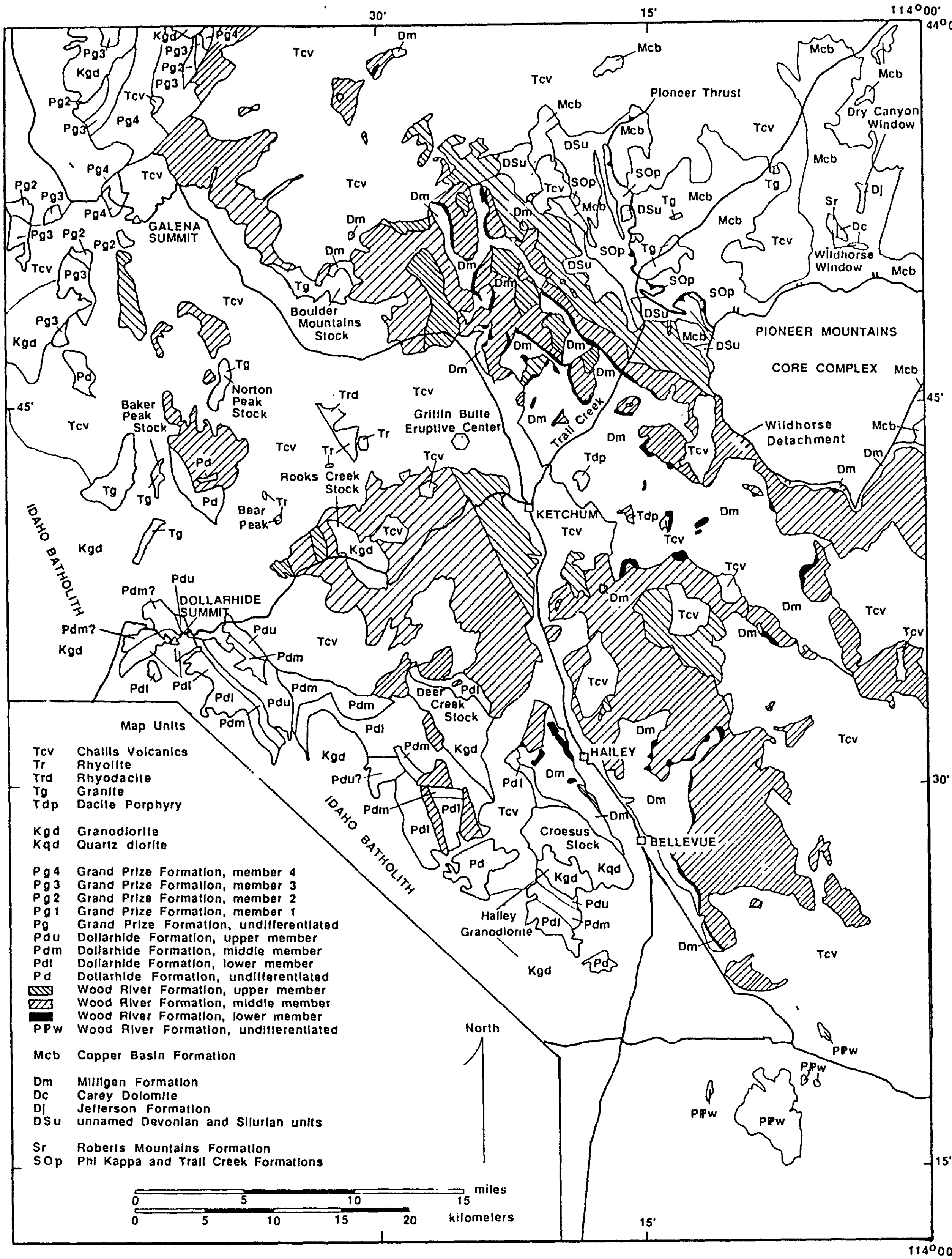

Figure 1. Geologic map showing outcrop areas of sedimentary units in the eastern part of the Hailey 1°x2° quadrangle. Map is modified from compilation by V.E. Mitchell, Idaho Geological Survey.

The mineral deposits of the Wood River area were first documented by Lindgren (1900). Other important reports on deposits in the eastern Hailey 2-degree quadrangle that include maps of individual mines and mineral districts include Umpleby and others (1930), Anderson and others (1950), Tuchek and Ridenour (1981), and Tschanz and Kiilsgaard (1986).

STRATIGRAPHY AND STRUCTURAL SETTING

Carbonaceous limestone, argillite, and siltstone were deposited in south-central Idaho intermittently from Ordovician through early Permian time. These rocks share lithologic similarity and structural complexity. Furthermore they are largely unfossiliferous. Most of the stratigraphic units host silver-lead-zinc deposits, and collectively have been termed the "Paleozoic black-shale mineral belt" by Hall (1985).

Various conflicting stratigraphic divisions for these rocks have been proposed in the literature. A summary of the stratigraphy of the eastern Hailey quadrangle, as currently defined, follows; for more detail refer to Skipp and Hall (1980), Dover (1983), Hall (1985), Hall and Hobbs (1987), Turner and Otto (1988), Link and others (1988), and Mahoney and Sengebush (1988). Note that the division of map units shown on figure 1 represents an amendment of the assignments made by Link and others (1988, fig. 18).

Ordovician and Silurian Phi Kappa and Trail Creek Formations

The Ordovician and Silurian Phi Kappa and Trail Creek Formations are imbricated in a series of thrust-bounded slices at the head of Trail Creek, east of Ketchum (fig. 1). Together, the formations are approximately 340 meters thick and consist of dark-gray graptolitic shale and argillite, banded siliceous metasiltstone, and fine-grained quartzite (Dover, 1983). The Phi Kappa and Trail Creek Formations lie structurally above the Pioneer thrust (fig. 1), which places early Paleozoic rocks on the Mississippian Copper Basin Formation. They lie structurally below the upper Wood River Formation on what we interpret to be detachment faults of Paleogene age.

Devonian Milligen Formation

The Devonian Milligen Formation is the host for rich Ag-Pb-Zn ores in the historically productive Minnie Moore and Triumph areas near Hailey and Bellevue (areas 11 and 17 of fig. 2). As described by Turner and Otto (1988), and shown diagrammatically in figure 3, the Devonian Milligen Formation contains several thousand meters of fine-grained locally carbonaceous strata which are almost totally devoid of recognizable macro- or microfossils. Sandberg and others (1975) describe the only fossil control available for the Milligen Formation. An informal stratigraphic division (fig. 3) includes 1) a lower member of argillite, locally chert- or sandstone-rich, 2) a middle member containing black argillite and limestone (Triumph argillite and Lucky Coin limestone) with lenses of coarse-grained sandstone (Cait quartzite), and 3) an upper member containing argillite with varying amounts of fine-grained sandstone and limestone (Independence sandstone). The base of the Milligen Formation has not been observed. The upper contact is an unconformity below the Pennsylvanian Hailey Conglomerate Member of the Wood River Formation or a fault against younger Pennsylvanian and Permian strata. In the areas of the Triumph and Minnie Moore Mines, the Milligen Formation occurs in northeast-vergent overturned folds with southwest-dipping limbs. Mineral deposits are confined to the middle member.

The published and unpublished maps of W.E. Hall and previous workers differ in assignment of strata to the Milligen Formation in the Wood River area (Triumph, Minnie Moore, and Bullion mineralized areas, 11, 17, and 15 of figure 2). Hall (1985) defined the Dollarhide Formation, which included rocks formerly assigned to the Milligen Formation west of Hailey and Bellevue, but only in the 1989 field season have workers on the Hailey CUSMAP project come to agreement about what is Dollarhide and what is Milligen. The stratigraphic assignments shown on figure 1 represent a consensus of recent workers in the area, including the present authors, R.S. Lewis, M.E. Ratchford, and Betty Skipp. The breakthrough which allowed recognition of these distinctions was subdivision of informal members within the Milligen Formation by Turner and Otto (1988).

Unnamed Silurian and Devonian Unit

Unnamed Silurian and Devonian rocks including black, siliceous to carbonaceous argillite and tan, calcareous to dolomitic siltstone are tectonically imbricated with the Phi Kappa and Trail Creek Formations at the head of Trail Creek east of Ketchum (fig. 1). Structural complexity and poor exposure prevented Dover (1981; 1983) and previous workers from dividing this sequence into formations. Limestones in the sequence have yielded Devonian and Silurian microfossils (Dover, 1981), which precludes chronocorrelation with the solely Devonian Milligen Formation, although the units are lithologically similar. The unnamed rocks probably accumulated in a similar depositional setting, if not within the same basin, as the Milligen Formation. Also present are dark quartzite, bioclastic limestone, and chert-quartzite granule to pebble conglomerate that closely resembles parts of the Mississippian Copper Basin Formation (Dover, 1983, p. 35). A dark argillite (designated DSa on the maps of Dover, 1981; 1983) is currently being explored for stratabound silver-lead-zinc mineralization in the Summit mineralized area of the northern Boulder Mountains (area 8 on fig. 2).

Mississippian Copper Basin Formation

The Mississippian Copper Basin Formation contains at least 3000 m of deep-water strata including (1) a lower interval of dark-gray argillite, siltite and conglomerate, (2) a middle interval of fine-grained limestone turbidites, and (3) an upper interval of chert-and quartzite-pebble and cobble conglomerate (Paull and others, 1972; Nilsen, 1977; Dover, 1983). The Copper Basin Formation crops out along the eastern edge of the Hailey 1°x2° quadrangle (fig. 1) and is contained in two thrust plates (the Copper Basin and Glide Mountain thrust plates of Dover, 1981; 1983). The Copper Basin Formation of the lower (Copper Basin) plate overlies Silurian and Devonian shelf carbonate rocks of the Dry Canyon and Wildhorse windows in the northeastern part of the map area (fig. 1). This contact is shown as a thrust by Dover (1981; 1983), although recent mapping by Betty Skipp shows it to be depositional in the Fish Creek Reservoir area in the adjacent Idaho Falls 1°x2° quadrangle (Link and others, 1988, p. 20). The Copper Basin Formation is thrust over finer grained coeval strata (Mississippian McGowan Creek Formation) east of the map area. The Copper Basin Formation is structurally

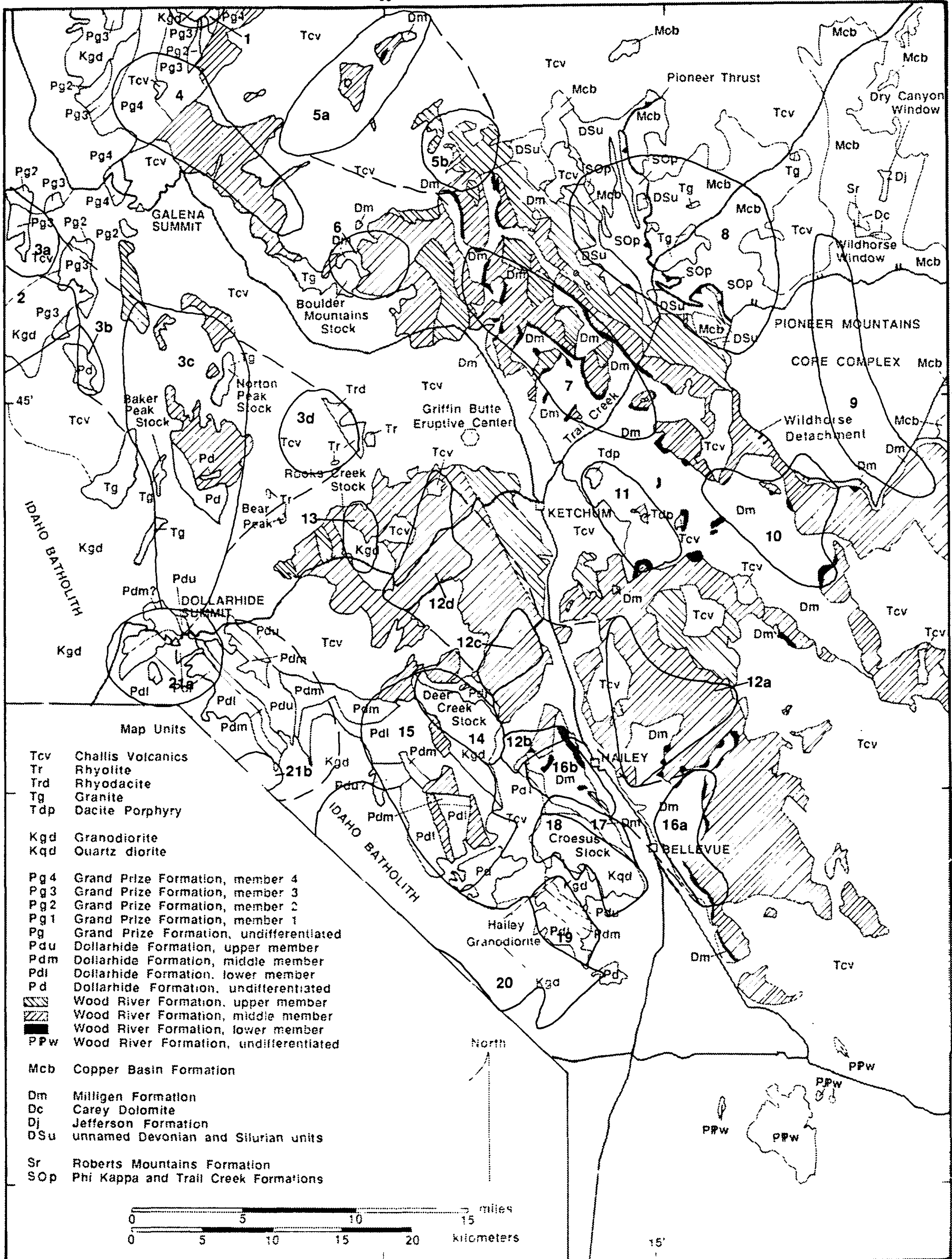

Figure 2. Map of mineral deposit areas in the eastern Hailey 1°x2° quadrangle. The geologic map of figure 1 is the base for this map. Mineral deposit areas include:)1) Washington Basin; (2) Marshall Peak; (3) Smoky Mountains; (a) Smiley Creek, (b) West Fork Big Smoky, (c) Norton-Baker Peaks, (d) Baker Creek; (4) Galena; (5) East Fork Salmon; (a) East Fork, (b) Ryan Peak; (6) Boulder Basin; (7) Lake Creek; (8) Summit; (9) Pioneer Dome; (10) East Fork Wood River; (11) Triumph; (12) Wood River; (a) Quigley Gulch, (b) Carbonate Mountain, (c) Greenhorn Gulch, (d) Warm Springs; (13) Rooks Creek Stock; (14) Deer Creek Stock; (15) Bullion; (16) Bellevue; (a) Slaughterhouse Gulch, (b) Vorberg Gulch; (17) Minnie Moore; (18) Croesus Stock: (19) Bunker Hill; (20) Hailey Gold Belt; (21) Carrietown-Buttercup, (a) Carrietown, (b) Buttercup.

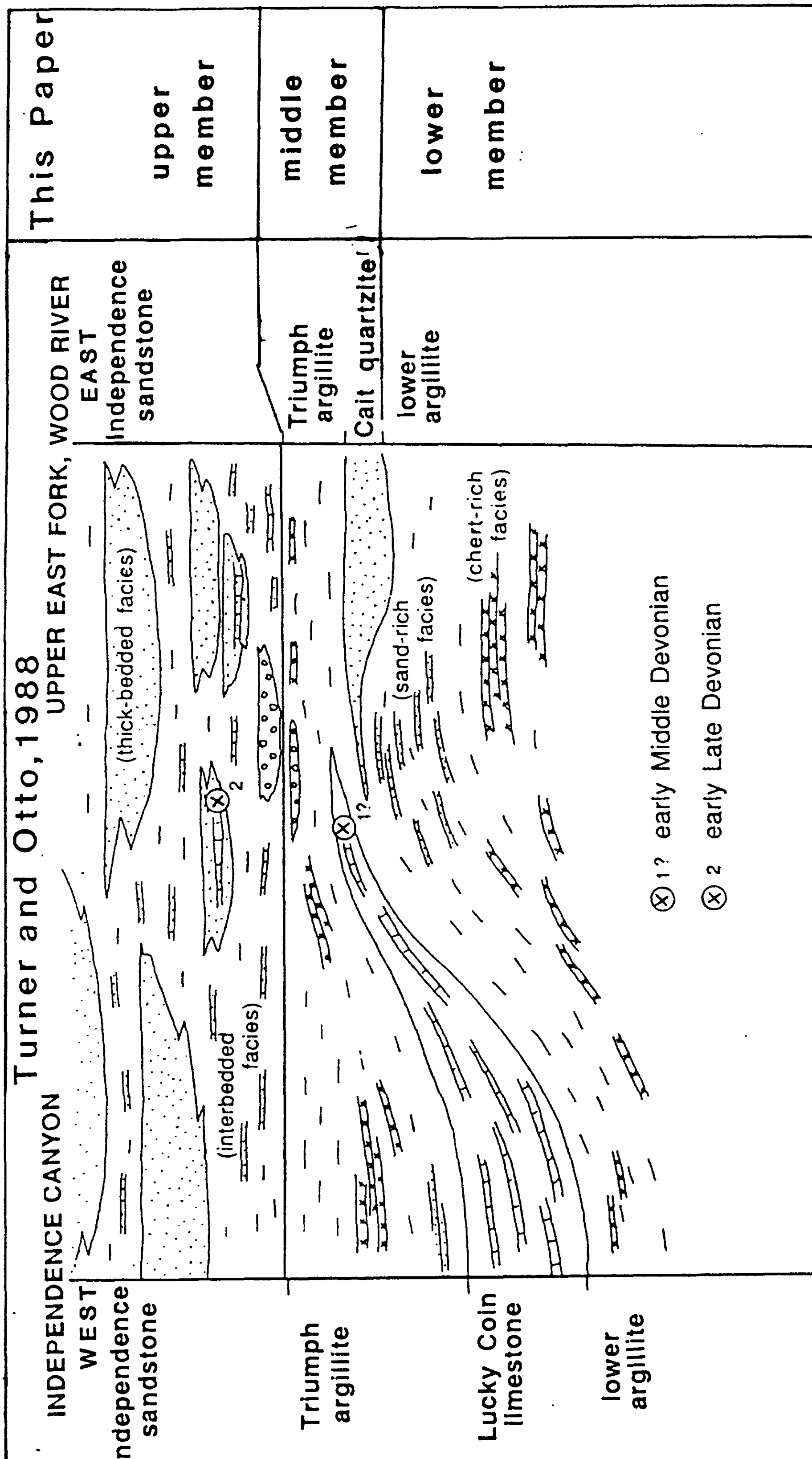

Figure 3. Diagrammatic stratigraphy of the Devonian Milligen Formation, from Turner and Otto, 1988. Informal stratigraphic terminology used in this paper is shown at the right of the figure.

overlain by the Phi Kappa and Trail Creek Formations and unnamed Silurian and Devonian unit along the Pioneer thrust (Dover, 1981; 1983). Lead-silver replacement mineralization is present in laminated calc-turbidites of the middle part of the Copper Basin Formation (Drummond Mine Limestone) in the Summit area (area 8 on fig. 2) and widespread on the Idaho Falls 1°x2° quadrangle to the east.

Rocks of the Pennsylvanian-Permian Wood River Basin

Hall (1985) recognized three coeval and lithologically similar formations in south-central Idaho: the Pennsylvanian-Permian Wood River, Dollarhide, and Grand Prize Formations, each of which contains sandy limestone and dark argillite. Link and others (1988, p. 25) proposed that these rocks were deposited in the epicratonic Late Paleozoic Wood River Basin. Each of these formations hosts mineral deposits.

Hall (1985) proposed that the three formations of the Wood River Basin belonged to distinct thrust complexes or tectonic stacks, and that their mutual boundaries were everywhere thrust faults of significant lateral displacement. Recent geologic mapping and measured sections (R.S. Lewis, P.K. Link, J.B. Mahoney, J.P. O'Brien, Betty Skipp, in preparation) demonstrate that boundaries between formations of the Wood River basin represent facies changes rather than structural contacts. In particular:

(1) the Dollarhide Formation is a western, carbonaceous facies of the Wood River Formation, with the contact defined by facies change west of the town of Hailey;

(2) the Grand Prize Formation is a northern, sandy facies of the Wood River Formation, with the contact defined by facies change along Pole Creek in the Galena area (area 4 on fig. 2). The Grand Prize Formation tends to be silicified because of its proximity to Cretaceous intrusions.

Wood River Formation

The Wood River Formation contains at least 3000 m of strata which we divide into three members (Link and others, 1988). This simplification of the seven or eight informal units of Hall and others (1974; 1978) facilitates geologic mapping. Some of the informal units of Hall undergo facies changes and pinch out across the outcrop belt.

The lower member of the Wood River Formation includes the basal Hailey Conglomerate and overlying bioclastic limestones of unit 2 of Hall and others (1974). These rocks represent braid-delta conglomerates succeeded by clear-water biostromal limestones (Burton, 1988). The middle member (units 3 through 6 of Hall and others, 1974) includes more than 1000 m of mixed carbonate-siliciclastic strata, and is dominated by thick-bedded micritic or siliceous sandstone. Siltite and silty limestone are subordinate. The middle member is interpreted to represent sediment grain flow and disorganized turbidite deposition on the outer portions of a carbonate apron (Burton, 1988). The upper member consists of more than 1000 m of sandy micrite and argillite (units 7 and 8 of Hall and others, 1974; 1978). The upper member is generally finer grained than the middle member, and contains finely laminated silty turbidites with abundant ichnofossils and soft-sediment deformation. It represents deposition on the outer portions of a carbonate apron and the adjacent basin (Burton, 1988).

There are a variety of mineral deposit types present in the Wood River Formation. The Hailey Conglomerate hosts zinc-bearing veins in the Lake Creek area (area 7 on fig. 2). Brittle sandstones of the middle member host vein mineralization in the Boulder Basin (Ratchford, 1989), Galena, and Lake Creek areas (6, 4, and 7 on fig. 2). Limestones of the middle member host replacement deposits in the Wood River area (12 on fig. 2). Carbonaceous siltstone of the upper member host shear zone vein mineralization in the Lake Creek area (7 on fig. 2).

Dollarhide Formation

Since definition of the Dollarhide Formation by Hall (1985), its recognition has been the subject of confusion for two reasons. First, parts of it are lithologically similar to the Milligen Formation, contain similar mineral deposits, and lack diagnostic microfossils which would allow the use of chronostratigraphy. Second, where the Dollarhide is intruded and tectonized by the Idaho batholith, it is locally strongly silicified and altered to a lighter color. This led to recognition of a distinct "Carrietown sequence" along the eastern margin of the batholith, which is mineralized in the Carrietown-Buttercup mining district (area 21 on fig. 2) (Darling, 1988). Whitman and Link (1989) have demonstrated that the "Carrietown sequence" contains silicified and contact metamorphosed rocks of the lower and middle members of the Dollarhide Formation, and is not in thrust contact below the Dollarhide as shown by Darling (1988) and Link and others (1988).

Stratigraphy of the Dollarhide Formation has been described by Wavra and others (1986) and Geslin (1986). The informal members of the formation have now been mapped by Betty Skipp, R.S. Lewis, J.P. O'Brien, and P.K. Link (in preparation). The formation contains over 2300 m of strata divided into three members. The lower member contains a variety of dark-colored, dominantly fine-grained, and locally calcareous strata including sandy limestone, argillite, laminated "banded" siltite and fine-grained sandstone. The lower member is distinguished by the presence of lenticular conglomerates and by lenticular sandy limestones. The conglomerates contain both intraclasts and extrabasinal chert and quartz pebbles, and may be distal equivalents of the basal Hailey Conglomerate of the Wood River Formation. The middle member of the Dollarhide Formation contains light-colored medium- to coarse-grained calcareous sandstone with local conglomerate. The middle member is identical to, and probably the western equivalent of, parts of the middle member of the Wood River Formation. The upper member of the Dollarhide Formation contains thick black argillites plus the same variety of lithologies as the lower member, including dark sandy limestone, laminated siltstones, and local conglomerate. The Dollarhide Formation is interpreted to represent outer carbonate apron and basinal depositional environments, similar to the Wood River Formation.

The lower member of the Dollarhide Formation hosts vein and replacement mineralization in the Bullion, Carrietown, Buttercup, and Deer Creek stock areas (areas 15, 21a, 21b, and 14 on fig. 2), and a laminated barite deposit in the northern Bullion area. Sheared vein and replacement mineralization is present in the upper member of the Dollarhide Formation in the Bunker Hill and Carrietown-Buttercup areas (areas 19, 21a, and 21b on fig. 2).

Grand Prize Formation

The Grand Prize Formation consists of over 2500 m of fine-grained calcareous strata which have been divided into four informal members by Mahoney and Sengebush (1988). Basal member 1 contains chert pebble conglomerate, graded beds of sandstone, and bioclastic limestone and is lithologically and stratigraphically similar to the lower member of the Wood River Formation. Member 2 consists of over 750 m of thick-bedded to massive calcareous sandstone, which decreases in grain size upward into banded fine-grained sandstone and siltstone of member 3. Member 4 contains dark carbonaceous siltstone and mudstone with intercalated sandy limestone and minor sandstone. The Grand Prize Formation is interpreted to represent sediment gravity flow and pelagic deposition (Mahoney and Sengebush, 1988).

Members 1, 2, and 3 of the Grand Prize Formation host gold skarns and polymetallic veins in Washington Basin (area 1 on fig. 2). Member 3 hosts a tungsten skarn in the Smiley Creek area (area 3a on fig. 2). Carbonaceous limestones of member 3 host replacement deposits in Pole Creek (Galena area, 4 on fig. 2). Limestones of member 4 also have potential for replacement mineralization in the Galena area.

GEOLOGIC HISTORY

The deformational and magmatic history of the eastern Hailey 1°x2° quadrangle is complex. Repeated syngenetic deposition of metals occurred through early and middle Paleozoic time. Several generations of faults, of Mesozoic through Neogene age, are present. At least two (Cretaceous and Eocene) magmatically driven hydrothermal systems have existed. Post-mineralization faulting occurs in all the major mining districts, and has dismembered the ore deposits. A brief synopsis of this complex history follows, with annotations of important events relating to mineralization.

Paleozoic

Base- and precious-metal enrichment was recurrent during early to middle Paleozoic sedimentation in the eastern Hailey quadrangle. Carbonaceous deep-water shales and limestones are present in strata from Ordovician through Permian age and are the ultimate source of the metals now found in the black-shale mineral belt (Hall and others, 1978; Hall, 1985; 1987b). However, no specific faults or radiometrically-dated Paleozoic magmatic activity have been identified positively to tie these metal concentrations to sedimentary exhalative processes.

Deposition of the Devonian Milligen Formation was closely followed by, or perhaps partly coeval with folding and metamorphism of the Antler orogeny (Turner and Otto, 1988). The Antler event produced axial planar cleavage and tight to isoclinal folds in the Milligen Formation in the Triumph area (Turner and Otto, 1988, p. 162-163). An Antler-age deformational fabric has not been identified in the Phi Kappa, Trail Creek, and unnamed Devonian and Silurian units. That the Milligen was part of the Antler highland is suggested by the fact that it has an inferred Antler-age structural fabric and by the presence of Milligen Formation clasts in conglomerates of the Mississippian Copper Basin Formation. Skipp and Hall (1975) and Nilsen (1977) suggested that the Milligen Formation was thrust eastward during the Antler orogeny and made up the Antler Highland in Idaho. This thrust, if it existed, has been overprinted by Mesozoic and Paleogene tectonism, and has yet to be identified (Dover, 1980).

Early Mississippian uplift of the Antler highland produced a source for a wedge of chert-rich flysch which becomes finer eastward. These strata comprise the Mississippian Copper Basin Formation, which was deposited in the foreland basin east of the Antler highland (Skipp and Hall, 1980; Dover, 1983). Black and gray chert and cleaved argillite clasts in the Copper Basin Formation were derived from the Devonian Milligen Formation. The source for light-colored quartzite clasts found in the Copper Basin Formation has not been identified.

