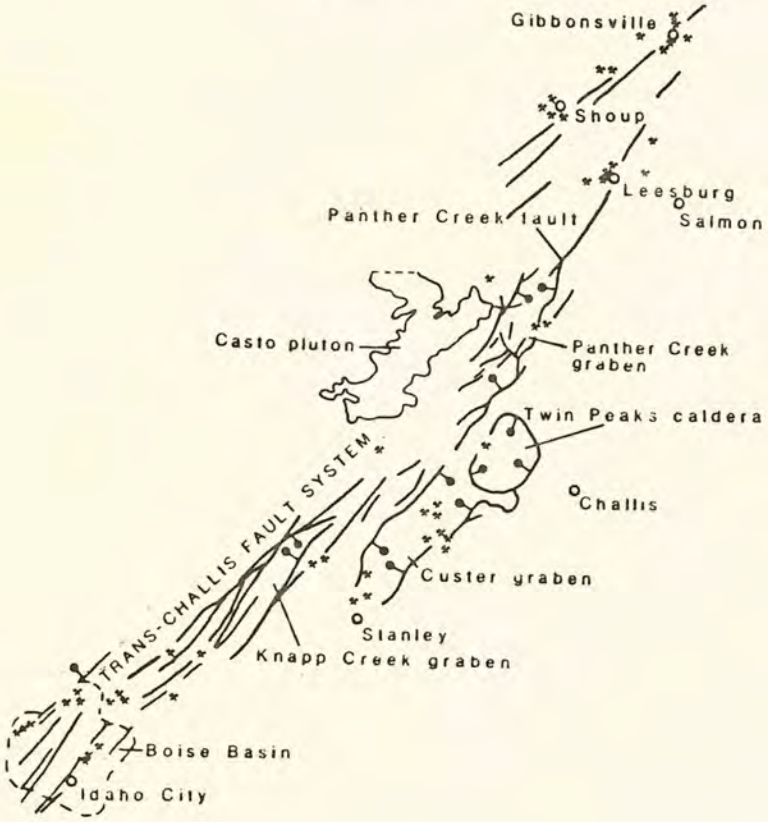
TOBACCO ROOT GEOLOGICAL SOCIETY

FIFTEENTH ANNUAL FIELD CONFERENCE

GEOLOGY AND ORE DEPOSITS OF THE TRANS-CHALLIS FAULT SYSTEM /GREAT FALLS TECTONIC ZONE



Salmon, Idaho
August 16-18,
1990

FALMA J. MOYE
EDITOR

GUIDEBOOK

OF THE

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FIELD TRIP THROUGH THE BLACKBIRD MINING DISTRICT
LEMHI COUNTY, IDAHO

BY

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INTRODUCTION

This field trip will review rocks of the Middle Proterozoic Yellowjacket Formation that contain a unique class of strata-bound Co-Cu-Au deposits. These deposits of the Blackbird district occur within a sequence of metasedimentary rocks that are well exposed in the canyon walls along Blackbird Creek. The first three stops on the trip will examine the host rock units approximately three miles along strike (southeast) from the main Blackbird mine area. The next two stops will be in similar rocks displaying some of the distal features of the mineralized stratigraphy about a mile from the main deposit zone. The final two stops will view ore lodes and their host rocks exposed at the Sunshine prospect and Blackbird open pit.

HISTORY

The first mining claims in the Blackbird district were located in 1893 and many of the present claims covering the Blackbird mine were recorded during the following three years. All of the original claims were held for gold, but in 1896 the discovery of high-grade copper deposits led to a new generation of claim staking. In 1899, a central group of 29 claims were bonded by the Blackbird Copper-Gold Mining Co. and for the next two years about 1400 feet of underground development work was completed. The claims were patented at the conclusion of this work.

Cobalt was first recognized in 1901 when John Beliel, a resident of Leesburg, staked 14 claims over erythrite-stained outcroppings located along the West Fork of Blackbird Creek. A few shipments of ore were extracted from these claims for mill and smelter tests. No further attempt was made to exploit the cobalt until 1917 when Union Carbide Corp. developed the Haynes-Stellite property. By 1920, a total of 55 tons of concentrate containing 17.75% Co were produced from 4000 tons of ore taken from the Haynes-Stellite mine.

Interest in cobalt was not renewed until World War II when development of the turbine engine created a strategic importance for the metal. A joint exploration program in the 1940's between the U.S. Bureau of Mines, U.S. Geological Survey, and the Howe Sound Co. delineated a large tonnage of cobalt-copper ore in two deposits. These formed the core of the 1.7 million tons of ore that were to be produced by the Calera Mining Co. (subsidiary of Howe Sound Co.) between 1951 and 1959. The average grade for these ores was 0.63% Co and 1.65% Cu. Gold was recovered from the copper concentrates which carried between 0.15 and 0.50 ounces per ton.

During the 1960's the property was operated on an intermittent basis by the Machinery Center Co. who extracted high-grade copper-gold ores. In 1967, Hanna Mining Co. began an exploration program at the mine and initiated a metallurgical research program. After several feasibility studies they concluded the project to be uneconomical at the existing metal prices in 1976.

In 1977, Noranda optioned the property from Hanna based on recognizing that the ores represented a type of stratabound deposit that could be very large. In 1978, an invasion of Zaire by Cuban-supported rebels from Angola threatened to disrupt the supply of cobalt to free-world markets – the price quadrupled to $25/lb. By 1982, Noranda's program had delineated a reserve of nearly eight million tons of mine-ready ore grading about 0.74% Co and 1.40% Cu. At least four new ore zones were identified by applying new geologic concepts regarding the genesis and strata-control of the mineralization.

Even though the district is considered to have excellent potential for hosting tens of millions of tons of high-grade cobalt ores, the present economic conditions do not warrant mine development. However, this resource represents an "insurance policy" that would supply primary cobalt to the U.S. should foreign sources be cut-off or even disrupted for extended periods.

REGIONAL GEOLOGY

The Blackbird district is located in a part of east-central Idaho that is clearly dominated by rocks of Proterozoic age (Figure 1). The Yellowjacket Formation represents the oldest unit in a nearly 50,000 feet thick sequence of metasedimentary rocks (Ruppel, 1975). The base of the Yellowjacket has never been observed though prevalent interpretations favor deposition on an older crust of Archean metamorphic rocks that were wrenched apart to form an intracratonic rift basin (Hughes, 1983).

The Yellowjacket basin represents a first-order tectonic trough formed by a structural network of basin-margin and intrabasin faults. A complex submarine topography across much of this tectonic basin predicted the style and rates of sedimentation for the Yellowjacket Formation. Most clastic materials were deposited in the main trough either on large submarine fan complexes or deltaic aprons that were frequently "drowned" by tectonic-foundering of the basin. The Yellowjacket rocks are often correlated, in a general sense, with similar looking lithologies in the lowermost sequences of the Belt Supergroup in northern Idaho and western Montana.

The Lemhi Group represents another thick accumulation of Proterozoic metasedimentary rocks in the region. They are considered to be younger than rocks in the Yellowjacket Formation (Ruppel, 1975). Ruppel (1978) also considers the Lemhi

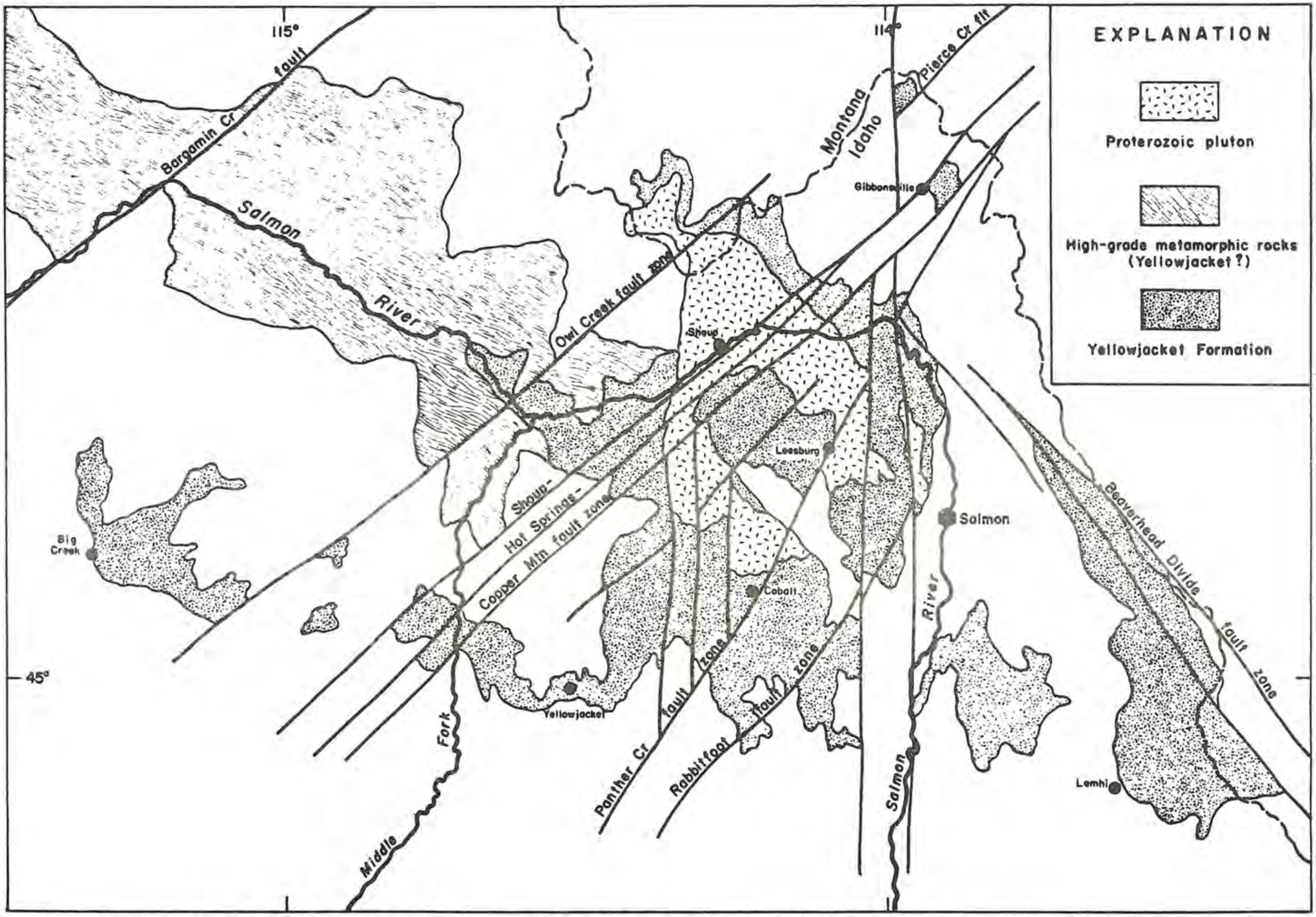

Figure I. Generalized geologic map showing the distribution of the Yellowjacket Formation in east-central Idaho.

Group to be allochthonous throughout much of the region, having arrived in their present positions by riding the Medicine Lodge thrust system. Regional correlations between the Lemhi Group and rocks of the Belt Supergroup have been proposed (Ruppel,1975), and would indicate an age between about 1,400 and 1,000 m.a.

Igneous rocks of widely variable compositions and ages are distributed throughout the region. The oldest recognized suite of igneous rocks are represented by mafic tuffs and their possible comagmatic feeder-dikes and sills that predominantly occur within the middle unit of the Yellowjacket Formation. Granitic rocks dating back to 1,370 m.a. were emplaced after an early-stage of metamorphism had been overprinted on the Yellowjacket (Evans and Zartman, 1981). Alkali granite and syenite were emplaced in the Yellowjacket during an apparent anorogenic phase in the Ordovician (Evans and Zartman, 1988). The next major episode of magma emplacement resulted in the formation of the Idaho Batholith (L. Cretaceous - E. Tertiary). The last important phase of magmatism culminated in the Eocene with the eruption of the Challis Volcanics and emplacement of their sublevel plutons.

The Yellowjacket rocks have been significantly affected by at least four major tectonic episodes:

1. Proterozoic basin development
2. Paleozoic doming
3. Mesozoic overthrusting
4. Cenozoic block-faulting

Each of these events created new structures and reactivated many of the older ones. Consequently, the region is cut by numerous faults that follow just about any direction on a compass. However, without much doubt, the most impressive tectonic feature is a northeast trending, trans-continental zone of faults called the trans-Challis fault system (Kiilsgaard and others, 1986). Some of the faults in this system probably represent structures that contributed to shaping the Yellowjacket basin. A large number of mineral systems are located along the extent of this tectonic zone, perhaps some of the oldest are to be found in the Blackbird district.

GEOLOGY OF THE BLACKBIRD MINE AREA

Detailed studies by Noranda geologists from 1978 to 1982 identified three significant lithostratigraphic units within the Yellowjacket Formation. Recognition of these units was the key to interpreting a) the environment of deposition, b) stratigraphic controls, and c) the basinal distribution of the various ore deposits in the Blackbird mine area. The distinct facies are shown on a generalized geologic map of the Blackbird district (Figure 2), and are graphically depicted on a stratigraphic column (Figure 3).

Yellowjacket Stratigraphy

Lower Unit. The oldest rocks in the Blackbird district are found in the lower unit of the Yellowjacket Formation. This unit is at least 10,000 feet thick and mainly represented by thin to very thin bedded argillite, siltite, and very fine-grained quartzite. The most distinctive facies within the lower unit consists of graded beds of argillaceous siltite and argillaceous quartzite. The original clay and silt rich beds are now dark olive-gray to nearly black in color; whereas, sandier beds are medium gray to pale greenish-gray.

Horizontal planar and wavy laminae are common sedimentary features while ripple cross-laminations are less common. Long, sinuous mudcracks filled with very fine sand and silt are frequently observed within the argillite beds in the upper one-third of the unit. Penecontemporaneous deformation features include convoluted laminae, load casts, intrastratal folds, mud "flames", and fluidal-textured intervals. Outcroppings of the lower unit of the Yellowjacket display phyllitic textures in the vicinity of major regional structures such as the Panther Creek fault.

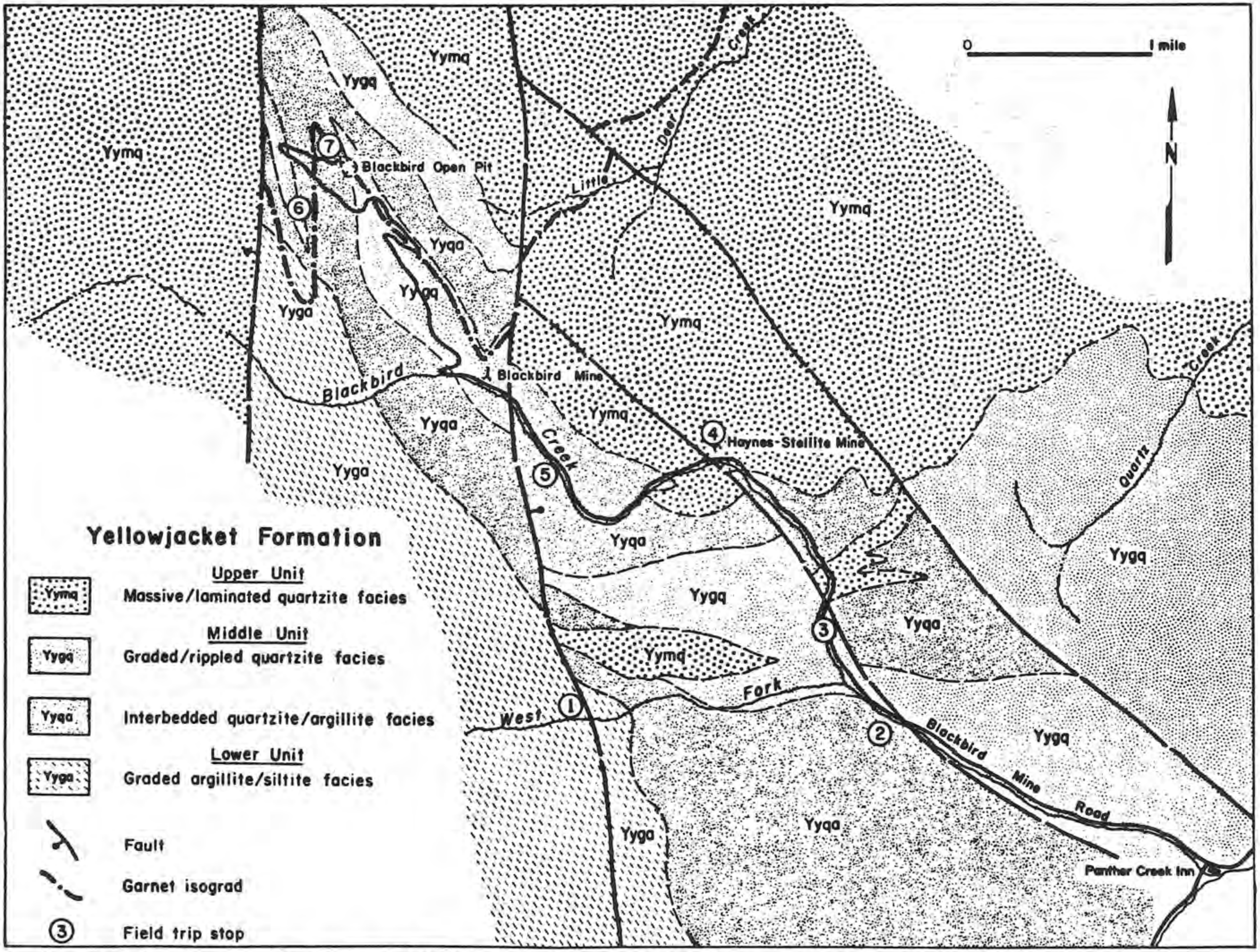

Figure 2. Generalized geologic map and field trip stops in the Blackbird mining district, Lemhi County, Idaho.