Late Pennsylvanian to Early Permian subsidence of the Wood River Basin produced a depositional site for large volumes (over 3000 m) of fine-grained quartzose sand and micritic mud. The provenance for both of these detrital fractions has not been proven, although Link and others (1988) suggest a northerly cratonic source for much of the sand fraction and Burton (1988) proposed an easterly source for the carbonate fraction. The basal conglomerate units present in Wood River Basin strata were locally derived, from subjacent early Paleozoic rocks and possibly recycled quartzite conglomerate of the Copper Basin Formation. Local anoxic depositional sites within the Wood River Basin may have been locations of base and precious metal concentration (lower and upper members of the Dollarhide Formation, and perhaps member 4 of the Grand Prize Formation).

Mesozoic

Thrusting and folding associated with the Late Jurassic and Cretaceous Sevier orogeny produced eastward overturned macroscopic folds, locally with axial planar cleavage, in all Paleozoic sedimentary rocks. A stack of older-over-younger thrust sheets with tens of kilometers of displacement exists to the east of the Wood River area (Skipp and Hait, 1977). The farthest westward major thrust is the Pioneer thrust which places Ordovician and Silurian Phi Kappa and Trail Creek Formations or the Devonian Milligen Formation over the Mississippian Copper Basin Formation (Skipp and Hall, 1975; Skipp and Hait, 1977; Dover, 1983; Link and others, 1988, fig. 3).

The presence of far-travelled Mesozoic thrust plates in the Wood River area has long been the accepted tectonic model. However, our recent mapping has recognized no major thrusts west of the Pioneer Mountains core complex and suggests that in the Wood River area, the primary effect of Sevier compressional deformation was to produce northeast-vergent overturned folds with local interbed shearing and boudinage. The axial regions of some of these folds are fractured and serve as sites for mineral deposition. Blind thrusts are inferred to core some of the folds, and a basal detachment thrust is required at depth in the Milligen Formation to accommodate the shortening observed at the surface.

The Wood River "Thrust"

The contact between the Milligen Formation and overlying Pennsylvanian-Permian strata of the Wood River Formation has long thought to be a regional thrust of major displacement (the "Wood River thrust" of Hall and others, 1978, and Dover, 1983). We interpret this contact to be a locally sheared unconformity across which rocks of varying competence have responded differently to Mesozoic compression and Paleogene extension (Skipp and others, 1986; Burton and others, 1989).

We recognize three geologic relations across the Wood River-Milligen contact:

(1) an unconformity across which there has been no shearing, with the basal Hailey Conglomerate of the Wood River Formation containing clasts of the subjacent Milligen Formation;

(2) an unconformity by thrusting during northeast-directed Mesozoic compressional deformation of the Sevier orogeny and across which the basal Hailey Conglomerate has been sheared, boudinaged, and injected by locally mineralized quartz veins;

(3) a Paleogene detachment fault (described below) which accommodated northwest-directed extension, and which places folded middle and upper parts of the Wood River Formation on Milligen Formation.

Idaho Batholith

Intrusion of the Idaho batholith followed folding of the Sevier orogeny and lasted from approximately 90 to 70 Ma (Kiilsgaard and Lewis, 1985; Lewis and others, 1987; Johnson and others, 1988). Two intrusive phases are recognized in the area west of Bellevue.

The early border phase was emplaced from about 90 to 80 Ma and produced quartz diorite of the Croesus stock about 88 Ma. The three satellite stocks east of the main batholith (Hailey granodiorite, Deer Creek stock, Rooks Creek stock, fig. 1) are also probably of this age. Lamprophyre dikes are found in association with the Croesus stock in the Minnie Moore area. Hydrothermal systems driven by these intrusions appear to have been the important mineralizing agents in the Bullion, Bellevue, Minnie Moore, Hailey Gold Belt, Bunker Hill, Deer Creek, Rooks Creek, and possibly, Triumph mineral deposit areas. A system of northwest-trending faults or shear zones developed during late stages of intrusion and were the sites of silver-lead-zinc mineralization.

The main phase biotite granodiorite was emplaced approximately 80 to 72 Ma and occupies the western edge of the area shown in figure 1. Mineralization of the Carrietown-Buttercup and Marshall Peak areas (21 and 2 on fig. 2) is associated with intrusion of the main phase batholith.

Synkinematic Cretaceous intrusion is probable in the Pioneer Mountains core complex. One hornblende-rich sample of foliated gneissose granodiorite yielded a hornblende K-Ar age of 67.6±1.6 Ma (Dover, 1983). Mesozoic deformation in the core complex is documented by O'Neill and Pavlis (1988).

Tertiary

During Paleogene time, widespread northwest-southeast extension occurred. Manifestations of this stress regime include:

(1) an erosion surface on the Idaho batholith above which the Eocene Challis Volcanics lie unconformably, indicating rapid uplift after emplacement;

(2) the Wildhorse detachment, along which the Pioneer Mountains core complex rose (Wust, 1986);

(3) other low-angle normal faults in the upper plate of the Pioneer Mountains core complex which may sole into the Wildhorse detachment. These

are the prominent low-angle faults previously mapped as thrusts in the Wood River area;

(4) high-angle northeast-trending normal faults which are present throughout the eastern Hailey 1°x2° quadrangle and which are parallel to the Trans-Challis fault system (Bennett, 1986). These northeast-trending faults served to localize eruptive and intrusive centers during Challis magmatism;

(5) eruption of the Challis Volcanic Group (Fisher and Johnson, 1987) from about 51 to 44 Ma. Eruption of intermediate composition volcanic rocks was accompanied by intrusion of hypabyssal plutons that range from diorite to granite in composition (Bennett and Knowles, 1985; Moye and others, 1988). In general, magmatism evolved from intermediate to silicic. The late phase of magmatism produced "pink granite" and rhyolite porphyry and is generally associated with mineralization. Extensive dike injection occurred marginal to some plutons. Challis magmatism was coeval with extensional faulting; precise age relations are not yet fully resolved.

Mineral deposit areas affected by hydrothermal cells generated during Challis magmatism and associated faulting include much of the Triumph, Galena, Boulder Basin, Lake Creek, Smoky Mountains, Summit, and East Fork Salmon River areas (areas 11, 4, 6, 7, 3, 8, and 5 on fig. 2).

Neogene tectonic activity in the Wood River area includes the development of northwest-trending Basin and Range normal faults which control the present topography. No remobilization of metals can be attributed to these faults, although they are responsible for modern locations of hot springs and possible metal remobilization and deposition at depth. These young faults also cut the mineral deposits, locally terminating minable ore shoots.

MINERAL DEPOSIT MODELS APPLICABLE TO THE EASTERN HAILEY QUADRANGLE

Several of the mineral deposit models compiled in Cox and Singer (1986) are applicable to the central Idaho black-shale mineral belt. The following discussion, while neither a rigorous coverage of the models nor a mine-by-mine review of each mineral district, is meant to summarize present understanding and set the stage for more comprehensive reports which will be completed as part of the final products of the Hailey CUSMAP project. Model numbers refer to the Cox and Singer (1986) review.

Skarn Deposits (Model 18)

Skarns have not been major producing ore bodies in the Wood River area, but are documented to occur in several places where limestones are intruded by either Tertiary and Cretaceous stocks. A gold skarn is developed in the Lucky Coin limestone of the Milligen Formation at the June Day Mine in Parker Gulch southeast of Ketchum. The intrusive rock is Eocene dacite porphyry. Tungsten skarn is hosted by members 1 to 3 of the Grand Prize Formation in the Ura Group prospects in Smiley Creek, (area 3a on fig. 2) hosted by Grand Prize Formation members 1 to 3. Gold skarns occur in the Washington Basin area (area 1 on fig. 2) where probable Cretaceous granodiorite intrudes limestone of Grand Prize Formation member 1 in the Black Rock claims. Skarns also occur in the Bunker Hill, Minnie Moore, Deer Creek, and Rooks Creek areas (19, 17, 14, and 13 on fig. 2) adjacent to the Cretaceous border phase plutons.

Polymetallic Replacement Deposits (Model 19a)

Several formations host replacement deposits. Complex polymetallic replacement ore in the Minnie Moore and Triumph mine areas is hosted by the Lucky Coin limestone member of the Milligen Formation. In the Carrietown-Buttercup area (21 on fig. 2) the host rock is limestone of the lower member of the Dollarhide Formation. In the Bunker Hill area (19 on fig. 2) the host is limestone of the upper member of the Dollarhide Formation. In the Boulder Basin area (6 on fig. 2) the replaced host is calcareous sandstone of the middle member of the Wood River Formation (Ratchford, 1989). The Phi Kappa mine of the Summit area (8 on fig. 2) contains Pb-Zn-Ag replacement deposits in the Drummond Mine Limestone Member of the Copper Basin Formation. In the northern Galena area (4 on fig. 2), calcareous sandstone of Grand Prize member 3 is replaced.

Limestone is the preferred host rock to be replaced, and so potential for such deposits exists in the Lucky Coin limestone of the Milligen Formation, the lower member of the Dollarhide Formation, Grand Prize Formation members 1, 3 and 4, the middle and upper members of the Wood River Formation, and in the Drummond Mine Limestone Member of the Copper Basin Formation.

Polymetallic Veins (Model 22)

A variety of vein-type mineralization occurs in the eastern Hailey 1°x2° quadrangle. In the East Fork Wood River, Bellevue, Minnie Moore, and Triumph areas (10, 16, 17, and 11 on fig. 2), veins contain Ag-Pb-Zn and are found in black argillite of the middle member Milligen Formation (Triumph argillite and Lucky Coin limestone). In the Bullion area (15 on fig. 2) veins are hosted by the lower member of the Dollarhide Formation. Veins are hosted by both upper and lower members of the Dollarhide Formation in the Carrietown-Buttercup area and by the upper member in the Bunker Hill area (19 on fig. 2). In the Wood River, Galena, Lake Creek, Washington Basin, and Boulder Basin areas (12, 4, 7, 1, and 6 on fig. 2) the host for mineralization is calcareous sandstone and silty limestone of the middle and upper members of the Wood River Formation and lower portion of the Grand Prize Formation.

The veins follow fault systems or shear zones of both Cretaceous and Paleogene age and are associated with both Cretaceous and Tertiary plutonism. In the Minnie Moore and Bullion areas veins strike northwest and dip southwest, filling a northwest-striking Cretaceous fault system. The veins were formed by a hydrothermal system driven by heat from the late Cretaceous border phase plutons. In Washington Basin (1 on fig. 2), polymetallic veins of probable Cretaceous age occur in northeast-trending shear zones or fractures localized along the axial plane of an east-vergent anticline. In the Galena district (4 on fig. 2), mineralized silver-lead-zinc veins follow Tertiary shear zones that strike northeast and dip northwest. In the Lake Creek district (7 on fig. 2), veins are found within northwest-striking normal faults in the upper plate of a Paleogene detachment fault. In both cases associated igneous rocks are dacite and rhyolite porphyry of Eocene age.

Siderite gangue characterizes the veins hosted by black shale (Minnie Moore, Triumph, and Bullion areas). Quartz gangue characterizes veins with sandstone host rock in the Galena, Washington Basin, and Boulder Basin districts.

Sedimentary Exhalative Pb-Zn Deposits (Model 31)

An ultimate sedimentary origin of the ore metals in the Wood River valley has been established from isotopic data by Hall and others (1978) and Howe and Hall (1985). However, whether any of the deposits present today are primary syngenetic stratiform deposits is unclear. Considering the complex tectonic, magmatic and hydrothermal systems operative since Paleozoic time, it is no wonder that the primary deposits are difficult to isolate. Recent exploration models in the Milligen Formation and correlative unnamed Devonian and Silurian argillite unit are built around the presence of Devonian third-order basins characterized by restricted circulation, syn-sedimentary faulting, and possibly Devonian igneous activity (Turner and Otto, 1988). A similar scenario of stratiform mineralization may have existed in the lower member of the Dollarhide Formation in the Bullion Gulch area, and also in Deer Creek where laminated (bedded?) barite deposits occur. Syn-sedimentary faulting is suggested by the presence of intraformational breccias in the lower and upper members of the Dollarhide Formation.

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VOLCANIC-HOSTED AND INTRUSION-RELATED MINERAL DEPOSITS IN THE HAILEY AND WESTERN IDAHO FALLS QUADRANGLES, IDAHO

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In the Challis volcanic field, epithermal precious metal mineralization was spatially and temporally associated with the waning stages of volcanism and cauldron subsidence and emplacement of sub-volcanic rhyolite intrusions. Examples of producing and developing mines include the Champagne Creek mine on the Idaho Falls 1°x2° quadrangle (fig. 1) and the Sunbeam, Grouse Creek and Thunder Mountain deposits on the Challis 1°x2° quadrangle.

The Challis Volcanic field, which covers about 25,000 km^2, developed during an extension-related Eocene (51-44 Ma) volcanic event. The eruptive history and compositional range of the entire Challis field are tripartite with (1) early effusive volcanism, dominantly andesite to dacite, followed by (2) explosive rhyodacite to rhyolite ash-flow tuff eruptions and formation of cauldron complexes, and culminating in (3) intrusion of late-stage rhyodacite to rhyolite domes and plugs. Volcanic rocks are compositionally diverse, but they can generally be characterized as high-K, calc-alkaline to alkaline suite; they include high-K basalt, andesite, dacite, and rhyolite as well as shoshonite, absarokite, and trachyte. Volcanism was focused along NE-trending extensional structures such as the trans-Challis fault system and sub-parallel structures to the southeast.

The southeastern Challis volcanic field, on the Hailey and Idaho Falls quadrangles, shows compositional and volcanic evolution similar to that of the rest of the field. Volcanism began at about 49 Ma with the onset of andesite volcanism from widely dispersed fissures and point sources. The lavas formed an extensive blanket and ponded in subsidence structures formed over vents. Voluminous intermediate lavas and ash-flow tuffs were erupted from 49-47 Ma. Vents for the intermediate lavas are largely obscured by younger volcanic deposits; however, the presence of NE-trending andesite and dacite dikes suggests that these lavas were extruded from fissure vents. Intermediate ash-flow tuffs were erupted from sources in the Lehman Basin and Alder Creek areas (fig. 1). Dissection of sub-volcanic terrane has exposed co-genetic dacitic to rhyolitic intrusive rocks that were emplaced late in the magmatic history of the region. In deeply dissected eruptive centers such as the Lehman Basin cauldron complex, silicic plutons intrude their own pyroclastic ejecta.

By virtue of juxtaposition of various structural levels by normal faulting and depth of erosion, the Challis volcanic field offers an opportunity to compare and contrast different levels in mineral depositing systems. On the Hailey and Idaho Falls 1°x2° quadrangles, we have identified three types of epithermal mineral deposits which may represent shallow, intermediate, and deep levels in epithermal systems, respectively. These are (1) **hot-spring deposits**, (2) **vein deposits**, and (3) **shallow intrusive-related deposits**.

The **hot-spring deposits** are represented by the Champagne Creek mine and prospect areas northwest of Mackay near Burma Road. In these systems, fluids rose along structurally controlled hydrothermal breccia zones to form siliceous sinter at the paleosurface. In the Champagne Creek area, mineralization was strongly controlled by north-trending structures; in the

Burma Road area, mineralization was localized along north- and northeast-trending structures related to subsidence of the Lehman Basin cauldron complex. Hydrothermal cells associated with each of these areas encompass several square miles of extensive propylitic and argillic alteration with locally intense silicification. Gold and silver occur in mineralized fissure fillings, silicified breccia zones, and disseminations and stockwork. Preliminary fluid inclusion studies from the Lehman Basin area indicate that hydrothermal fluids had low salinities (<3 wt percent) and low to moderate temperature (250-300 °C). Ore mineralogy varies among deposits; however precious metals are commonly associated with pyrite, tetrahedrite, argentite, electrum, and silver sulfosalts as well as other sulfides. Gangue is dominantly cryptocrystalline quartz with crustiform, banded and drusy textures. Indicator elements for gold and silver ore include As, Sb, and Hg; locally, Bi and Mo may be indicative of precious-metal deposits.

Vein deposits are recognized in the Lehman Basin area northwest of Mackay, the Lava Creek district, and the Baker Creek area north of Ketchum. Mineralized veins in the Lehman Basin and Lava Creek areas are strongly controlled by north- to northwest-trending structures and intersections of those structures. In all three areas, veins are spatially associated with complex intermediate to silicic intrusions. Mineralization took place during multiple brecciation and silicification events and was hosted by intrusive rock and the volcanic country rock. Gold and silver are associated with dominantly pyritic zones; other associated sulfides and sulfosalts include chalcopyrite, molybdenite, tetrahedrite, galena, and sphalerite; the presence of rough-textured free gold in panned concentrates taken from Lehman Creek suggests that free gold might occur in that vein system. Gangue is cryptocrystalline to coarsely crystalline quartz. Gold and silver are highly anomalous as are associated trace elements, including As, Sb, Hg. Vein systems are thought to represent intermediate levels of epithermal systems that have been exposed by high-angle normal faulting.

Shallow intrusive-related deposits include porphyry-type stockwork and contact-related deposits. Shallow porphyry systems are found in the Baker Creek area and in the Lehman Basin intrusion complex. In both of these areas, widespread propylitic and argillic alteration are associated with stockwork deposits; locally, the rocks have been flooded with silica. Shallow, porphyry-type alteration occurs in and around silicic intrusions and associated intermediate to silicic dike swarms. Contact-related deposits include skarn and replacement deposits. Skarn deposits associated with the Mackay granite have produced base metals and are currently being studied for their gold skarn potential. Replacement deposits are associated with shallow intrusions and dike swarms in the Muldoon district where Paleozoic limestone hosts lead and zinc deposits.

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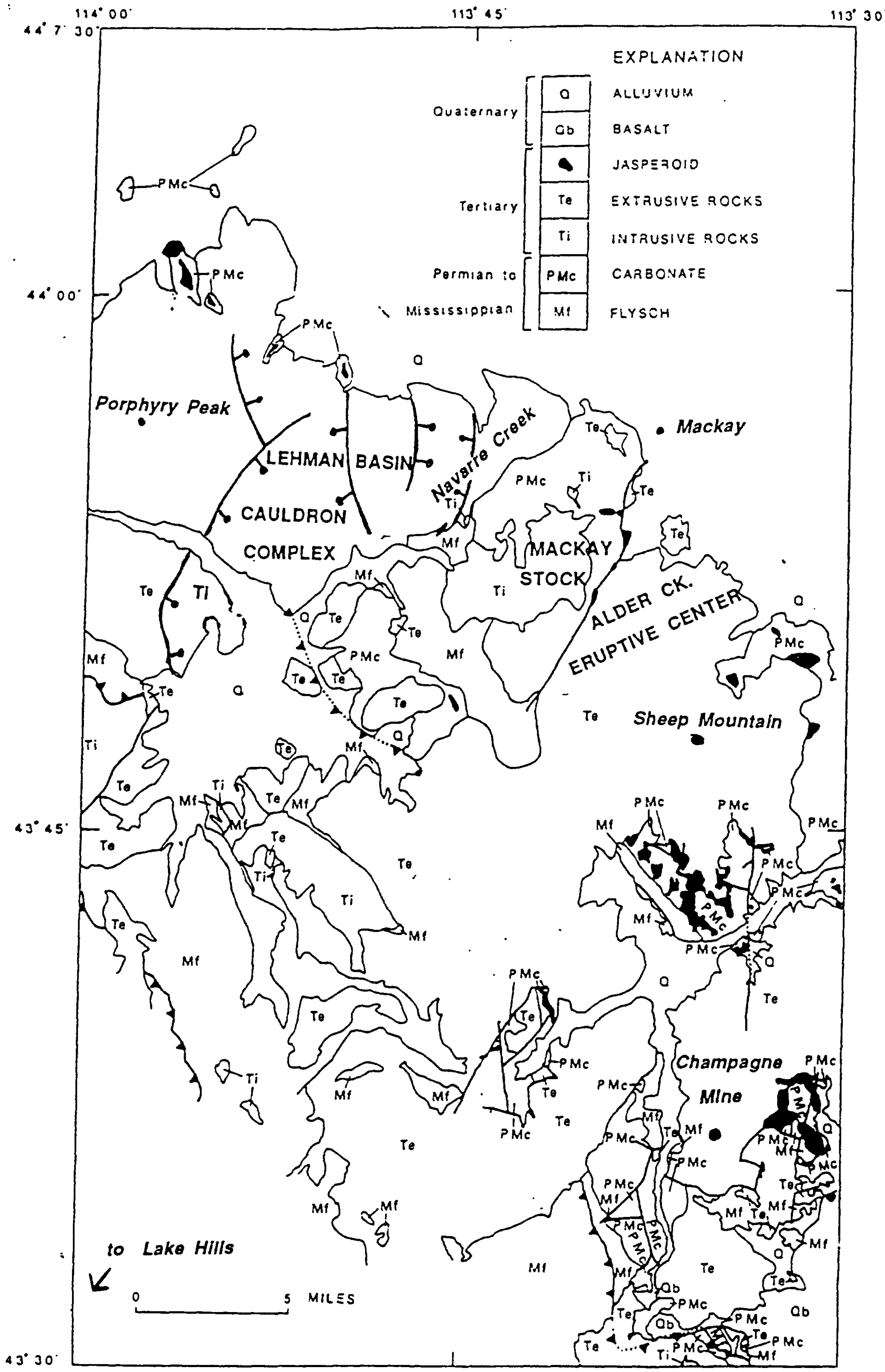

Figure 1. Generalized geologic map of the western Idaho Falls 1°x2° quadrangle.

THE GEOLOGY AND ORE DEPOSITS OF THE IDAHO BATHOLITH IN THE HAILEY 1°x2° QUADRANGLE, IDAHO

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The Idaho batholith (Atlanta Lobe) and related rocks in the Hailey 1°x2° quadrangle in Idaho host over 750 mines and prospects located in 30 mining districts. Several of these districts are being actively explored for low-grade, bulk-minable, precious metal deposits.

The geology of this part of the batholith (Atlanta Lobe) is relatively simple. Cretaceous plutonic rocks include large bodies of granodiorite (most common variety) and 2-mica granite. An area containing numerous pegmatite dikes is located between House Mountain and Smoky Dome. The Tertiary is represented by plutons of granite (pink anorogenic granite), granodiorite, and diorite. These Tertiary plutonic rocks have hypabyssal equivalents, including major dike swarms of rhyolite and rhyodacite-dacite. Challis volcanics are relatively sparse in the western batholith but increase in amount and thickness to the east. Metasediments of unknown correlation and age occur at House Mountain, Chimney Peak, and in a number of small roof pendants.

Older structures (Cretaceous?) in the batholith include a major low-angle fault at the base of House Mountain which may be the decollement for one of the large thrust sheets in eastern Idaho. The southern part of the Atlanta lobe is cut by northeast-oriented faults (parallel to the trans-Challis fault system, TCFS) related to Eocene extension and northwest-oriented faults formed during basin and range extenson in the Miocene. The basin and range faults form topographic breaks typical of basin and range topography. The combination of northeast and northwest faults has broken the batholith into a series of rhomboid blocks.

Mining districts in the batholith in the Hailey map area having over 100,000 ounces of gold production include: Boise Basin (contains eight districts and is the largest gold-producing area in Idaho), the Yuba District (Atlanta Hill), the Red Warrior District (Rocky Bar), the Camas District (Hailey gold belt), and the Neal District. Many of the deposits in these districts are in northeast-oriented shear zones.

Recent work on the Hailey CUSMAP project indicates that deposits in Atlanta Hill, the Hailey gold belt and Rocky Bar may be Paleocene or late Cretaceous in age. Other districts, such as Boise Basin, are Eocene or younger in age and related to structures in the TCFS. Granitic plutons and rhyolite dikes are important in some of these deposits, including mines in the Neal District, Boise Basin, and the Dixie District. In contrast, Rocky Bar and the Hailey gold belt contain few Tertiary rhyodacite-dacite dikes and almost no rhyolite. No rhyolite dikes are known near the Atlanta lode, but a swarm of northeast-trending rhyolite dikes cross the area immediately north of Atlanta Hill and Eocene granite may underlie mineralized zones of the hill.

A few small prospects are located along basin and range faults like the Iron Mountain-Deer Park fault. These may be Miocene or younger in age but none have had any substantial production.

LAKE CREEK MINERALIZED AREA, BLAINE COUNTY, IDAHO

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ABSTRACT

The Lake Creek mineralized area is on the west slope of the Boulder Mountains, northwest of Ketchum, Blaine County, Idaho. Structurally-controlled silver-lead-zinc mineral deposits occur in fine-grained calcareous rocks of the Pennsylvanian-Permian Wood River Formation. Deep water argillaceous rocks of the unconformably underlying Milligen Formation are the probable source of the metals. Large-scale east-vergent overturned folds in both formations were formed during the Late Mesozoic Sevier orogeny. In Paleogene time, these folds were cut by gently west-dipping, top-to-the-northwest, oblique-slip normal faults with N60°W mean transport direction. These detachment faults, previously interpreted to be thrust faults, are the result of crustal thinning in the upper plate of the Pioneer Mountains metamorphic core complex.

Mineral deposits in the Lake Creek area occur in two geologic settings. The highest grade ore concentrations are in vein-filled fissures and replacement deposits along shear zones in the hanging wall of the Lake Creek detachment fault. Lower grade deposits are not related to the detachment, but are found in bedding-parallel brecciated pods above the locally sheared unconformity between the Milligen and Wood River Formations. The dominant ore minerals are zinc and lead carbonates, sphalerite, galena, and chalcopyrite. Gangue minerals include quartz, calcite, hematite, and pyrite. Hydrothermal activity driven by the Challis magmatic episode (about 51 to 44 Ma) was probably responsible for the remobilization and concentration of metals. A new exploration model, based on the recognition of mineralization in the hanging wall of low angle detachment faults, can be applied to an extensive area of the Boulder and Pioneer Mountains.

LOCATION

The Lake Creek mineralized area (fig. 1) is about 10 mi northwest of Ketchum in Blaine County, Idaho, on the west flank of the Boulder Mountains. Five inactive mines and numerous prospects are in the Lake Creek, Eagle Creek, and Trail Creek drainages. The most extensive workings are in the Lake Creek drainage and include the Homestake, Long Grade, High Grade, Lake Creek, and Price Group properties (Numbers 1 through 5 on figs. 1 and 2). Access to the area is from gravel roads up Lake, Eagle, and Trail Creeks. Steep northeast-trending valleys and intervening tree-covered ridges characterize the topography where elevation ranges from 7,000 to 9,500 ft.

PREVIOUS WORK

Umpleby and others (1930), Service and Kellum (1956), and Tuchek and Ridenour (1981) described the ore deposits of the area. The geology was described in general by Dover (1981) and Link and others (1988), and in detail by Burton (1988).

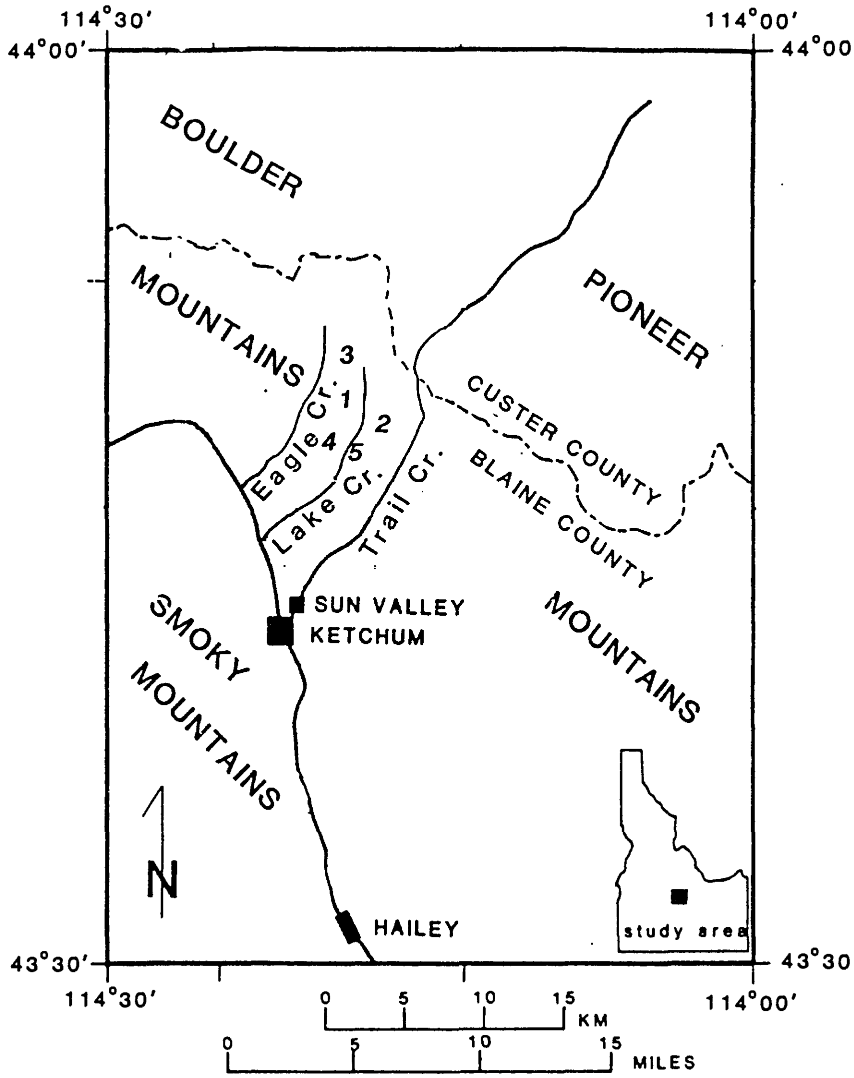

Figure 1. Location of Lake Creek mineralized area, Blaine County Idaho. Numbered mines include: (1) Homestake, (2) Long Grade, (3) High Grade, (4) Lake Creek, (5) Price claims.

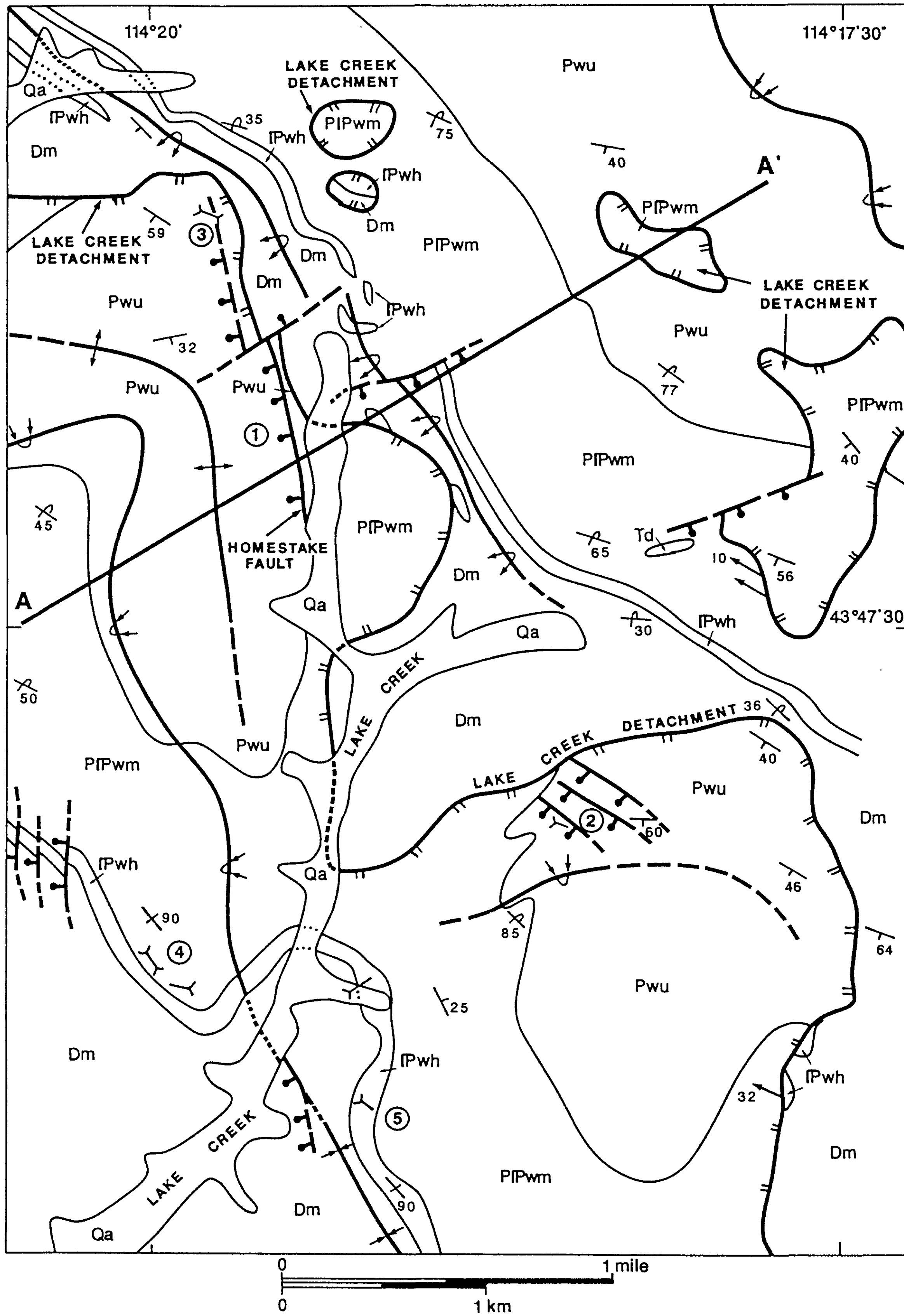

Figure 2. Geologic map of the headwaters of Lake Creek, Rock Roll Canyon quadrangle (after Burton, 1988). Locations of mines shown with numbers: (1) Homestake, (2) Long Grade, (3) High Grade, (4) Lake Creek, (5) Price claims.

LEGEND

QUATERNARY	Qa		ALLUVIUM	
	Unconformity			
PALEOGENE	Td		DACITIC DIKES	
PERMIAN – PENNSYLVANIAN	Pwu	**WOOD RIVER FORMATION**	UPPER MEMBER	
	PIPwm		MIDDLE MEMBER	
	IPwh		HAILEY CONGLOMERATE MEMBER	
	Unconformity			
DEVONIAN	Dm		MILLIGEN FORMATION	

CONTACT

PALEOGENE DETACHMENT FAULT

HIGH ANGLE NORMAL FAULT

OVERTURNED SYNCLINE

SYNCLINE

OVERTURNED ANTICLINE

ANTICLINE

30 STRIKE AND DIP OF BEDDING

85 STRIKE AND DIP OF OVERTURNED BEDDING

32 TREND AND PLUNGE OF FAULT STRIAE

ADIT

TRENCH

1 **HOMESTAKE MINE**

2 **LONG GRADE MINE**

3 **HIGH GRADE PROSPECT**

4 **LAKE CREEK MINE**

5 **PRICE MINE**

GEOLOGIC SETTING

Terrane Description

Paleozoic rocks of the central Idaho black-shale mineral belt (Hall, 1985) are exposed in the Lake Creek area. These include the Devonian Milligen Formation (argillites, cherts, fine-grained dolomitic sandstones) and overlying Pennsylvanian-Permian Wood River Formation (sandy limestones, calcareous and siliceous argillite, minor conglomerate, and fossiliferous limestone). Contact between the formations is a locally sheared unconformity.

The Paleozoic rocks are deformed into northeast-vergent overturned folds having tight anticlines and open synclines (figs. 2 and 3). Bedding dips steeply to the southwest on both limbs of these folds. Tight overturned anticlines are thought to be fault propagation folds cored by small displacement splay faults above a basal thrust. This thrust is inferred to exist at depth, but is not exposed in the Lake Creek mineralized area. Depth-to-detachment calculations, using the method of Woodward and others (1985), indicate that a basal thrust fault in the lower part of the Milligen Formation (about 2,000 ft above sea level) is necessary to explain the amount of shortening observed in the megascopically folded Wood River Formation. Both the folds and the inferred thrust are Mesozoic structures formed during the Sevier orogeny.

Northwest-striking oblique-slip normal faults (detachments) with shallow (10° to 30°) southwest dip cut the overturned folds. These faults generally place younger rocks of the upper and middle members of the Wood River Formation on older strata of the lower Wood River and Milligen Formations (fig. 3). Slip indicators on the faults indicate that the hanging-wall moved to the northwest. The faults are corrugated or curviplanar in cross section (fig. 3). Thus the faults range locally from normal to strike-slip. They are related to Paleogene extension in the upper plate of the Pioneer Mountains Core complex (Wust, 1986; Burton and others, 1989). Within the hanging wall of the shallowly dipping detachment faults are northwest-striking, moderately to steeply southwest-dipping normal faults which probably are listric and which sole into the underlying detachments.

At the headwaters of Lake Creek (fig. 2), the Lake Creek detachment fault strikes northwest and places upper and middle members of the Wood River Formation on folded rocks of the Milligen and Wood River Formations below (fig. 3). In most places, this results in a younger-over-older relationship. The Lake Creek fault was recognized by Umpleby and others (1930, p. 65) and Dover (1981) who mapped it as a Mesozoic thrust. Burton (1988) recognized that the Lake Creek fault was extensional, based on consistent N. 50° to 65° W. kinematic indicators (mainly fault striae indicating last movement to the northwest), and the long-recognized (Umpleby and others, 1930, p. 66) truncation of northeastward overturned folds by the fault. Klippen of the detachment occur on the top of the ridge east of the Homestake Mine, and in several places to the west between the ridge and the valley floor along the southwest-dipping slope. In these excellent exposures of the fault surface, the Lake Creek detachment exhibits a curviplanar geometry with large scale sinusoidal undulations. The axes of these corrugations are parallel to the northwest transport direction. Because of the pre-existing folds, the fault cuts both up and down section in rocks of the hanging wall.

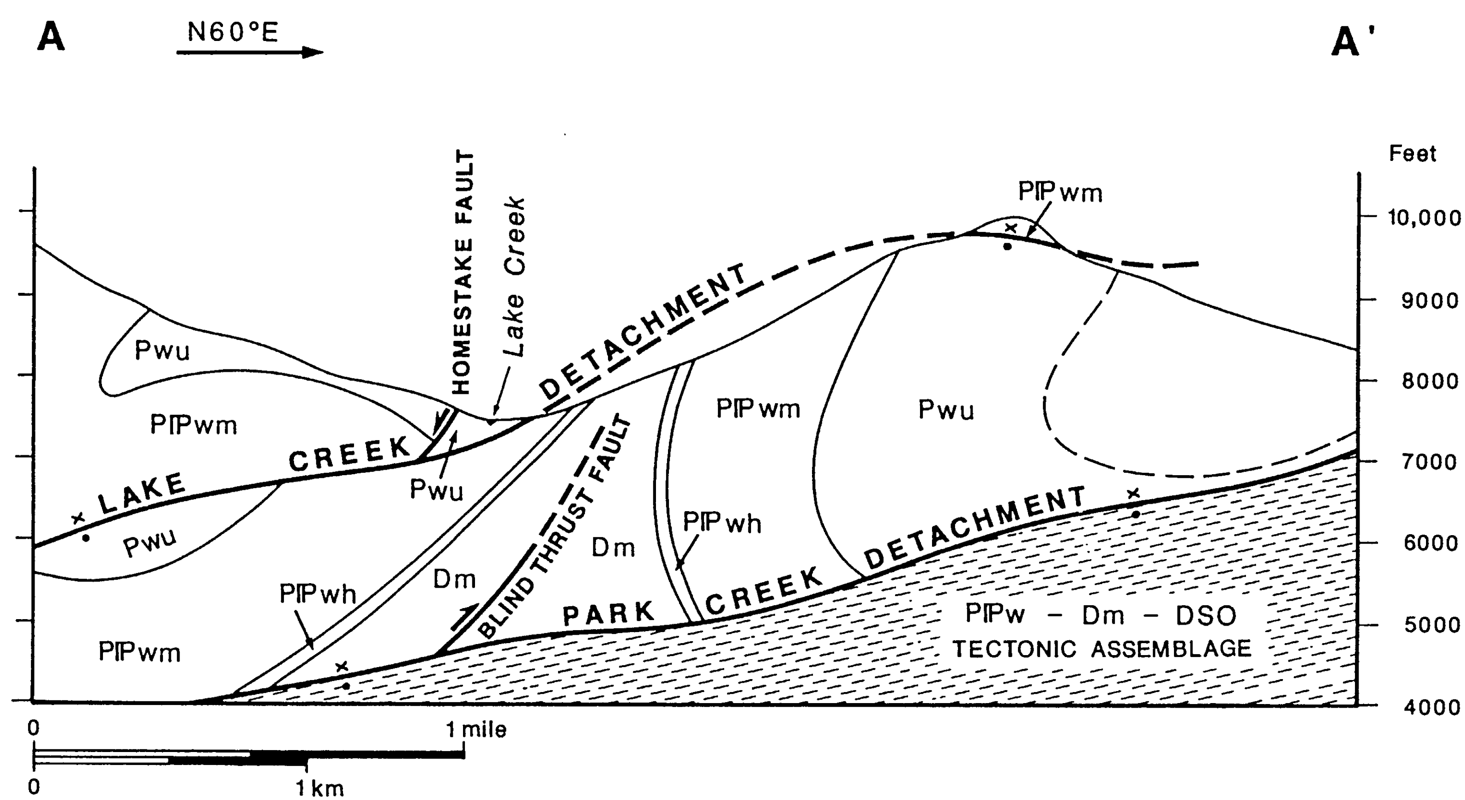

Figure 3. Geologic cross section (N60E). The Homestake Mine is in a shear zone of the Homestake Fault, a high angle normal fault synthetic to the underlying Lake Creek detachment fault. Motion on the Lake Creek detachment fault is oblique-normal slip with hanging wall toward N60°W. The fault displays a curviplanar geometry with axes of corrugations parallel to the movement direction. Tight anticlines in the Milligen and Wood River Formations are thought to be thrust fault propagation folds. Geologic symbols are the same as in figure 2.

Tightly folded and cleaved gray argillite of the Devonian Milligen Formation in the footwall of the Lake Creek detachment is suspected to be the major source of the ore metals in the Lake Creek mineralized area. The metals were probably remobilized from the Milligen Formation and deposited in the Wood River Formation by hydrothermal convection cells probably associated with dacite porphyry dikes which are present throughout the Lake Creek area. These intrusions are part of the regionally extensive Challis magmatic episode (about 51 to 44 Ma) (Moye and others, 1988; Fisher and Johnson, 1987).

Host Rock Characteristics

The host rock in the Homestake mine is carbonaceous silty micrite and calcareous mudstone of the upper and middle members of the Wood River Formation (fig. 4). Although the largest deposits are in fissure veins, replacement of calcareous wall rock occurs, with the wall rock composition appearing to control the site of deposition of ore minerals. Non-calcareous sandstone and quartzite are generally not mineralized.

The host rocks in the Lake Creek and Price Group prospects are conglomerate and limestone of the lower part of the Wood River Formation (fig. 4). Mineral deposits occur within a few tens of meters of the underlying Devonian Milligen Formation.

Alteration

Alteration of the host rock is limited. Iron-stained gossan occurs in a thin zone around the Lake Creek and Price Group deposits. Breccia and gouge in the mineralized fissures are bleached and have discontinuous wisps of hematite stain.

GEOLOGIC SETTINGS FOR MINERAL DEPOSITS

Two geologic settings are known to host mineral deposits in the area: (1) steeply southwest-dipping normal faults host vein-filled fissures or replacement deposits (Homestake, Long Grade, and High Grade properties) and (2) the sheared unconformity between the Milligen Formation and overlying Hailey Conglomerate member of the Wood River Formation hosts pods of mineralized rock (Lake Creek and Price Group claims).

Homestake Mine, Long Grade, and High Grade Prospects

The Homestake Mine is near the headwaters of the west fork of Lake Creek on the Rock Roll Canyon 7.5′ Quadrangle (fig. 2). The mine is located in secs. 3, 4, and 5, T. 5 N., R., 18 E., and in secs. 32 and 33, T. 6 N., R. 18 E., Warm Springs Creek Mining District, Blaine County, Idaho. Carbonaceous argillite and silty micrite of the upper member of the Wood River Formation host mineral deposits along a major northwest-striking (N. 26° W.) shear zone (the Homestake fault, figs. 2 and 3) that dips 45° to 75° southwest (Tuchek and Ridenour, 1981, p. 223). The shear zone can be traced on the surface for about 1 mi and can be inferred from discontinuous exposure for over 3 mi. The shear zone accommodated S. 70° W. extension and is synthetic to the underlying Lake Creek detachment fault, which forms a dip slope north of the mine in the northeast headwaters of the creek (figs. 2 and 3).

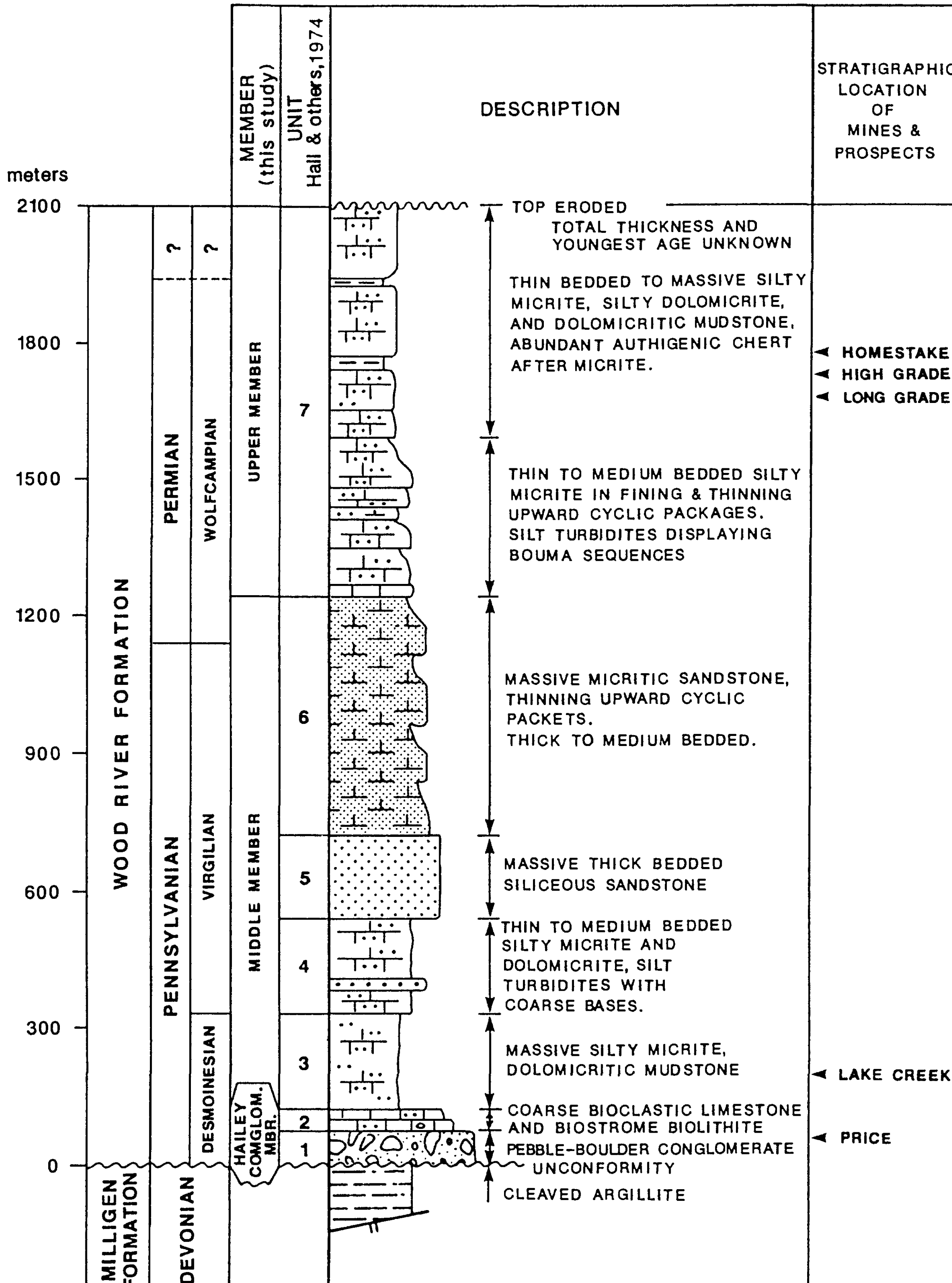

Figure 4. Stratigraphic section of the Wood River Formation after Burton (1988). Stratigraphic location of each mine is shown.

Primary ore minerals are zinc and lead carbonates, mainly smithsonite, calamine, and cerussite. Altered remnant galena occurs in the wall rock. On the margins of mineralized veins, the sheared, altered, brecciated, and veined country rock contains quartz, calcite, and hematite. Ore shoots range in thickness from 3 in to 7 ft and are from 10 ft to more than 100 ft in length. Production records from 1923 to 1961 indicate that the average ore grade from fissure filling and replacement deposits was 6.5 oz Ag, 4.5 percent Pb, and 9.5 percent Zn (Tuchek and Ridenour, 1981).

Ore occurs preferentially where the host rock is carbonaceous silty micrite. Ore does not occur where the shear zone cuts quartzite or siliceous mudstone of the middle member, Wood River Formation. The workings are described by Tuchek and Ridenour (1981) who estimate that at least 730,000 tons of submarginal reserves of Ag-Pb-Zn exist in unexplored portions of the shear zone.

The Long Grade mine is east of Lake Creek, about 1.2 mi south of the Homestake workings (fig. 2). Five adits expose northwest-trending, breccia-filled mineralized shear zones which cut gray silty mudstone and tan calcareous sandstone of the upper member of the Wood River Formation (fig. 4). The shear zones occupy the hanging wall of the Lake Creek detachment, and are directly analogous in geologic setting to the Homestake structure (Tuchek and Ridenour, 1981, p. 231). The zones strike N. 10° to 30° W. and dip 30° to 50° SW. Ore minerals include lead and zinc carbonates and minor amounts of galena, sphalerite, and pyrite in a gangue of calcite, hematite, and crushed limestone wall rock. Tuchek and Ridenour (1981) estimated that the Long Grade mine has 32,500 tons of submarginal reserves of Pb-Zn.

The High Grade prospect (Number 3 of figs. 1 and 2) is near the ridge top between Lake and Eagle Creeks, along the northward extension of the Homestake shear zone (Tuchek and Ridenour, 1981, p. 229). The controls on mineralization are analogous to those at the Homestake mine.

The Homestake, Long Grade and High Grade shear zones are confined to normal faults synthetic to the underlying Lake Creek detachment which formed during Paleogene rise of the Pioneer Mountains core complex. Thus the mineral deposits are probably Eocene or younger in age.

Lake Creek and Price Mines

The Lake Creek Mine (Number 4 on figs. 1 and 2), described in detail by Tuchek and Ridenour (1981), is about 800 ft northwest of Lower Lake Creek Lake, 900 ft above the valley floor. The workings follow a zinc-bearing and calcite-filled shear zone 0.3 to 3.5 ft thick within silty micrite of the middle member of the Wood River Formation (unit 3 of Hall and others, 1974) (fig. 4). The shear zone strikes N. 40° W., and dips from 80° SW. to vertical. The shear zone is parallel with the strike of beds in the country rock and probably represents bedding-parallel shear produced by Mesozoic folding.

The Price Group mine (number 5 on figs. 1 and 2) is about 1,300 ft southeast of Lower Lake Creek Lake, 1,000 to 1,600 ft above the valley floor. Veins are developed in calcite-filled shear zones up to 1,000 ft long which strike N. 40° W. and dip from 80° SW. to vertical. The host rock is conglomerate of the Hailey Conglomerate Member of the Wood River Formation. Umpleby and others (1930, p. 196) note that mineral deposits occur as pods along irregular zones of brecciation, joints, and discontinuous fractures.

The mineral deposits consist of pods located along the sheared unconformity between the Hailey Conglomerate Member of the Wood River Formation and the underlying Milligen Formation. They are most likely associated with differential shear between beds of different competence during folding.

Because the host structures of the Lake Creek and Price Group deposits are Mesozoic in age, mineralization could have occurred either in the late Mesozoic (probably associated with regional Cretaceous plutonism) or in Eocene time, coincident with mineralization in the Homestake, Long Grade, and High Grade prospects. The small size and podlike nature of this type of mineral deposit limit its value for future mineral development.

EXPLORATION MODEL

The recognition (Link and others, 1988; Burton and others, 1989) of Paleogene detachment faulting in the northern Boulder Mountains, and the presence of the Homestake and Long Grade deposits in normal faults confined to the upper plate of such detachments suggests a new exploration model for silver-lead-zinc deposits in the region. Previously, these deposits were thought to be associated with Mesozoic compressional structures. In the Lake Creek area, however, only small uneconomic deposits are hosted by structures that formed during the Mesozoic.

The Lake Creek deposits are similar in tectonic setting to the silver-lead-zinc deposits in the southern Trigo Mountains of Yuma County, Arizona (Garner and others, 1982), and to the gold deposits at the Picacho Mine, Imperial County, California (Liebler, 1988). In these places mineralization occurs in the hanging wall of extensional detachments. An areally extensive network of vertically stacked detachment faults occurs around the south, west, and northwest sides of the Pioneer Mountains metamorphic core complex, an area that comprises most of the Boulder and Pioneer Mountains.

Throughout the Pioneer and Boulder Mountains region, the following conditions, favorable for mineralization, exist:

(1) a source of metals is available from metamorphosed Devonian and older Paleozoic strata;

(2) Mesozoic compressional structures exist which localize ore deposits by serving as low-permeability barriers; these structures serve to increase the stratigraphic separation of the Paleogene extensional faults which cut them;

(3) low-angle Paleogene detachment faults are vertically stacked in the upper plate of the Pioneer Mountains core complex and associated shear zones may have served as sites for formation of vein and replacement ore;

(4) the presence of folds of Mesozoic age and Eocene hypabyssal intrusions suggest that remobilization and concentration of metals by hydrothermal convection cells could have occurred during Cretaceous compression or in association with Eocene hypabyssal intrusion and volcanism;

(5) calcareous fine-grained sedimentary rocks are available as hosts for mineralization.

The Lake Creek detachment model offers an alternative to existing mineral deposit models in south-central Idaho. The extensive outcrop exposure of the Boulder and Pioneer Mountains will facilitate mineral exploration.

ACKNOWLEDGMENTS

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GEOLOGY, PETROLOGY, AND GEOCHEMISTRY OF THE EOCENE INTRUSIVE SUITE OF THE MACKAY PEAK AREA, SOUTH-CENTRAL IDAHO

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INTRODUCTION AND REGIONAL SETTING

Preliminary geological and geochemical results are given from Eocene intrusive rocks exposed in the Mackay Peak area, Custer County, Idaho, located in the northwest corner of the Idaho Falls 1°x2° quadrangle (fig. 1). The 57 km^2 field area lies within the White Knob Mountains in the southeastern portion of the Challis volcanic field. The Challis volcanic field (51-44 Ma) was emplaced during an episode of intense Eocene magmatic activity (55-40 Ma) and is part of a regionally extensive Eocene volcanic belt in the northwestern United States (Moye and otherss, 1988). The Mackay Peak area has been dissected to subvolcanic levels, providing excellent exposures of a cogenetic intermediate-to-felsic intrusive rocks.

Intrusive rocks of the Mackay Peak area occur as intermediate-to-silicic, northeast-trending, elongate plutons and dike swarms which intrude the carbonates of the Mississippian Copper Basin Formation and White Knob Limestone. Nelson and Ross (1968) classify the intrusive suite as: (1) quartz monzonite, (2) granite, (3) leucogranite porphyry, (4) quartz latite, (5) rhyolite, and (6) porphyritic rhyolite.