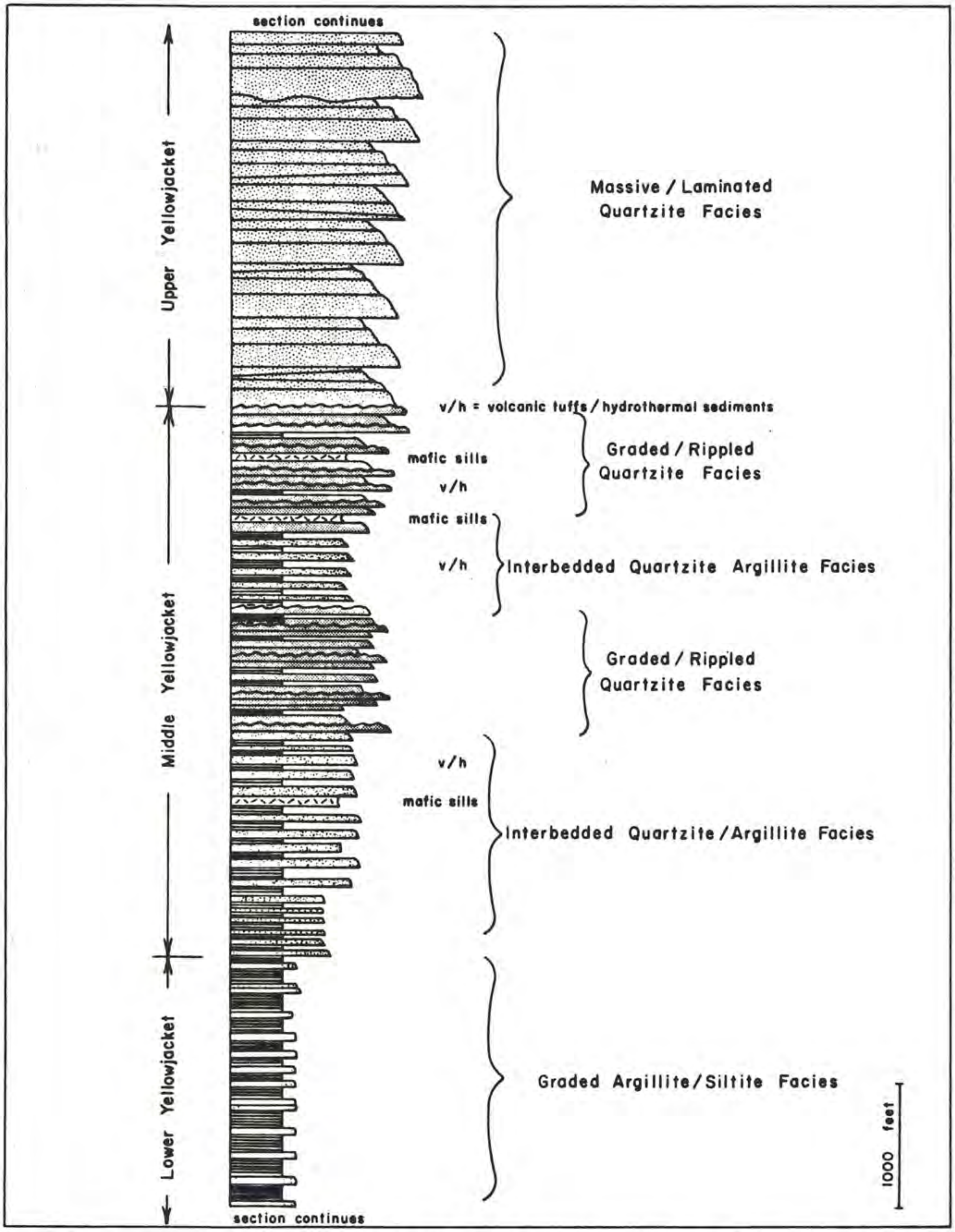

Figure 3. Partial stratigraphic column for the Yellowjacket Formation in the Blackbird mining district, Lemhi County, Idaho.

Deposition of the lower unit mainly occurred during very quiet basinal conditions. Periodic pulses of very fine sand and silt deposited during traction along the bed were followed by muddy materials being dropped from a suspended column of sediment. This style of sedimentation typically represents distal basin plain turbidites or prodelta muds that accumulate near the outer fringes of prograding submarine fans.

Middle Unit. The middle unit of the Yellowjacket Formation in the Blackbird district is at least 4,000 feet thick. Two different facies are generally recognized within this unit. An interbedded quartzite and argillite facies overlies rocks of the lower unit. This facies grades both upwards and laterally into a graded and rippled quartzite facies. In the Blackbird mine area this change is repeated in two megacycles.

The interbedded quartzite and argillite facies is characterized by thin to medium beds of fine-grained quartzite alternating with generally thin to very thin beds of sandy and silty argillite. Bed thickness for the quartzite units is very uniform over large areas. Low-angle cross-bedding is commonly observed in quartzite beds that appear to rest on slightly undulating scour bases. Grading within the alternating beds of quartzite and argillite is not as prominent as that normally seen in the other facies. Other common features in the interbedded quartzite and argillite facies are mud cracks, wavy and planar laminations, pseudonodules, and discontinuous ripples. Sedimentation of this facies seems to have occurred in an almost steady, rhythmic pattern.

The graded and rippled quartzite facies of the middle unit appears to be both vertically and laterally gradational with the interbedded quartzite and argillite facies. This facies is also dominated by very thin to medium bedded quartzite layers that are interbedded with very thin argillites. However, numerous current-rippled bedding surfaces and ripple cross-laminated bedding sections characterize this facies. Climbing, erosional, and starved ripples are also common throughout the quartzite layers of this facies. In addition to the prominent ripple marks, the beds are also noticeably graded.

Mud cracks formed by "stretching" of the argillite beds down the basinal slope are common sedimentary features. Water-escape structures formed by rapid deposition and compaction of the graded and rippled quartzite facies are seen in almost every outcrop. Other early deformation features are intrastratal folds, slump folds, flame structures, ball-and-pillow structures, microfaults, and fluidized textures. Strata of this facies seem to be warped almost everywhere they are exposed.

Deposition of the middle unit of the Yellowjacket Formation apparently took place under frequently fluctuating energy conditions. Periodic sediment fluxes were transported downslope mainly by turbidity currents. When a pulse of sand deposition waned, mud deposition gradually resumed. Occasional bottom traction currents winnowed the fines and rippled the sand to produce flaser structure. Rapid deposition and burial inhibited dewatering of the sediment column causing quicksand movement and pull-apart of the argillaceous units during compaction. This unit is representative of medial to distal turbidites deposited in the lower fan subenvironment of a submarine fan complex, or delta-front sands and muds deposited in a prograding, fluvial-dominated deltaic system.

Upper Unit. The massive, laminated quartzite facies is probably the most unmistakable unit of the Yellowjacket Formation. This facies forms the upper unit in the Blackbird district and achieves a thickness of at least 5,000 feet. Medium to thick beds of fine to medium-grained quartzite with local thin to very thin interbeds of argillaceous siltite and argillite dominate this facies. Most beds are quite uniform and very continuous though wedge-shaped beds are also present. The monotonous repetition and blocky weathering of these beds are a distinctive feature of this facies.

Parallel, horizontal laminations and tangential, low-angle cross-laminations are the most common sedimentary features in the upper unit. Another primary sedimentary feature is the undulatory scour surface at the base of many sand beds. Some rippled beds are found in the lower portions of the massive, laminated quartzite facies, and some small rip-up clasts of argillaceous sediment have also been observed in the same intervals. Large-scale hummocky beds occur well within this unit and indicate the possibility of periodic storm-wave action on the sediment bed.

Soft-sediment deformation is uncommon in the upper unit with creep-folds and small, compaction-faults being noted on occasion. Faintly visible dish structures occur within some massive quartzite beds, especially in the first deposited several hundred feet of the unit.

Overall, the upper unit consists of coarsening and thickening upward megasequences that may be 1,000 to 2,000 feet thick. Deposition of the unit is the result of strong and fairly continuous current activity. Traction carried most of the sandy sediment on the bed abd deposited it under conditions of high flow regime. The massive, laminated quartzite facies is typical of the mid-fan subenvironment of deepwater submarine fans, or distributary mouth bars of delatic systems.

Igneous Rocks

Mafic tuffaceous rocks principally located within the middle unit of the Yellowjacket Formation are believed to be the oldest igneous rocks in the Blackbird district. Zircons from a tuff in the middle unit yielded apparent U-Pb ages of about 1,670-1,700 m.a. (Hahn and Hughes, 1984), but should probably be considered a maximum age due to possible inherited radiogenic lead (Nash and Hahn, 1986). A minor amount of tuffaceous material is present in the lowermost portions of the upper unit of the Yellowjacket.

Mafic tuffs are usually biotite-rich, have high alkali contents, and locally contain high concentrations of iron-carbonate. Original textures have been effectively masked by the coarsely recrystallized fabric imparted to these rocks during metamorphism. Individual tuff beds probably do not have good lateral continuity, but tuffaceous intervals in the middle unit of the Yellowjacket can be mapped continuously over more than 5,000 feet of strike length in the main Blackbird mine area (Nash and Hahn, 1986). Hydrothermal (chemical) sediments are also a constituent of the tuffaceous stratigraphy in the Blackbird district.

Mafic dikes and sills are quite common in the middle unit of the Yellowjacket (Figure 3). The compositions of the mafic dikes and sills is similar to that of mafic tuffs (Nash and Hahn, 1986), and these rocks may represent a possible comagmatic event that occurred along a northwest trending zone in the Yellowjacket basin. Several small occurrences of carbonatitic rock associated with breccias located in the West Fork drainage, and lamproitic diatremes located along Blackbird Creek, may also have been coeval with the mafic magmatism.

Tertiary mafic and felsic dikes were emplaced along fault zones cutting the Yellowjacket rocks in the Blackbird district. Normally the wallrocks show very little or no effects of being intruded by these rocks.

Structure

Much of the structure in the Blackbird district formed during the synsedimentary tectonic activity that was instrumental in developing the Yellowjacket basin. Soft-sediment folds, compaction-faults, dewatering foliations, and retextured beds occur along north-northwest trending "lines" that indicate the probable positions of Yellowjacket basin growth faults (Hughes, 1983). These growth faults and their subsidiary structures probably controlled the location of vents for mafic tuff eruptions and chemical sedimentation that occurred in the Blackbird mine area.

Several large regional faults can be mapped through the Blackbird district (Figure 2). Northwest trending faults that follow the course of Blackbird Creek may represent more recently reactivated zones along old structures that were initiated during evolution of the Yellowjacket basin. Major facies changes across these structures suggest that they were originally normal, down-to-basin growth faults. More recent (Laramide?) offsets imply strike-slip and perhaps scissors-like movements along these faults.

Steep north trending faults bracket the Blackbird mine area, and are responsible for the horst-like block that contains the main mine workings (Figure 2). Eocene age dikes located along some of these north-south faults indicate they formed during the initial stages of Tertiary regional extension. Offset of the garnet isograd by one of the north trending faults indicates a post-metamorphic displacement of at least several thousand feet.

Northeast trending faults of the trans-Challis fault system are present at the boundaries of the Blackbird district. The Panther Creek and Big Deer Creek faults are the largest of these structures. Shorter, en echelon segments of the fault system are also present in the Quartz Creek and Little Deer Creek drainages.

Co-Cu-Au Deposits

The Co-Cu-Au deposits of the Blackbird district can be generally classified as syngenetic, stratabound metal enrichments. They have undergone some selective metal reconcentrations and ore mineral recrystallization during at least two subsequent metamorphic events. The most important type of cobalt deposit is conformable and closely associated with mafic tuffs and hydrothermal sediments within the middle unit of the Yellowjacket Formation. Fine-grained cobaltite and coarser-grained chalcopyrite and pyrite are the principal ore minerals in most lodes. However, several smaller occurrences of siliceous hydrothermal sediments that contain mostly fine-grained cobaltite, with very little pyrite and locally significant chalcopyrite, have only recently been described (Nash and Hahn, 1986). The dominant ore suite of elements for the Blackbird deposits include Co, Cu, Au, Bi, and REE; whereas, strongly associated elements are Fe, As, S, and B (Nash and Hahn, 1986).

Discordant ore lodes are not common in the Blackbird mine area. The only one of any significant size appears to have formed where ore fluids had flooded into a dilation zone along a fault. This fault may have been a conduit for feeding ore fluids into the hydrothermal vents on the basin floor. On the otherhand, the apparent scarcity of discordant ore lodes suggests that feeder zones are neither well developed nor easily preserved in the thick pile of clastic rocks. An alternate hypothesis that may also explain the lack of epigenetic feeder structures calls on the present stratabound ore lodes to have been mechanically transported away from those structures. The association of some ore lodes with layers of retextured, "quick" sediments suggests this to be a viable alternative.

Cobaltite-bearing, tourmaline and quartz cemented breccias are common in the upper 1,000 feet of the lower unit, and the lower 1,000 feet of the middle unit of the Yellowjacket. Similar breccias are also found in the remainder of the middle unit and lowermost portions of the upper unit, but not with the same frequency. The spatial distribution of this type of cobalt deposit differs from the conformable types in the main Blackbird mine area. The vast majority of cobalt-bearing tourmaline breccias have been found in a 1.5 mile wide, eastward trending zone whose axis is about 1.5 to 2.0 miles south of the main mine area.

Many of the breccias either contain some mafic tuffaceous material and probable hydrothermal sediment components, or are very closely associated with Yellowjacket rocks that contain them. Tourmaline and quartz have pervaded into the surrounding sedimentary rocks, especially the sandier layers. Rusty-colored weathered surfaces associated with the outcropping breccias also denote the presence of finely disseminated and fracture-controlled pyrite in their vicinities. The generally very low copper contents in the tourmaline breccias contrasts with the cobalt-copper ores of the Blackbird mine area, and therefore probably forces a different genetic process for their origin.

The cobaltite-bearing tourmaline breccias have very irregular shapes that can vary from thin (inches), discordant vein-like occurrences to thick (tens of feet), conformable lenses. Most are linear and change from conformable to discordant shapes along strike as well as down dip. The matrix to clast ratio is generally high indicating the transport of breccia material was probably in a semi-fluid to plastic medium. Fluidal-textured rocks that encase many breccias point to the rapid evacuation of fluids from the sedimentary pile during breccia formation. They also denote a prelithification origin for the breccias.

The cobalt ores mined at the Haynes-Stellite mine in the early 1900's were extracted from unusually high-grade zones in one of the tourmaline breccia-type deposits.

Metamorphism

A high geothermal gradient accompanying Yellowjacket basin evolution and sedimentation can account for the hydrothermal fluids that transported metals into the basin and formed the Blackbird ore lodes. This process probably occurred during the same metamorphic heating event (1,600 - 1,800 m.a.) that has been recognized in the Archean rocks of southwestern Montana. The anomalous heat-flow in the Yellowjacket basin should have accelerated many of the diagenetic processes, and probably caused some early-stage metamorphic reactions to take place in the Yellowjacket sediments before they reached complete lithification.

Metamorphism that accompanied the emplacement of 1,370 m.a. granitic rocks (Evans and Zartman, 1981) less than three miles from the Blackbird mine has overprinted an earlier, biotite grade, regional metamorphic assemblage. This earlier assemblage may represent the advanced stages of the geothermal metamorphism that occurred during Yellowjacket sedimentation. The garnet isograd from the earlier metamorphic event is shown on Figure 2, with the garnet-bearing rocks occurring north of the isograd.

Regional metamorphism that attended the emplacement of the Idaho Batholith does not appear to have significantly affected the rocks in the Blackbird district. Some overprinted retrograde effects may have resulted during this Late Cretaceous to Tertiary event.

The net effect of all metamorphic events on the rocks in the Blackbird mine area was to change the original sandstones, siltstones, and mudstones into much denser, recrystallized quartzites, siltites, and argillites. Some of the mafic tuff-bearing intervals in the middle unit of the Yellowjacket show a more dramatic change to coarser-grained biotite-garnet-chloritoid granofels (Nash and Hahn, 1986). Recrystallization and annealing of grain boundaries

FIELD TRIP STOPS

The field trip through the Blackbird mining district will begin at the mouth of Blackbird Creek. All but the first two stops will be along the main road leading to the Blackbird mine.

Stop 1

Park at the tailings dam on the West Fork and walk westward along an old mine road for approximately one mile. Bold exposures along the north side of the West Fork of Blackbird Creek drainage are mainly within rocks of the middle unit of the Yellowjacket. Proceed to the first stop which is within the graded argillite and siltite facies of the lower unit. Here we note the style of sedimentation and some of the common sedimentary features observed in this unit. These are examples of the sediments that were probably deposited on a basin plain environment near the outer fringe of a submarine fan complex.

The Conicu prospect is located just uphill (north) from this stop. It represents the first cobalt discovery in the Blackbird district, and was made by John Beliel in 1901. The mineralization is associated with a tourmaline breccia that occurs near the base of the interbedded quartzite and argillite facies of the middle unit. Along the return to the vehicles we will get a closer look at one of these types of cobalt occurrences.

Return to vehicles and proceed up the Ludwig Gulch road.

Stop 2

Examine outcrops of the interbedded quartzite and argillite facies in the middle unit of the Yellowjacket Formation. These exposures are quite typical for most of this facies where it has not been involved in the mineralization processes that occurred in the Blackbird mine area. Note the sedimentary features that are common in this facies, and the apparent "unaltered" look of the rocks. We will see these same rocks at Stop 5 where they will have a somewhat different appearance caused by their closer proximity to the mineralization at the Blackbird mine. The rocks in these outcrops represent clastic sedimentary materials that were deposited in the lower fan subenvironment.

Stop 3

The graded and rippled quartzite facies exposed along the Blackbird mine road in this area represents a storehouse of sedimentary features. Some of the more important features to note here are the apparent energy conditions that prevailed in the basin during deposition of this facies. Note also the abundant display of top-and-bottom indicators in these beds.

The graded and rippled quartzite facies probably represents more energetic pulses of sedimentary materials into the basin than those observed in the interbedded quartzite and argillite facies. The stratigraphic and lateral distribution of these rocks suggests that a major cycle of submarine fan progradation had reached this area only to wane or completely die-out while sedimentation of the interbedded quartzite and argillite resumed. The rocks of the graded and rippled quartzite facies at this stop were probably deposited in the upper portions of the lower fan subenvironment, or distal parts of the mid-fan subenvironment.

The argillaceous component in these rocks probably contains some of the tuffaceous material that is common in the Blackbird mine area. Resedimentation of this fine-grained material probably occurred over broad areas in this part of the basin.

Selected sampling of this material would reveal an anomalous amount of cobalt and copper in these rocks.

Stop 4

Park at the old road that leads to the Haynes-Stellite mine. The adit to the mine is a short distance up the road which was built across a large pile of talus.

The Haynes-Stellite mine is located within rocks of the upper unit of the Yellowjacket. This unit is represented by the massive, laminated quartzite facies which is very well exposed in the outcroppings at this locale. Note the much greater increase in overall bed thicknesses and the dominance of sandy clastic materials over that of the argillaceous component. Wedge-shaped beds can be seen in the outcroppings above the adit, and large-scale wavy bedforms are also present. The massive, laminated quartzite facies was probably deposited in the upper portions of the mid-fan subenvironment.

The cobalt mineralization at the Haynes-Stellite is associated with a tourmaline breccia. The breccia zone also contains a significant amount of mafic tuff that can be identified by its high biotite content and the presence of large white-mica porphyrotopes. The cobaltite in the breccia is very fine-grained and difficult to identify even with a hand-lense. The pink stains on the mineralized rocks is cobalt-bloom, or erythrite — $Co_3(AsO_4)_2 \cdot 8H_2O$.

The view across the canyon to the southeast shows a good exposure of a tourmaline breccia that has been injected upward through the Yellowjacket rocks. Note that the basal portion changes to being conformable with the enclosing rocks. The shape of this breccia body resembles many of the mud-lump diapirs that are found in the Mississippi River delta. Some of the tourmaline breccias may have developed from processes similar to those that generated the mud-lump diapirs.

Stop 5

The interbedded quartzite and argillite facies exposed here is less than a mile from the Blackbird mine. The effects of overprinted mineralization processes is quite apparent in these rocks. Note the pervasive distribution of tourmaline and the silicified nature of the beds, especially the sandier layers. This characteristic is thought to be the result of hydrothermal fluids that moved laterally from their sources in the most porous and permeable units. Several ages of migration are usually indicated; the pervasive type seen here, and younger cross-cutting tourmaline-quartz veinlets. Mafic tuffs are also present and sequences with fluidized textures occur throughout this area.

Stop 6

Park at the reclaimed waste dumps from the Blackbird open pit and walk to the Sunshine prospect.

The Sunshine ore lode was trenched and drilled by Noranda in 1979-81. It represents an unusually high-grade zone of cobalt mineralization that averages about 1.10% Co over a strike length of at least 1,500 feet. The mineralization is associated with a siliceous (cherty) exhalite that is contained within a zone of garnet-rich biotite granofels. Chloritoid metacrysts are also present. These rocks represent beds of siltite and quartzite that contain layers of mafic tuff and hydrothermal sediment components. These rocks have been intensely sheared and foliated just west of this zone due to their proximity to one of the major north trending faults. Note we have crossed the garnet isograd and the cobalt grains in the mineralized rocks are coarser and more easily identified than in the breccia at the Haynes-Stellite mine.

Most of the Sunshine lode has a low copper content that averages about 0.25% Cu. However, gold is a very significant constituent of this zone, and bismuth and rare earth elements are also abundantly present. Tourmaline is common in some very thin layers associated with the mineralized zone. Proximity to a hydrothermal vent(s) may be indicated by the high cobalt and gold content of this lode, and the apparent lateral zoning of Co:Cu (Nash and Hahn, 1986).

Walk to the edge of the Blackbird open pit and take a birds-eye view of the mineralized Yellowjacket stratigraphy. Return to the vehicles and drive down to the entrance to the open pit.

Stop 7

Care should be exercised at this stop not to climb on the loose and unstable walls of the open pit.

The Blackbird open pit was excavated in the late 1950's on the Blacktail ore zone which consists of a series on en echelon lodes hosted by rocks in the interbedded quartzite and argillite facies of the middle unit of the Yellowjacket. The gradational contact with the overlying graded and rippled quartzite facies passes through the pit area.

The mineralization in the open pit is typical of many of the ore lodes in the Blackbird mine. The individual lodes are largely conformable with the enclosing beds of quartzite and argillaceous siltite. Note that the clastic sedimentary rocks exhibit abundant soft-sediment deformation structures and that numerous layers of mafic tuff are present in the sequence. Some portions of the ore lodes have brecciated textures that probably formed when the muddy, tuffaceous layers were being compacted under the weight of the overlying sandier sequences.

The Blackbird open pit is located along the trace of the garnet isograd. Note that much of the ore material consists of coarse-grained pyrite, chalcopyrite, and some pyrrhotite. Cobaltite and arsenopyrite generally are finer-grained than the other ore minerals. Even though metamorphism has modified the silicate mineralogy, its effect on the ore minerals was mainly recrystallization and only local-scale remobilization of some phases. Overall chemical compositions of the ores are probably close to that of the pre-metamorphic mineral zones.

Return to vehicles. End of this field trip.

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TOBACCO ROOT GEOLOGICAL SOCIETY TOUR

of the

BEARTRACK PROJECT
Mackinaw Mining District
Lemhi County, Idaho

MERIDIAN GOLD COMPANY
FMC GOLD COMPANY

Ed Bartels
Ian Douglas
Gary Van Huffel
Steve Busby

ABSTRACT

and

ROAD LOG

August 17th and 18th, 1990

ABSTRACT

The Beartrack claim block area is dominated by two Proterozoic units. The Yellowjacket Formation (Yy) is a clastic metasedimentary which is intruded by a quartz monzonite pluton (Yqmp), also of Proterozoic age. An alkalic granite to syenite stock of Ordovician age outcrops on the south west side of the property. Tertiary flows, lithic tuffs, and dikes cover the older rocks along a structurally controlled basin paralleling Napias Creek. At least two ages of glacial deposits mantle the bedrock in the Napias Creek drainage south of Leesburg and in the higher valley west of Napias Creek.

All of the Proterozoic sedimentary rocks have undergone a regional metamorphic event prior to the intrusion of the 1370 M.Y. porphyritic quartz monzonite. The sediments have been raised to biotite grade, greenschist facies. Broad open folding of the sediments accompanied this event.

Five different fault sets are observed in the study area: northwest, eastwest, northeast, thrust faults, and north-south. Northwest and east-west faults are Proterozoic growth faults reactivated in the Cretaceous and Teritary. Cretaceous low angle faults thrust migeoclinal sediments eastward over Archean basement rocks during the compressional phase of Laramide tectonics. A klippe of this thrust sheet is exposed on the property. Northeast faults are part of a major, first order crustal shear that may have been active since the Precambrian. The shear zone is called the trans-Challis trend in Idaho. The Panther Creek Fault is a major fault segment within this zone. North-south faults control outcrop and distribution of Cenozoic rocks in the Leesburg Basin.

The Beartrack gold mineralization is an epigenetic, structure-controlled deposit hosted by middle Proterozoic metaclastic and alkalic intrusive rocks. Alteration and mineralization is primarily restricted to lower plate Proterozoic rocks along structures. The northeast-trending faults are the primary control of mineralization. Gold, with As, Sb, Hg, Mo, Pb and minor Cu was deposited along a dilatant segment of the northeast-trending Panther Creek Fault. The zone is characterized by an extensive zone of stockwork veining within a zone of quartz-sericite-pyrite alteration. Zones of silica flooding and silicified intrusive hydrobreccias also occur within the ore zone. Hydrothermal k-spar and sericite has been dated at 68 M.Y. (Evans and Snee, USGS) from Ar/Ar systemics, suggesting a genetic relation to Idaho Batholith intrusion. A regional geophysical high near Camp Creek suggests an intrusive at depth, a potential magmatic source for mineralizing fluids and/or a heat source for convecting connate fluids.

The Beartrack mineralization was re-discovered in 1986 by Bob Perry, then a Canyon Resources Corporation geologist. Meridian Gold Company entered into a joint venture agreement with Canyon/Minex in July of 1987. Over 400 rotary holes and 52 core holes have been completed through 1989. The indicated geologic reserve reported to date is approximately 45 million tons containing, depending on cutoff grade used, 1.0 to 2.1 million ounces of gold.

BEARTRACK PROJECT TOUR

ROAD LOG

The Beartrack Project road log begins at the Napias Cr. bridge on USFS Rd. 21 and extends to the Arnett Cr. Rd. (mile 3.7). The Arnett Cr. Project log begins at this point, going up Arnett Cr. and returning via the same route. The Beartrack log resumes at the Arnett Cr. - Napias Cr. road junction, (mile 3.7) on the Beartrack log. Refer to Figure 1 for geology and road log stops.

MILE	
0.0	**START** Napias Cr. bridge on USFS Rd. 21. Turn right on USFS Rd. 242 going up Napias Cr.
0.7	**STOP 1** Proterozoic Y quartz monzonite porphry. (Yqmp). Orbicular, coarsely porphryitic, quartz monzonite intrusive. Relatively fresh at this outcrop. This unit hosts the North Orebody at Beartrack.
1.1	Dacite lithic crystal tuff (Tdlt) moderately welded. Angular fragments of dacitic tuff and minor quartzite in a welded glass matrix. Approximately 10% quartz and feldspar crystals. This unit forms many of the steep walls along the east side of Napias Creek between Pony and Phelan Creeks. Volcanics within the Napias Cr. basin generally strike north and dip 10to 30 degrees east.
1.6	Rhyolite lithic crystal tuff (Trlt.) unwelded, crystal poor, glassy. Approximately 20 % of rock volume consists of sericitized Proterozoic Yellowjacket Formation (Yy). Lowermost volcanic unit exposed at Beartrack.
1.8	Note dredge tailing, to right. Extensive placering activity from this point all the way to Leesburg.
2.1	Passing klippe of Proterozoic Big Creek Formation which is thrust over Trlt. Left side of road.
2.6	USFS historic ranger station. Located to left. Active during the early 1900's.
2.85	Trlt to left and Tdlt to right on east side of Napias Creek.
3.3	Note old placer water ditch. Extensive water diversions throughout basin.

3.5 Trlt/Qg contact to left. Note spire of Tdlt outcropping on east side of Napias Cr. Quaternary geology in basin is complex. At least two periods of glaciation, possibly three. Major ice loboes followed Camp, Moose, Hornet, Rabbit, Pony and Missouri Creeks, and represent the older period of glaciation (Qg,older). The till is almost entirely quartzite (Yy). Younger deposits (Qg) are composed of stratified boulders of quartz monzonite and quartzite with volcanic ash and peat interbeds. No distinction is made between glacial deposits and alluvial deposits on Figure 1.

3.55 Arnett Creek bridge

3.70 Arnett Creek road. Arnett Cr. Project tour starts here.

5.35 Proposed Beartrack Project heap leach pad site to left on bench above road.

6.05 Leesburg cemetery up slope to the left.

6.2 Weather station installed by Meridian to gather baseline meteorologic data for EIS.

6.25 **Stop 2** Leesburg town site. Leesburg is on the National Register of historic sites. A brief presentation of regional and project geology will take place here.

6.40 Proposed mine office facility to right.

6.60 Proposed crusher site.

6.70 Crossing over proposed South orebody south east highwall. Lath to right marks eventual pit highwall. Centerline of pit will parallel the Panther Creek fault approximately 300 feet to the left.

6.85 Altered Proterozoic quartz monzonite prophry (Yqmp) exposed in drill pad to left.

6.90 **Stop 3.** Walking tour through South orebody. Vans will meet group up the road. Will see altered, mylonitized, Yqmp; altered Yellowjacket (Yy) and intrusive hydrobreccia.

7.25 Approximate northern highwall of South pit. Proposed heap leach pit will be approximately 2100 feet long, 1000 feet wide and 750 feet deep upon completion.

7.35	Trace of Panther Cr. fault is evidenced by break in slope seen to left. Exposure over the South orebody was very poor; surface mapping and sampling was limited to float.
7.75	Kirkpatrick or Gold Dust Mine area down slope to right. Mine produced very limited ore from bull quartz vein occurring along the fault. The bull quartz vien is largely barren of gold within the South orebody. Active exploration/development drilling is underway in the Middle area to your left.
7.85	Crossing Panther Creek fault going from Yqmp into hanging wall Yy.
7.90	Crossing Wards Gl. Placer gold was first discovered in 1866 at the mouth of Ward's Gl. 1 1/2 miles downstream. South end of the North pit highwall will daylight in this area.
8.0	Crossing Panther Cr. fault, passing from Yy into Yqmp. Fault passes through the prospect pit to left.
8.05	Driving over southernmost Canyon/Minex discovery holes in North orebody.
8.20	**Stop 4.** Walking tour of a portion of the North orebody. Will see progressively stronger alteration of Yqmp including sericitization of feldspar, graded destruction of biotite, mylontization, Panther Cr. fault and Yy hanging wall alteration.
8.25	Definition drilling in progress to right to determine southeast limit of mineralization.
8.45	Monument hill. This is the widest portion of the North area mineralization. The proposed pit is approximately 800 feet wide at this point. The trace of the Panther Cr. fault is approximately 100 feet to the left.
8.85	The first Canyon/Minex discovery hole was drilled at this location.
9.0	Northernmost limit of proposed heap leach pit. The Panther Cr. fault continues on to the northeast.
9.3	Moose Creek bridge
9.4	USFS Rd. 300. End of road log.

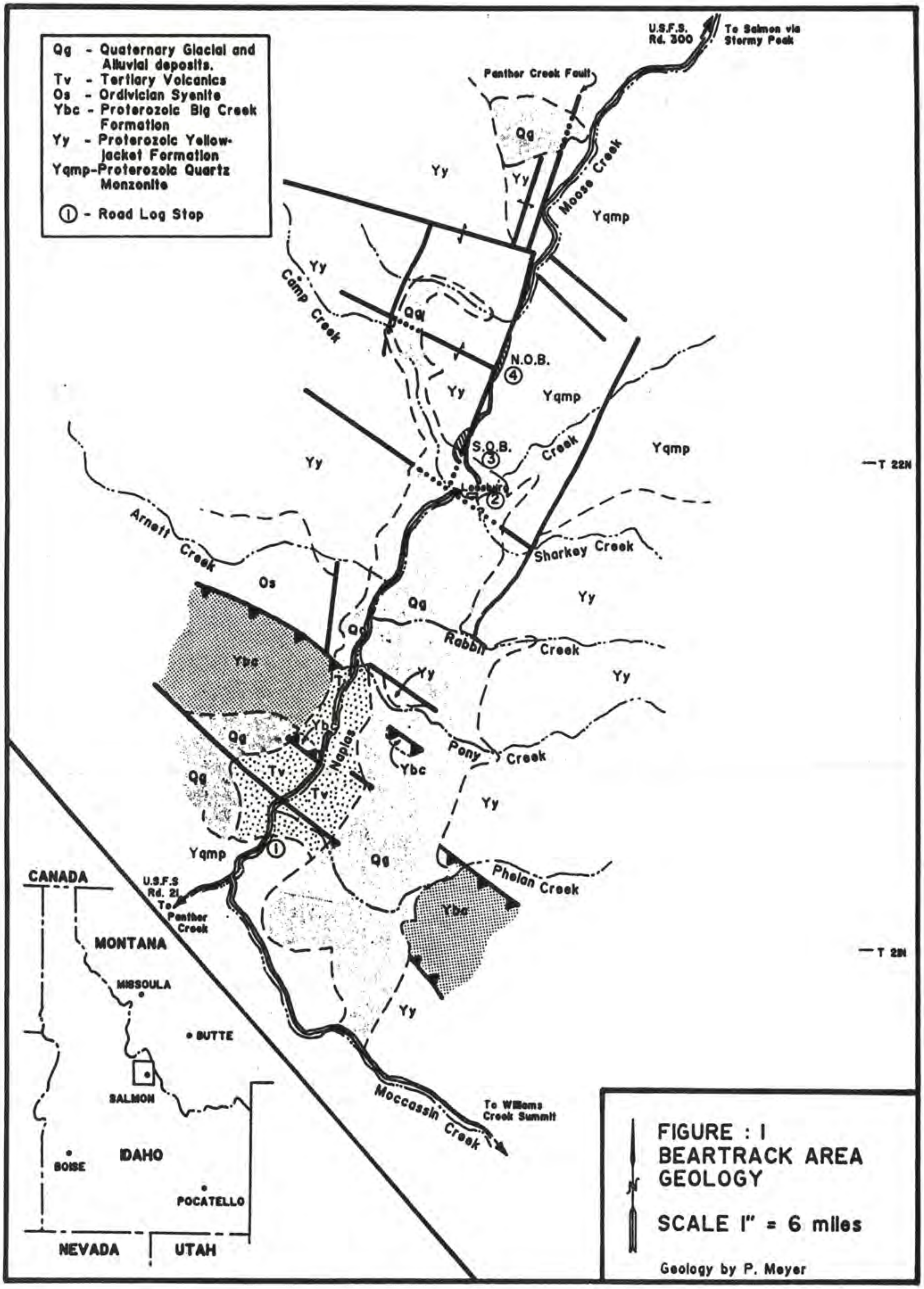

FIGURE : I
BEARTRACK AREA
GEOLOGY

SCALE I" = 6 miles

Geology by P. Meyer

Geology and mineralization of the Arnett Creek Project, Lemhi County, Idaho.

KEITH REEVES Arnett Creek Joint Venture, Salmon, ID 83467.
RON CASTAGNE Arnett Creek Joint Venture, Salmon, ID 83467.

The Arnett Creek Project is a gold exploration project encompassing 20 square miles within the Mackinaw Mining District of Lemhi County, Idaho. The property is currently controlled by the Arnett Creek Mining Company, an unincorporated joint venture between American Gold Resources Corporation and Meridian Gold Company. Meridian Gold Company is the designated operator.

The occurance of widespread anomalous and ore-grade gold values has led to the definition of several exploration targets on the project. These areas are in various stages of evaluation ranging from the Haidee area, where extensive drilling activity has occurred, to the Jureano Basin, where limited stream sediment sampling has indicated anomalous gold values.

PAST PRODUCTION

Historic producing lode gold mines on the Arnett Creek project included the Haidee Mine, the Italian Mine, the Little Chief Mine and the Thomson (Dutch John) Mine. Additionally, most of the stream drainages within the Arnett Creek watershed have been heavily placered.

The Italian Mine, located in 1892, was one of the major lode producers within the Mackinaw District, having produced $175,000 (approx. 8,400 ounces) of gold by 1913 (Umpleby, 1913). Gold was produced from a fissure vein, however, conflicting reports exist concerning the orientation of the vein. Umpleby (1913) characterized the mineralization at the Italian as being a fissure vein contained within a 5000 foot long, 300 foot wide fracture zone in the host biotite granite. This fracture zone was reported to strike north-south and dip 45 degrees to the west (Umpleby, 1913). An earlier report indicates that the vein was striking northwest and dipping to the northeast (Anonymous, 1892). Umpleby (1913) also reported that the metallized granite within the fracture zone contained stockwork style mineralization and assayed between two and four dollars (0.1 to 0.2 ounces) per ton. Bell (1903) reported that a 10 stamp mill was operating at the Italian Mine. In 1909-1910 a 30 stamp mill was constructed. By 1921, the Italian Mine was idle (Campbell, 1921).

The Haidee Mine developed a four foot wide fissure vein striking north-south, dipping 57 degrees to the west (Umpleby, 1913), and assaying up to 20 dollars (0.96 ounces) per ton (Bell, 1903). The vein was discovered in a flat, swampy bench area that precluded easy development by sinking a shaft, therefore, a 3,000 foot tunnel was planned to permit development (Bell, 1903). The development tunnel was driven 1,100 feet but was lost to poor ground and excessive caving (Umpleby, 1913). This development tunnel intersected a 90 foot zone that assayed four dollars (0.2 ounces) per ton (Bell, 1904). In addition to the underground development, an eluvial placer deposit adjacent to the Haidee mill was worked successfully each spring (Umpleby, 1913).

The Thomson Mine probably corresponds to and area now referred to as the West Italian Pit. Bell (1903) reports that two to four dollar (0.1 to 0.2 ounce) per ton material was being mined from a quarry at the Thomson Mine. This may be the area where the 4,000 ton test sample of the metallized granite reported by Umpleby (1913) was taken in 1910. This test sample was said to have milled an average of 2.25 dollars (0.1 ounces) per ton.

Very few historical references to the Little Chief Mine have been found. Bell (1904) reports that the Little Chief Mine was developing a two and one-half foot pay streak of fine milling ore.

GEOLOGY

Outcrop is relatively scarce on the property. An estimated 75 to 80% of the property is covered by overburden, thus making mapping interpretation difficult. The central portion of the property is dominated by the Arnett Creek stock. The stock is an enlongate body with a WNW trend intruded into the Proterozoic Y Yellowjacket Formation. The Yellowjacket Formation is unconformably overlain by the Proterozoic Y Big Creek (Hoodoo) Formation.

Stratigraphic Relations

Proterozoic Y Sediments

Yellowjacket Formation

The Yellowjacket Formation is represented on the property by a gray to greenish gray, fine-grained, argillaceous quartzite. Parallel laminations and occasional ripple cross-laminations are the main sedimentary structures observed within this argillaceous quartzite. The contact of the Yellowjacket Formation with the overlying Big Creek Formation is not exposed on the property, however, several previous studies suggest that the contact is a thrust fault (Lopez, 1981; Evans and Zartman, 1988).

Big Creek Formation

The Big Creek Formation on the Arnett Creek project is typified by a white to light gray, medium-grained quartzite. Ripple marks and cross-bedding constitute the observed sedimentary structures within this unit.