Quartz monzonite, granite, and leucogranite porphyry comprise the plutonic bodies. Quartz monzonite is intruded by the granite, and all intrusive phases are cross-cut by dikes of rhyolite and porphyritic rhyolite, indicating that the intrusive rocks became progressively more felsic with time. Rhyolite dikes intrude all levels of the volcanic/plutonic complex and represent the waning phases of magmatic activity in the southeastern Challis volcanic field. Small, northeast-trending, late-stage basic andesite dikes (approximately 53 percent SiO_2) are randomly interspersed within the predominantly felsic dike swarms. The preferred orientation of the intrusive bodies along a northeast trend expresses the control exerted by regional, northeast-trending Eocene extensional structures.

In the Mackay Peak area, base metal skarns are developed within the northeastern contact aureoles of the granite and leucogranite and the White Knob Limestone. Skarns are locally found along the contact zone between the porphyritic rhyolite and the White Knob Limestone.

PETROGRAPHY

The intrusive suite has a phenocryst assemblage dominated by quartz, potassium feldspar, and plagioclase with minor biotite and hornblende. Accessory minerals are apatite (inclusions in biotite), sphene, clinopyroxene, and Fe-oxides. Zircon is found in trace amounts in the granite and leucogranite porphyry.

Rapakivi texture and spheroidal, embayed, smoky quartz phenocrysts characterize the granite, leucogranite porphyry, and porphyritic rhyolite. The granite and leucogranite porphyry often display myrmekitic texture. The rhyolite dikes are dominated by an aphanitic groundmass (composing about 85% of the rock mass) and contain small phenocrysts of smoky, dipyramidal quartz

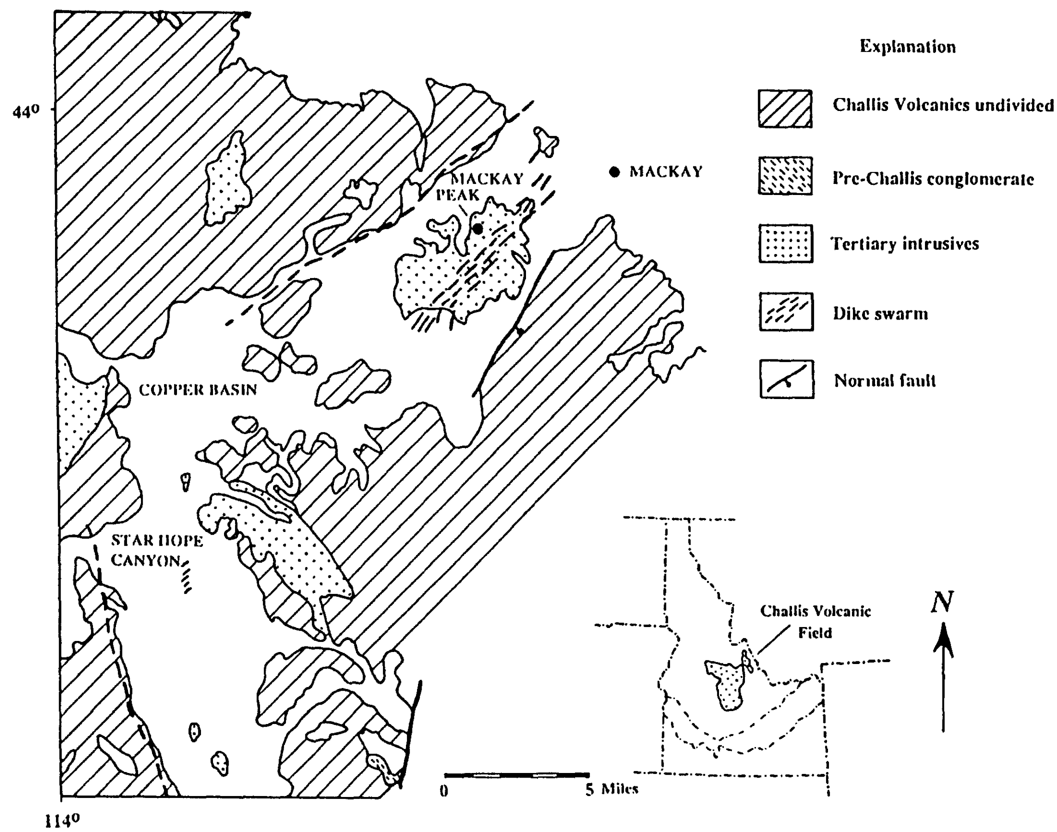


Figure 1 : Location map of Mackay Peak area, northwestern portion of the Idaho Falls 1° x 2° quadrangle

and minor biotite. The rhyolite dikes have undergone a significant amount of silicification and argillic alteration.

MAJOR AND TRACE ELEMENT GEOCHEMISTRY

Major and trace element variation within the rock suite can be explained by fractional crystallization of the observed phenocryst minerals. If the concentrations of highly compatible major elements show distinct variation within a comagmatic rock suite, the main process through which the melt evolved was fractional crystallization; if the concentration of these major elements were to have remained relatively constant, the main mechanism of evolution would have been partial melting. The inverse relationships Fe_2O_3 and MgO to SiO_2 (figs. 4 and 5) is explained by the crystallization of biotite, hornblende, pyroxene, and Fe-oxides from the parental magma. Fractionation of apatite explains depletion of P_2O_5 (fig. 7), and the fractionation of opaque Fe-Ti oxides, together with sphene, explain depletion of the melt in TiO_2 (fig. 2). Fractionation of feldspars will deplete the melt in Al_2O_3 (fig. 3), while the crystallization of potassium feldspar will deplete the melt in K_2O (fig. 6).

The evolution of the melt through fractional crystallization is supported by the behavior of trace elements with large *Kd* values (specifically Sr, Ni, Cr, and V). If the concentrations of elements with high *Kds* show significant variation (figs. 8, 9, and 10), then the main process through which the melt evolved was fractional crystallization, rather than partial melting (Robb, 1983; Hanson, 1978).

Feldspar fractionation is suggested by the marked decrease in Sr versus Rb (the most incompatible trace element), since plagioclase and potassium feldspar both have high *Kd* values for Sr (fig. 10). Potassium feldspar and biotite have high affinities for Ba, and their fractionation explains the positive correlation between Sr and Ba (fig. 11). Figure 9 shows the relationship of Zr and Ba versus Rb; the sequestering of Ba during potassium feldspar and biotite fractionation explains the decrease in Ba when correlated with Rb content. The uniform content of Zr throughout the rock suite negates using zircon as an incompatible element or fractionation index.

Assuming perfect crystal fractionation (Rayleigh Law behavior and $D_{bulk}=0$ for Rb, approximately 46 percent fractionation of major minerals (quartz, potassium feldspar, plagioclase, biotite, and hornblende) would explain the geochemical evolution of the rock suite during evolution of the melt. Despite its compatibility with biotite, Rb is the most incompatible element in relation to other silicate minerals within the intrusive suite and so was chosen as an index of fractionation. Because biotite is a relatively minor mineral phase, it is felt that crystallization of biotite would have only negligible impact on the use of Rb as a fractionation index.

CONCLUSION

Intrusive rocks of the Mackay Peak area, Custer County, Idaho, comprise an Eocene hypabyssal complex. The intrusions are part of an intermediate to silicic suite, emplacement of which was controlled by northeast-trending structures parallel to the trans-Challis fault system. Field relationships indicate that initial intrusive activity was intermediate in composition and

Fig. 2

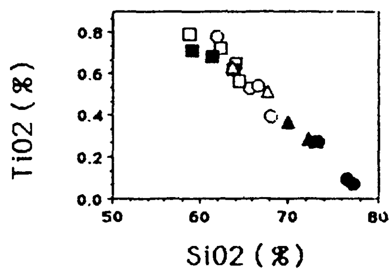


Fig. 3

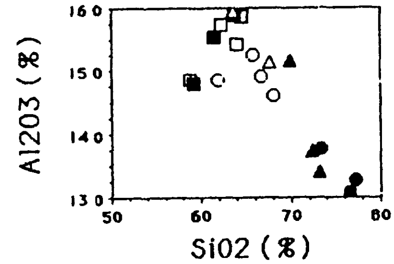


Fig. 4

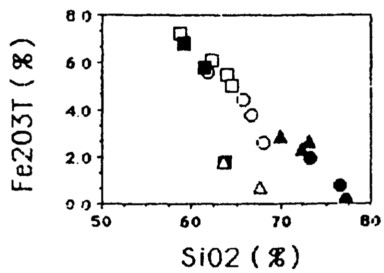


Fig. 5

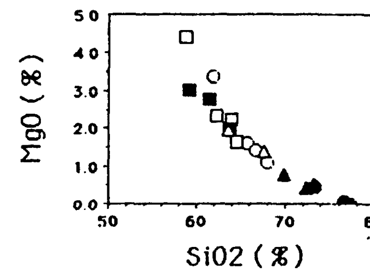


Fig. 6

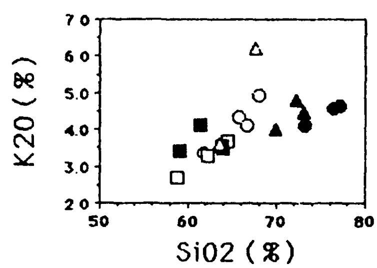


Fig.7

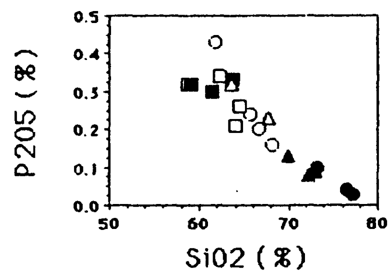


Fig. 8

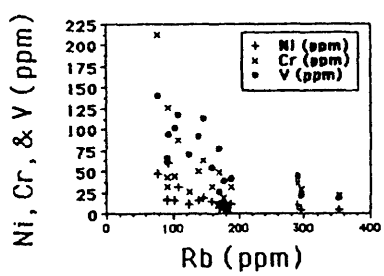


Fig. 9

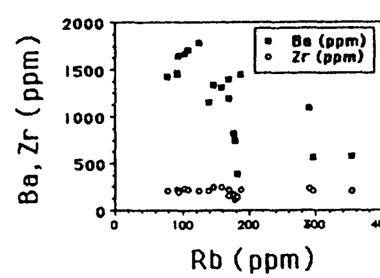


Fig. 10

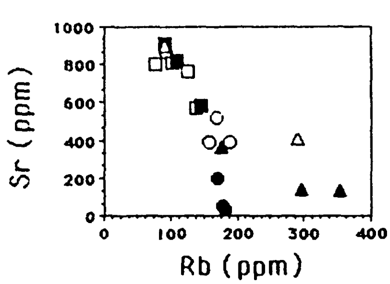
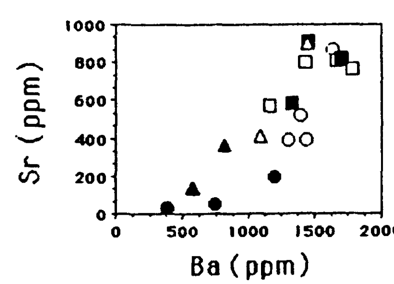


Fig. 11



○ Por.Rhyolite ● Rhyolite □ Qtz.monzonite ■ Qtz.latite △ Leucogranite ▲ Granite

Figure 2 - Figure 11 :

Geochemical plots of bulk rock major and trace element data derived from the intrusive suite from the Mackay Peak area.

became progressively felsic with time; emplacement of rhyolite dikes represent the waning stage of intrusive activity. Geochemical and petrographic data suggest that compositional and mineralogical variation of the rock suite is best explained by fractional crystallization of approximately 50% major minerals (quartz, potassium feldspar, plagioclase, plus minor biotite and hornblende).

ACKNOWLEDGMENTS

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GEOCHEMICAL EXPRESSION OF POORLY EXPOSED METALLIZATION IN THE SOUTHEASTERN CHALLIS VOLCANIC FIELD, IDAHO--EMPHASIS ON BIOGEOCHEMICAL METHODS

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Biogeochemical and geochemical exploration techniques were used in the Lava Creek mining district, north of Craters of the Moon National Monument, and in the Baker Creek area near Ketchum to assess the potential for new metallic mineral occurrences. Stream sediments, soils, heavy-mineral concentrates, and several species of plants, principally big sagebrush (*Artemisia tridentata*), were collected in reconnaissance-scale, drainage-basin surveys. Biogeochemical methods can be especially effective where surface expression of mineralization is poor.

The Lava Creek area was selected because of its historical importance as a silver producer with byproduct gold. Most of this area is underlain by Eocene Challis Volcanics that have been so hydrothermally altered that good outcrop is extremely limited. Hydrothermal alteration east of Baker Creek chiefly affects dacitic lava; a central zone of intense argillic alteration is surrounded by a broader zone of propylitic alteration. The Baker Creek area showed no permanent evidence of mineral prospecting when the survey was made, but it has geological characteristics that favor the existence of an epithermal precious-metal system.

At Lava Creek, initial results showed geochemical anomalies in sagebrush which suggested that a possible host for epithermal, precious metals was the Mississippian McGowan Creek Formation (chiefly turbidites), peripheral to the formerly active mines. Follow-up sampling of sagebrush, stream sediments, and soil defined a source of a Hg-Ag-Au-As-Sb anomaly in a fault breccia near a poorly exposed dacite porphyry intrusion.

In the Baker Creek area, heavy-mineral concentrates all contain abundant barite, even beyond the zone of propylitic alteration. Botryoidal pyrite was also found, but it is restricted to the alteration zones. Visible gold and pyromorphite, a lead phosphate, were found in heavy-mineral concentrates from a small drainage near a stockwork-veined outcrop in the central zone of argillic alteration. A grab sample of the outcrop contained 50 ppm Au, 1.5 ppm Ag, and 50 ppm Mo. The maximum concentration of Au in nearby stream sediments and upland soils was only 2 ppb, but aquatic mosses from within the argillically altered zone contained as much as 64 ppb Au in the ash. This intensely altered zone is characterized by anomalous levels of Ag, Mo, Pb, Cd, and Zn, and, to a lesser extent, Au, Bi, Mn, As, Sb, and Tl in the various media samples. The geochemical signature suggests a Ag- and Mo-rich, low-temperature, epithermal type of mineralization.

GEOLOGY AND ORE DEPOSITS OF GARFIELD CANYON, BLAINE COUNTY, IDAHO

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Garfield Canyon is located in the Muldoon Pb-Ag-Zn mining district on the western boundary of the Idaho Falls 1°x2° quadrangle (fig. 1). Hosts for mineralization are the Devonian Carey Dolomite and the Mississippian Copper Basin Formation, predominantly the Drummond Mine Limestone member, which are interpreted to be part of the Copper Basin thrust plate. The Carey Dolomite lies unconformably beneath the Copper Basin Formation, which is broadly folded and east-dipping.

Intruding the Copper Basin Formation is the Garfield Stock, a 3 km^2, Eocene, porphyritic biotite-pyroxene quartz monzonite. The Muldoon stock, 1/2 km southeast of the Garfield stock, yields a K-Ar age of 49 Ma. Geophysical evidence suggests that the Garfield stock and Muldoon stock are joined at depth; therefore, the Garfield stock is assumed to be 49 Ma. The Garfield stock becomes coarser toward the center. Mafic minerals and plagioclase are abundant at the border, but decrease toward the center, with increasing quartz. The border of the stock is characterized by subhedral, equigranular (0.5 mm - 1.0 mm) mafic minerals (40 percent) including biotite, hornblende, and pyroxene, with smaller (0.5 mm) tabular plagioclase (60 percent). Quartz is sparse. Toward the center of the stock, subhedral mafic minerals (25 percent) increase in size from 0.5 mm to 3.0 mm. Plagioclase (30 percent) is commonly zoned and is 3.0 mm to 4.0 mm. Anhedral quartz becomes abundant (43 percent) and is 0.5 mm in size. Potassium feldspar is common (2 percent) but is less than 0.5 mm in size. Cutting the stock and sub-parallel to bedding are numerous northwest-trending rhyolite dikes. Mafic dikes, both mineralized and unmineralized, are less abundant and do not have clearly defined structural control.

Three sets of oblique-slip faults are recognized: (1) NW bedding-parallel, (2) N50°E, and (3) N80°E. Bedding-parallel faults are probably Mesozoic in age and formed as a result of broad folding, since they are deformed with bedding. Actual faulting is localized, and orientations of folded stylolites indicate that movement occurred in a reverse direction. The N50°E and N80°E faults offset the stock and rhyolite dikes and are therefore younger than 49 Ma. N50°E faults are predominantly joints with local minor offset up to 4 m and are subparallel to regional NE-trending trans-Challis fault system. The N80°E faults appear to be regionally extensive, with possibly a great amount of offset. Age relations between N50°E and N80°E faults are equivocal.

Mineralization appears to be exclusively controlled by the intersection of bedding-parallel faults with N50°E and N80°E faults. Reverse movement of bedding-parallel faults created open spaces which acted as sites for later mineral deposition. Therefore, mineralization appears to be local and of high grade. Mineralization is best exposed at the Eagle Bird Mine. At this location, 1 m by 0.5 m pods of pyrrhotite, galena, sphalerite, and scorodite extend down dip of a bedding-parallel fault where it intersects a N50°E fault. There appears to be a rudimentary zoning of sulfides, with chalcopyrite and pyrite more abundant closer to the stock. Although mineralization is spatially associated with the stock, there also is an abundance of rhyolite dikes intruded along the bedding strike; therefore, it is questionable as to

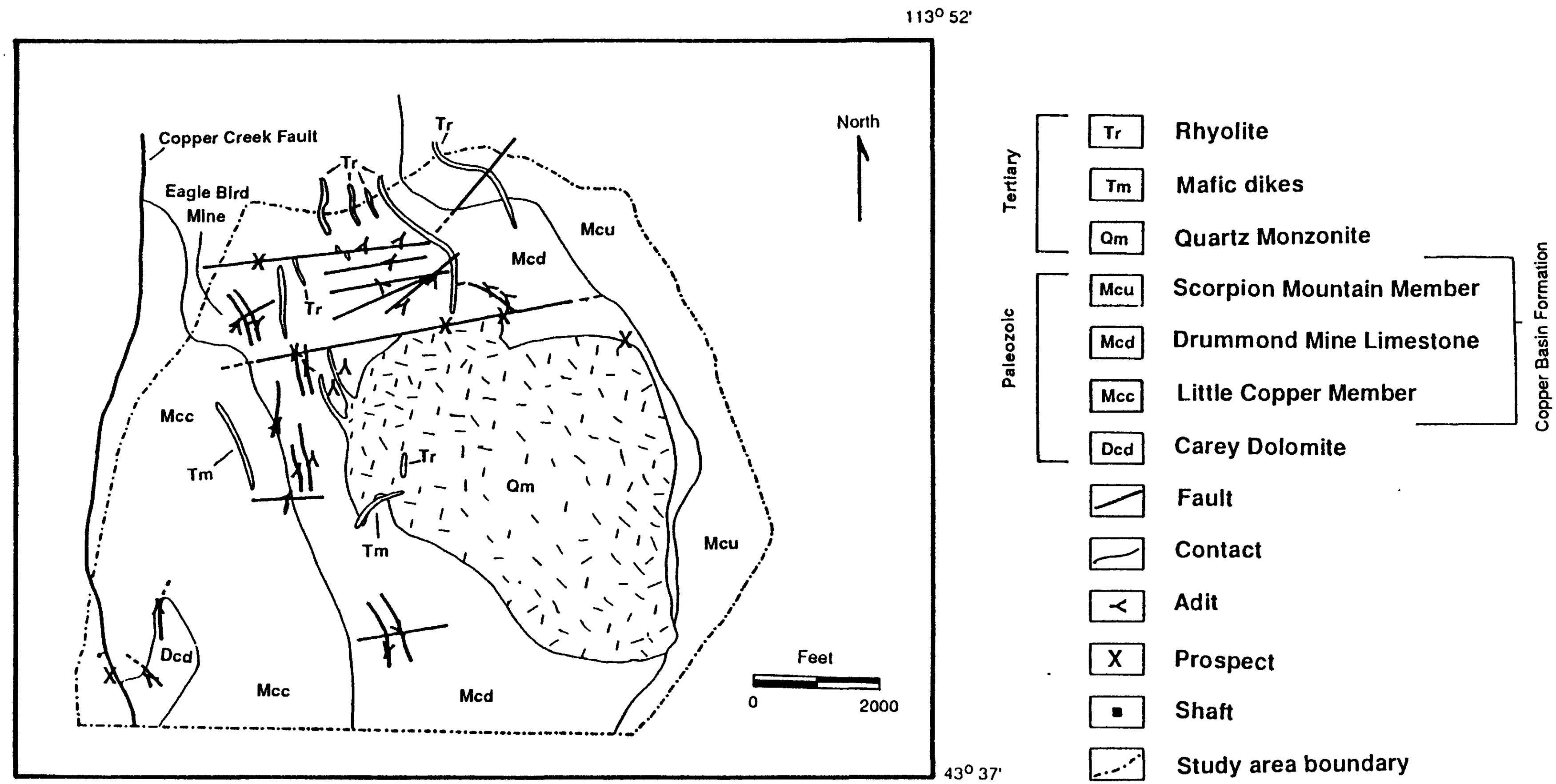

Figure 1. Generalized geologic map of Garfield Canyon area.

whether the stock or rhyolite dikes provided the hydrothermal influence for mineralization. Potential for low-grade, high-tonnage deposits is unknown at this time, but since mineralization is fault controlled, it is considered to be low.

Future laboratory work will include: major-, minor- and trace-element geochemistry, with thin sections to characterize the intrusive suite, ICP analyses and polished sections to characterize the mineralization suite, and fluid inclusion studies to determine the temperature of emplacement of mineralization.

Support for this project has been provided by the Idaho Geological Survey Idaho Initiative, the U.S. Geological Survey Hailey 1°x2° CUSMAP, the Idaho State Board of Education, a Grant-in-Aid of Research from Sigma Xi, and by the Idaho State University Graduate Student Research Committee.

GEOLOGY AND MINERAL DEPOSITS OF THE WASHINGTON BASIN AREA, CUSTER COUNTY, IDAHO

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INTRODUCTION

Washington Basin is located at the southern terminus of the White Cloud Peaks in the southeastern corner of Custer County, Idaho. The basin is a spectacular glacial cirque with a southeasterly aspect, and is drained by Washington Creek, a tributary at the headwaters of Germania Creek (fig. 1). The terrain is extremely rugged, with elevations ranging from 3187 m (10,519 ft) on Washington Peak to 2545 m (8,400 ft) at the east end of the cirque. In this report, the term "Washington Basin area" refers to both Washington Basin itself and to Germania Basin, located directly to the south. The basin is bounded on the north by the headwaters of Fourth of July Creek, on the west by Champion Lakes Basin, on the east by Washington Lake Creek, and on the south by Germania Creek. The area is reached by following the Pole Creek road 22 km from State Highway 75, then continuing 3.5 km along a rough jeep trail into the basin itself. The Washington Basin area is within the Sawtooth National Recreation Area, and is withdrawn from new mineral entries; current exploration is limited to existing claims. The White Cloud Peaks adjacent to Washington Basin are currently under review for inclusion in the White Cloud-Boulder Wilderness Area.

The Washington Basin area is situated on the boundary between the Hailey and Challis 1°x2° quadrangles. The geology of the area has never been described in detail. The geology of Washington Basin was briefly described in conjunction with the regional geology of the White Cloud Peaks by Ross (1963), Fisher and others (1983), Sengebush (1984), Hall (1985), and by Hall and Hobbs (1987). The geologic interpretations presented in this report represent a substantial revision of the ideas presented in these earlier works.

The mineral deposits of the Washington Basin area were described by Van Noy and others (1986), who included the Washington Basin area in the Germania Creek Mining District. Their deposit descriptions have been incorporated into this report.

HISTORICAL PERSPECTIVE

The Washington Basin area has been the site of intermittent exploration and mining activity since the late 1800s. The first mining claims were located in 1879, with the majority of mining activity occurring in the 1890s and early 1900s. Other periods of notable activity were the mid-1930s and the early 1970s. Over 329 mining claims have been located in the area, of which 10 are patented lode claims (Van Noy and others, 1986).

Although records of production are limited, the major producers in the area were apparently located on the northwestern flank of Bible Back Mountain (the Idahoan and Old Bible Back claims), and at the southwestern end of Washington Basin (Black Rock claims) (fig. 2). Quantitative estimates of the volume of ore removed are nonexistent, but monetary values (1915 prices) exceeded $250,000 from Bible Back Mountain, and $125,000 from the head of

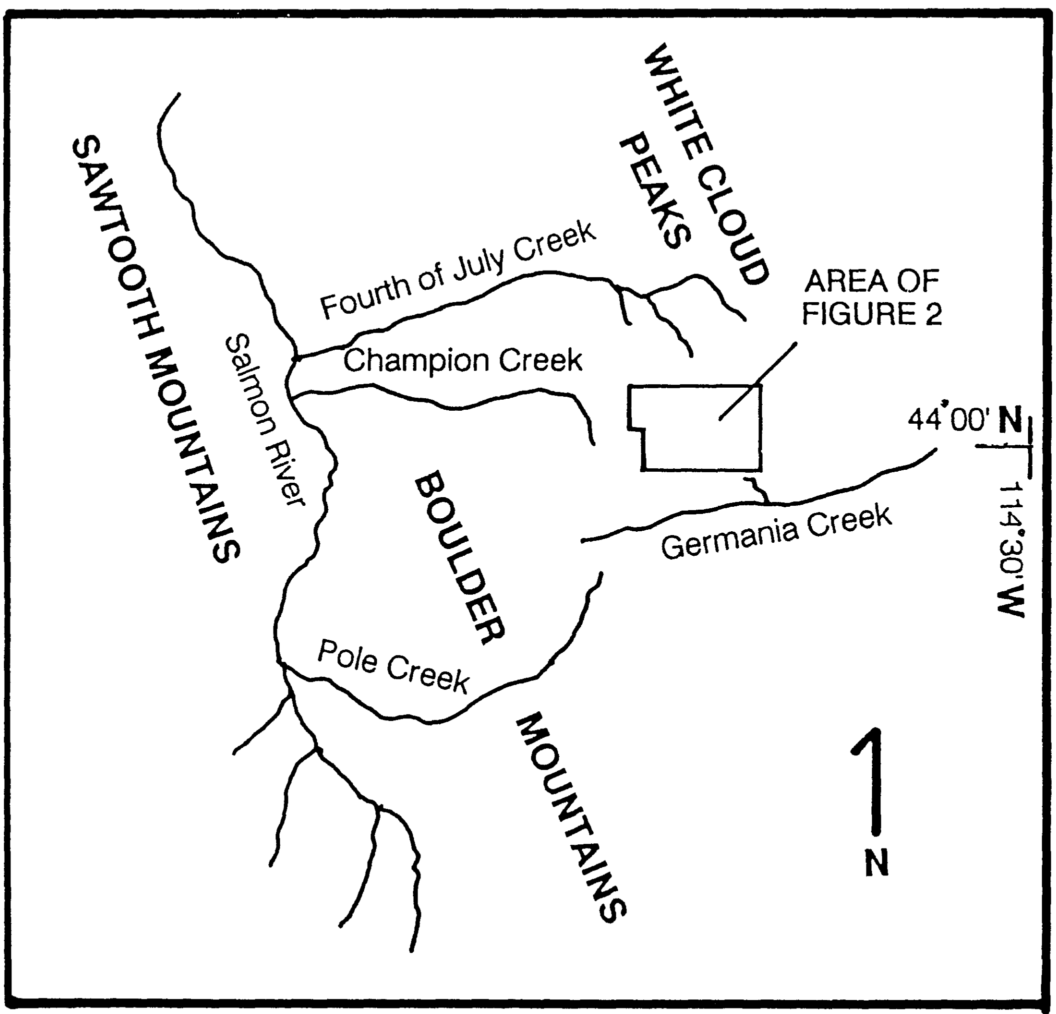

Figure 1 - Geographic Location Map of the Washington Basin Area

Washington Basin (Umpleby, 1915). The production in the Washington Basin area has been primarily in lead-silver ores, although some gold values have been reported (Van Noy and others, 1986). Small amounts of copper and zinc also may have been produced.

Exploration activity has been sporadic since the main period of activity in the early 1900's. The Washington Basin area has potential for minable lead-silver ores adjacent to the well-documented northeast mineral trend, and there is high potential for gold skarn deposits. Other reported commodities include copper, zinc, antimony, tungsten, bismuth, tellurium, and selenium (Van Noy and others, 1986). The U.S. Bureau of Mines reports that between 25,000 and 100,000 tons of submarginal ore-grade material (1974 prices) may exist in the Washington Basin area, and that there is a potential for the discovery of minable ore shoots in the area. The area is currently the target of extensive gold exploration activities, including an ongoing drilling program. Environmental concerns will significantly impact any proposed development plan, and may hinder future exploration in the area.

GEOLOGIC SETTING

Bedrock in the Washington Basin area consists of complexly folded and faulted Paleozoic strata, which have been intruded by at least two different igneous stocks. The mineral deposits in the area display a strong structural control, and at least two separate mineralizing events are believed to have occurred.

Recent mapping has resulted in significant revision of previously postulated stratigraphic and structural models. Following sections describe the geology of the area in terms of stratigraphy, structure, and igneous intrusions; mineral deposits are described at the end of this report.

STRATIGRAPHY

Washington Basin is underlain by three Paleozoic sedimentary units: the Paleozoic Salmon River assemblage, the Pennsylvanian-Permian Wood River Formation and the Lower Permian Grand Prize Formation (fig. 2). Mineral deposits are hosted by each of the units in the Washington Basin area, although the most important host is the Grand Prize Formation, which underlies the majority of the basin.

The sedimentary rocks in the area have been assigned to variety of formations in previous reports, including the Pole Creek Formation (Fisher and others, 1983), the Wood River Formation (Tschanz and others, 1986), and the Grand Prize Formation (Hall, 1985; Mahoney and Sengebush, 1988).

Salmon River Assemblage

The Salmon River assemblage in the Washington Basin area includes dark carbonaceous argillite, limestone, siltstone, and minor sandstone exposed on the ridge north of the basin (fig. 2). The unit extends northward from Washington Basin, underlying all of the Phyllis Lake cirque, and continuing north across Fourth of July Creek. The Salmon River assemblage forms the core of a south-plunging, east-vergent anticline, the dominant structure of the area. In Washington Basin, the southern boundary of the assemblage forms a

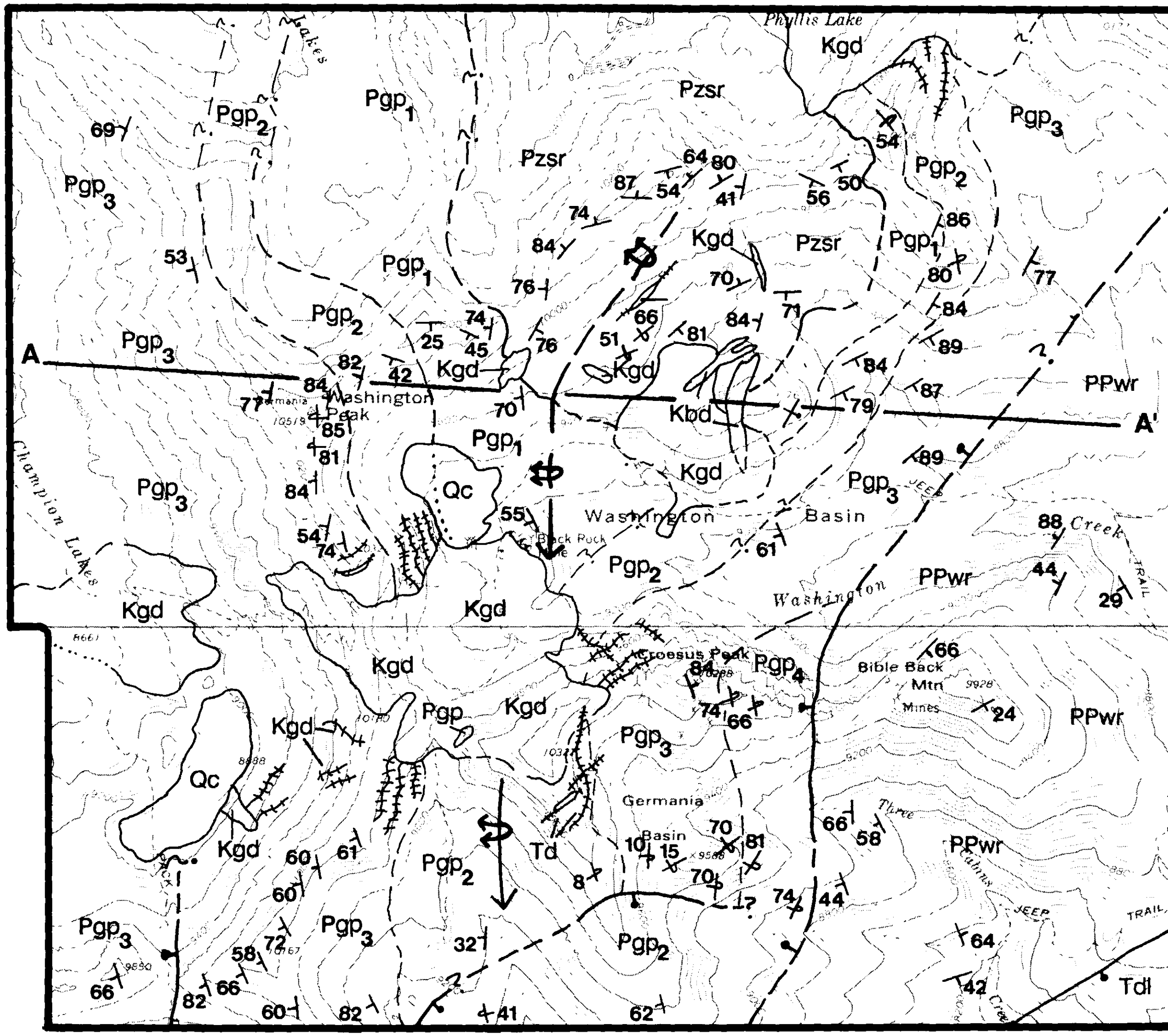

Figure 2 Geologic Map of the Washington Basin Area, Custer County, Idaho

KEY

Qc	Quaternary colluvium
Tdl	Tertiary dacite lava
Td	Tertiary dacite porphyry dikes
Kbt	Cretaceous(?) biotite tonalite
Kgd	Cretaceous biotite granodiorite
Pgp_1	member 1
Pgp_2	member 2
Pgp_3	member 3
Pgp_4	member 4
PPwr	Pennsylvanian-Permian Wood River Formation
Pzsr	Paleozoic Salmon River assemblage

Members 1–4: Permian Grand Prize Formation

overturned anticline

strike and dip

contact

fault

concave-north map pattern with the overlying Grand Prize Formation (fig. 2). The lower contact of the Salmon River assemblage is not exposed in this area.

The Salmon River assemblage is primarily composed of thin-bedded, dark-gray to black, carbonaceous argillite that locally is iron-stained and commonly has a phyllitic sheen. Locally, cleavage nearly parallel to bedding is well developed. The dark argillite is intercalated with medium-bedded, blue-gray sandy limestone, thin-bedded parallel to cross-laminated, fine-grained sandstone, and thin-bedded, dark-gray siltstone. The unit is highly deformed, and contains mesoscopic folds, cleavage, quartz veining, and fractures which obscure original bedding relationships. The Salmon River assemblage is locally contact metamorphosed to calc-hornfels, and is commonly silicified.

The age of the Salmon River assemblage is problematic, with age determinations including Late Cambrian, Devonian, and Late Mississippian. The assemblage currently is designated simply as Paleozoic (Hall and Hobbs, 1987). The Salmon River assemblage in Washington Basin includes similar sedimentary facies as the Devonian Milligen Formation to the south, and occupies a similar structural position (Link and Mahoney, this volume). The carbonaceous units within the assemblage commonly host mineral deposits, and the assemblage contains numerous mines and prospects throughout the White Cloud Peaks.

Grand Prize Formation

The Lower Permian Grand Prize Formation unconformably overlies the Salmon River assemblage in the Washington Basin area. The Grand Prize Formation is at least 2500 m thick, and is subdivided into four informal members, all of which are exposed in the Washington Basin area. A complete stratigraphic description of the formation can be found in Mahoney and Sengebush (1988).

The nature of the contact between the Grand Prize Formation and the Salmon River assemblage has been the subject of dispute among workers in the region. Hall (1985) and Hall and Hobbs (1987) describe the contact as a thrust fault, with the Grand Prize Formation thrust over the Salmon River assemblage along an intensely brecciated fault zone. Sengebush (1984) and Mahoney and Sengebush (1988) suggest that the contact is actually a sheared unconformity, and that the brecciated zone is a tectonically deformed clast-supported conglomerate within the lowest member of the Grand Prize Formation that was originally deposited unconformably on the underlying Salmon River assemblage.

Examination of the contact on the ridge north of Washington Basin strongly supports the sheared unconformity hypothesis. In this location, the contact consists of matrix-supported (locally clast-supported) conglomerate and intercalated coarse-grained quartzite overlying black argillite with a low-angle structural discordance. The existence of intercalated quartzite beds, sedimentary structures, including graded bedding, and the presence of subrounded black argillite clasts strikingly similar to the underlying Salmon River assemblage within the conglomerate beds suggest deposition of the lowest member of the Grand Prize Formation on top of the Salmon River assemblage. The preferred clast orientation and brecciation within the conglomerate is the result of shearing along the unconformable contact during megascopic folding in the Mesozoic.

In the Washington Basin area, the Grand Prize Formation is contained in both limbs of a south-plunging, east-vergent anticline (figs. 2, 3). In the

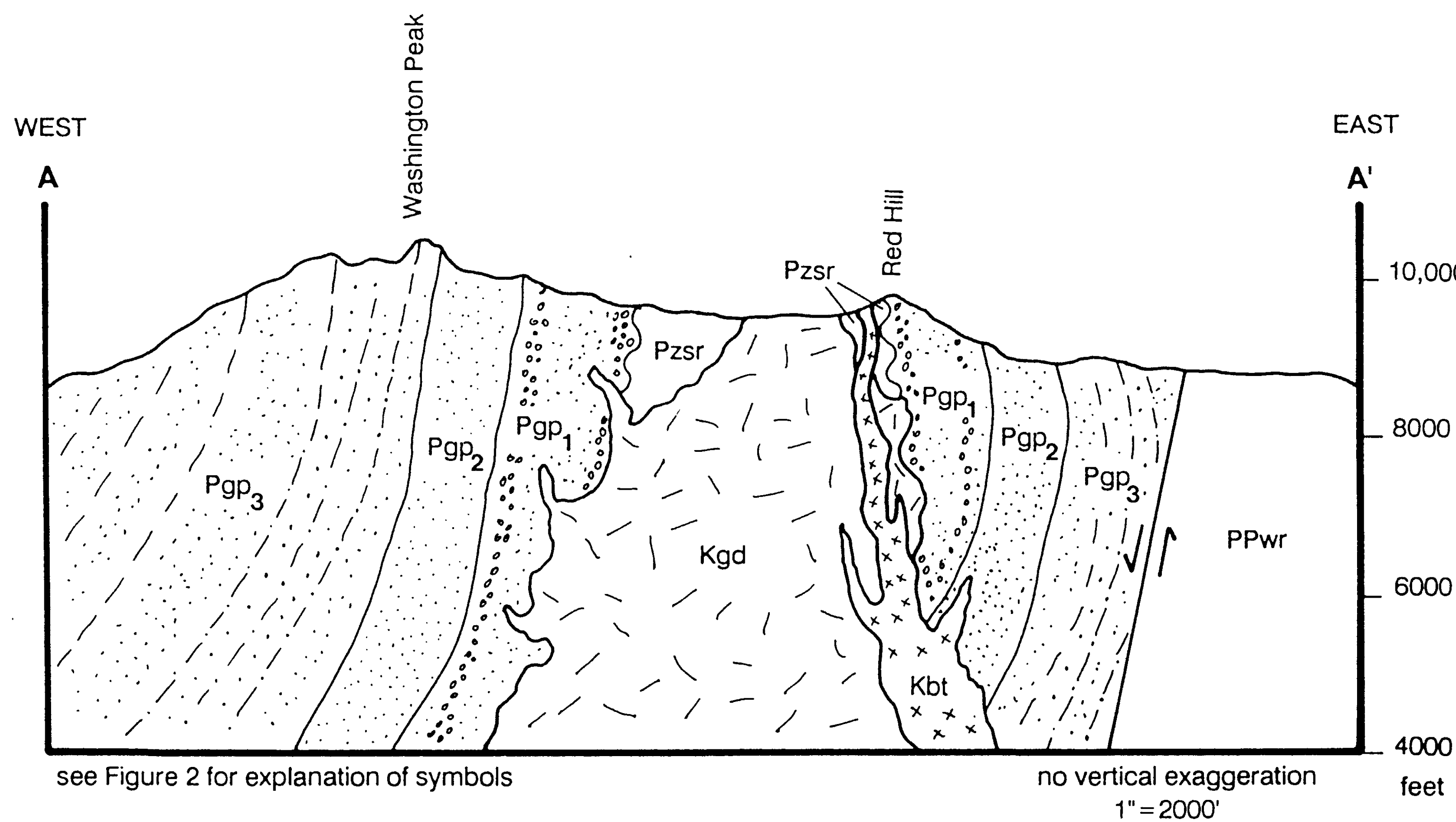

Figure 3 - Geologic Cross-section of the Washington Basin Area

western limb, member 1 of the Grand Prize Formation consists of approximately 125 m of matrix- to clast-supported conglomerate with minor intercalated coarse-grained quartzite overlain by approximately 250 m of thin- to medium-bedded, locally parallel to wavy laminated, fine-to medium-grained quartzite. Overlying this quartzite is approximately 100 m of clast-supported conglomerate, which is overlain by approximately 25 m of blue-gray sandy limestone. In the western limb of the anticline, member 2 of the Grand Prize Formation is in conformable contact with the underlying member; the contact is immediately west of the peak of elevation 10,500 ft on the north side of the basin. Member 2 consists of about 350 m of thick-bedded to massive, fine-grained, calcareous quartzite intercalated with minor thin- to medium-bedded, parallel- to cross-laminated, fine-grained sandstone and dark-gray calcareous siltstone. Member 2 has a gradational contact with overlying member 3; the contact is placed on the eastern end of Washington Peak (fig. 2).

Member 3 of the Grand Prize consists of thin-bedded, fine-grained sandstone and dark siltstone comprising distinctive banded couplets, 15-45 cm, which are well exposed in the cirque headwall. Intercalated with these banded couplets are thick-bedded to massive quartzite beds up to 25 m thick. The thickness of member 3 in the Washington Basin area is uncertain due to structural thickening by a fault in the Champion Lakes cirque (fig. 2). Geographically, the member underlies the majority of the Champion Lakes cirque; the upper contact with member 4 is exposed below the ridge west of Champion Lakes. In this area, member 3 contains a much higher percentage of sandstone and quartzite than previously reported in the Grand Prize Formation (Mahoney and Sengebush, 1988).

In the Washington Basin area, member 4 is exposed in the eastern limb of the anticline, on the east side of Croesus Peak. The first three members of the Grand Prize are sporadically exposed in the eastern limb of the fold, due to obliteration by post-folding intrusions (fig. 2). Member 4 of the Grand Prize Formation consists of thin- to medium-bedded carbonaceous limestone, siltstone, and mudstone, and forms the distinctive dark unit on the east side of Croesus Peak. The fine-grained carbonaceous siltstones and limestones of member 4 contain abundant sedimentary structures, including graded bedding, cross-laminae, parallel laminae, and soft-sediment deformation features.

The Grand Prize Formation has been assigned an Early Permian age, based on conodonts collected from the type section, west of Washington Basin (Hall, 1985). This age determination is poorly constrained, however, and better biostratigraphic control is a primary goal of ongoing investigations.

In Washington Basin, the Grand Prize Formation is locally contact metamorphosed to calc-silicate hornfels, and is commonly bleached and silicified. The Grand Prize Formation hosts mineral deposits in members 2, 3, and 4 in both Washington Basin and in Germania Basin to the south; mineral deposits occur in polymetallic veins in fracture systems and in skarn deposits. The Grand Prize Formation contains numerous mines and prospects throughout the White Cloud Peaks and in the northern Smoky Mountains.

Wood River Formation

The Wood River Formation consists of more than 3000 m of calcareous sandstone and sandy limestones, subdivided into three informal members (Link and others, 1988). The formation is widely exposed south of Washington Basin, where it comprises a series of fault blocks on the eastern end of Washington and Germania Basins (fig. 2). Only the middle member of the Wood River Formation is exposed in the area, most notably on Bible Back Mountain, where it is host to polymetallic veins in shear zones.

In the Washington Basin area, the middle member of the Wood River Formation consists primarily of thick-bedded to massive fine-grained calcareous sandstone, intercalated with minor thin- to medium-bedded, blue-gray sandy micrite and calcareous siltstone. The unit is extensively brecciated, making a thickness determination difficult.

The age of the middle member of the Wood River Formation is well-constrained as Late Pennsylvanian to Early Permian, and it is interpreted as a lateral equivalent of member 2 of the Grand Prize Formation. South of Washington Basin, the middle member of the Wood River Formation is gradationally overlain by member 3 of the Grand Prize Formation, suggesting that the Grand Prize Formation is a northern, more sandy facies of the Wood River Formation (Link and Mahoney, this volume, fig. 1).

The Wood River Formation is in high-angle fault contact with the Grand Prize Formation in the saddle west of Bible Back Mountain (fig. 2). On Bible Back Mountain, the Wood River Formation is metamorphosed to a siliceous quartzite, and is extensively fractured and iron stained (limonite); the formation is cut by numerous north-trending shear zones which host polymetallic lead-silver veins. This fractured and iron-stained unit is in low-angle fault contact with underlying, essentially unfractured and unaltered Wood River Formation; this relationship is well-exposed on the north face of Bible Back Mountain.

STRUCTURE

The dominant structure in the Washington Basin area is a northeast-trending asymmetric east-vergent anticline which has folded the Salmon River assemblage and Grand Prize Formation (fig. 3). This anticline and its associated hinge-parallel fractures have exerted a strong control on subsequent mineralization (fig. 4). The Salmon River assemblage forms the core of the anticline, which plunges at a moderate angle (~35°) into Washington Basin (fig. 2). The west limb of the anticline consists of steeply southwest dipping to near vertical Grand Prize Formation in unconformable contact with the underlying Salmon River assemblage. Competency differences between the two units at the unconformity has resulted in ductile deformation of the upper portions of the Salmon River assemblage and intense brecciation and shearing in the more competent lower member of the Grand Prize Formation. The eastern limb of the anticline consists of near vertical to overturned (steeply southwest dipping) beds of the Grand Prize Formation. The eastern limb of the fold is difficult to recognize due to widespread igneous intrusions, resulting in a lack of continuous exposure. The structural orientation of the limb is easily recognized on the east side of Croesus Peak, where members 3 and 4 of the Grand Prize Formation are overturned, dipping west, and young to the east (fig. 3).

Numerous northeast-trending high-angle faults are exposed throughout the Washington Basin area (fig. 2). The majority of the faults are believed to have minor displacement (tens of meters), and result in structural thickening. Two high-angle faults in the area are believed to have significant displacement (tens to hundreds of meters). The fault in the saddle east of

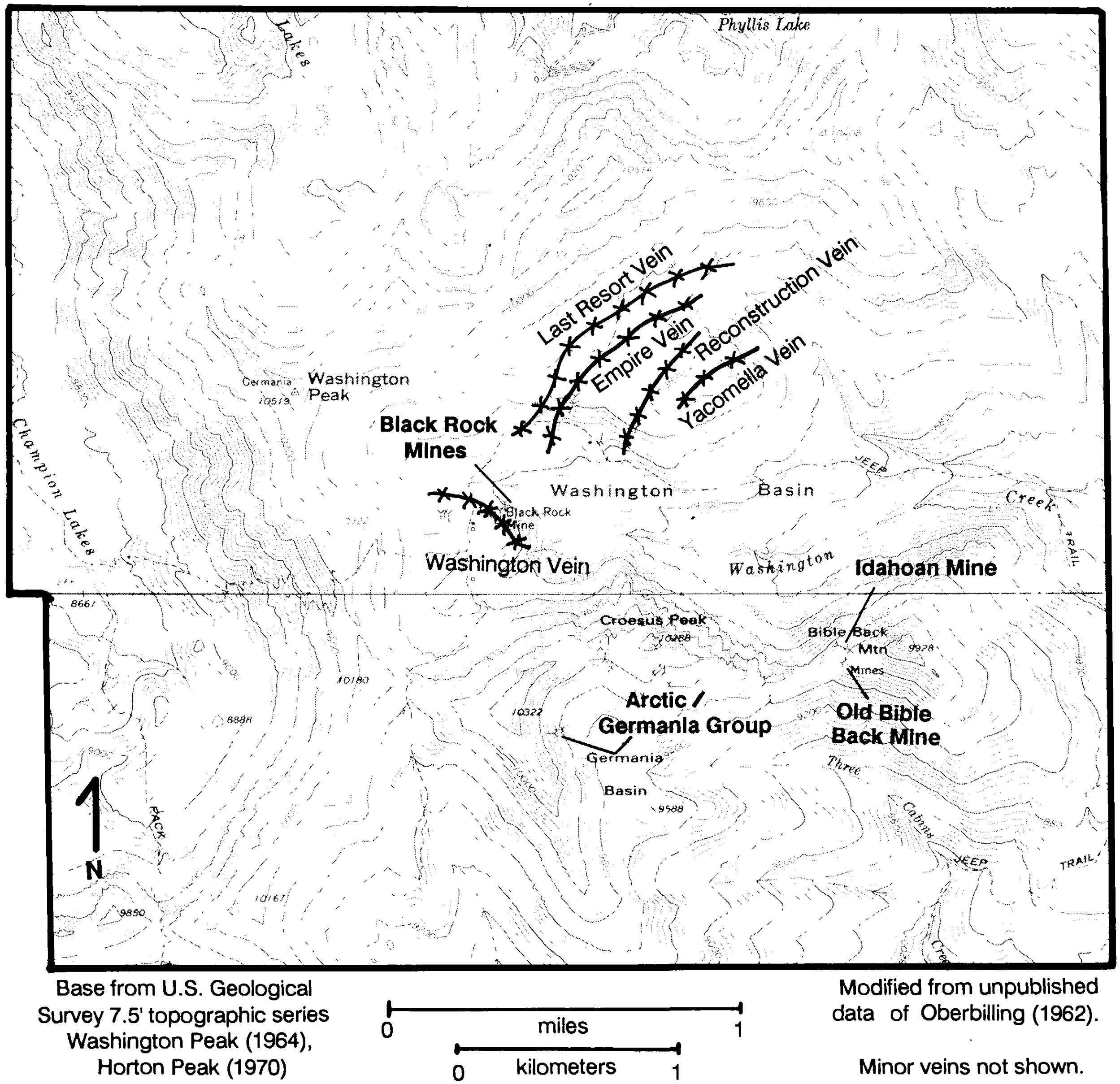

Figure 4 - Mineralized Structures of the Washington Basin Area

Bible Back Mountain dips steeply to the west, and juxtaposes Grand Prize Formation (member 4) to the west with Wood River Formation (middle member) to the east. This fault has been reported to be a large northeast-trending reverse fault, although criteria by which sense of movement was determined is not disclosed (Van Noy and others, 1986). I interpret the fault as a normal fault (down on the west) of significant displacement, based on the juxtaposition of heavily fractured, sheared, and iron stained quartzite of the middle member of the Wood River Formation to the east with the less deformed upper member of the Grand Prize Formation to the west. Several mineralized veins parallel this fault on the north side of Washington Basin, and there is potential for discovery of additional mineralized zones along this fault (Van Noy and others, 1986).

The second fault of significant displacement in the Washington Basin area is located to the southeast of Germania Basin, where a northeast-trending normal fault places the middle member of the Wood River Formation against andesitic lavas of the Challis Volcanic Group (fig. 2). The lack of Paleozoic sedimentary units and the existence of over 500 m of volcanics to the east of this fault, together with the absence of volcanics on the west side of the structure, suggest that the fault has a displacement of at least 1000 m.

A prominent low-angle structure is exposed on the north face of Bible Back Mountain, where massive quartzite and calc-silicate rocks of the middle member of the Wood River Formation structurally overly thin- to thick-bedded quartzite, siltite and limestone of the middle member of the Wood River Formation. The structure strikes to the northeast, and dips about 18° to the southeast (Van Noy and others, 1986). Both sides of the structure are heavily brecciated and fractured, but the upper plate is much more silicified and iron stained than the lower plate. Van Noy and others (1986) describe the low-angle structure as a thrust fault. However, the brittle nature of the fault zone, the structural discordance between the upper and lower plates (up to 70 degrees), and uncertain age relations between the two plates argue against a thrust fault designation, and suggest the fault may be a low-angle normal fault. More work, including sense-of-movement indicators, is needed before this fault can be accurately classified.

IGNEOUS INTRUSIONS

The Washington Basin area has been subjected to at least three separate intrusive episodes. The largest intrusion is a porphyritic biotite granodiorite that is exposed in the southwestern and central portions of Washington Basin, and along the ridge west of Croesus Peak. The granodiorite occurs as large stocks in these areas, but also forms dikes on the north side of Croesus Peak, and in the headwall of the Washington Basin cirque.

The biotite granodiorite is medium- to coarse-grained, has a grussy weathered appearance, and locally contains phenocrysts of alkali feldspar 3-10 cm long. The porphyritic nature of the granodiorite decreases with depth, and is believed to be a metasomatic replacement phenomenon near contact zones. The granodiorite intrudes the Grand Prize Formation, and locally contains roof pendants of the formation. Randomly oriented aplite dikes are common in the granodiorite. The granodiorite is locally iron stained near contact zones, particularly in exposures on the west side of Red Hill (fig. 2). The granodiorite is believed to be Cretaceous in age, based on proximity to the Cretaceous White Cloud Stock, and because it is cut by Tertiary dacite pophyry dikes. The emplacement of the granodiorite may have been structurally controlled, as the intrusion is located in the hinge region of the asymmetric anticline.

The second stage of intrusion is characterized by a fine-grained biotite-rich quartz diorite dike that intrudes the porphyritic granodiorite and the Grand Prize Formation on Red Hill (fig. 2). The dike is approximately 10-15 meters wide on Red Hill, and splits into three separate dikes on the north side of Red Hill. The dike is extremely biotite-rich, with fine-grained biotite comprising over 50 percent of the unit. The dike is presumed to be Cretaceous in age, and is believed to be a late-stage intrusion associated with the porphyritic granodiorite. However, the biotite diorite dike cross-cuts the porphyritic granodiorite, and may be as young as Tertiary in age. The biotite quartz diorite is associated with replacement mineralization in the Grand Prize Formation on the north side of Red Hill, and farther to the east in Pole Creek.

Dacite porphyry dikes intrude both the Grand Prize Formation and the Cretaceous granodiorite in the head of Germania Basin. Emplacement of the dikes was primarily along the contact between the granodiorite and the Paleozoic rocks. The dacite porphyry is believed to be Eocene in age. A mafic dike of uncertain age cuts the quartzite on the west end of Bible Back Mountain, and is spatially related to mineralization; similar lamprophyric dikes elsewhere in the White Clouds are interpreted as Cretaceous in age.

MINERAL DEPOSITS

In the Washington Basin area, mineral deposits are hosted by members 1, 2, and 3 of the Grand Prize Formation, by the middle member of the Wood River Formation, and by the Cretaceous porphyritic granodiorite. Emplacement of mineral deposits was structurally controlled, as shown by mineralization of north-trending shear zones on Bible Back Mountain, and by the presence of the northeast-trending (fold-hinge parallel) mineralized veins in Washington Basin (fig. 4). Alteration of the host rocks is pervasive, including oxidation (i.e. iron staining) in both the igneous and sedimentary hosts, and sericitic alteration in the igneous host. Alteration products include limonite, hematite, goethite, and sericite (Van Noy and others, 1986). Van Noy and others (1986) provide a complete description of mines and prospects in the Washington Basin area, and information from their report is included in the following descriptions.