Intrusive Rocks

Alkali-feldspar granite

Approximately two-thirds of the Arnett Creek stock is composed of a light gray, medium- to coarse-grained, porphyritic, hypidiomorphic granular alkali-feldspar granite. The coarse-grained granite typically contains 40 to 50% subhedral, perthitic alkali feldspars, 15 to 20% anhedral plagioclase, 20 to 30% anhedral quartz, 2 to 8% anhedral to subhedral biotite, and 1 to 3% opaque minerals (dominantly magnetite or ilmenite). Large (1 to 4 cm) porphyryblasts of perthitic alkali feldspars rimmed by sodic plagioclase (rapakivi texture) are relatively common, indicating that this alkali-feldspar granite may correlate with the Proterozoic Y granite of the Leesburg / Panther Creek area. The medium-grained equivalent of this alkali-feldspar granite possesses a similar mineralogy, however, the phenochrysts exhibit what could be described as an incipient rapakivi texture. In the medium-grained granite, the plagioclase-rimmed, alkali feldspar phenocrysts are smaller (0.5 to 1 cm) and often less rounded. Complete gradation between these two textural types can be found on the property.

Alkali-feldspar syenite

The western one-third of the Arnett Creek stock is composed of a medium gray to dark gray, medium- to coarse-grained, hypidiomorphic granular alkali-feldspar syenite. The syenite is composed dominantly of equigranular alkali feldspars, containing greater than 85% subhedral, perthitic alkali feldspars, less than 5% anhedral quartz, and 5 to 10% anhedral to subhedral biotite. U-Th-Pb analyses of zircon from this syenite indicates an Ordovician (492 m.y.) age (Evans and Zartmen, 1988).

Granite

At the Italian Mine, a white to light gray, fine-grained, allotriomorphic granular granite intrudes the alkali-feldspar granite. The Italian granite is composed of 55 to 60% anhedral, perthitic alkali feldspar, 35 to 40% anhedral quartz, 1 to 2% anhedral biotite, and 2 to 3% opaque minerals (magnetite or ilmenite).

Diorite

Fine-grained porphyritic mafic dikes composed dominantly of anhedral biotite with 15% 0.5cm lath-shaped plagioclase phenocrysts intrude into both the alkali-feldspar granite and the Italian granite.

Age Relationships

The age relationships of the intrusive rocks of the Leesburg / Arnett Creek area have been the subject of much debate (see, for example, Bennett, 1980 and Evans and Zartman, 1988). Most discussions of the age of the Arnett Creek stock make the assumption that the stock is a single intrusive body. On the basis of field relationships and textural evidence, we believe that the stock is a composite intrusive body formed by three or more separate intrusive events. The main body of the stock is an alkali-feldspar granite with many petrographic similarities to the Proterozoic Y granites of the Leesburg / Panther Creek area. No absolute age dates are available for this granite. On the eastern end of the stock, this alkali-feldspar granite has been intruded by an alkali-feldspar syenite. U-Th-Pb zircon analyses indicate a lower Paleozoic (Ordovician) age for the syenite (Evans and Zartman, 1988). At the Italian Mine, a younger equigranular granite intrudes into the alkali-feldspar granite. Additionally, numerous mafic dikes intrude into the alkali-feldspar granite.

Structure

Although the contacts between the intrusives and the metasediments could be structural in nature, the widespread occurrence of assimilated fragments of Yellowjacket fragments within the alkali-feldspar granite suggests that intrusive contacts are most likely. To the south of Arnett Creek, a thrust fault contact between the Yellowjacket Formation and the Big Creek Formation has been suggested (Evans and Zartman, 1988), however, the lack of any exposure of this contact on the property has prevented confirmation of this theory.

The WNW trending Arnett Creek fault system is the most obvious structural trend on the property. This system is composed of two linear features. The Arnett Creek drainage represents one of the linear features and a slope break 500 to 2000 feet south of Arnett Creek represents the other. Intense brecciation within the Big Creek Formation on the western edge of the property marks the location of the convergence of these two features. Many of the placer deposits on the project are contained between these two linears, however, the relationship between mineralization and the Arnett Creek fault remains uncertain. Additionally, the relationship or the placer activity to bedrock Au values is uncertain.

Mineralization at Haidee, Little Chief, Little Chief Extension, and Italian all exhibit some relationship to north-trending fractures. This association of mineralization with north-trending fracture systems probably reflects an influence of the Trans-Challis fault system upon mineralization controls.

Alteration and Mineralization

Sericitic alteration of the host intrusive's feldspars is prevalent across much of the property. In the initial stages, this sericitization preferentially attacks the plagioclase rim of the zoned feldspars. With progressive alteration, the host rock feldspar and biotite are converted to a pale to dark green sericite. Quartz-pyrite veining is common throughout the property. At Haidee, limited wall rock silicification comprises a portion of the wall rock alteration. At Little Chief, Little Chief Extension and Italian wall rock silicification is more extensive than at Haidee.

Mineralization is most closely associated with quartz veining, wall rock silicification, disseminated iron oxides and iron oxide fracture fillings. At Haidee, visual keys to mineralization are often subtle. At Little Chief, Little Chief Extension and, to a lesser degree, Italian, mineralized zones are more readily apparent. At these three areas, the stronger wall rock silicification provides a better visual key to mineralization.

ROAD LOG

0.0 Begin log at the Junction of the Napias Creek Road and the Arnett Creek Road.

0.3 Passing over the approximate contact between the Tertiary Challis Volcanics and the Proterzoic Y Yellowjacket Formation.

0.4 The iron-stained, weakly bleached outcrop of Yellowjacket on the right carries weakly anomalous Au values and anomalous As values. Arnett Creek is below to the left. Note that placer activity has occurred along the length of Arnett Creek.

1.2 Passing over the approximate contact between the Proterzoic Y Yellowjacket Formation and the Ordovician alkali-feldspar syenite.

1.25 STOP ONE: This outcrop of the Ordovician alkali-feldspar syenite is reportedly the location for the U-Th-Pb age date of 492 m.y.

1.35 To the left across Arnett Creek, bench placering has cut to the underlying syenite outcrop. Syenite also crops out in road cut. The bench placers on the opposite bank continue for the next 0.3 mi.

1.85 Rapps Creek Bridge. Approximate boundary of the Arnett Creek Project. Junction of Arnett Creek and Rapps Creek is 200 feet downstream. Arnett Creek passes to the south side of the ridge. The road follows Rapps Creek to the north of the ridge for a short distance. The Rapps Creek placers are on the right for the next 0.5 mi.

2.3 Diorite crops out in Rapps Creek by the cabin.

2.4 Inlier of Yellowjacket Formation in the road cut. Syenite crops out in Rapps Creek below to the right.

2.95 A shallow trench dug in the road at this point exposed the alkali-feldspar granite. A quartz vein in the trench produced euhedral smokey quartz crystals up to 8 cm in diameter and up to 10 cm in length.

3.1 The road to the right leads to the North Italian target area. Quartz-pyrite veining in the North Italian area is hosted by a weakly argillized, weakly iron stained granite. Both the granite and the quartz-pyrite veins are mineralized.

3.4 Periphery of the Italian Mine. Granite dominantes the dumps of the numerous pits and adits above and below the road.

3.5 STOP TWO: The main dumps of the Italian Mine are on both sides of the road. The main haulage adit is on top of the dumps to the right. Two large caved stopes are 100 feet above the dumps.

The remains of the Italian Mill (circa 1909-1910) are below the road to the left. Granite and diorite are the dominant rock types on the dumps. Historic and modern placer activity parallel Arnett Creek for the next 0.6 mi. Silicification, quartz veining, and disseminated iron oxides accompany mineralization at the Italian Mine.

3.8 West Italian Pit. Younger granite intrudes the Proterozoic granite. The granite is strongly silicified parallel to the joint planes. This is possibly on of two historic mines, either the Thomson Mine or the Old Dutch John Mine.

3.9 Davis Gulch enters from the right (north). For the next 0.7 miles abundant prospect pits and adits explore the hillside. Exploration by the 'old-timers' using hydraulic excavation is evident. Ditches above supplied water which was used to erode the hillside to expose bedrock. This area has abundant quartz veining with anomalous Au values.

4.1 The South Arnett Creek target area is directly across the creek to the south (clever name). This target area is a portion of the larger Porcupine target area.

4.3 The Goff Brother's placer is to the left across the creek. This placer was worked well into the 1920's.

4.85 The placer operation to the right is concentrated near the junction of Arnett Creek and Sheenan Gulch. Sheenan Gulch was once the site of a substantial reservoir providing water to the placer activities along Arnett Creek. Sheenan Gulch was also the site of intense placer activity.

4.9 Haidee Junction. The access to the Haidee Mine area climbs through an old eluvial placer cut. Watch for rapakivi textures in the granite.

5.2 Switchback to the Haidee Mill site and the Haidee hydraulic placers. The Lower Haidee Adit, driven to develop the Haidee vein, is approximately 300 feet below the road.

5.45 STOP THREE: Haidee Mine. A long stop is planned here (lunch?). Mineralization at Haidee is widespread. Prevailing fractures trend NNE, NE and WNW. Dips are variable throughout the area. Sericitization, argillization, silicification, disseminated iron oxides, and quartz-pyrite veining accompany mineralization. Dumps and mill site to the left. Caved adit to the left.

5.5 On trenches excavated in 1989. Caved stope to the right. Directly over the 'Haidee Main Zone.'

5.55 Haidee Shaft site to the left.

6.4 Haidee Junction again.

6.5 Road to the left accesses Sheenan Gulch and the Porcupine target area. Mineralization at Porcupine is typified by thin quartz-iron oxide filled fractures. Strong sericitization and moderate silicification accompany mineralization. The Porcupine target area is located along a major WNW trending lineament that probably represents the Arnett Creek fault.

6.88 STOP FOUR: Little Chief Mine. The hillside cut above to the right extends up to the Haidee Mine. In the underground workings a series of quartz veins striking NNW and dipping 30 to 40 degrees to the west are mineralized. Later faulting with a WNW trend and near vertical dips has associated quartz-biotite veining.

6.9 WALK FROM STOP FOUR: Bottom of Little Chief Extension access. Intense silicification and associated quartz-pyrite veining (now oxidized) in the host granite are associated with a NNW striking and 50 degree east dipping fracture system.

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DESCRIPTIVE ROAD LOG
Northeastern Leesburg Basin - Bobcat Project Area

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Starting at the junction of USFS 061 and USFS 023

(0.0) - (1.7) You are currently following the general trace of the main (Trans-Challis) Panther Creek fault as it passes through the Middle Proterozoic granitic rocks. To the east, on Racetrack Meadows, the upper reaches of the Moose Creek placer activity can be occasionally glimpsed. Iron stained and sheared roadcrops along this stretch of the route are indicative of the extensive splay faulting off the major zone of movement.

(1.7) Turn north on to USFS 406 past the Moose Creek Diversion. You will note the extensive placer tailings along the route. Early miners diverted the course of the Moose Creek drainage into Dump Creek in order to obtain additional water for placer mining. The extensive erosion from diversion of the larger stream created a siltation problem in the Salmon River to the north. To the immediate west of you is the man made replacement for the original stream channel, constructed in the 1970's.

(1.7) - (3.2) After crossing the cattle guard continue on to the road fork and enter right hand fork of the road. To the east is the Dump Creek volcanic block basin and to the west is Precambrian granite. The Dump and Moose Creek placers produced a combined total of over 200,000 ounces of gold. This work was carried out using hand labour and latter by means of hydraulic mining.

(3.2) Stop 1 Hydraulic pit and slump area. The large pit in front of you was initially created by hydraulic monitors directing high pressure water at the base of this area. Continued down cutting by the diverted water course further weakened the toe of the area and created major slumping. The rocks you see exposed are Challis age volcanics, mainly of dacitic composition. This is the margin of the dump Creek Block basin with its boundary defined by the Panther Creek fault to the west. Mineralization contained along the face of the outcrop is mainly base metal with only minor gold and silver. It appears to be the product of a late stage epithermal system of late Eocene age. Further down Dump Creek, siliceous sinters and acid sulfate alteration can be seen at a former uranium prospect. Other mineralization related to this Challis volcanic event occurs in the switchbacks to your northeast.

The roads to the northeast, on Napoleon ridge, were constructed to allow drill access to the Bobcat porphyry copper molybdenum system. This Eocene, 49 my, copper occurrence contains 60 million tons of 0.40% Cu. as indicated by Cominco drilling in 1980. Drill logs suggest that a sheeted dike swarm channelling along northeast and northwest zones of structural weakness is host to two phases of mineralization and at least five phases of intrusion. Precious metals values are typical for those found in granodiorite hosted copper-molybdenum systems with gold between 0.002 - 0.010 opt and silver from 0.01 - 1.9 opt. Copper values from drilling returned 0.37% over 326 ft., including 111 ft. @ 0.56, with associated molybdenum.

Due east, where the road crosses the saddle on Napoleon Ridge, is a large shear zone in Precambrian quartzite. This area contains, an open ended soil geochemical anomaly for gold and lead that is 900 feet wide and 4000 feet long. The 1989 drilling program by the Corona/Formation Joint Venture resulted in one hole at the margin of this zone which returned a grade of 0.03 opt Au. from surface to 100 ft. in depth. Age dating of the sericite from this drill hole indicated an age 69 my, similar to those received for Beartrack. As with other Cretaceous age mineralization in the area, surface leaching, to a depth of 15 feet was noted. Follow up work on this discovery will proceed this year by the Gold Fields/Corona/Formation JV.

(3.2) - (4.7) Return to main road.

(4.7) - (6.2) You are currently passing through gulch placers derived from Jackass Ridge and are crossing the buried contact between Precambrian granites and quartzites.

(6.2) - (7.8) The roadcrops on the switchbacks you are transiting now are grey fine grained arkosic sandy siltstones of the Lemhi group. As you approach the crest of the Salmon River Mountains the ridge to your north is composed of metamorphosed and mylonitized Yellowjacket Formation. The structural relationship on the ridge appears to indicate an older over younger thrusting, presumed to be related to the Medicine Lodge thrust. However, metamorphic alteration of the Yellowjacket from this location makes stratigraphic correlation difficult, relict textures, where present, suggest that this sequence is composed of upper middle Yellowjacket volcanic sediments. A northeasterly trending shear zone beneath the thrust has been extensively silicified.
Precious metals mineralization from this zone is contained in, mylonite replacement, veins and quartz stockwork veining.

(7.8) Turn at ridge crest on to USFS 020

(7.8) - (7.9) Proceed up the road to the tailings pile on the left hand side of the road.

(7.9) Stop 2 Sims Mine

This is the most easterly workings of the Sims Mine which has produced over 4,000 ounces of gold from shallow workings spread all along the slope to the west. The host rock at this point is thrust Yellowjacket meta-volcanics located above the shear. Field relationships from this location indicate that mineralization was penecontemporaneous with thrusting, which correlates well with the general timing assumed for the Medicine Lodge thrusting and age dates from Beartrack and Napoleon ridge mineralized zones. Initial 1988 drilling to test the potential for thrust hosted mineralization showed that away from the deep seated fault structure little ore grade mineralization has been emplaced. Gold bearing materials are confined to large quartz veins carrying visible free gold with grades of between 1 - 5 opt. Sampling and mapping of old trenches indicates that the mineralized shear varies in width from 50 - 100 feet along a strike of 3500 feet. Airborne geophysical surveys have suggested the presence of a buried magnetically positive intrusive to the north, between this target and the Lange Mine area.

You will note that all of the workings, including several adits, some over 500 feet in length, are near surface such that their collapse has left long trench like profiles on the hill side. It appears from old exploration that the highest grade materials were primarily found in vertical veining at the thrust to shear contact. Since this contact is primarily dip slope, the old timers proceeded in roof level drifts to mine coarse gold.

(7.9) - (8.0) Return to main Stormy Peak Road, USFS 023, turn left

(8.0) - (11.5) As you descend to the southeast on USFS 023, on the east is the Bird Creek Project area of American Gold Resources. In addition to small historic gold mines in this area, extensions of several of the splays of the Panther Creek fault have been noted. Work is currently in progress to determine the bulk minable potential of the targets contained within these claims.

(11.5) Crossing Wallace Creek and entering Formation Capitals Wallace Creek Project area.

On the hill to the south and it's basin beyond, the Tom Bigby shear has been extensively placered. This shear contains alteration for over 1000 feet of width near its junction with the main Sharkey Creek fault. Mineralization occurs internal to this zone as large scale pods of from 50 - 100 feet in width with strikes varying from 700 - 1500 feet. Grades range from 0.01 to 0.15 opt gold within the pods. The structural zone with the greatest promise occurs near the junction of the Big Vein and the Tom Bigby shear, where former trenching and road cuts indicate a width of 200 feet with a strike of over 2500 feet.

(11.7) - (11.8) Return to Stormy Peak road and turn right.

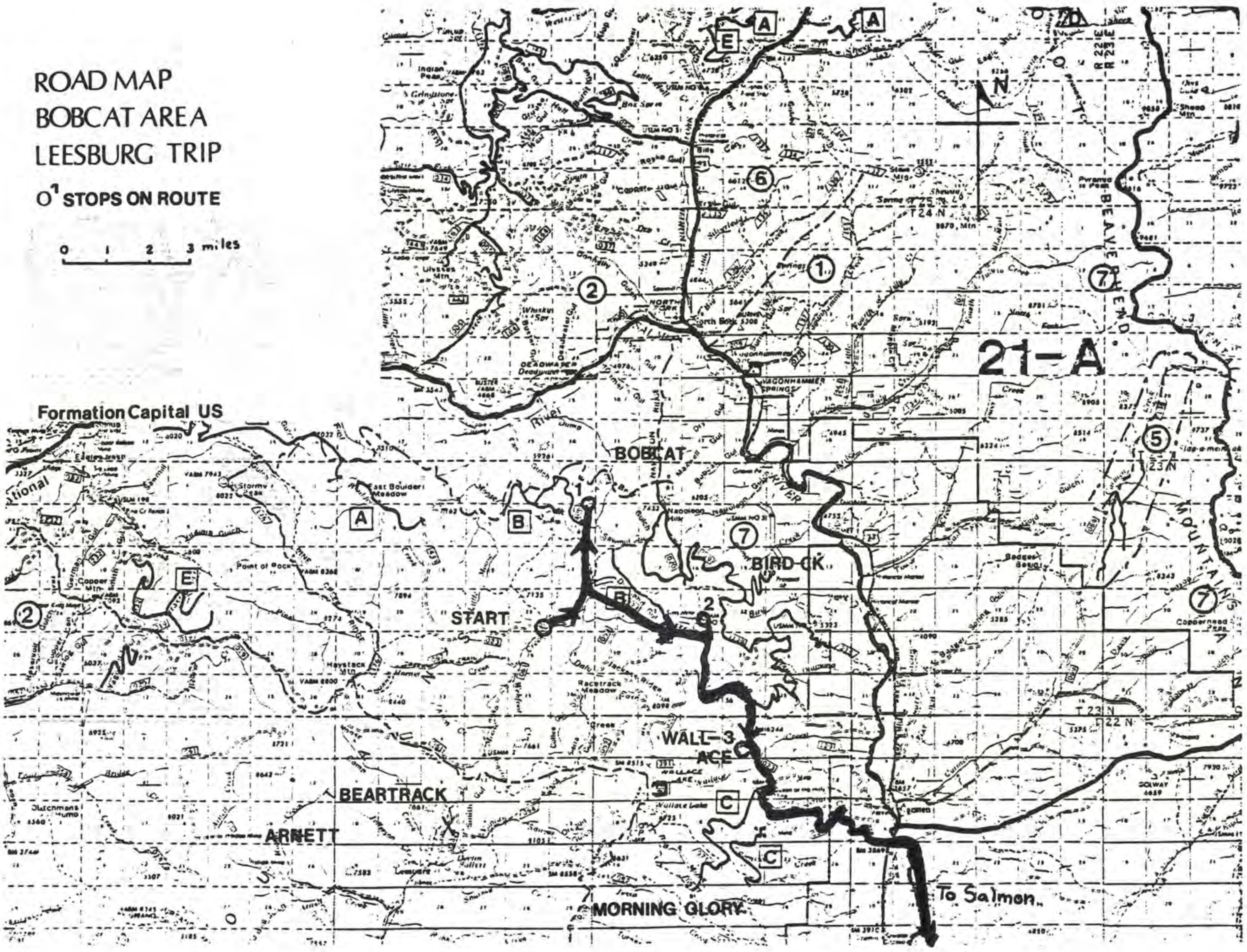
(11.8) - (14.9) As you travel along USFS 023 you will note placer tailings, water ditches, prospect pits and other evidence of historical small scale mining activity. To the east are a group of small prospects called the Tendoy group.