Mineralization in the Washington Basin area is of three types: (1) skarn-type mineralization associated with intrusion of Cretaceous porphyritic granodiorite; (2) shear zone-hosted polymetallic veins in sedimentary rocks; (3) replacement deposits in quartzitic rocks.

SKARN MINERALIZATION

Mineralization is present along the contact between the Cretaceous porphyritic granodiorite and the intruded Grand Prize Formation. Intrusion of the granodiorite resulted in contact metamorphism of members 2 and 3 of the Grand Prize Formation, and mineralization of both the Paleozoic sedimentary rocks and the granodiorite itself. Examples of this type of mineralization are at the Black Rock mine in the southwest portion of Washington Basin and at the Germania and Arctic Groups in the head of Germania Basin.

At the Black Rock mine, mineralization consists of a northwest-striking, southwest-dipping quartz-pyrite mineralized zone that cuts both granodiorite and the Grand Prize Formation (Van Noy and others, 1986). The mineralized zone is approximately parallel to the intrusive contact. Sulfide minerals are contained in pods up to 0.3 m in diameter and 0.6 m in length in the granodiorite, and in tactite zones in the Paleozoic sedimentary rocks (Van Noy and others, 1986). In the granodiorite, the wall rock is sericitized, and disseminated pyrite is common; the Paleozoic rocks are contact metamorphosed to calc-silicate hornfels. The entire contact zone is heavily iron stained. Sulfide minerals include pyrrhotite, arsenopyrite, pyrite and minor sphalerite. Assays of the vein averaged 0.005 ounces of gold/ton; 1.3 ounces of silver/ton; 1.27 percent lead; and 0.32 percent antimony (Van Noy and others, 1986).

In Germania Basin, the Germania and Arctic Groups explore mineralized veins that are approximately parallel to the contact between the porphyritic granodiorite and members 2 and 3 of the Grand Prize Formation. The vein material is predominately quartz and leached pyrite, and is therefore vuggy and iron stained. Reported sulfide minerals include pyrite, galena, sphalerite, pyrrhotite, and arsenopyrite (Van Noy and others, 1986). Assay values from the vein quartz contain an average of 0.06 ounces of gold/ton; 8.2 ounces of silver/ton, and 6.14 percent lead (Van Noy and others, 1986). Numerous Tertiary dacite porphyry dikes cut the mineralized area; these dikes are iron stained, and the possibility of associated mineralization cannot be discounted.

SHEAR ZONE-HOSTED POLYMETALLIC VEINS

There are two varieties of shear zone or fracture-hosted polymetallic veins in the Washington Basin area: (1) northeast-trending quartz-pyrrhotite-pyrite veins in Washington Basin; and (2) north-trending shear zones containing finely disseminated, oxidized sulfide minerals on Bible Back Mountain. In addition, a thick brecciation zone associated with the low-angle structure on Bible Back Mountain contains sulfide minerals throughout its extent (Van Noy and others, 1986).

The most prominent mineralized structures in the Washington Basin area are the northeast-trending vein structures on the floor of Washington Basin. These shear zones and fracture systems are parallel to the fold axis of the asymmetric east vergent anticline that dominates the structure of the area, and are probably the result of tensional stresses in the axis of the fold (figs. 2, 4). The vein system is easily recognized by heavily oxidized sulfide minerals near the surface. The veins consist primarily of quartz and pyrrhotite, with lesser amounts of pyrite, sphalerite, and galena. Small amounts of scheelite, telluride minerals, and jamesonite are also reported (Van Noy and others, 1986). Samples assayed from the Empire Vein, in the center of Washington Basin, averaged 0.01 ounces of gold/ton, 0.59 ounces of silver/ton; 0.28 percent lead, 0.40 percent arsenic, and less than 0.01 percent WO_3 (Van Noy and others, 1986). Minor amounts of disseminated gold have been reported from wall rock near the vein systems. Both the Grand Prize Formation and the porphyritic granodiorite are hydrothermally altered and mineralized with finely disseminated sulfide minerals along the fracture system, suggesting post-intrusion mineralization, probably the result of late-stage hydrothermal circulation of metal-bearing fluids associated with the Cretaceous pluton.

The second type of polymetallic vein system consists of north-trending shear zones cutting the heavily brecciated middle member of the Wood River Formation on Bible Back Mountain. The shear zones trend north, are nearly vertical, and are generally 0.5 to 1.5 m in width. Visible sulfide minerals are rarely observed, but the entire mountain is pervasively iron stained (limonite), and heavily oxidized gouge is common in the shear zones. Both quartz and calcite stringers are present. Samples of shear zone analyzed by the U.S. Bureau of Mines averaged 0.08 ounces of gold/ton, 13.7 ounces of silver/ton, 21.0 percent lead, and 0.4 percent antimony (Van Noy and others, 1986). Other assays revealed minor amounts of copper and zinc.

The low-angle structure exposed on the north face of Bible Back Mountain is mineralized. It yielded much of the production from the Washington Basin area (Umpleby, 1915; Van Noy and others, 1986). The brecciated fault zone varies from 1-1.5 m in thickness, strikes to the northeast, and dips about 18° to the southeast (Van Noy and others, 1986). The fault zone is highly oxidized, and iron-staining is common; visible sulfide minerals are rare. Van Noy and others (1986) report small, irregular replacement-type ore shoots within the fault zone that consist of small amounts of pyrite, galena, and jamesonite in a quartz and calcite gangue. Samples from the mineralized zone collected by the U.S. Bureau of Mines averaged 0.06 ounces gold/ton, 12.6 ounces of silver/ton, 4.27 percent lead, 1.02 percent zinc, 0.58 percent copper, and 0.45 percent antimony. The U.S. Bureau of Mines estimates that approximately 8,600 tons of paramarginal resources exist within the old workings on Bible Back Mountain, and that an additional 30,000 tons of paramarginal resources may exist in unexplored portions of the fault zone.

REPLACEMENT-TYPE DEPOSITS

A small replacement-type deposit is exposed in a bulldozer cut on the north side of Red Hill. The deposit consists of nearly complete replacement of massive quartzite, probably from the Grand Prize Formation, with massive sulfide minerals. The ore zone is highly oxidized and iron stained, and the deposit has a distinct odor of sulfur. Within the ore zone, massive sulfide material, including galena, has completely replaced the interstitial material between the fine to medium sand-sized quartz grains comprising the quartzite. Vein mineralization is absent, although small stringers of quartz are associated with the deposit. The replacement deposit apparently grades into unaltered sedimentary rock, although exposure in the area is poor. The bulldozer cut is exploratory, no evidence of production exists. The replacement deposit is located at the north end of a biotite quartz diorite dike associated with the main quartz diorite on Red Hill, and is believed to be genetically related to that intrusion.

GENESIS/ORE CONTROLS

The primary controls on mineral deposits in the Washington Basin area are steep or shallow shears or fractures associated with the anticline. The most prominent deposits in the area are mineralized veins that parallel fractures along the axial plane of the asymmetric east-vergent anticline. The anticline formed in response to northwest- to southeast-directed compressive stresses during the Sevier orogeny; northeast-trending fractures formed parallel to the fold hinge. The anticline apparently controlled the emplacement of the Cretaceous porphyritic granodiorite, and hydrothermal systems driven by this intrusion remobilized metals from the Paleozoic sedimentary rocks, and deposited them within northeast-trending fractures. The intrusive event also resulted in skarn mineralization in the calcareous Paleozoic country rock. The most probable source of the metals is the Paleozoic Salmon River assemblage, which forms the core of the anticline, and is exposed at, or occurs at depth beneath each of the mineralized areas.

Post-intrusive fault activity is probable, as shown by shearing of the granodiorite and possible local shearing of the mineral deposits, with the resultant development of abundant sulfide-rich fault gouge. Renewed faulting led to the development of the north-trending shear zones on Bible Back Mountain, and to the development of the large north- to northeast-trending normal faults found in the southeastern portion of the area. This renewed activity may also have controlled the emplacement of the biotite quartz diorite on Red Hill. The amount of mineralization associated with this second stage of activity is uncertain, but is believed to be of a smaller magnitude than the initial stage.

Future exploration activities should concentrate on investigating the prominent northeast-trending vein system found in Washington Basin. If the veins are indeed localized along fold hinge fractures, the likelihood that veins continue at deeper levels to the south under Germania Basin, is quite high. The potential also exists for additional skarn mineralization along the contact between the porphyritic granodiorite and the Paleozoic sedimentary rocks, particularly in Germania Basin. The deposits on Bible Back Mountain are believed to be of insufficient grades to warrant further exploration.

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GEOLOGY AND MINERAL DEPOSITS OF THE VIENNA MINING DISTRICT, BLAINE AND CAMAS COUNTIES, IDAHO

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INTRODUCTION

The Vienna Mining District of south-central Idaho is located at the southern end of the Stanley Basin, north of Ketchum, Idaho (fig. 1). The district is situated in the northern Smoky Mountains, directly south of the Sawtooth Mountains, in Blaine and Camas Counties. The majority of mining activity has been located in the headwaters of Smiley and Beaver Creeks, although a number of mines and prospects are located in the surrounding drainages. The most successful workings are located in deep glacial cirques, suggesting that the level of erosion may be an important factor in the discovery of economic deposits.

HISTORICAL PERSPECTIVE

The Vienna Mining District was discovered in 1879, and the majority of mining activity occurred from 1880-1888. Approximately 1 million dollars (1915 prices) of silver ore was removed from the district during this initial period of activity (Umpleby, 1915). Significant production has not occurred since this initial period. Exploration has been intermittent since the late 1800's, with a brief increase in activity in the 1960's, when 79 claims were filed in the district (Van Noy and others, 1986). A small amount of production occurred in the early 1980's, when a number of the existing mine dumps were processed by a local mill. Exploration, including a drilling program and geophysical analysis, is currently being conducted by Rothschild Mining Company at the Webfoot property in the headwaters of Smiley Creek; production is anticipated, but is contingent on silver prices.

GEOLOGIC SETTING

Lithologic Units

The Vienna District is situated on the southeastern edge of the Atlanta Lobe of the Cretaceous Idaho batholith. Biotite granodiorite of the Idaho batholith underlies virtually the entire district. The mineral deposits in the district, with the exception of a small skarn deposit west of Frenchman Creek, are hosted by the biotite granodiorite.

The biotite granodiorite is medium to coarse grained, and is primarily composed of quartz, plagioclase, microcline, and biotite. The unit is locally porphyritic, with alkali feldspar phenocrysts up to 8 cm in length. The porphyritic nature of the biotite granodiorite is believed to be the result of potassium metasomatism. Thin (2-20 cm), randomly oriented aplite and pegmatite dikes locally cut the granodiorite. The origin of the aplites and pegmatites is uncertain, they may be a late stage differentiate of the Cretaceous batholith, or may be related to the Tertiary pink granite of the

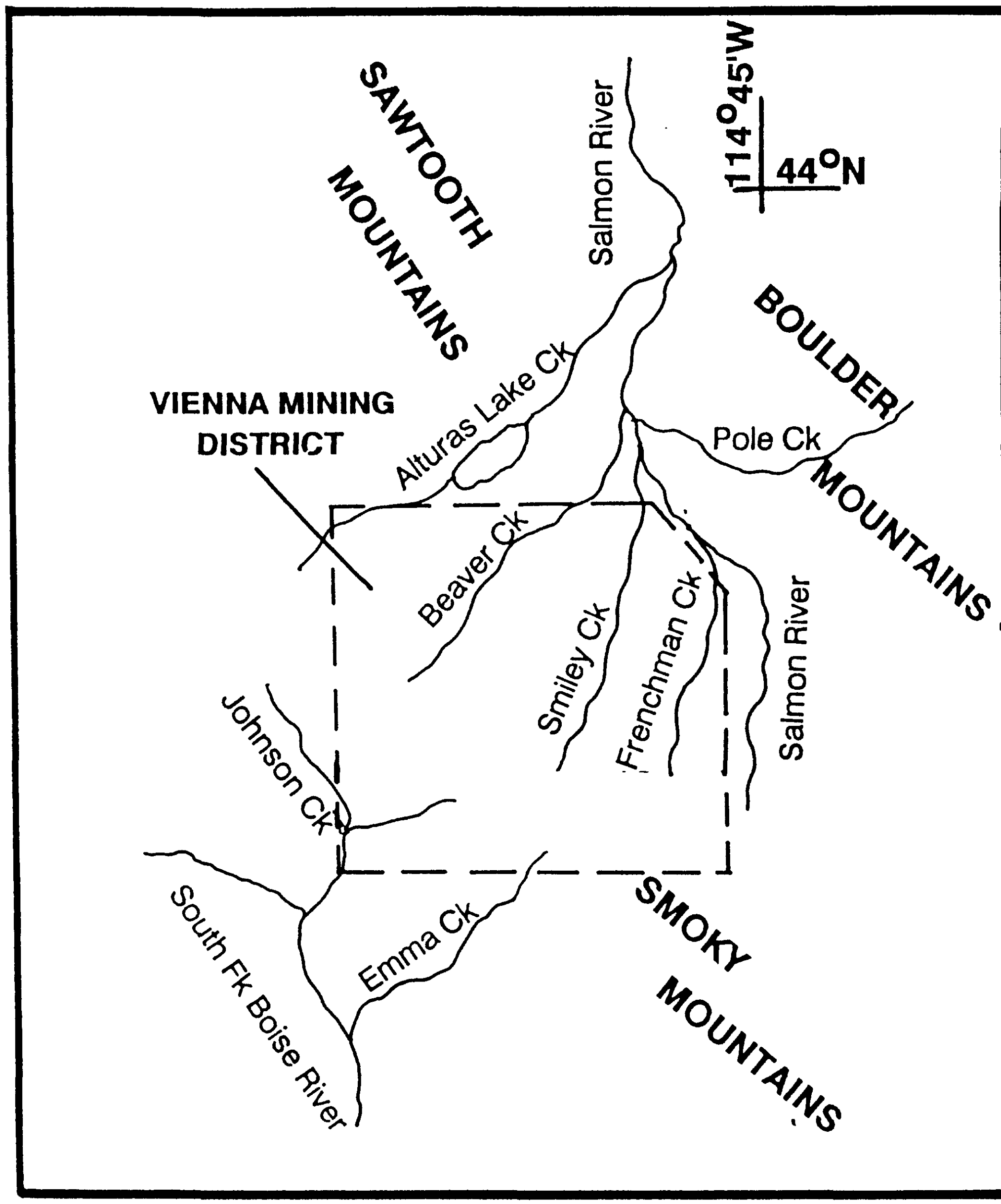

Figure 1 Geographic Location Map of the Vienna Mining District

Sawtooth batholith, exposed 8 km to the north, or to a smaller Tertiary intrusion exposed 5 km to the south.

The biotite granodiorite contains a prominent joint system, which is well exposed in the headwaters of Beaver Creek. The dominant joint set strikes N35-50°W, and dips steeply to the north. The geometry of the joint system varies markedly, from planar, parallel joint sets to curvilinear joint patterns. Previous workers have suggested that mineral veins in the Vienna district are confined to the joint sets, but it is now recognized that the vein system is subparallel to the joint sets (Ballard, 1922). Joint sets are locally mineralized adjacent to mineralized shear zones.

The biotite granodiorite intrudes calcareous sandstone, siltstone, and sandy limestone of the Lower Permian Grand Prize Formation in the eastern half of the district. The Grand Prize Formation has been contact metamorphosed to calc-silicate hornfels, and locally contains abundant wollastonite, diopside, and tremolite. The metamorphosed Grand Prize Formation hosts tungsten skarn deposits on the northern edge of the district, east of Beaver Creek (fig. 1). Small roof pendants of sedimentary rock are exposed along ridges in the southeastern portion of the district.

Andesitic to dacitic lavas and associated volcaniclastic sediments of the Eocene Challis Volcanic Group unconformably overlie both the Cretaceous biotite granodiorite and the Permian Grand Prize Formation in the eastern portion of the district. The volcanic rocks comprise a thick package (over 500 m) of andesitic to dacitic lava flows, tuff breccias, and volcanogenic sandstone and siltstone. The volcanic rocks are locally in fault contact with the granodiorite, where the rocks are preserved in downthrown fault blocks adjacent to north- and northeast-trending normal faults.

Northeast-trending Tertiary dacite porphyry dikes cut the biotite granodiorite throughout the Vienna district. The consistent northeast trend suggests that dike emplacement was structurally controlled. Chilled contacts and minor alteration zones characterize the intrusive contact between the dike rocks and the granodiorite, and there is no evidence of associated mineralization.

Thin (0.2-0.4 m) lamprophyric dikes of uncertain age locally cut the biotite granodiorite. The lamprophyric dikes apparently cut mineralized veins, although the relation between mineralized structures and the dikes is commonly ambiguous.

Structure

The dominant structures in the Vienna District are northeast-trending (N50-75°E) normal faults of significant displacements ranging from 10's to 100's of meters. The normal faults cut the Paleozoic sedimentary rocks, the Cretaceous biotite granodiorite, and the Eocene Challis Volcanic Group. The faults apparently controlled the emplacement of the dacite porphyry dikes, and are probably coeval or slightly older than the Tertiary dikes. There is a strong correlation between the trend of the faults and the location of the mines and prospects in the Vienna District, suggesting the faults may be genetically related to economic mineralization.

The most important structural control for mineralization in the Vienna District is the abundant shear zones that cut the granodiorite. The shear zones are characterized by two distinct trends: northwest-southeast and east-west. The shear zones are near-vertical or dip steeply to the north. On the surface, the shear zones are iron-stained (limonite) and locally brecciated; below ground brecciation is common within the shear zones and intense chloritic and sericitic alteration extends 1-3 m away from the shear zones. Evidence of shearing and brecciation is more pronounced along the east-west-trending shear zones (Ballard, 1922). Both sets of shear zones are mineralized, although the east-west set is normally of higher grade. Ore shoots occur locally at shear zone intersections.

A prominent north-trending normal fault juxtaposes Cretaceous granodiorite and the Challis Volcanic Group west of Smiley Creek. The fault is nearly vertical, and has an offset of at least 500 m, based on the thickness of the volcanic rocks on the downthrown eastern block of the fault. The fault apparently cuts shear zones in the granodiorite, and is believed to be associated with Basin and Range-age extension.

MINERAL DEPOSITS

Two distinct types of mineralized veins occur in the Vienna District: shear-zone-hosted quartz-silver sulfide ribbon veins and shear-zone-hosted quartz-sericite-pyrite-galena veins. Mineralogical and textural evidence suggests that the two types of veins are of different ages, and that the ribbon veins represent the earlier mineralizing event.

Quartz-Silver Sulfide Ribbon Veins

The quartz-silver sulfide ribbon veins consist of roughly defined bands of massive white quartz and aphanitic dark-gray silver sulfide minerals, primarily pyrargyrite and proustite. Tetrahedrite, ruby silver, argentian stibnite, and argentite ("black metal") are locally abundant (Ballard, 1922; Shannon, 1971). The arsenic and antimony content of the ores varies considerably across the district (Shannon, 1971). Pyrite is noticeably absent within the competent ribbon veins.

The ribbon veins pinch and swell noticeably along strike, and occur as tabular lenses. Vein width varies from 0.1 to about 2 m, with alteration zones extending 1-3 m from the veins. The ribbon vein system is subparallel to shear zones, and is most well developed in the east-west shear system. Alteration adjacent to the veins consists of intense sericitization adjacent to the veins, and chloritization near the outer edges of the altered zones.

The ribbon veins represent a hypogene mineralization event characterized by repeated silicification and deposition of aphanitic sulfide minerals. However, sulfide mineral paragenesis and primary alteration features associated with the ribbon veins are difficult to determine due to the strong overprint of the second mineralizing event. The ribbon veins have been intensely sheared and brecciated by reactivation of the shear zones after the first mineralizing event.

Quartz-Sericite-Pyrite-Galena Veins

The second vein system in the Vienna District is a shear-zone-hosted quartz-sericite-pyrite-galena vein system that consists of arsenopyrite, galena, sphalerite, and minor stibnite in a quartz, sericite, siderite, and pyrite gangue. The arsenopyrite reportedly contains minor amounts of gold (Shannon, 1971). Pyrite, and to a lesser extent, arsenopyrite, are disseminated throughout the shear zone and surrounding wall rock alteration zone. The sulfide minerals occur as irregular patches within the intensely sheared gouge zone, and the gangue material primarily consists of irregular, elongated grains of quartz together with abundant sericite. Locally, sulfide minerals occur in massive lenses up to 3 m in length.

The quartz-sericite-pyrite-galena vein system occurs within the same shear zones as the ribbon vein system, and significantly overprints the earlier event. The ribbon veins are strongly sheared and brecciated, although small unsheared segments of the ribbon veins do occur as inclusions within the intensely sheared and mineralized gouge zone. The gouge zone contains abundant sericite and minor chlorite, and the wall rock is chloritized and sericitized within 1-3 m of the shear zone. Quartz-sericite-pyrite-galena mineralization occurs along both shear zone systems in the Vienna District, and ore shoots of this assemblage locally occur at shear zone intersections. Minor mineralization also occurs along joint planes near joint/shear zone intersections.

Mineral Deposit Age and Genesis

The quartz-sericite-pyrite-galena vein system is clearly younger than the ribbon vein system as shown by: (1) brecciation of the ribbon ore within the lead-silver ore system; (2) pyrite and arsenopyrite are disseminated throughout both the shear zone and the alteration zone, yet notably absent in competent (i.e. impermeable) ribbon ore; (3) replacement of the primary (ribbon) quartz by the "second crop" of minerals, including sericite, siderite, and sulfide minerals (Ballard, 1922). Mineralization at shear zone intersections (ore shoots), at shear zone/joint plane intersections, and along joint planes suggest that the second mineralizing event was emplaced by a larger hydrothermal system than that which deposited the ribbon veins, which are confined to the center of the shear zones.

Our current working hypothesis involves two distinct periods of mineralization. We believe the first period of mineralization emplaced the ribbon vein system. The ribbon vein system is mesothermal in origin, and involved the multiple injections of siliceous, metal-bearing fluids into east- and northwest-trending shear zones. These shear zones are subparallel to the prominent joint system, are younger than the joint system, and may be the result of Late Cretaceous isostatic readjustment of the granodiorite intrusion. The metal-bearing fluids may be a late-stage hydrothermal system associated with the Cretaceous biotite granodiorite. However, the age of the ribbon veins is not well constrained. The ribbon veins obviously post-date the granodiorite, and are cut by lamprophyre dikes in the Webfoot mine (Ross, 1927). Lamprophyre dikes are believed to be Late Cretaceous in age elsewhere in the region (R.G. Worl, personal commun., 1989). We propose that the ribbon ore is probably Late Cretaceous in age, although it may be as young as Early Tertiary.

The second mineralization event emplaced the quartz-siderite-pyrite-galena vein system. We believe this system is epithermal in origin, and is Late Tertiary in age. The Vienna District is cut by numerous northeast-trending normal faults associated with the Tertiary trans-Challis fault system. We believe that these faults caused reactivation of the existing east- and northwest-trending shear zones, which resulted in intense shearing and brecciation of the ribbon ore. After faulting, both the dacite porphyry dikes and nearby pink granite were intruded. Metal-bearing hydrothermal systems associated with these intrusions, most likely the pink granite intrusions, emplaced the quartz-siderite-pyrite-galena veins along the reactivated shear zones.

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STRATIGRAPHY, PETROLOGY, AND STRUCTURE OF EOCENE VOLCANIC ROCKS IN THE NAVARRE CREEK AREA: SOUTHEASTERN CHALLIS VOLCANIC FIELD, IDAHO

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Detailed mapping of the southeastern Challis volcanic field in the Navarre Creek area has resulted in the compilation of stratigraphic, petrologic, and structural data of rocks of Eocene age in this region.

Voluminous lava flows of intermediate composition, silicic ash-flow tuffs and extensive silicic intrusive rocks are characteristic of the Challis Volcanics in the area. Thin sheets (0-40 m) of volcaniclastic sediments are intercalated throughout the stratigraphic sequence. The stratigraphic units in the Navarre Creek area, from oldest to youngest, are: (1) lower latite lavas: composed of 0-300 m of latite, trachyte and andesite lava flows; (2) tuff of Cliff Creek: 0-150 m of latitic ash-flow tuff; (3) upper andesite lavas: 0-100 m of andesite lava flows; (4) tuff of Cherry Creek: 0-250 m of trachytic ash-flow tuffs; (5) upper latite lavas: composed of 0-200 m of latite, andesite, dacite, and trachyte lava flows; (6) porphyritic rhyolite intrusion; (7) upper dacite lavas: 0-200 m of dacite lava flows.

Phenocryst assemblages in the intermediate rocks include plagioclase, sanidine, biotite, hornblende, and quartz. Latite and trachyte contain primarily plagioclase and biotite phenocrysts with pyroxene present in the latite.

Eighteen bulk rock analyses for major- and trace-elements range from 58 to 68 wt. percent silica and include high-K andesite and dacite, latite, trachyte, and rhyolite, according to the classification of Peccarillo and Taylor (1976). All rocks have high concentrations of the large-ion lithophile elements Ba, Sr, and Rb, and low abundances of P, Y, Ni, and Cr. Hence, all rocks are thought to be differentiates of more primitive parental magma, or partial melts of the continental crust.

Normal faults in this area trend predominately north-northwest. Northeast-trending normal faults do occur, but less frequently than the north-northwest faults. These structures commonly pre-date the rhyolite intrusives and upper dacite lavas, which date movement along these structures as Eocene.

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GEOLOGY AND ORE DEPOSITS OF THE ROOK'S CREEK STOCK, BLAINE COUNTY, IDAHO

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The Rook's Creek stock, located 16 km west of Ketchum, Idaho, is an approximately 16 km^2 stock of Cretaceous age (92.2 Ma). The stock (fig. 1), which is interpreted as one of several high-K outliers of the Idaho batholith, is composed of two major phases. The earliest main phase is a medium-grained, equigranular biotite granodiorite of Cretaceous age. Intruded into the center of the granodiorite are two lobes of megaporphyritic granite to granodiorite of probable Cretaceous age. The stock is also intruded by numerous northeast- to east-trending dactic dikes of probable Eocene age and a small dacitic stock of probable Eocene age.

There are two distinct types and possible ages of mineralization associated with the Rook's Creek Stock. These are: (1) contact-related replacement deposits of possible Cretaceous age, and (2) intrusive-hosted, vein mineralization of possible Tertiary age.

Replacement-type mineralization occurs at three locations in the Rook's Creek area. Replacement deposits are located where the stock is in contact with calcareous siltstones and sandstones and silty limestones of the Pennsylvanian-Permian Wood River Formation. Three of the major historic mines in the study area (Ontario, Lucky Boy, and Blue Kitten) are replacement deposits, although they differ greatly in character. The Blue Kitten mine is located at a contact of biotite granodiorite and micritic limestone. The predominant ore mineral at the Blue Kitten is galena with quartz gangue in a host of recrystallized and silicified limestone. The Lucky Boy Mine is located at a contact of biotite granodiorite and silty limestone. Ore minerals at the Lucky Boy Mine include predominantly arsenopyrite with some sphalerite and galena. Gangue is mainly quartz and pyrite. Very little alteration of host rocks was observed at the Lucky Boy Mine, but iron-oxide staining and gossan are common. The Ontario Mine area is situated along the contact of biotite granodiorite and micritic limestone to calcareous sandstone of the Pennsylvanian-Permian Wood River Formation. In addition to replacement deposits, further mineralization may have been the result of emplacement of numerous, small felsic dikes along the contact. Alteration in the Ontario Mine area is predominantly sericitic, with a quartz/sericite/pyrite mineral assemblage. Ore mineralogy varies from pyrite veins in sandstone to polymetallic veins, characterized by an arsenopyrite/galena/sphalerite mineral assemblage, hosted in granodiorite.