(14.9) You are now passing the workings of the Queen of the Hill Mine which lies on the north side of the Canyon. The Silver Queen shaft's dump is highly visible. Work began on this mine in 1892 from the upper shaft and progressed until 1900 when an adit at the 400 foot level was driven 1500 feet to intersect the original workings. From 1902 - 1914 the mine continued to produce at an annual rate of 15,000 tons per year with an average head grade of 0.25 opt gold. After that period the mine has been only intermittently active with short periods of small production. Estimates of total production range from 25,000 to 75,000 ounces of gold, with the upper end of the scale appearing the most likely from stope evidence. Three small 10 - 20 foot wide northeast shears are exposed in the workings and contain mylonite replacement, quartz veining and quartz stockworks. Formation's work to date has defined a potential for intersection of these structures at a depth of 100 feet below the level of the present workings.

(14.9) - (17.1) To the south are the workings of the Silver Bromide claims. For those interested in mining machinery a visit to the mill on this property is a must. Ray Smith has built a gravity separation mill which uses Wifely table batteries to separate coarse gold from ore, all from scratch over the last 10 years. With a 2 ton per day capacity and a hand built crusher, this plant was drawn from designs current in the late 1800's.

(17.1) - (18.1) Passing the turn off for the Silver Bromide and continuing downhill, the complete Vista of the Beaverhead Mountains is displayed. You will cross the contact between the Precambrian and the Miocene Carmen Creek conglomerates. These sediments are brilliant red.

(18.1) Off the hill and back to town, via the fairground. At the junction with Highway 93 turn south for Salmon.



ROAD MAP
BOBCAT AREA
LEESBURG TRIP

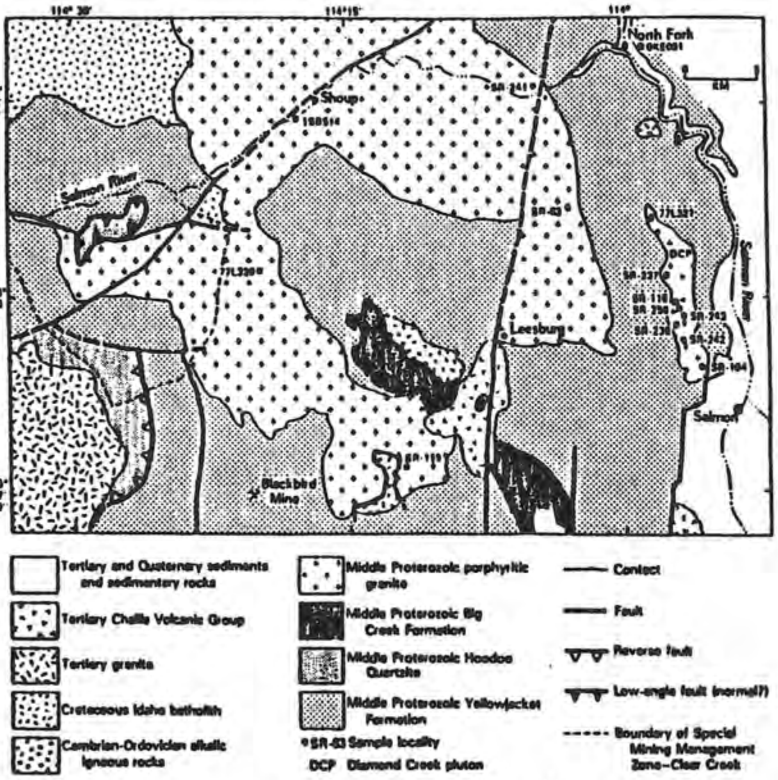


Figure 1. Simplified geologic map of part of the Salmon River Mountains between Salmon, Idaho, and the Middle Fork of the Salmon River (located immediately west of the area shown). The area mapped by Evans (1981) is south of 45°15' and east of 114°15'. Data sources include Evans (1981), Connor and Evans (1986), Lund and others (1983a), Mitchell and Bennett (1979), Lopez (1982), Ruppel and others (1983), and unpublished data of J. J. Connor, E. B. Ekren, and K. V. Evans.

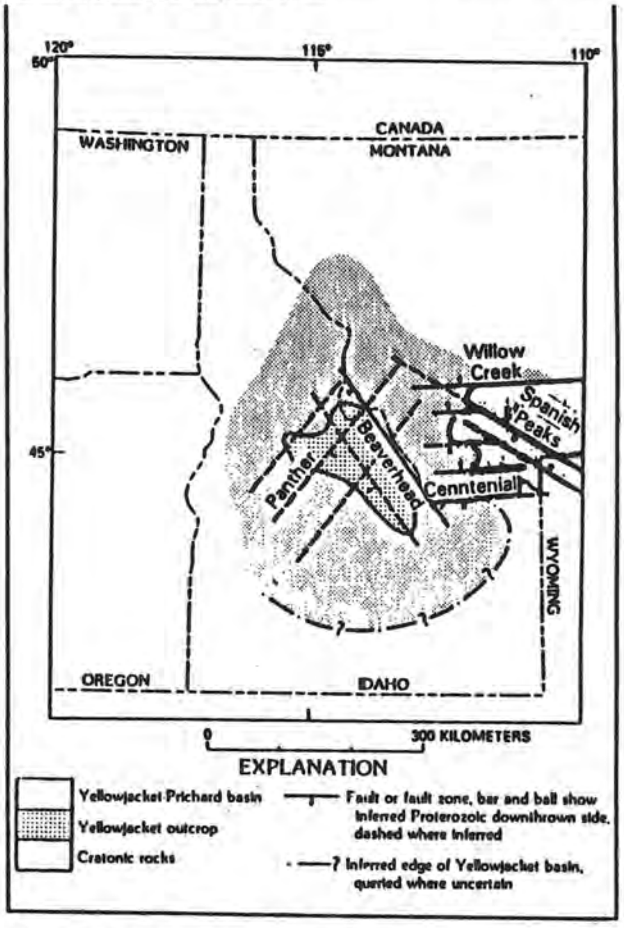
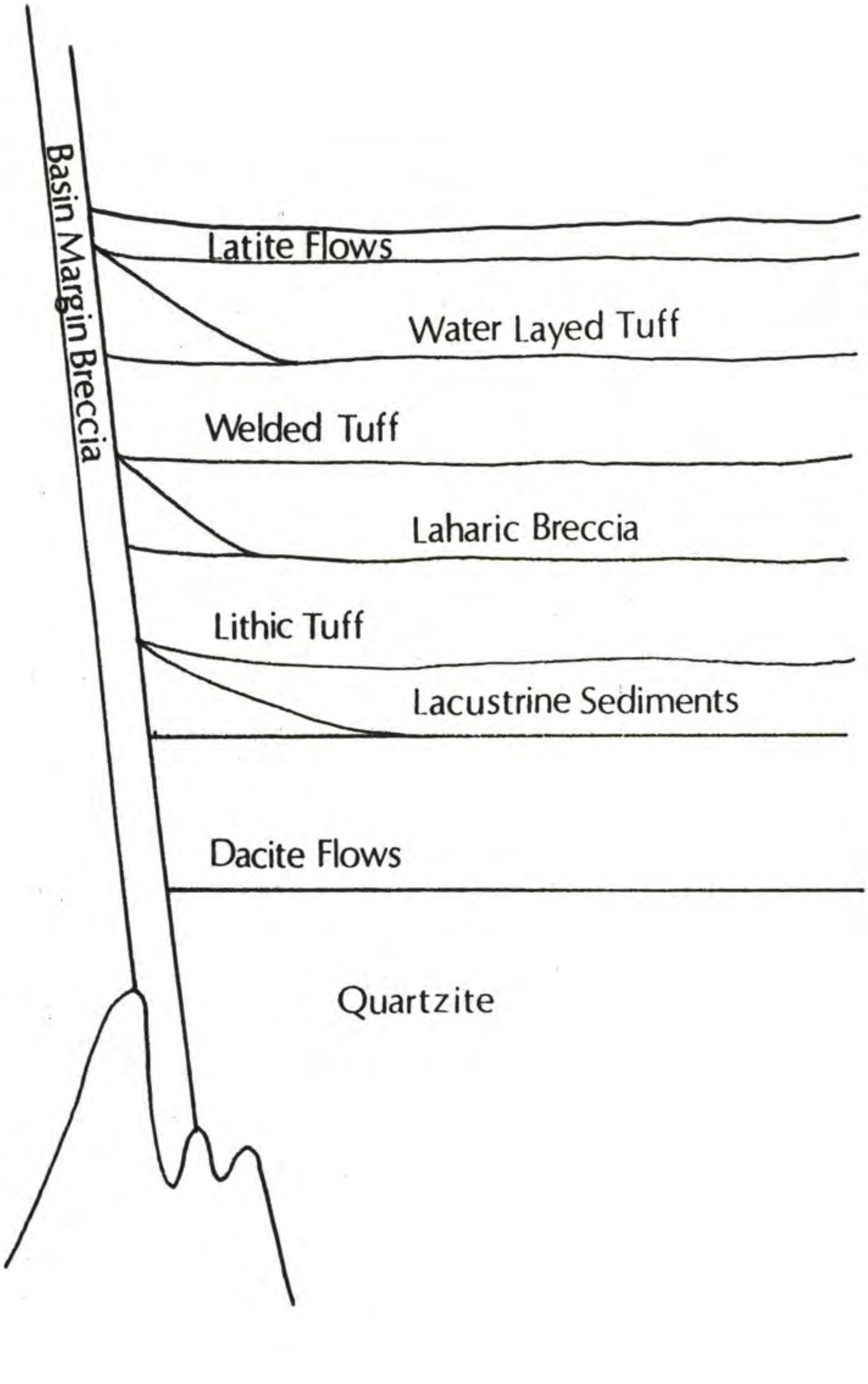
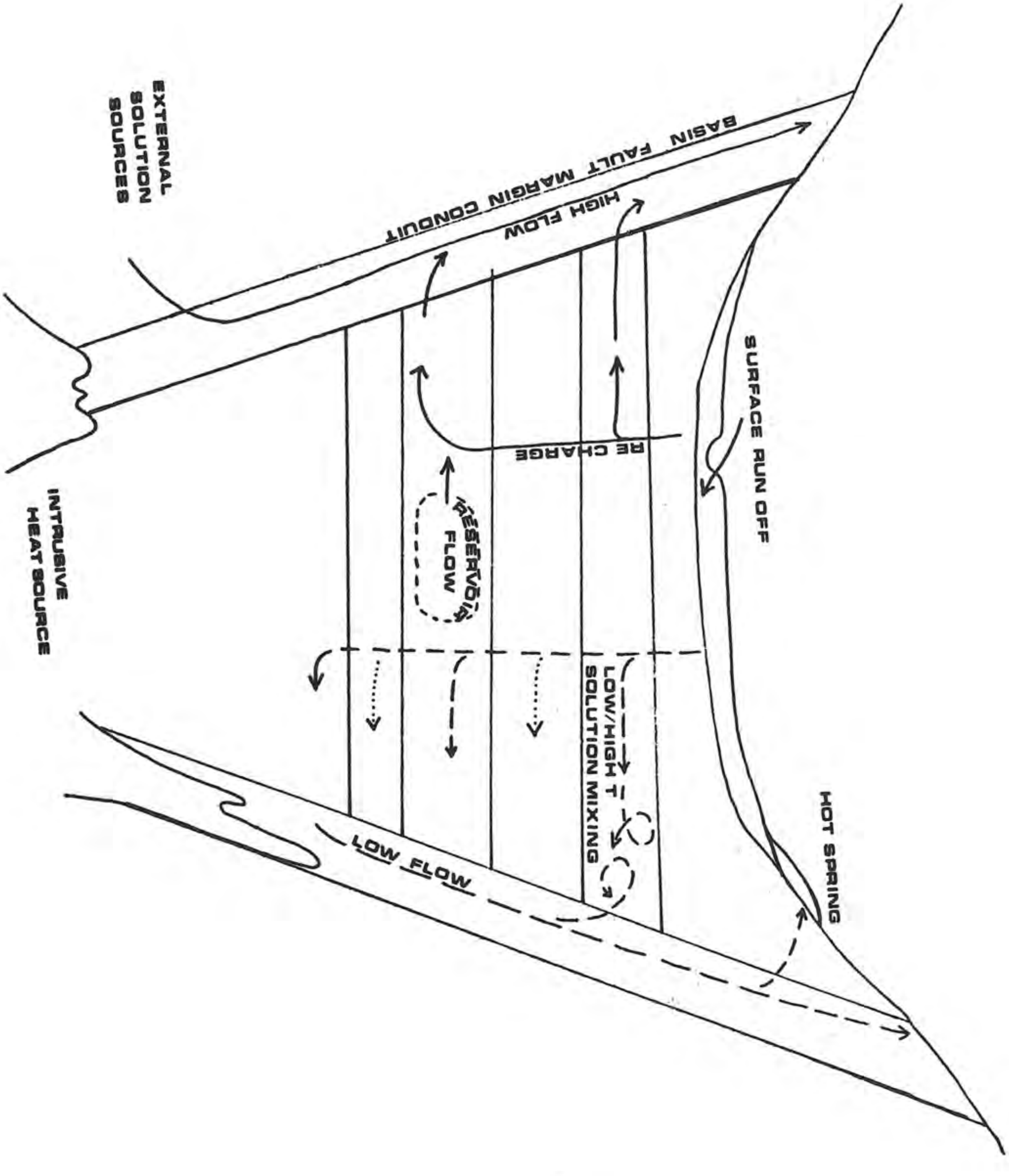
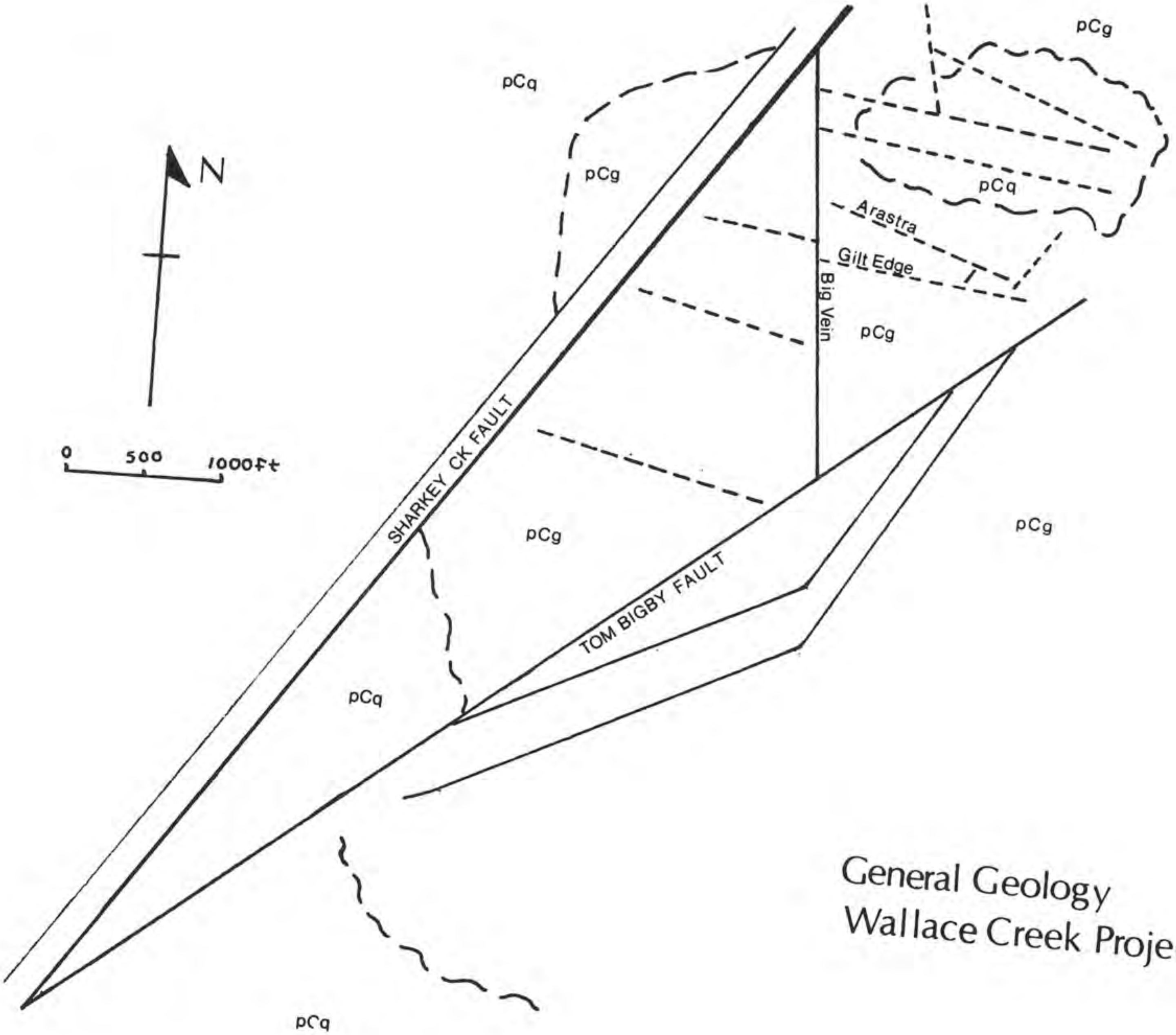


Figure 2—Sketch map of the early middle Proterozoic Yellowjacket depositional basin.


Basin Margin Breccia
Latite Flows
Water Layed Tuff
Welded Tuff
Laharic Breccia
Lithic Tuff
Lacustrine Sediments
Dacite Flows
Quartzite


EXTERNAL SOLUTION SOURCES
BASIN FAULT MARGIN CONDUIT
HIGH FLOW
SURFACE RUN OFF
RE CHARGE
RESERVOIR FLOW
INTRUSIVE HEAT SOURCE
LOW/HIGH T SOLUTION MIXING
HOT SPRING
LOW FLOW



General Geology
Wallace Creek Project

FIELD TRIP GUIDE TO THE SUNBEAM DEPOSIT, YANKEE FORK DISTRICT, IDAHO

KURT ALLEN, Grouse Creek Mining, Inc, and Boise State/Idaho State University
FALMA J. MOYE, Dept Of Geology, Idaho State University

INTRODUCTION

The Sunbeam deposit is located in the Yankee Fork Mining District, Custer County, Idaho, which was the site of placer and lode mining operations from 1873 until 1952 when the Yankee Fork dredge shut down operations. The district produced more than $12,000,000 in gold, silver, copper, lead and zinc, mostly before 1912.

The Sunbeam deposit contains 3.2 MT of ore at a grade of 0.082 opt Au for a total of 262,400 ounces of gold. A total of 129,539 feet of drilling in 533 exploration and definition drill holes has been completed on the Sunbeam deposit to date. Grouse Creek Mining Inc. is presently in the permitting stage and is going through a Supplemental Environmental Impact Statement as well as applying for all of the necessary permits for the operation. The operation will employ conventional open pit mining and milling practices. The milling process consists of grinding-agitation leach-CCD.

The Sunbeam deposit is hosted by and genetically related to Challis Volcanic rocks within a major northeast trending volcano-tectonic zone, the Custer graben which is part of the trans-Challis fault system.

GEOLOGY OF THE CHALLIS VOLCANIC FIELD

The Eocene Challis volcanic field contains compositionally varied volcanic and cogenetic intrusive rocks that are distributed over a 25,000 km^2 area in central Idaho. Outliers also occur south of the Snake River Plain in southwestern Idaho. The Challis field is the largest, and most compositionally and lithologically diverse of all the Eocene volcanic fields in the Pacific Northwest. In addition, it is deeply dissected and subvolcanic complexes are exposed.The northern part of the Challis volcanic field, where the Sunbeam deposit is located, is dominated by silicic cauldron complexes and is known primarily from work by the U.S. Geological Survey for the Challis 1 x 2 degree CUSMAP (McIntyre and others, 1982; McIntyre, 1985).

The Challis Volcanic Group (Fisher and Johnson, 1987) of central Idaho is exposed over an area of about 25,000 km^2 and forms the largest subregion of Eocene volcanic rocks in the northwestern United States. The rocks were deposited on an irregular prevolcanic terrain underlain by Precambrian crystalline rocks and Belt Supergroup, Paleozoic sedimentary rocks, and Mesozoic Idaho Batholith. Stratigraphic relations in the Challis volcanic field are complex as a result of deposition on an irregular surface, coeval graben subsidence, and pre-, syn-, and post-volcanic block faulting (Hardyman, 1981; McIntyre and others, 1982; Fisher and others, 1983; McIntyre, 1985; Ekren, 1985). Subsequent erosion has deeply dissected the volcanic field, giving excellent exposure of both volcanic deposits and subvolcanic rocks (Bennett, 1980; Fisher and others, 1983; Kiilsgaard and Lewis, 1985; Hardyman and Fisher, 1985

The entire Challis volcanic field was active for about 10 Ma, but most volcanism occurred during a period of only 5 million years from 50 to 45 Ma. Regional stratigraphy, volcanic styles and bulk compositions of volcanic rocks are similar in the northern and southern Challis field. Early volcanism, beginning at about 51 Ma, was dominated by intermediate to mafic lava flows and tuff breccias that were deposited on an irregular topographic surface of Precambrian to Cretaceous rocks. Vents are difficult to identify, but general field relations suggest that eruptions occurred from numerous central-vent stratovolcanoes and that fissure-fed lava flows were emplaced into northwest trending graben. The early intermediate volcanic rocks have a broad compositional range, including high-K basalt, shoshonite, absarokite, high-K andesite and latite, but there is no apparent systematic age or spatial relation among these lava types.

This was followed by eruptions of voluminous intermediate to silicic ash-flow tuff eruptions which were primarily focused along the trans-Challis fault system. In the northern part of the field, voluminous explosive and effusive eruptions deposited dacitic to rhyolitic ash-flow tuffs and lava flows and resulted in the concurrent subsidence of large cauldron complexes, calderas and northeast-trending volcano-tectonic depressions occurred in the northern part of the field from 48.4 to 45 Ma. (McIntyre and others, 1982). Explosive volcanism began earlier in the southern part of the field at 48.7 Ma but was less voluminous and persisted for a shorter time interval until about 47 Ma. Similar but smaller volcanic subsidence structures developed in the southern part of the field.

The final stage of igneous activity which characterizes the entire field involved the emplacement of dacitic to rhyolitic domes and the effusion of minor silicic lava flows. These dome flow complexes and hypabyssal intrusions were emplaced throughout the life of the field but the main period of activity was from about 48 to 45 Ma; the youngest intrusions yield radiometric ages of 40 Ma. Emplacement of intrusions was controlled by cauldron structures and northeast-trending regional faults. Epithermal mineral deposits are spatially associated with this late igneous activity.

Early Mafic-to-Intermediate Effusive Volcanism

Beginning at about 51 Ma, early volcanism was dominated by mafic-to-intermediate lava flows and tuff breccias that were deposited on an irregular topographic surface of Precambrian to Cretaceous rocks. Mafic-to-intermediate lavas and associated autoclastic tuff breccias are best exposed in two areas: the southeastern Challis and northeastern Hailey 1x2-degree quadrangles, and the west-central portion of the Idaho Falls 1x2-degree quadrangle (Muldoon 15-minute quadrangle). Vents are difficult to identify, but field relations generally suggest that central-vent stratovolcanoes and eruptive fissures were the sources of lavas emplaced into northwest-trending graben. In areas such as the southeast corner of the Challis 1x2-degree quadrangle, vent locations were controlled by north/northwest-trending structures (McIntyre and others, 1982). The volume of material erupted from any one source was small, and the few isolated central-vent complexes which have been recognized are generally small, covering a few square kilometers. A well-exposed example of a large, central-vent complex is the Sheep Mountain eruptive center in the southeastern part of the Challis 1x2-degree quadrangle, where dike swarms and a central zone of hydrothermal alteration are surrounded by intermediate lava and pyroclastic rocks. The volcanic rocks dip uniformly to the north rather than radially, and dikes trend dominantly north-northeast, suggesting that sheetlike lava flows were later tilted by intrusion and faulting, rather than construction of a large stratovolcano.

The rarity of central-vent lithofacies indicators within thick accumulations of sheetlike, mafic-to-intermediate lava flows indicates that the lavas were largely fissure-fed and formed extensive volcanic plateaus of sheetlike lava flows having meters to tens of meters individual thickness, aggregate thicknesses up to 1km, and covering tens to hundreds of square kilometers. We refer to such accumulations as intermediate volcanic plateaus. Thick (>1km) accumulations of sheetlike, andesitic and dacitic lava flows occur over large areas along the North Fork of the Big Lost River and the East Fork of the Salmon River. Within this extensive lava field, the lava flows dip uniformly to the northeast at about 15 degrees and there is no volcanic-facies evidence for steep-sided stratovolcanoes.

The early intermediate volcanic rocks have a broad compositional range, including high-K basalt, shoshonite, absarokite, high-K andesite and latite. Basalts, absarokites and shoshonites contain phenocrysts of olivine and clinopyroxene, with minor plagioclase and amphibole. Fresh rocks are generally black to dark gray, and weather reddish brown. Intermediate lavas include andesites and latites with phenocrysts of clinopyroxene, plagioclase and hornblende. Dacites and trachytes are strongly porphyritic rocks with combinations of plagioclase, hornblende, biotite, clinopyroxene and orthopyroxene phenocrysts. Dacites are gray-green on fresh surfaces, and tan to gray on weathered surfaces.

Mafic-to-intermediate lava flows erupted largely from about 51 to 48 Ma and dominate the lower part of the volcanic section. In the northern part of the Challis field, K-Ar geochronometry indicates that volcanism began around 51.1 Ma with eruption of dacite lava, and eruption of andesitic lava began at about 50.3 Ma (McIntyre and others, 1982). In the southern part of the field, $^{40}Ar/^{39}Ar$ geochronometry indicates that the onset of effusive volcanism occurred 48.8 ± 0.17 Ma in the Porphyry Peak area, and K-Ar geochronometry indicates volcanism began at 49.4 ± 0.7 Ma in the southernmost part of the field (Betty Skipp, pers. comm., in: Moye and others, 1988). Thus, radiometric data suggest that the basal, mafic-to-intermediate lavas were deposited during about a 2-million-year time span in the northern Challis field, and during less than a million years in the southern part of the field. This is consistent with the observation of few erosional surfaces or interbedded volcaniclastic sedimentary rocks between lava flows. Sporadic effusion of mafic-to-intermediate lava flows continued throughout Challis volcanism, but became volumetrically unimportant by the time cauldron-forming eruptions began.

Cauldron-Related Rocks

Early mafic-to-intermediate volcanism was followed at about 49 Ma by eruptions of voluminous, intermediate-to-silicic, ash-flow tuffs from large volcanic-subsidence structures. We use the term "cauldron complex" to describe related, commonly elongate volcanic structures that developed on extending crust. When overall geometry and collapse features are poorly constrained or absent, we use the more general term "eruptive center."

A cauldron is a large, volcanic-subsidence structure that may be bounded by a zone of arcuate faults, but is not as well-defined as a classic circular caldera. Cauldron complexes are generally shallow, elongate eruptive basins without well-developed collapse structures, and may include linear structures such as half-graben. Source regions are inferred from volcanic lithofacies of the associated deposits, including rapid lithofacies changes from thin, outflow sheets of ash-flow tuff, to thick, cauldron-filling ash-flow tuffs; the size and distribution of lithic fragments; increasing number of individual flows in compound cooling units as the vent area is approached; and close association of petrographically and geochemically similar intrusive rocks. The general rarity of plinian-fall deposits beneath ash-flow tuffs suggests that the Challis ash flows were vented from cauldron-related, volcano-tectonic fissures rather than from central vents, since the latter would likely produce sustained, vertical eruption columns.

Northern Challis volcanic field

In the northern Challis field, voluminous, dacitic-to-rhyolitic ash-flow tuffs and lava flows erupted about 48-45 Ma from the Van Horn Peak and Thunder Mountain cauldron complexes, and the Twin Peaks caldera (Figure 1) (McIntyre and others, 1982). The Custer and Panther Creek grabens are related volcano-tectonic depressions that formed during crustal extension and magma withdrawal. Cauldron-forming eruptions began in the northern part of the field with the eruption of the 48-Ma tuff of Ellis Creek from the Van Horn Peak cauldron complex, and continued until 47 Ma, with eruption of the Sunnyside Tuff and collapse of the Thunder Mountain caldera (Leonard and Marvin, 1982).

Van Horn Peak cauldron complex. The Van Horn cauldron complex as defined by McIntyre and others (1982) and Ekren (1985) is a northwest-elongate, elliptical volcanic subsidence structure approximately 70 km x 40 km in size. Although the original Van Horn Peak volcanic structure has been obscured by younger cauldron forming and intrusive events, Ekren (1985) describes a series of ash-flow tuff eruptions related to the cauldron. The Van Horn Peak cauldron complex is the earliest and largest volcanic structure in the Challis field, with northwest-elongate dimensions of 70 x 40 km. The most voluminous eruption deposited the dacitic tuff of Ellis Creek, a compound-cooling-unit, welded tuff which is up to 1500m thick within the cauldron, 300m thick in outflow sheets, and flowed 35 km from the cauldron boundary. Lithic fragments in the cauldron fill are up to 1 meter in size, and those in the outflow sheet are up to 3 cm in size. McIntyre and others (1982) suggest that eruption of the tuff and collapse of the cauldron were simultaneous because there is such a thick intracauldron fill. The Van Horn Peak cauldron complex continued to form by sector eruptions which produced cauldron filling tuffs but produced only minor outflow sheets. Eruption of the tuffs of Eightmile Creek, Camas Creek, and Black Mountain occurred over a rapid interval, possibly within 500,000 years.

The earliest event deposited the lithic-rich rhyodacitic tuff of Corral Creek and formed the Corral Creek collapse segment of the cauldron. This unit is confined largely to the area around the collapse segment.

The largest volume eruption (>400 km^3) related to the main cauldron collapse was the eruption of the tuff of Ellis Creek. Ekren (1985) inferred boundaries of the original cauldron collapse structure based on thickness of intracaldera fill of the tuff of Ellis Creek. The intracauldron fill is as much as 1500 m thick; outflow sheets extend as much as 30 km from the cauldron and are up to 300 m thick. The tuff of Ellis Creek is a crystal-rich rhyodacite tuff with little mineralogic variation through the section; the crystal assemblage includes plagioclase, biotite, hornblende, and rare clinopyroxene and quartz.

The culminating ash-flow eruptions associated with the cauldron complex filled the subsidence block. These units were derived from local sector eruptions and filled in, near, or over their vents. The tuff of Eightmile Creek and the tuffs of Camas Creek and Black Mountain are rhyodacitic crystal tuffs with variable amounts of lithic and pumice fragments. The crystal assemblage of the tuff of Eightmile Creek is similar to the tuff of Ellis Creek , except the former contains potassium feldspar, more quartz, and less mafic minerals. The tuffs of Camas Creek and Black Mountain have a crystal assemblage dominanted by plagioclase and pyroxene with lesser biotite and hornblende.

After formation of the initial Van Horn Peak cauldron, sector eruptions from trap-door structures occured. Ekren (1985) identified near vent facies of the Castle Rock sector based on inferred high temperatures suggested by rheomorphic folding and thick accumulations and multiple cooling units within the collapsed sector with few thin outflow sheets. The northeast-trending Custer graben and Panther Creek graben developed in response to rifting and co-eval magma withdrawel. Both have been inferred as source regions based primarily on thickness of ash-flow tuff within the structures. Sector eruptions produced local flow lobes which accumulted in volcanic subsidence structures.

Twin Peaks caldera. The Twin Peaks caldera is the only caldera in the entire Challis field with definitive circular ring fractures (Hardyman, 1985; Hardyman and Fisher, 1985). It is a fault-bounded, circular depression containing intracaldera ash-flow tuff and megabreccia from slope failure of the inner caldera walls. The Twin Peaks caldera formed about 45 Ma with eruption of the tuff of Challis Creek and represents the final stage of the cauldron-forming eruptions in the northern Challis field.

The Twin Peaks caldera formed about 45 million years ago and represents the final, major ash-flow-tuff-forming eruption in the Challis Volcanic field. The 20-km diameter caldera formed within, and is considered to be part of the Van Horn Peak cauldron complex. The caldera is divided into 3 main structural blocks, each of which involved collapse and volcanism. The eruptions produced the rhyolitic tuff of Challis Creek, which is about 800m thick within the caldera, and generally less than 30-40m-thick for outflow sheets that extend up to 40 km from source. The tuff of Challis Creek is densely welded, crystal-rich tuff with 25 to 30 % phenocrysts of bipyramidal quartz and alkali feldspar, and minor plagioclase and pyroxene. It contains two cooling units, separated by thin ash-flow tuffs and discontinuous epiclastic units.

Silicic-to-Intermediate Intrusions and Flow/Dome Complexes

During the final stage of igneous activity, dacitic-to-rhyolitic domes, small silicic lava flows, and silicic-to-intermediate intrusions were emplaced across the entire field. Although dome-flow complexes and associated hypabyssal intrusions were emplaced throughout the life of the Challis volcanic field, the main period of activity was from about 48 to 45 Ma and the youngest intrusions yield radiometric ages of 40 Ma. Intrusions tend to occur along cauldron structures and northeast-trending regional faults. Epithermal mineral deposits are spatially associated with the late-stage, silicic igneous activity (McIntyre and Johnson, 1985; Moye and Hall, 1988)

In the western Challis volcanic field, large, silicic-to-intermediate, hypabyssal complexes with roof pendants of pre-Tertiary granitic and sedimentary rocks, and Challis volcanic rocks, are exposed. The hypabyssal intrusions include the pink and gray porphyries of McIntyre and others (1982). The pink porphyry corresponds to the pink granite described by Bennett (1980) and includes the Sawtooth granite, the Casto Pluton, and the Mackay stock. The granite typically contains smokey quartz and conspicuous pink potassium feldspar. The gray porphyry is characterized by abundant phenocrysts with varying porportions of plagioclase, hornblende, biotite and clinopyroxene.

In the eastern Challis volcanic field, dissection is less deep, and dome/flow complexes intruding their own ejecta are the subaerial counterparts of silicic-to-intermediate hypabyssal intrusions. Silicic intrusions are rhyolitic to trachytic in composition and range from microphyric to strongly porphyritic; dominant phenocrysts include sanidine, biotite, plagioclase, quartz and rare hornblende. Across the entire Challis volcanic field, rhyolitic intrusions are generally the youngest igneous rocks.

Small plugs and flow/dome complexes of intermediate composition are volcanic/hypabyssal equivalents of the gray porphyry of McIntyre and others (1982).

GEOLOGY OF THE SUNBEAM DEPOSIT.

The Sunbeam deposit is a structurally controlled epithermal gold-silver system associated with late-stage emplacement of rhyolitic dome complexes in the Custer graben (Figure 1). The Sunbeam rhyolite, Tr, flow-dome yields an age of 45.8 ± 2.3 Ma (McIntyre and Johnson, 1985) and intruded through a volcanic sequence consisting of, from oldest to youngest: 1) Ta- andesite, 2) Tpt- propylitically altered pyroclastics, and 3) Tp- massive pyroclastics (Figure 2,3,4). Following intrusion of the rhyolite, deposition of carbonaceous black shales, volcaniclastic mudstones, siltstones and and sandstones comprising unit Tbt began. Both the rhyolite flow-dome and the massive pyroclastics are hosts to mineralization at the Sunbeam deposit.

Structure plays an important role in localization of mineralization at the Sunbeam deposit. Regionally, three structural trends occur and are, from oldest to youngest: 1) N25-40E; 2) N30-45W; and 3) N50-85E. Within the Sunbeam deposit, there are two distinct structural trends: 1) N50-60E and N20-40 W (Figure 5). The dominant and youngest set of structures is the N50-60 E set; this set is also the dominant structural control on mineralization .

ROAD LOG

STOP ONE: Preachers Cove Mill turnoff

VIEW OF VOLCANIC STRATIGRAPHY IN CUSTER GRABEN.

This stop affords a view of the volcanic stratigraphy and structure along the southeastern margin of the Custer graben. Silicic pyroclastic rocks which were thought to have erupted from a source within the graben are tilted to the east along the eastern border fault of the graben. This fault juxtaposes Challis Volcanics in the hanging wall with Paleozoic sedimentary rocks of the Wood River assemblage and Cretaceous granite of the Idaho Batholith in the footwall.

The mill is owned and operated by U.S. Antimony.

STOP TWO (OPTIONAL): YANKEE FORK DREDGE

This dredge operated on the Yankee Fork for a total of eight years from 1940- 1952 with total production of 30,000 ounces. Today the dredge is maintained as a museum and tourist attraction.

STOP THREE: SUNBEAM DEPOSIT, THE MILL TUNNEL

This stop offers a view of the old Golden Sunbeam mill and the main adit area. Ore was mined from several levels of adits and dropped into the glory hole which provided acess to the mill. The original mining operation extracted ore from high-grade bonanza veins which had an average grade 0.300 opt Au. The original Golden Sunlight Mine produced approximately 9,000 ounces of gold and unknown quantity of silver with a gold to silver ratio of about 2.5 to 1.