A large altered and mineralized area within the Rook's Creek stock is characterized by a predominantly sphalerite and galena ore mineral assemblage with a quartz-pyrite-sericite alteration envelope extending several meters from the veins. A N40°E trend of the veins in this area suggest a possible association with the northeast-trending trans-Challis fault system of Eocene age.

Field observations have shown that two separate granodioritic to granitic phases were emplaced during the Cretaceous. Emplacement of these plutons was probably responsible for the formation of replacement deposits in the Rook's Creek area. Shearing during the Eocene Challis volcanic event was probably responsible for the emplacement of numerous dacitic dikes and the formation of vein mineralization along northeast-trending shear zones. Additional

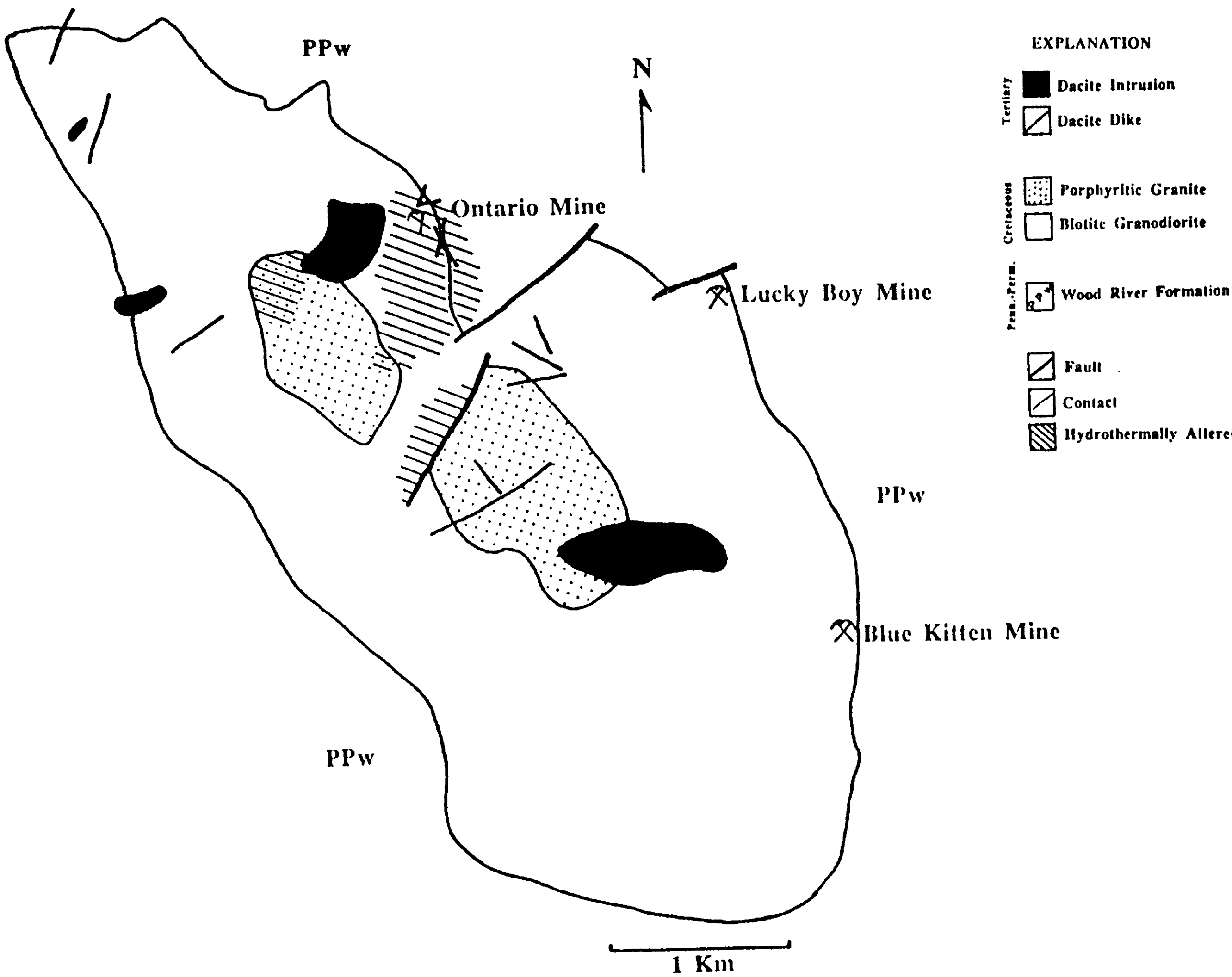

Figure 1. Generalized geologic map of the Rook's Creek Stock.

laboratory studies should provide information that will allow the construction of petrogenetic and ore deposition models to resolve the geologic history of the area.

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PB-ZN-AG-SB-SN-AU EPITHERMAL VEIN SYSTEMS IN BOULDER BASIN, BLAINE AND CUSTER COUNTIES, IDAHO

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In the Boulder Basin, south-central Idaho, a Tertiary hypabyssal-volcanic complex intruded metasedimentary and sedimentary rocks of the Devonian Milligen Formation and Pennsylvanian to Permian Wood River Formation. Principal mine workings are in the lower plate of the Boulder Basin thrust and are confined to northeast-trending, southeast-dipping Tertiary faults and brittle shear zones. These structures cut massive quartzite of unit 6 in the Wood River Formation, as well as andesite and dacite porphyry intrusions (unit numbers are those of Hall and others, 1974). Rhyolite porphyry dikes crosscut and intrude into these mineralized structures. Because all three porphyries are Eocene age, the faulting and mineralization in the principal mine workings is wholly confined to the Eocene. The ore deposits consist of lead-zinc-silver-antimony-tin-gold epithermal vein systems that are hosted primarily within the massive quartzite of the Wood River Formation with negligible disseminations in the adjacent andesite and dacite porphyry intrusions.

Thirty-eight highly mineralized samples were collected from mine dumps and underground workings and were analyzed for thirty-one elements by a six-step semiquantitative spectrographic analysis. Representative ore samples were examined by reflected light microscopy and by electron microprobe. The primary ore mineral is argentiferous galena with submicroscopic inclusions of freibergite(?) and pyrargyrite(?). These minerals are associated with varying amounts of pyrite, sphalerite, chalcopyrite, bornite, hematite, limonite, malachite, covellite, and cerussite(?). Thin sections of wall rock adjacent to mineralized structures, and polished thin sections of ore with interstitial gangue minerals of quartz and calcite indicate that sericitic alteration and silicification are closely associated with mineralization.

No discrete tin sulfide minerals were identified; however, thirteen samples appear to have anomalous concentrations of tin based on the geochemical analyses. The analyses from the electron microprobe indicate that tin is within the crystal structure of galena and sphalerite. Gold was not detected in this study and this may be a consequence of the low detection limitations for the six-step semiquantitative spectrographic analysis that was employed. Umpleby (1915) reported that shipments of ore from the Boulder Basin around 1915 contained 0.32 to 0.62 ounces of gold per ton.

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REGIONAL STRATIGRAPHY, PHYSICAL VOLCANOLOGY, AND GEOCHEMISTRY OF THE SOUTHEASTERN CHALLIS VOLCANIC FIELD

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INTRODUCTION AND REGIONAL SETTING

This abstract presents results of geologic field studies for the southeastern Challis Volcanic field in the western half of the Idaho Falls 1°x2° quadrangle (fig. 1).

The Challis Volcanic field developed during a regionally extensive Eocene (51-44 Ma) volcanic event in the northwestern United States. The eruptive history and compositional range of the entire Challis field is tripartite with (1) early dominantly andesitic to dacitic effusive volcanism, followed by (2) explosive dacitic to rhyolitic ash-flow tuff eruptions and formation of cauldron complexes, and culminating in (3) intrusion of late stage dacitic to rhyolitic domes and plugs. Volcanic and intrusive rocks are compositionally diverse and generally can be characterized as high-K calc-alkaline. The suite includes high-K basalt, andesite, dacite, and rhyolite as well as shoshonite, absarokite, latite, and trachyte.

STRATIGRAPHY

In comparison to the rest of the Challis Volcanic field, the southeastern part shows a similar compositional and volcanic evolution, but has a smaller volume of volcanic deposits that were erupted over a shorter time interval (49-47 Ma). The volcanic and intrusive units in chronological order, include (fig. 2): (1) as much as 400 m of **andesitic lavas and tuff breccias (Tal)** erupted in the Sheep Mountain and Antelope Creek areas [dated at 49.4±0.7 Ma (K-Ar, biotite)]; as much as 1000 m of andesitic lava in the North Fork of the Big Lost River drainage; (2) as much as 550 m of **latitic tuff of Antelope Creek (Tac)** in the Sheep Mountain and Antelope Creek areas; (3) as much as 470 m of **dacitic dome/flow complexes and tuff breccias (Tdl)** in the Sheep Mountain and Antelope Creek areas [dated at 48.93±.26 Ma ($^{40}Ar/^{39}Ar$, biotite)]; (4) as much as 700 m of **latitic, andesitic, and dacitic lavas and tuff breccias (Tll)** from Porphyry Peak east to Navarre Creek [dated at 48.8±.17 Ma ($^{40}Ar/^{39}Ar$, biotite)]; (5) as much as 250 m of **latitic tuff of Cliff Creek (Tcl)** in the Alder and Navarre Creek areas; (6) as much as 100 m of **andesitic lava (Tau)** in the Navarre Creek and Sheep Mountain areas; (7) as much as 200 m of **rhyolitic flow/dome complexes (Trl)** in the Lehman Basin and Porphyry Peak areas; (8) varied thicknesses of vent facies (1000 m) and outflow sheet (200 m) of **trachytic tuff of Cherry Creek (Tch)** from Porphyry Peak to Antelope Creek [dated at 48.87±.51 Ma ($^{40}Ar/^{39}Ar$, sanidine); 49.03±.88 ($^{40}Ar/^{39}Ar$, sanidine)] and intrusion of the consanguineous **Boone Creek stock and similar aphyric rhyolitic domes (Tri)** [dated at 48.86±.13 Ma ($^{40}Ar/^{39}Ar$, biotite)]; (9) as much as 100 m of **latitic to dacitic tuff of Little Lake Creek (Tlc)**; (10) as much as 300 m of dominantly **latitic lavas and tuff breccias (Tlu)** in the Lehman Basin and Navarre Creek areas; (11) porphyritic **rhyolitic intrusions (Trpi)** in the Lehman Basin and Navarre Creek areas; (12) as much as 300 m of regionally extensive **dacitic flow/dome complexes and tuff breccias (Tdu)**; and

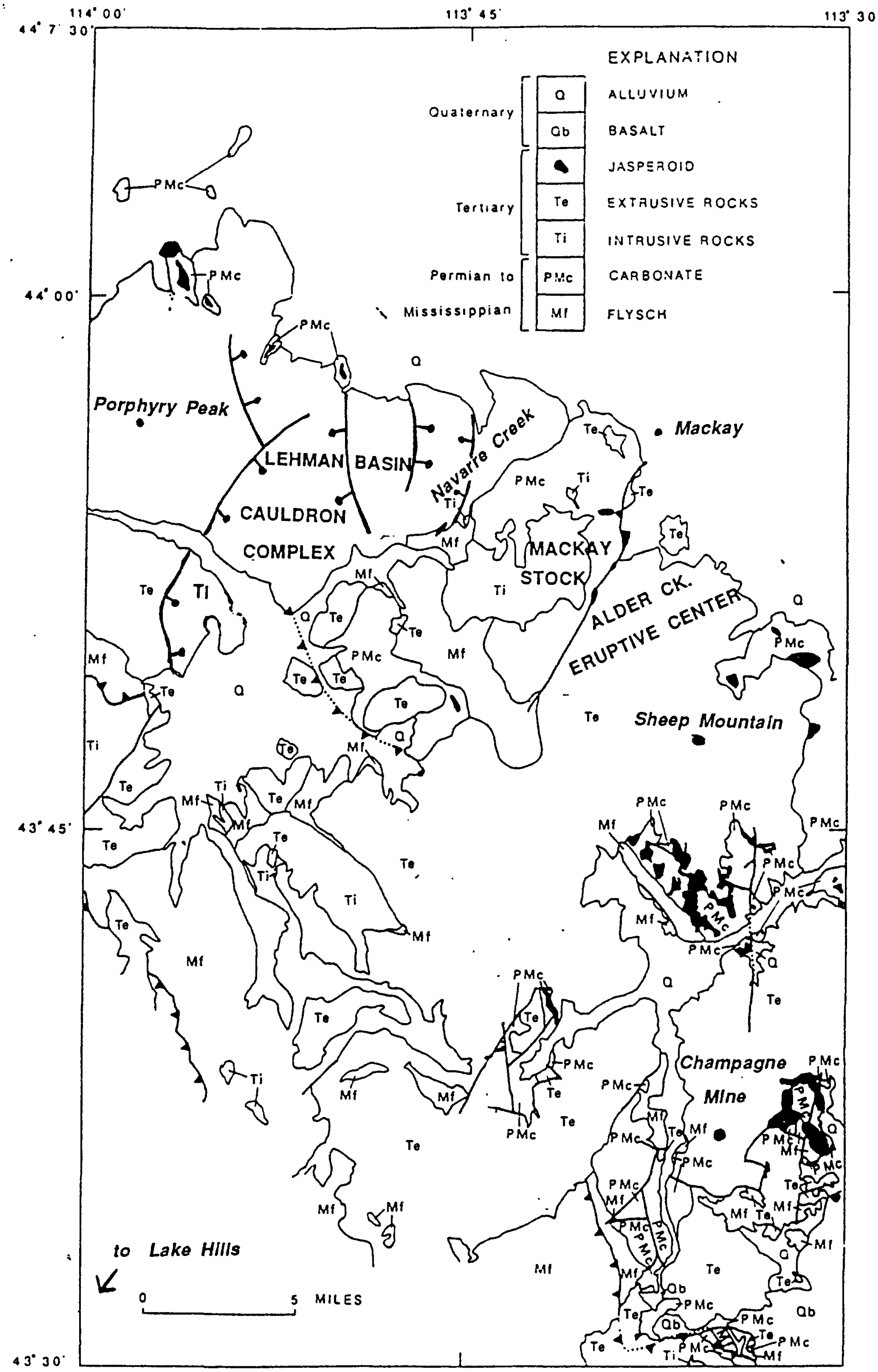

Figure 1. Generalized geologic map of the western Idaho Falls 1°x2° quadrangle.

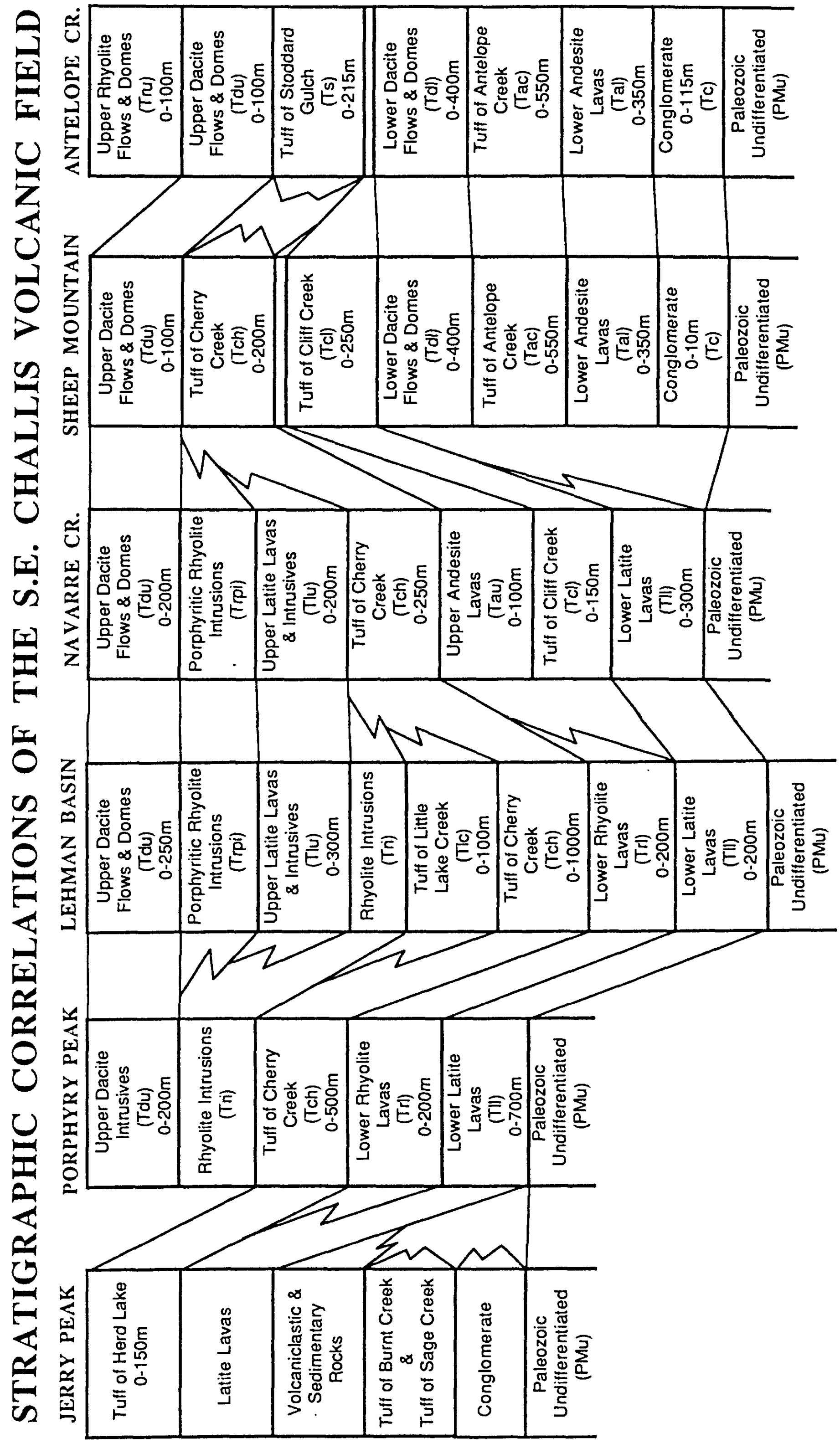

Figure 2. General stratigraphic correlations of southeast Challis Volcanic field. Jerry Peak section is from McIntyre (1982). Antelope Creek section is from Skipp (1988, 1989).

(13) rhyolite flow/dome complexes and dike swarms (Tru) in the Antelope Creek area [dated at 47.81±.22 Ma ($^{40}Ar/^{39}Ar$, biotite); 47.44±.16 ($^{40}Ar/^{39}Ar$, sanidine)].

REGIONAL CORRELATIONS

Regional correlations of the volcanic stratigraphy of the southeastern Challis field with areas to the north and west are possible primarily by using radiometric age dates. Only the volcanic section in the Lehman Basin and Porphyry Peak areas can be lithologically correlated to volcanic units defined to the north on the Challis 1°x2° quadrangle. In the Lehman Basin area, the upper cooling unit of the tuff of Cherry Creek grades upward into a rheomorphic, lava-like ash-flow tuff which can be correlated with the tuff of Herd Lake, exposed in the southeast corner of the Challis 1°x2° quadrangle. McIntyre and others (1982) postulated that the tuff of Herd Lake had a source from a southerly direction from the Idaho Falls 1°x2° quadrangle. The tuff of Stoddard Gulch in the Antelope Creek area may be correlative with crystal-lithic ash-flow tuffs in the Lake Hills area north of Carey, Idaho.

STYLES OF VOLCANISM

Volcanism within the southern part of the Challis field began at about 49 Ma with effusion of andesitic lavas from widely dispersed fissure and point source centers. Within this extensive andesitic field, there is little physical evidence of steep-sided stratavolcanos. Rather, andesitic lavas formed extensive volcanic plateaus composed of multiple, sheet-like flows with an aggregate thickness up to 1000 m. These andesitic plateaus are thought to have formed topographic barriers which impeded widespread distribution of later ash-flow sheets.

Andesitic volcanism was followed from 49-47 Ma by the eruption of voluminous intermediate lava and ash-flow tuffs. Vents for the intermediate lavas are largely obscured by younger volcanic deposits; however, the presence of NE-trending andesite and dacite dikes and lack of volcanic facies indicative of steep-sided cones suggest that these lavas were also extruded from fissure vents rather than stratavolcano complexes. Intermediate ash-flow tuffs were erupted from sources in the Lehman Basin and Alder Creek areas. Explosive volcanic activity resulted in catastrophic collapse of numerous inferred trap-door segments within a volcanic-tectonic zone. The Alder Creek eruptive center produced the smaller volume tuff of Cliff Creek. Eruption from the Lehman Basin cauldron complex and subsequent catastrophic roof collapse produced more than 200 km^3 of lithic-rich, trachytic tuff of Cherry Creek that extends more than 30 km south and 40 km north from the source. More than 1000 m of ash-flow tuff accumulated within the vent region; lithic fragments are up to 1 m in size. Near-vent, outflow facies of the tuff of Cherry Creek are as much as 500 m thick in the Porphyry Peak area. Distal outflow facies in the Sheep Mountain area, 30 km south of the inferred vent, are up to 200 m thick and contain lithic fragments up to 8 cm in size. Lithic fragments of the tuff of Cherry Creek are predominantly aphyric rhyolite which is indistinguishable from later rhyolite intrusions such as the Boone Creek stock. The eruption of the tuff of Cherry Creek appears to have demolished and incorporated the partly solidified roof of a shallow silicic magma chamber. Soon after explosive activity, the Boone Creek stock, a shallow, silicic intrusion of almost batholithic proportions intruded its own pyroclastic ejecta. Locally, the roof of the intrusion is manifest as an intrusion breccia as much as 200 m thick, exposed over an area of about 3 km^2. In other areas, the intrusion was a warm mass which partially fused and assimilated the tuff of Cherry Creek. Throughout the Lehman Basin cauldron complex, numerous rhyolitic intrusions were emplaced at shallow levels and many probably vented to the surface.

GEOCHEMISTRY

The compositional diversity of igneous rock types in the Challis field precludes use of any model which invokes simple fractional crystallization model from a single parental magma. Mafic to intermediate rocks, including absarokite, latite, and Hi-K basalt and andesite, are inferred to be derivatives from Hi-K basic andesites and shoshonite by fractional crystallization or crystal accumulation processes. Silicic rocks include trachyte, Hi-K dacite and rhyolite and are inferred to be crustal melts.

STRUCTURAL SETTING

Structural styles in the southeastern Challis field are not dominated by NE-striking high-angle, large displacement faults as have been mapped in the area of the Custer graben to the north. Rather, both NNW- and NE-striking faults played important roles in the focus of volcanic centers as well as development of Eocene topography. Early andesites were deposited in NNW-trending topographic lows. This structural grain might have been a relict of pre-Challis Mesozoic deformation or might have been related to incipient Eocene extension. Localization of mafic volcanic centers along NNW-trends suggests that at least some of the NNW structure is Eocene in age; however, other NNW-trending basins present during nascent volcanism are unequivocally not fault-bounded. NNW-trending faults or joints continued to be active throughout the magmatic event as indicated by their importance in the localization of epithermal mineral deposits. NE-trending structure is best expressed by the orientation of late-stage silicic dike swarms such as those associated with the Mackay stock. The smaller volume of igneous activity and structural displacement suggest that this part of the Challis field did not undergo the same amount of extension as the trans-Challis fault system to the north.

MINERAL DEPOSITS

Gold and silver deposits of the Challis field include a variety of volcanic-, volcanic-sediment- and shallow-intrusive hosted epithermal deposits related to the end stages of silicic volcanism which was focused by volcano-tectonic structures such as the Lehman Basin cauldron complex. Actual deposit types include mineralized fissure fillings, silicified breccia zones and disseminations and stockwork. Wallrock alteration includes silica flooding and stockwork, and widespread argillization and propylitization. Oxygen isotope work in the Custer graben (Criss and others, 1985) and the Lehman Basin area indicate that hydrothermal alteration is related to large-scale convective movement of meteoric waters, most likely heated by shallow intrusions or high heat flow related to extensional tectonics. Preliminary fluid inclusion studies from the Lehman Basin area indicate that hydrothermal fluids were low salinity (<3 wt. percent) and low to moderate temperature (250-300 °C).

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GEOLOGIC CONTROLS OF MINERALIZATION IN SEDIMENTARY ROCKS, WESTERN IDAHO FALLS 1°x2° QUADRANGLE

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INTRODUCTION

The Paleozoic sedimentary rocks (fig. 1) of the western half of the Idaho Falls 1°x2° quadrangle are known hosts for two types of mineral deposits: skarn deposits and polymetallic vein deposits, including fissure fillings and replacements along fractures and bedding. Known mineral deposits in this area, thought to be Tertiary in age, are the result of either contact metamorphism and metasomatism or deposition from hydrothermal solutions. The physical and chemical characteristics of the Paleozoic sedimentary rocks indicate a potential for as yet undiscovered, sediment-hosted, jasperoid-associated precious-metal deposits.

MINERAL DEPOSITS

Three major mining districts, Alder Creek, Copper Basin, and Lava Creek, and numerous other prospects are in the Mackay area (fig. 2). Most of the deposits were discovered during the 1880's when rich silver-lead ores were mined throughout central Idaho. Many properties in the area have been active intermittently since their discovery, but have not produced substantial amounts of additional ore.

Two types of mineral deposits, skarn and polymetallic veins found in Paleozoic sedimentary rocks, have been exploited in the Mackay area. Metals in these deposits may have been in part remobilized out of Paleozoic sedimentary country rocks. Mudstone and siltstone of the Mississippian McGowan Creek Formation that were deposited in a deep restricted marine basin may be metal-enriched (Erdman and others, 1988). Fluids associated with Tertiary magmatic activity could have remobilized some of these metals into the present skarn deposits, though the primary metal source was magmatic. Hydrothermal systems associated with late stages of volcanism and intrusion probably formed the polymetallic vein deposits of this area.

GEOLOGY

Geologic Setting

Geology of the Mackay area consists mainly of Paleozoic carbonate and clastic rocks intruded by Tertiary plutons and hypabyssal bodies and overlain by Eocene Challis Volcanics (fig. 1). Paleozoic sedimentary rocks in the area range in age from Early Mississippian to Early Permian and consist of the McGowan Creek Formation, the White Knob Limestone, and a group of limestones collectively called the carbonate bank (includes the Middle Canyon, Scott Peak, South Creek, Surrett Canyon, Bluebird Mountain, and Snaky Canyon Formations). Isolated outcrops of older Devonian and Ordovician rocks are exposed south of Timbered Dome (Skipp and others, 1989).

Plutonic bodies are multiphase granitic intrusions composed mainly of porphyritic granite, granodiorite, quartz diorite, leucogranite porphyry, (Nelson and Ross, 1968; 1969a, b) and quartz monzonite (Dover, 1981). Numerous dikes of quartz latite, rhyolite, and porphyritic rhyolite cut all rock types and are commonly found in swarms parallel to northwest- and northeast-trending fracture systems (Nelson and Ross, 1969a, b). Andesitic to rhyolitic flows, breccias, and tuffs of the Challis Volcanics are widespread. Several north-northwest and northeast-trending high-angle faults cut the rocks of this area and probably acted as conduits for mineralizing fluids. Jasperoid and jasperoid breccia, mainly siliceous replacements of limestone, locally form prominent outcrops near the faults.

HOST ROCKS

McGowan Creek Formation

The McGowan Creek Formation (Lower Mississippian; Sandberg, 1975) was deposited in a foreland basin (Skipp and others, 1979) and consists of distal thin-bedded turbidite, calcareous siltstone, mudstone, and minor silty limestone (Skipp, 1988; 1989). It is exposed near Grouse and near Timbered Dome (fig. 1). Sandstone and siltstone of this formation have been silicified locally, forming jasperoid bodies. Recent biogeochemical studies by Erdman and others (1988) suggest that the McGowan Creek is locally metal-enriched. Although no mineral deposits have been located in the McGowan Creek Formation, it may be a source of metals that were remobilized by hydrothermal solutions.