Minor production occurred between 1879, when Sunbeam Mountain was first staked and 1903, when Sunbeam Mines, Inc. built a mill on the property. Betwen 1903 and 1909, $315,000 of gold (~ 9,000 ounces) was produced from the Golden Sunbeam claims on Sunbeam hill (Lockard and Rice, 1971, p.13). The Sunbeam mine was closed in 1911 and has never since reopened, except for sporadic cyaniding of mill tailings from 1911 through 1935. The Sunbeam mill last operated in the summer of 1947.

At this stop the main rock type exposed is the massive pyroclastic unit Tp. It is argillically altered as is typical within the main ore zone.

STOP FOUR: BASAL PYROCLASTIC UNIT

The basal welded ash-flow tuff unit (Tpt) is exposed in roadcut at this stop. The unit is typically a crystal rich ash-flow tuff with abundant lapilli and block sized green, flattened pumice fragments. This unit is generally propylitically altered producing a mineral assemblage of chlorite, calcite and celadonite clay at the expense of siliceous volcanic glass; it is also locally argillized. The ash-flow tuff typically contains phenocrysts of quartz and alkali feldspar and is not a host for ore in the Sunbeam deposit.

The basal welded ash-flow tuff blanketed an irregular Eocene topographic surface. Subsequent block faulting and/or erosion produced an irregular surface on top of the ash-flow tuff prior to deposition of other units (Hahn, 1989).

STOP 5. MIDDLE BENCH

The middle bench exposes both the southern and northern part of the Sunbeam deposit. In addition both of the rock types which host ore can be seen at this site. At the southern end of the deposit, the host for ore is the Sunbeam rhyolite porphyry intrusion (Tr) which is characterized by phenocrysts of quartz and argillically altered alkali feldspar in an aphanitic groundmass which is variably altered. In cross section, the rhyolite forms a funnel shaped intrusion (Figure 4). It has an age of 45.8 ± 2.0 m.y. based on K-Ar geochronmetry.

At the northern end of the deposit , the Sunbeam rhyolite (Tr) and massive pyroclastic (Tp) are exposed by the blast wall. Both units host ore at this locality.. Tp is characterized as fine to coarse grained primarily welded to non-welded air-fall and ash-flow tuff, interbedded with pyroclastic breccias. The unit is light to dark gray and contains up to 3 percent subhedral to anhedral quartz phenocrysts and argillically altered plagioclase and alkali feldspar phenocrysts. Locally the rock has a spongy or vuggy appearance due to significant amounts of relict pumice which have been obliterated by alteration.

In addition to the two rock types, this stop also offers the opportunity to examine clay rich, mineralized fault zones. These are often high-grade ore zones and contain visible gold

STOP 6 : WEST END FAULT ZONE

The west end fault zone juxtaposes altered rhyolite and pyroclastic rocks of the Sunbeam deposit against unaltered sediments and upper pyroclastics of the Grouse deposit (Figure 3). The fault appears to have had pre-mineralization to post-mineralization movement. In general the fault is a is a N 25-30 E trending normal fault which bends to N60 to 70 E trend. The total offset on the fault is interpretated to be about 300 feet with the east side down.

STOP 7. (optional) FLANK OF ESTES MOUNTAIN.

This stop offers the opportunity to view the Sunbeam deposit from afar and gain a better appreciation of the vertical extent of the mineralizing system and the eventual layout of the mine operation.

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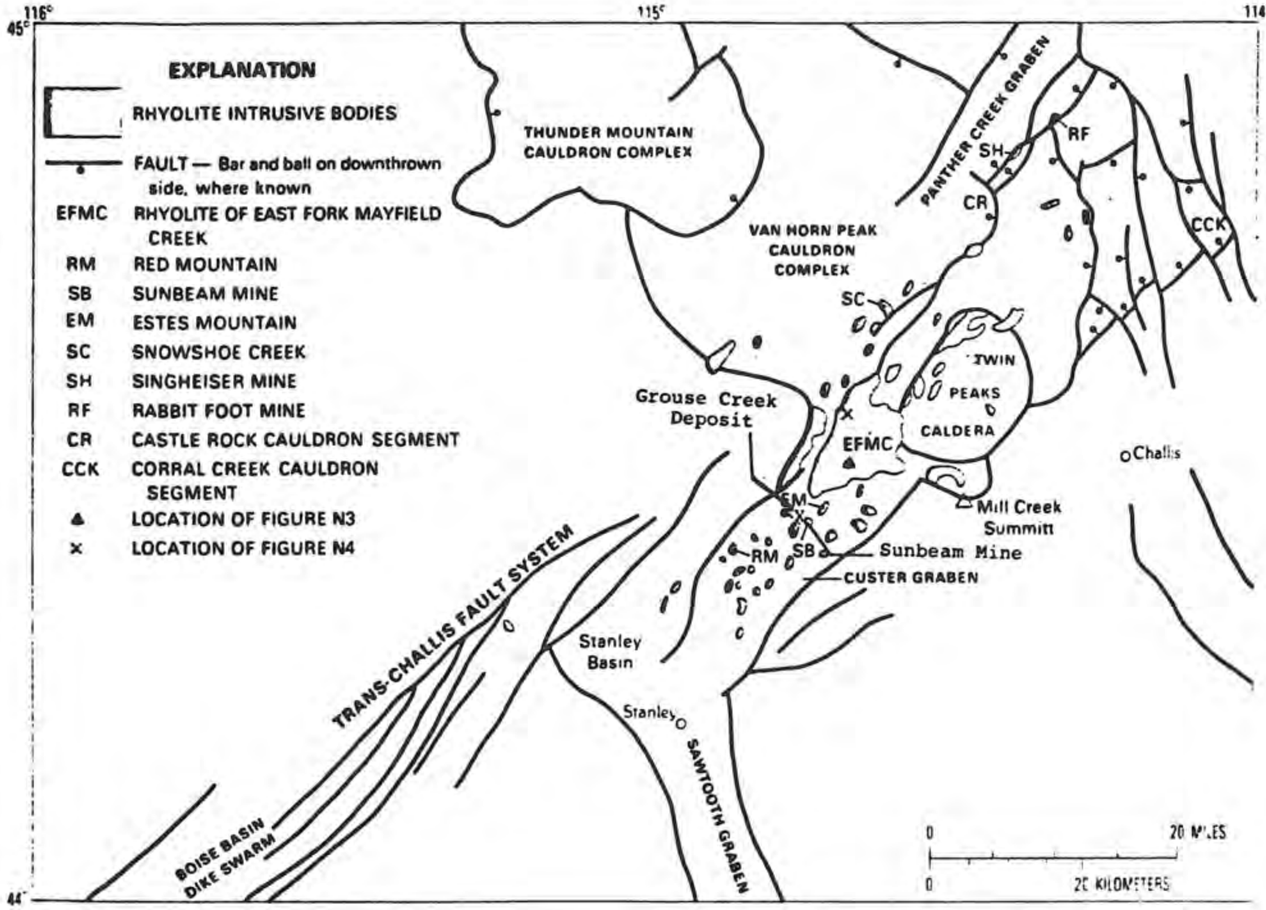


Figure 7. Index map of the Challis quadrangle showing the location of the Grouse Creek deposit and the Sunbeam Mine within the Custer Graben. Taken from Hardyman and Fisher, 1983, p.169.

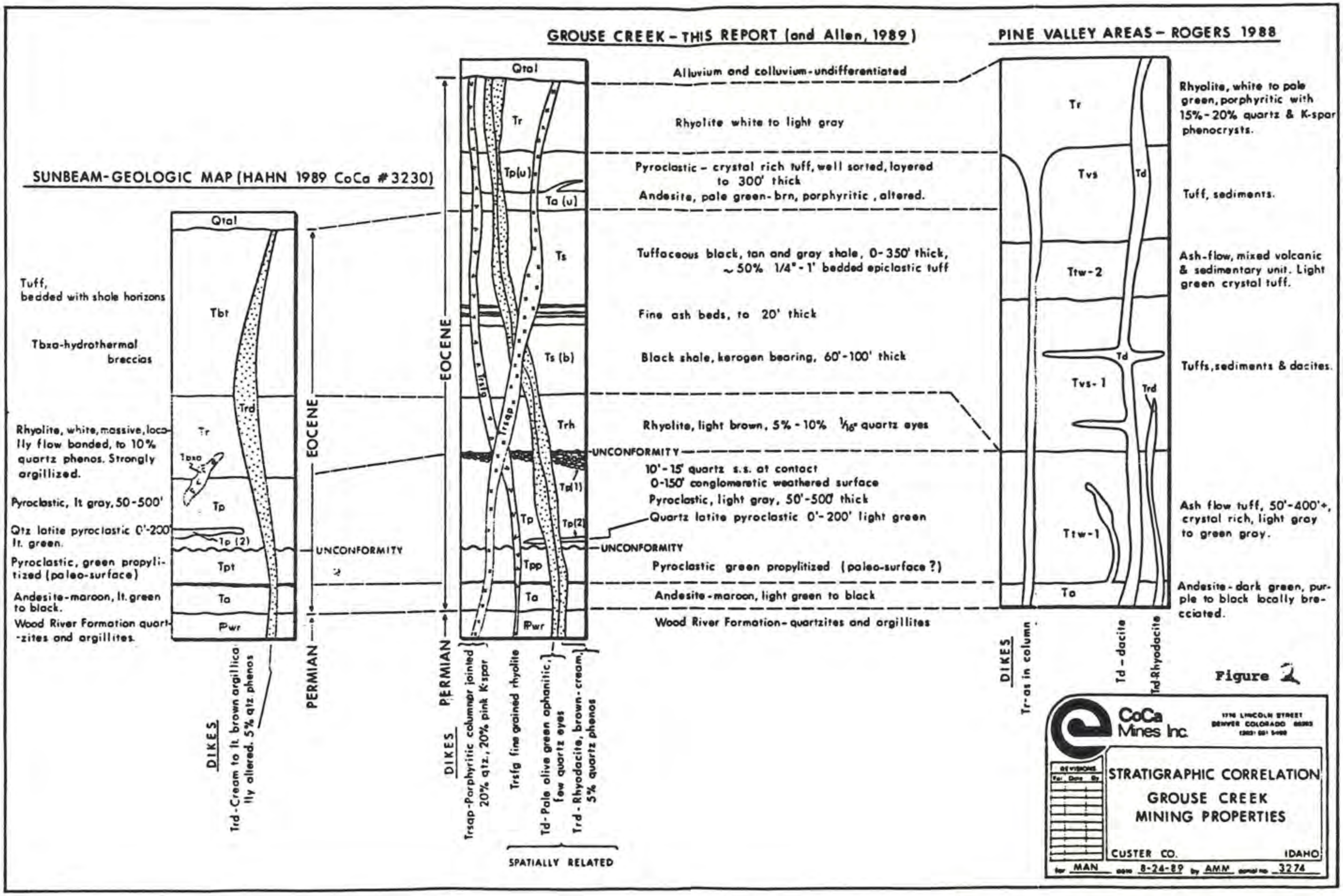
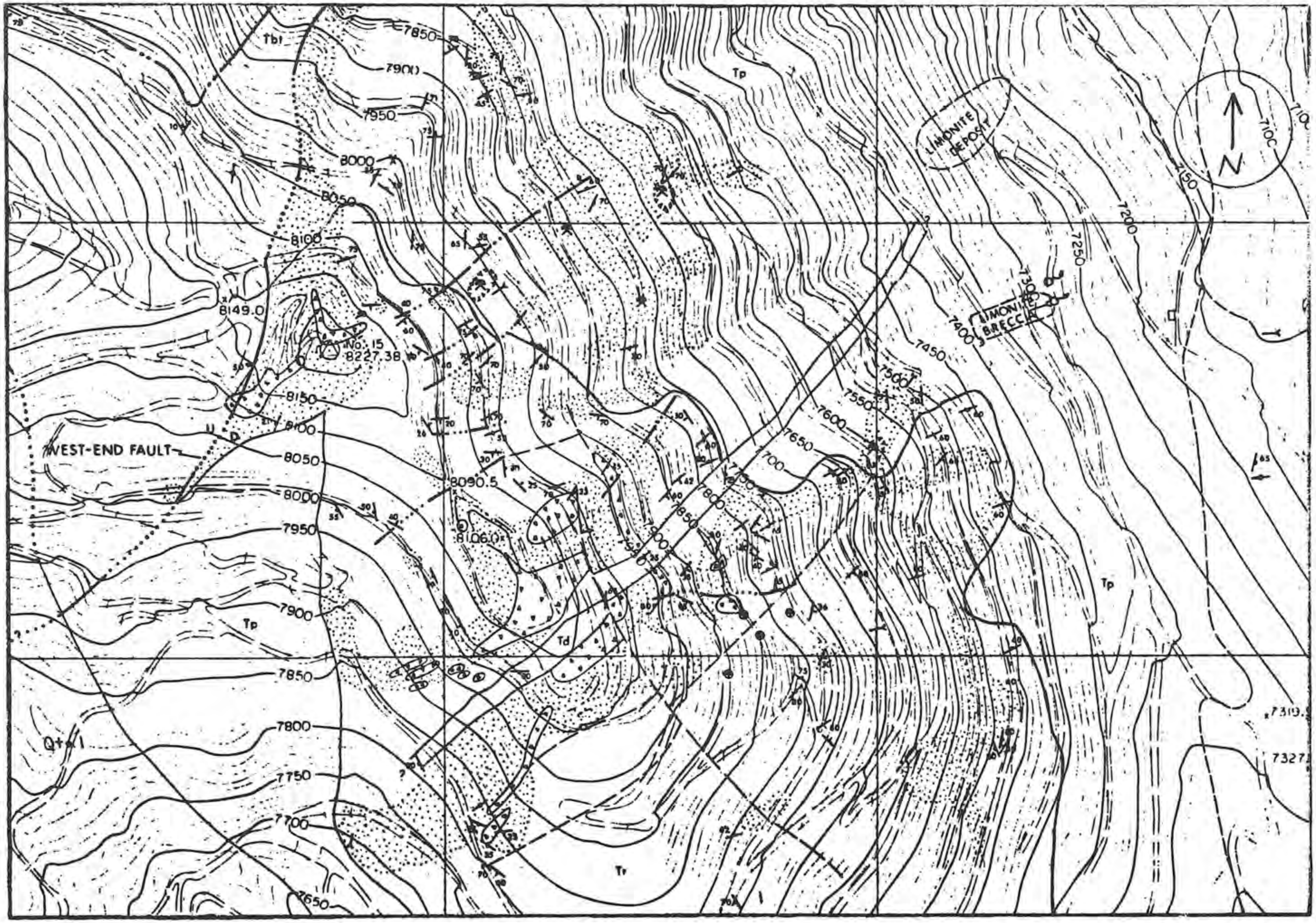


Figure 2

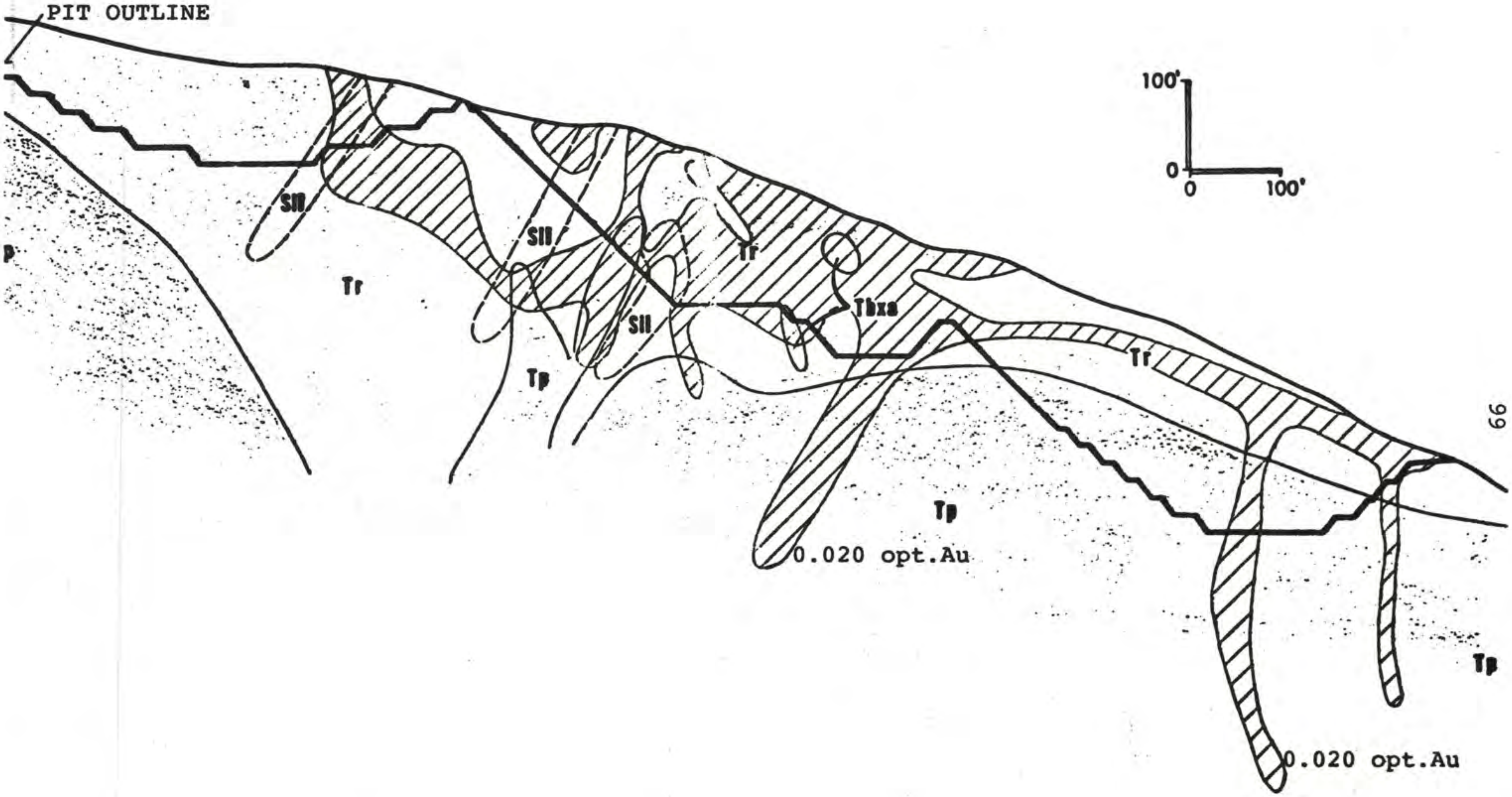
STRATIGRAPHIC CORRELATION
GROUSE CREEK
MINING PROPERTIES