White Knob Limestone

The White Knob Limestone (Mississippian) consists of light- to medium-gray fossiliferous limestone with interbedded chert- and quartz-pebble conglomerate and sandstone in its upper part. It gradationally overlies the McGowan Creek Formation and was deposited in a subtidal to intertidal turbulent marine environment (Skipp and others, 1979). The White Knob is exposed in the White Knob Mountains west of the town of Mackay.

Known mineral deposits associated with the White Knob Limestone include skarn deposits and polymetallic vein and replacement deposits of the type found in the Copper Basin and Alder Creek Mining Districts (fig. 2). These deposits formed as the result of the intrusion of Tertiary-age stocks and related dikes into the White Knob Limestone. Magma from the stocks may have supplied iron, magnesium, silica, and some sulfide minerals. Contact metamorphism and metasomatism resulted in the formation of skarn; irregular-shaped ore bodies were found in carbonate rocks near the contact with the stock or in xenoliths in igneous rocks. Skarn deposits were mined for copper, lead, zinc, silver, gold, tungsten, and molybdenum and consist of chalcopyrite, pyrite, pyrrhotite, magnetite, galena, fluorite, scheelite, molybdenite, sphalerite, and specularite in addition to calc-silicate minerals, calcite, and quartz (Umpleby, 1917).

Polymetallic vein deposits, also found in these districts, are mainly replacements along fractures and bedding planes of the White Knob Limestone; local bedding planes, faults, and fracture zones controlled the location of the ore bodies. The ore minerals, argentiferous galena, pyrite, chalcopyrite, and sphalerite occur as disseminations, clots, and lenses. Many of the deposits were worked only in the oxidized zones where iron and manganese oxides, cerussite, and cerargyrite are common.

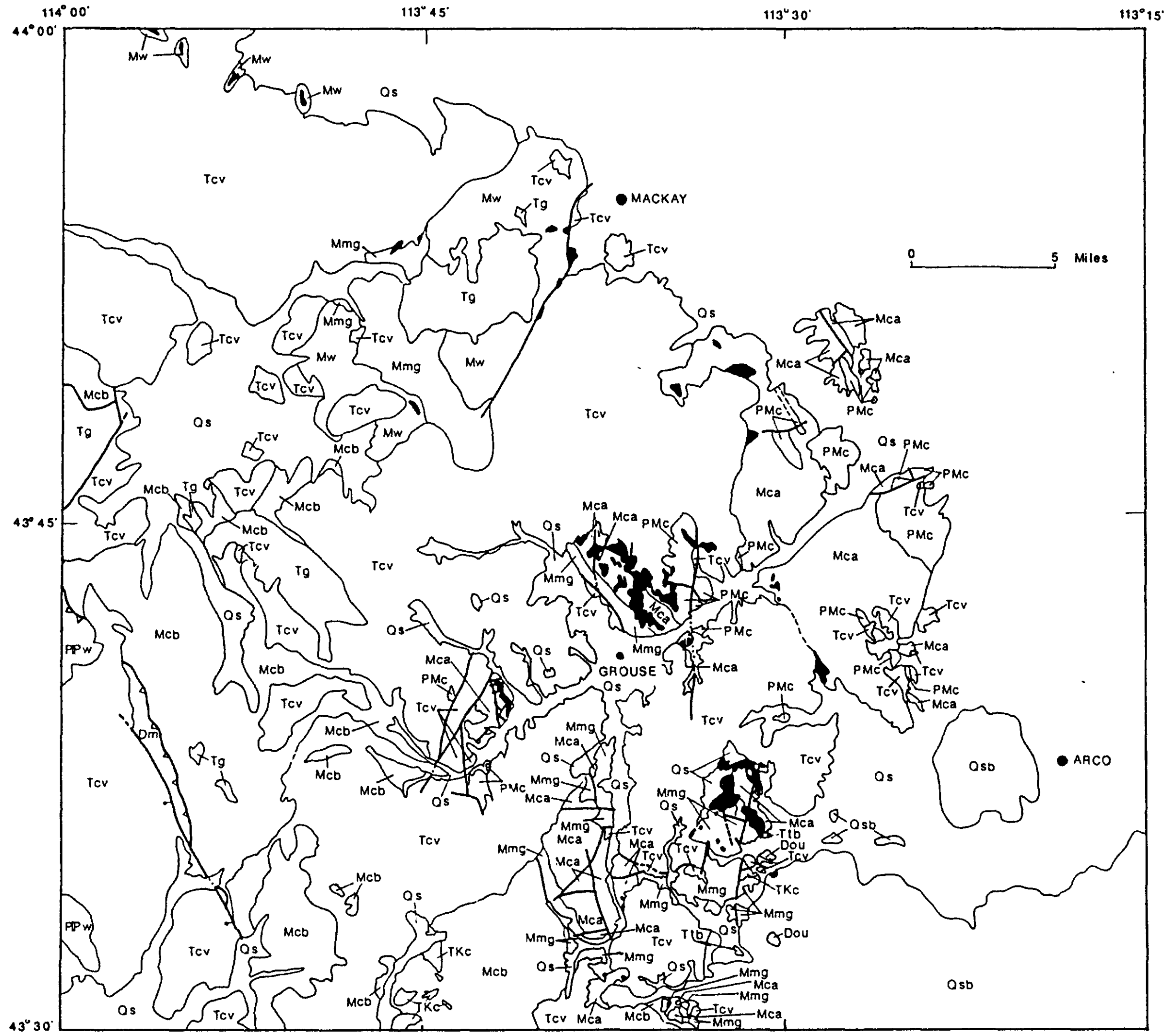

Figure 1. Generalized geologic map of the Mackay area, south-central Idaho (adapted from unpublished mapping of Worl and others).

EXPLANATION OF MAP UNITS

Qs	Alluvium, gravel, etc. (Quaternary)
Qbs	Basalt
	Jasperoid
Ttb	Tuff of Blacktail (Tertiary)
Tcv	Challis Volcanic Group (Tertiary)
Tg	Granite, granodiorite (Tertiary)
TKc	Basal Challis Conglomerate and
	Smiley Creek Conglomerate (Tertiary and Cretaceous)
PPw	Wood River Formation (Lower Permian to Middle Pennsylvanian)
PMc	Primarily carbonate rocks, including
	Snaky Canyon Formation (Lower Permian to Upper Mississippian)
	Bluebird Mountain Formation (Upper Mississippian and Lower Pennsylvanian)
Mca	Carbonate units, including
	Surrett Canyon Formation (Upper Mississippian)
	South Creek Formation (Upper Mississippian)
	Scott Peak Formation (Upper Mississippian)
	Middle Canyon Formation (Upper and Lower Mississippian)
Mw	White Knob Limestone (Upper Mississippian)
Mmg	McGowan Creek Formation(Lower Mississippian)
Mcb	Copper Basin Formation(Upper and Lower Mississippian)
Dm	Milligen Formation (Devonian)
Dou	Undifferentiated units, including
	Carey Formation (Upper Devonian)
	Fish Haven Dolomite (Upper Ordovician)
	Kinnikinic Quartzite (Middle Ordovician)
	Summerhouse Formation (Middle Ordovician)

Contact, dashed where inferred

Fault, dashed where inferred, dotted where concealed, ball on downthrown side

Thrust fault

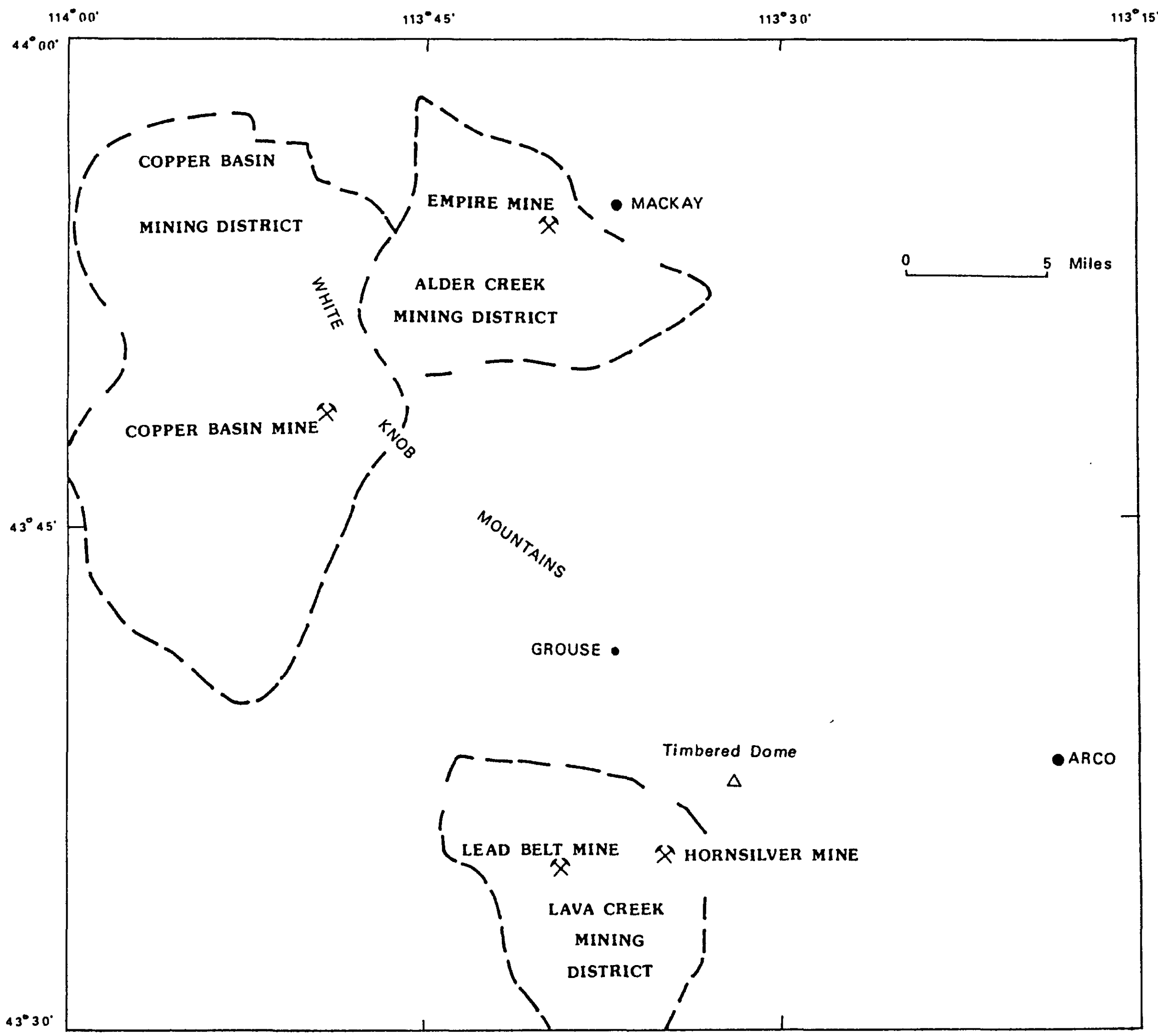

Figure 2. Map showing location of mining districts near Mackay, Idaho.

Carbonate Bank Limestones

The Lower and Upper Mississippian Middle Canyon Formation and the Upper Mississippian Scott Peak, South Creek, and Surrett Canyon Formations are carbonate bank and forebank deposits across which a flood of fine-grained sand, the Bluebird Mountain Formation, transgressed in latest Mississippian time (Skipp and others, 1979). The Snaky Canyon Formation (Upper Mississippian to Lower Permian) is a shallow-water carbonate bank deposit consisting mainly of sandy limestone and calcareous sandstone. These formations, which comprise the carbonate bank are exposed north and south of Antelope Creek.

Known mineral deposits associated with the carbonate bank limestones are mainly silver, lead, and zinc produced from replacement-type veins. Location of the ore is controlled by favorable bedding planes, fracture zones, and faults. In the Lead Belt mine area of the Lava Creek District ore deposits are mainly replacements in limestone and shale along north- and northeast-trending faults.

Jasperoid

All of the host rocks have been silicified locally to jasperoid or jasperoid breccia. Major fault systems probably acted as conduits for silica-bearing solutions that replaced the host rocks. These solutions may also have been metal-bearing in which case there could be large low-grade precious-metal deposits within or near the jasperoid bodies (Soulliere and others, 1988; Wilson and others, 1988; Worl and others, 1988). Jasperoid mainly replaces the limestone of the White Knob Limestone and carbonate bank formations near high-angle faults or volcanic rocks, but also replaces sandstone and shale of the McGowan Creek Formation (Skipp, 1988 and 1989; skipp and others, 1989). The degree of silicification varies from patchy areas to dense, massive jasperoid. Many jasperoid bodies show at least two generations of brecciation and silicification. Quartz and calcite veins are common, and minor barite, fluorite, jarosite, and stibiconite are present along fracture surfaces. Weathered surfaces are stained with hematite and limonite.

STRUCTURE

Intermediate to silicic plutonic and hypabyssal intrusive rocks are locally present along high-angle fault systems. The jasperoid bodies are also commonly aligned along high-angle regional faults and were probably formed from solutions moving along and outward from these faults. Southworth (1988) recognized significant northeast- and northwest-trending lineaments at a local and regional scale in the Idaho Falls 1°x2° quadrangle. He further noted that the northwest-trending faults are often offset by northeast-trending faults and that the intersection of these trends may be potential zones of mineralization.

POTENTIAL FOR SEDIMENT-HOSTED, JASPEROID-ASSOCIATED DEPOSITS

Many jasperoid bodies in the Mackay area are similar to jasperoid in the sediment-hosted precious-metal deposits of Nevada (Wilson and others, 1988).

Recent geochemical analyses of jasperoid and jasperoid breccia samples indicate that the suite of elements Ag, As, Au, Hg, Sb, and Tl is consistently present, commonly in anomalous concentrations (Soulliere and others, 1988; Wilson and others, 1988). These elements are characteristic of the upper parts of hydrothermal systems, hot-springs-type precious metals deposits, and sediment-hosted disseminated precious-metal deposits. The presence of this suite of elements suggests that some, if not most of the jasperoid in this area formed through the replacement of sedimentary rocks by silica and associated elements that were carried in hydrothermal solutions. Trace metals in jasperoid, together with known vein-type metal deposits along high-angle faults indicate that some of the solutions were metal-bearing. These solutions may have also formed precious-metal deposits in altered but nonsilicified country rock near jasperoid, in silicified country rock, or in veined and brecciated jasperoid.

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GEOLOGIC CONTROLS FOR CARBONACEOUS LIMESTONE- AND SHALE-HOSTED SILVER-LEAD-ZINC-COPPER DEPOSITS IN THE COPPER BASIN FORMATION, PIONEER MOUNTAINS, IDAHO

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Mississippian carbonate and clastic rocks of the Copper Basin Formation in south-central Idaho host many vein, replacement, skarn, and stratabound deposits of silver-lead-zinc (± copper). In several places, the silver-base metal deposits are associated with stratabound veins and replacements of barite. The locations of these deposits are governed by the presence of carbonaceous pelitic source rocks, chemically-reactive calcareous trap rocks, concordant and discordant structures that provided channelways for solutions, and intrusive rocks that provided heat and fluids for metasomatism. Known deposits with significant resources are confined to the southern and northern Pioneer Mountains (fig. 1).

The Copper Basin Formation consists of siliceous and calcareous sedimentary rocks that were deposited in an elongate marine trough or foreland basin between the Antler orogenic highlands to the west and the craton to the east. The formation includes two coeval but thrust-bounded sequences, juxtaposed along the Glide Mountain thrust: the upper, Glide Mountain plate, consisting of carbonate-poor, shallow marine to partly terrigenous basin-margin facies assemblages, and the lower, Copper Basin plate, consisting of marine basinal facies assemblages. Most mineral deposits occur in the basinal assemblages, which are designated the Little Copper and Drummond Mine Limestone members, and consist chiefly of argillaceous and micritic turbidites, respectively.

All known exposures of the Copper Basin Formation are allochthonous, having been transported eastward along several generations of thrusts that developed principally during the Cretaceous Sevier orogeny. In most places, the Copper Basin plate consists of a broadly-folded east-dipping homocline, whereas the Glide Mountain plate is characterized by pervasive tight folding with well-developed axial plane cleavage. The more intense deformation of rocks in the upper plate may reflect their longer tectonic transport from an original position farther west than rocks in the lower plate. Faults parallel to bedding are common in rocks of the Copper Basin plate. These bedding-plane faults likely formed during the Sevier orogeny also, but provided "prepared ground" for later mineralization.

Late Cretaceous and (or) early Tertiary uplift, erosion, and deposition of the Smiley Creek Conglomerate preceded onset of Challis volcanism about 49 Ma. The Eocene magmatic event included coeval volcanism and emplacement of hypabyssal stocks and dikes. Both sequences of the Copper Basin Formation are intruded by porphyritic stocks, dike swarms, and single dikes. Stocks are strongly-discordant porphyritic hornblende-biotite (± pyroxene) quartz monzonite, with no internal fabric. The stocks are elongated in a nothwest-southeast direction. Dikes are texturally and compositionally more diverse, ranging from quartz porphyry (rhyolite?) to diabase (and possibly lamprophyre). Dikes and country rock locally are intensely altered. In a few places, dikes intrude the stocks and are inferred to be feeders to flows of the Challis Volcanics. However, the stocks also are Eocene, and probably are coeval with earlier Challis volcanism.

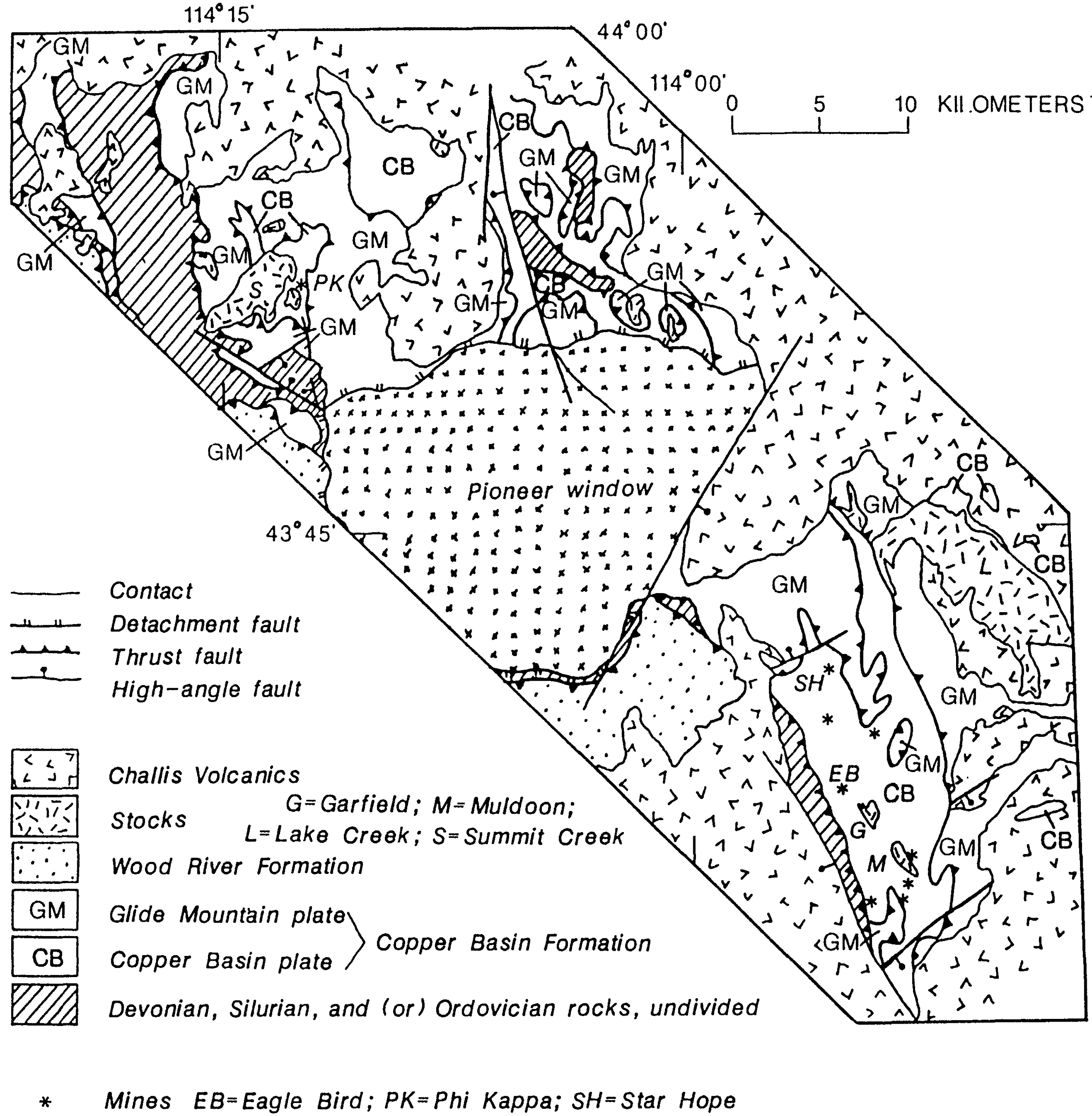

Figure 1. Geologic setting of selected mines in the Copper Basin Formation

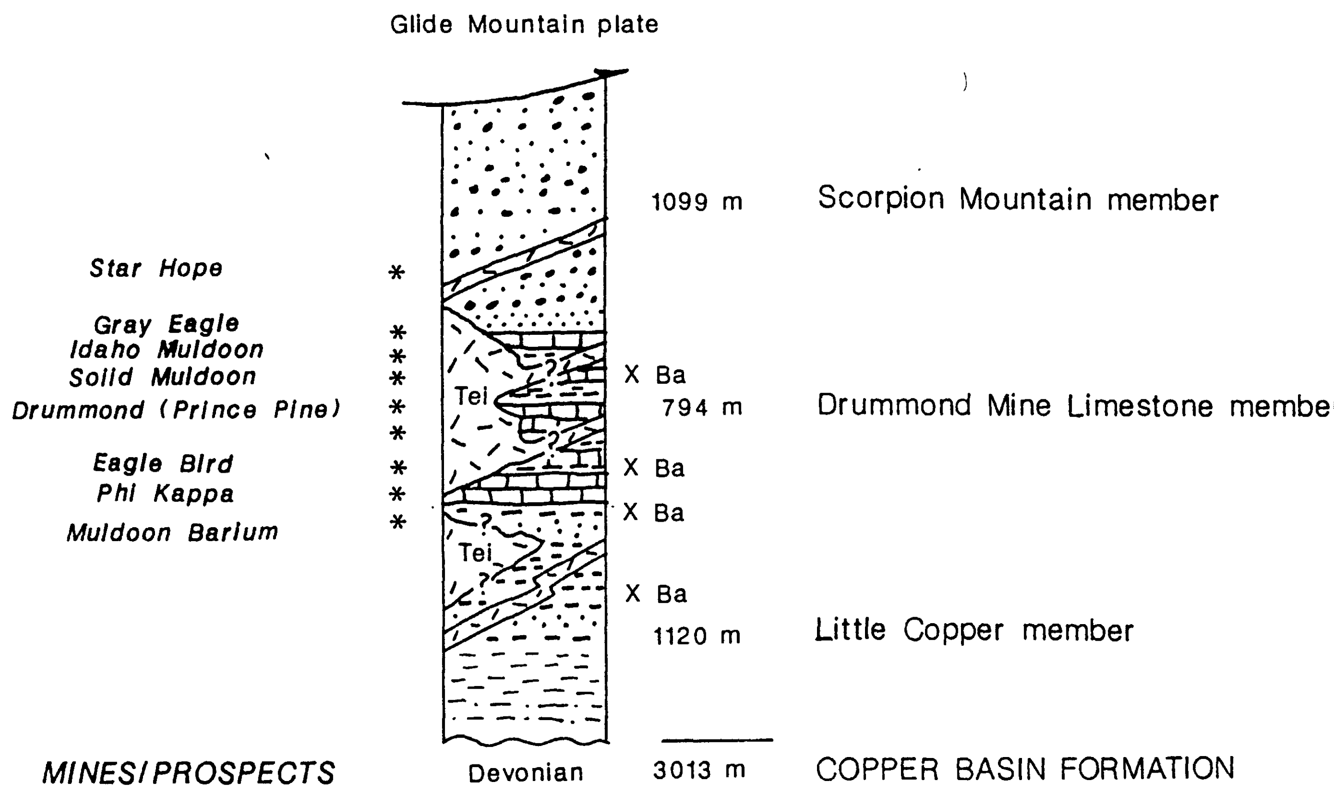

Figure 2. Schematic stratigraphic section of Copper Basin Formation, Copper Basin plate

Steep faults cut both sequences of the Copper Basin Formation in many places. They tend to follow two orientations, N. 10 W. to N. 30 W. and N. 50 E. to N. 70 E., which also are the dominant trends of Eocene dikes that cut the Copper Basin. The ENE-trending faults and dikes are roughly parallel to the trans-Challis fault system and likely formed during Eocene regional extension and magmatism. NNW-trending faults generally have been attributed to Neogene Basin and Range extension, but their rough parallelism to trends of some Eocene dikes may indicate an earlier ancestry.

Silver-base metal deposits of the Little Wood River, Copper Basin, and Alto districts have been known since the 1880s, when many of the highest grade properties had significant production. Although subsequent mining has been sporadic, U.S. Bureau of Mines records indicate some production in every decade since 1901. Numerous high-grade veins and replacements remain; however, they generally are very lenticular and of small tonnage. Exceptions include large stratabound replacement deposits at the Eagle Bird and Phi Kappa mines, where inferred resources are much larger. The deposits exhibit a stratigraphic control that may indicate some syngenetic metal deposition--an implication that lower grade resources may be relatively widespread in the Copper Basin Formation.

Dark argillaceous rocks in the lower part of the formation (particularly the Little Copper member) are significantly enriched in barium and boron, and slightly enriched in cobalt, chromium, manganese, lead, and zinc. They may have served as sources of metals for mineral deposits that are located higher in the stratigraphic sequence. Stratabound barite occurs as recrystallized, light-colored, bedding-parallel stringers and lenses at several horizons in the Little Copper, as well as in the overlying Drummond Mine Limestone member (fig. 2).

Impure calcareous clastic rocks in the middle of the formation (particularly the Drummond Mine Limestone member) host nearly all silver-base metal deposits in the Copper Basin Formation. In general, deposits are stratabound at several stratigraphic levels (fig. 2); in detail, they are localized or enriched where bedding-plane and cross-cutting faults intersect. Many deposits contain coarse-grained sulfide and calc-silicate minerals in contact with unrecrystallized strata suggestive of distal Zn-Pb skarns.

The specific influence of intrusions on formation of the silver-base metal deposits is equivocal. The two largest deposits (Phi Kappa and Eagle Bird) are located in the Drummond Mine Limestone within 1 km of quartz monzonite stocks (fig. 1, 2). However, unmineralized host rocks show little or no contact metamorphism. Other deposits with similar mineralogy and textures are located much further from exposed plutons, but all host rock sequences are intruded by scattered to abundant dikes. In many places, the Copper Basin Formation is extended 10 percent or more by dikes intruded either along the regional strike (NNW) or transverse to it (ENE). Whether or not the dikes drove hydrothermal systems, they apparently followed structures of the same orientation as those used by the fluids responsible for metal enrichment.

Little information is available on gold contents in the silver-base metal deposits, but the presence of gold is noted in most reports, and is detected in most samples that have been analyzed recently. Production records show a range from traces to about 0.2 g/t of ore shipped. In deposits of the Little Wood River district, a crude zoning of sulfide minerals is present, with pyrite, pyrrhotite, arsenopyrite, and chalcopyrite becoming more common closer to the Garfield and Muldoon stocks, and to the position of an inferred buried stock beneath the Star Hope mine. It is possible that, whereas silver contents can be expected to decrease near the stocks, gold grades might be sufficiently higher to warrant exploration, particularly where faults or fractures cut enclosing sedimentary rocks. It also is possible that gold grades may be marginally higher in non-calcareous strata in the Drummond Mine Limestone member, which locally comprise perhaps as much as twenty-five percent of the unit.