CUSTER CO. IDAHO

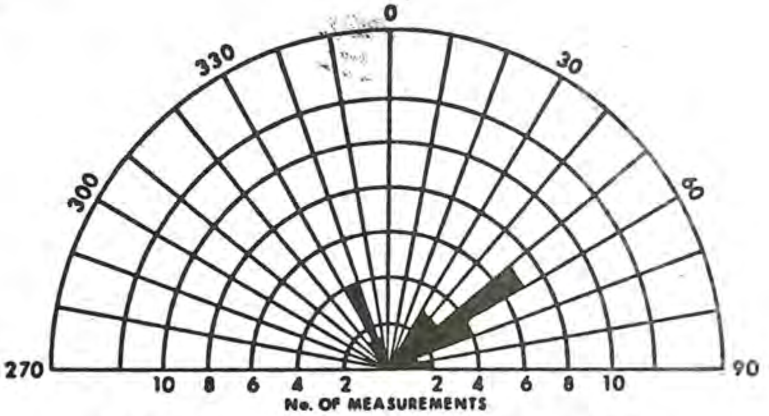
for MAN date 8-24-89 by AMM serial no. 3274



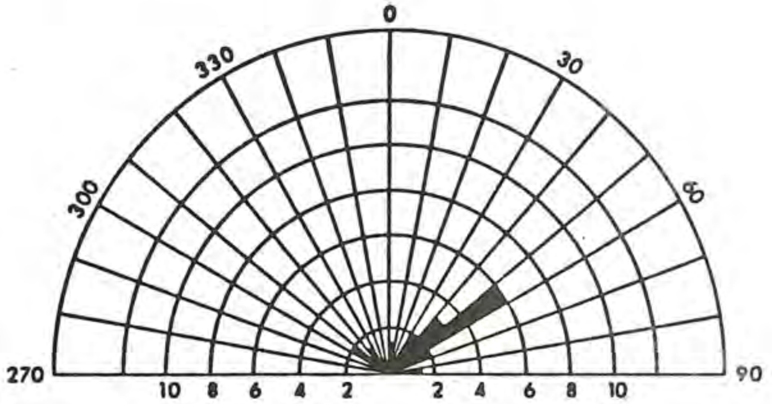
SUNBEAM DEPOSIT GEOLOGIC MAP Fig 3



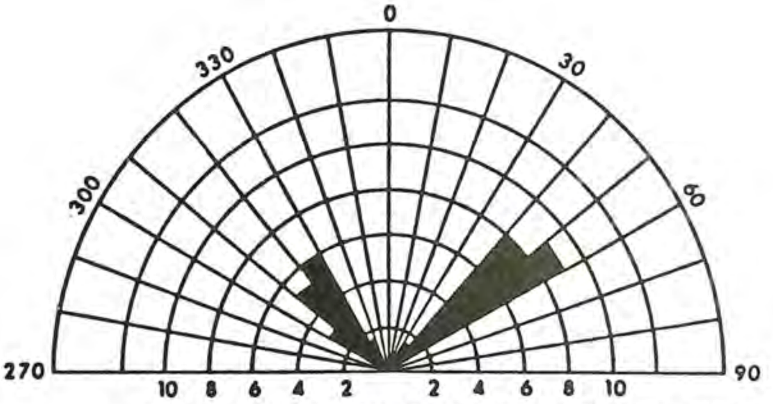
SUNBEAM DEPOSIT GEOLOGIC CROSS SECTION #10 Fig 4



SILICIFICATION/BRECCIATION ON GEOLOGIC MAP



SURFACE FAULTS ON GEOLOGIC MAP



TREND SURFACES OF 50 FT. COMPOSITE
DRILL HOLE ASSAY DATA

FIGURE 6

CoCa Mines Inc.

[illegible] LINCOLN STREET
DENVER COLORADO [illegible]
[illegible]

REVISIONS		
For	Date	By

GROUSE CREEK PROJECT

STRUCTURAL ANALYSIS
SUNBEAM DEPOSIT

CUSTER CO. 5 IDAHO

for GAH date 8·24·89 by AMM serial no. 3274

Major Extensional Events in Idaho since the Cretaceous: An Overview

by Earl H. Bennett, State Geologist, Idaho Geological Survey, Morrill Hall, University of Idaho, Moscow, ID 83843
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The 1980's were an exciting time for geologists in Idaho. During the past decade, two periods of major crustal extension have been recognized that have had a dramatic impact on the geologic history of Idaho and the rest of western North America. Advances in isotopic age dating, improved paleomagnetic and other geophysical techniques, and most important, a considerable amount of new geologic mapping, have answered some questions about regional extension and, as with all good science, raised many more.

Major mapping programs completed as part of the U.S. Geological Survey's Conterminous United States Mineral Appraisal Program (CUSMAP) in the Hailey, Challis, Idaho Falls, Dillon, and Wallace 2-degree quadrangles have contributed substantially to our understanding of the regional geology of the central part of the state. Ongoing work in the Elk City and Hamilton 2-degree quadrangles by the Idaho Geological Survey and the USGS continues to uncover major clues about regional extension. As a result of this work and contributions by many others, two major periods of extension have been identified; a Late Cretaceous-Eocene extension (includes the Challis event) and a Miocene to Recent extension (basin and range event). Although I have included the late Cretaceous as part of the Eocene extensional episode, the late Cretaceous structures are enigmatic and it is possible that these formed during the waning stages of Mesozoic compression.

Geologic mapping has provided important information about the Challis Volcanic event (including volcanics, bimodal plutons and related hypabyssal dike swarms), the Idavada-Snake River Plain volcanics, and the Columbia River Basalt. In Idaho, Eocene extension is confined between the Lewis and Clark line to the north and the Eastern Snake River Plain to the south. Basin and range extension is most pronounced south of the Trans Challis Fault System (TCFS) and west of the strontium isotope line in western Idaho, Oregon, and Washington. The strontium line separates accreted terranes to the west from continental crust to the east. In general, Cretaceous structures trend northerly, Eocene structures northeasterly, and basin and range structures are oriented northwest or north-northwest. However, many of these structures may be older than either extensional period and have reactivated to accommodate changing crustal stress patterns.

During the past decade, a number of features related to major Late Cretaceous-Eocene extension were discovered or studied in detail. These include the Challis Volcanic field, the TCFS, the Dillon lineament, the Bargamin Creek fault, the Pioneer Core Complex, the House Mountain decollement, the Bitterroot lobe of the Idaho batholith, the Newport fault, and many others extending from the Kaniksu batholith in the Idaho panhandle into central British Columbia.

Basin and range structures believed to be Miocene to Recent in age, have been mapped from the TCFS southwards into Utah and Nevada and along the western part of the state to the Olympic Wallowa lineament (OWL) in Washington. Basin and range extension includes the suspected progression of the Yellowstone hot spot across the eastern Snake River Plain with voluminous bimodal rhyolite-basaltic volcanism, and the eruption of the Columbia River Basalt. Typical basin and range structures are the Lost River, Lemhi, and Beaverhead ranges, and intervening valleys in eastern Idaho. Similar structures have now been mapped westwards across the southern part of the Atlanta lobe of the batholith, as far as and including the western Snake River Plain. Recent seismic activity such as the 7.3 magnitude Borah Peak earthquake in eastern Idaho and earthquakes in western Idaho indicate that basin and range extension is active today.

Age dating by Ar40/Ar39 and supporting field work has given relatively precise dates for a number of mineral deposits in Idaho. Some of these including Atlanta Hill, Rocky Bar, the recently discovered Beartrack deposit, and the vein deposits in the Elk City-Buffalo Hump area are probably late Cretaceous-Paleocene. As noted the relationship of the late Cretaceous deposits and structures to extension or compression is not well understood. The fact that virtually all of the Eocene to Recent precious metal deposits in the Idaho batholith are probably related to extension has provided impetus to understanding the mechanics of how the extension process works. Many deposits in the TCFS are Eocene in age including the important mines in Boise Basin and the Yankee Fork mines. Basin and range related ore deposits range in age from Miocene to perhaps very recent and include Delamar, the Blackpine district, mercury-gold deposits near Weiser, and a number of epithermal disseminated gold deposits in southeastern Oregon.

The Challis Volcanic field was long thought to be a volcanic arc related to subduction. In fact, there is almost a one to one correlation between major plate changes in the Pacific over the past 70 million years and igneous events in Idaho which suggests some association between plate tectonics and extension. Recently, it has been proposed that crustal extension occurs, not as a result of subduction, but as a consequence of overthickened continental crust. Analogous to the problem of subduction versus overthickened crust in the Eocene is the question; How much of the basin and range extension in Idaho is related to the Yellowstone hot spot and how much to subduction? Like most controversial ideas in geology the real answer is probably a complex story involving subduction, crustal overthickening, and hot spot activity. The answer to these questions may well be found in the 1990's with undoubtedly a few more surprises in store for all of us.

The trans-Challis fault system, its structure and relationship to Tertiary igneous activity and mineral deposits

by
Thor H. Kiilsgaard

The trans-Challis fault system was first identified during geologic mapping of the Challis 2-degree quadrangle, work done as part of the U.S. Geological Survey CUSMAP program. Since then, USGS work in the Hailey 2-degree quadrangle has shown faults of the system to extend southwest to beyond the western border of the Hailey quadrangle, and compilation from existing maps indicates northeast extension of the system at least to the Montana border (Kiilsgaard and others, 1989). Known length of the system in Idaho is at least 270 km: width of the main part of the system is about 30 km, but the total width of the northeast-trending fault system is unknown. The fault system is aligned with and appears to be a continuation of the Great Falls lineament, as proposed by O'Neill and Lopez (1983), who projected the lineament northeast from Idaho across Montana.

Where studied in the Challis and Hailey quadrangles, the trans-Challis consists of a broad system of northeast-trending, subparallel, high-angle faults and broad zones of crushed and altered rocks, aligned grabens, breccia pipes, eruptive centers, and roughly aligned intrusive rocks of Tertiary age. The fault system and roughly aligned intrusive rocks appear to be part of broad pattern of Tertiary crustal extension, along which faulting has been active intermittently since Precambrian time.

Many epithermal precious metal-deposits in central Idaho show a spatial relation to the trans-Challis fault system and appear to be genetically related to structures of the system and to Tertiary rocks that have been intruded along these structures (Kiilsgaard and Bennett, 1985). Some of the deposits trend northeast, aligned more or less along faults of the system, whereas other deposits trend at divergent angles to the system. Of particular significance are a number of veins that strike northwest and appear to follow tensional fractures induced by differential strikeslip movement along the regional northeast-trending faults. Veins in the Custer graben cut volcanic rocks of Eocene age (McIntyre and Johnson, 1985). In the Boise Basin, gold-bearing veins cut rhyolite dikes (Ballard, 1924), which, in turn, intrude plutonic rocks of Eocene age (Anderson, 1947). The veins can be no older than Eocene. Widespread silicification along faults and argillic and sericitic alteration of the wallrock in mineralized localities appear to have been derived from regional hydrothermal systems that were heated by Tertiary intrusive rocks. Extensive fracture patterns along the trans-Challis fault system provided access for the hydrothermal solutions that carried mineral components and formed the epithermal mineral deposits.

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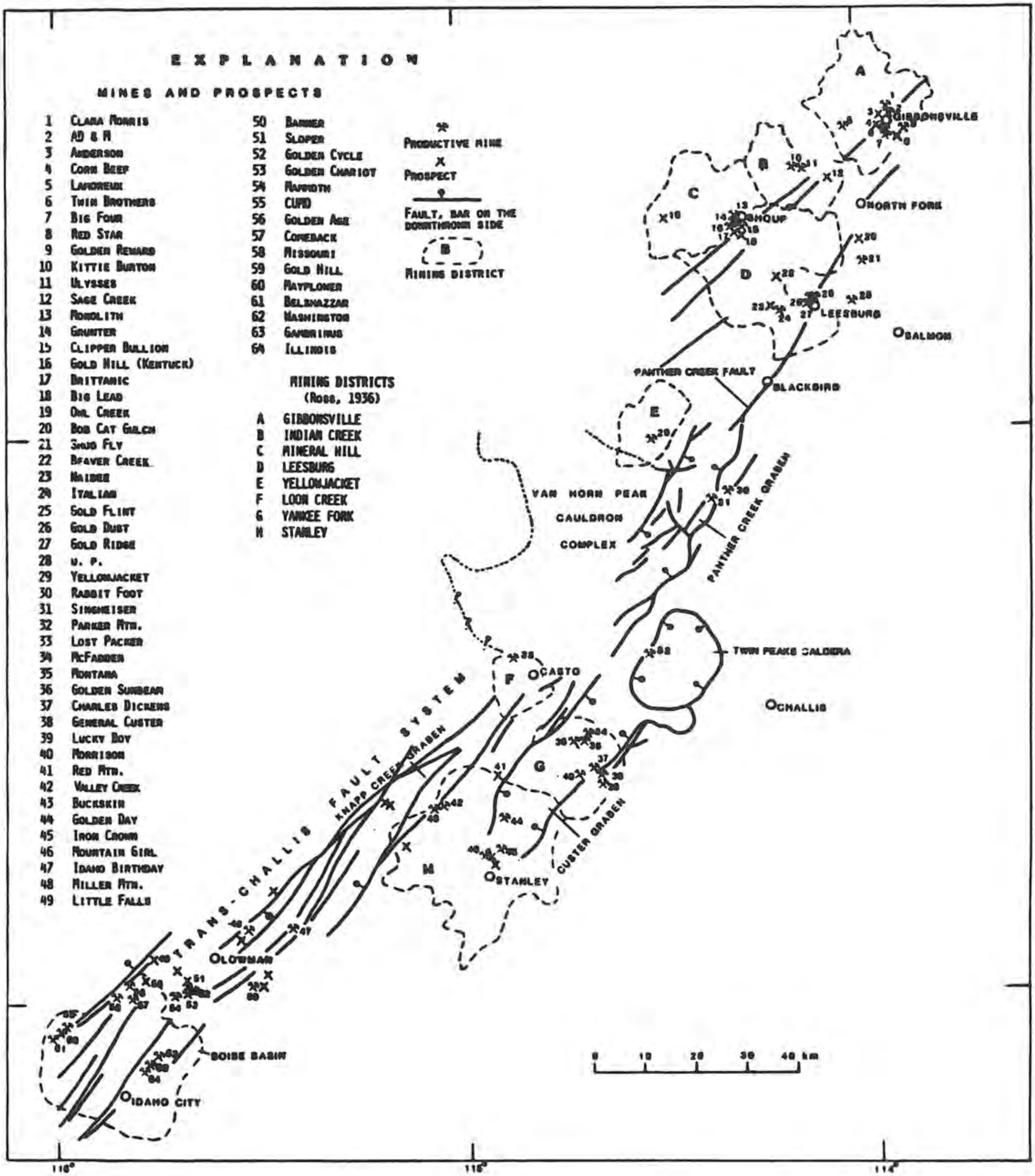

Figure 1.-- PRECIOUS-METAL MINES AND PROSPECTS OF THE TRANS CHALLIS FAULT SYSTEM

THE GREAT FALLS TECTONIC ZONE
EAST-CENTRAL IDAHO AND WEST-CENTRAL MONTANA

ABSTRACT

by
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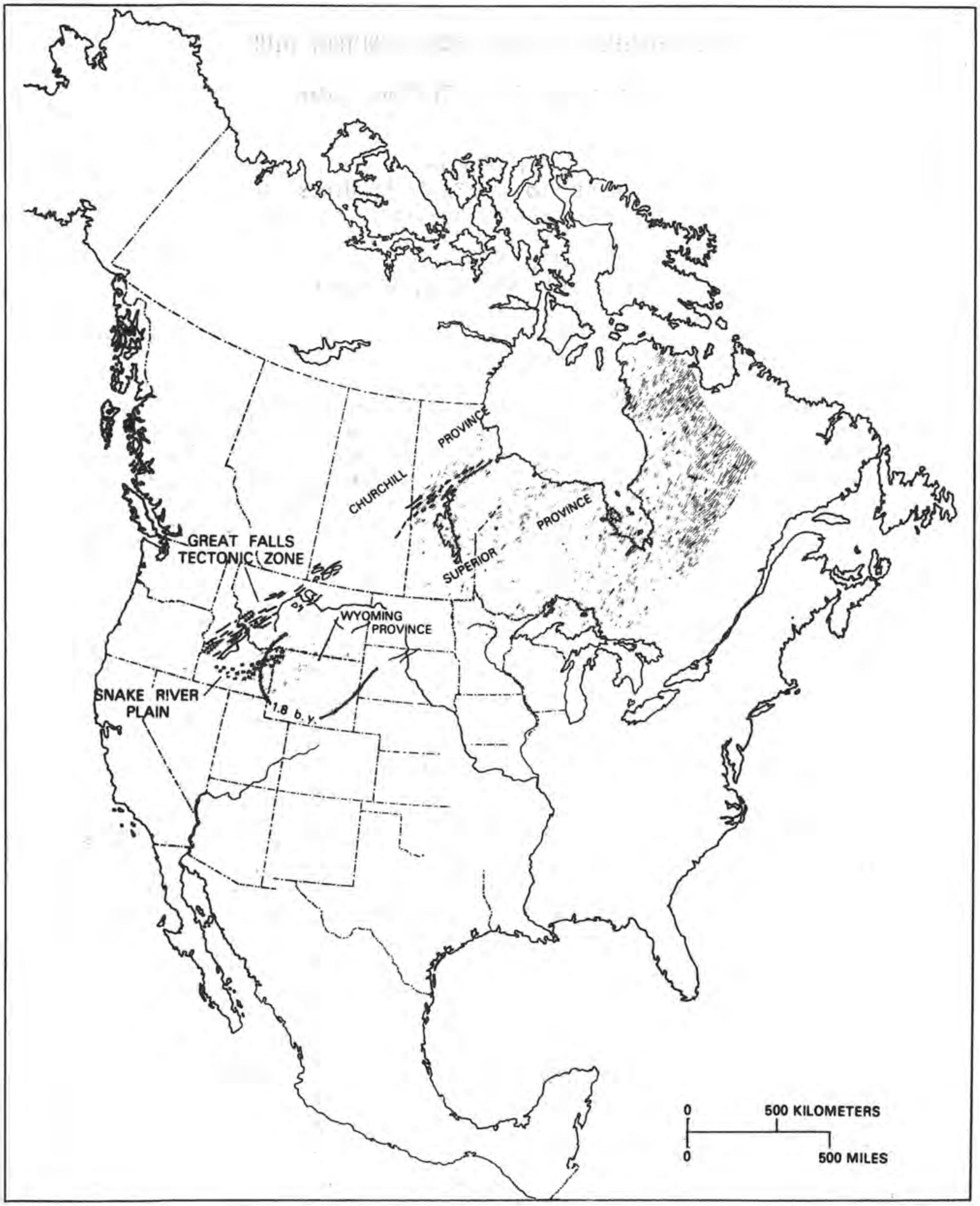
and
David A. Lopez
North American Resources Company, Billings, Montana

The Great Falls tectonic zone is a belt of diverse northeast-trending geologic features that can be traced from the Idaho Batholith in the Cordilleran Miogeocline, across thrust-belt structures and basement rocks of west-central and southwestern Montana, through cratonic rocks of central Montana, and into southwesternmost Saskatchewan, Canada. Geologic mapping in east-central Idaho and west-central Montana has outlined a continuous zone of high-angle faults and shear zones. These structures (1) extend more than 150 km (93 mi) northeastward from near Salmon, Idaho toward Anaconda, Montana; (2) extend southwestward as the Custer Graben and associated faults; (3) had recurrent movement from Middle Proterozoic to Holocene time; (4) controlled the intrusion and orientation of Late Cretaceous to early Tertiary dike swarms; and (5) controlled the uplift and orientation of the Anaconda-Pintlar Range. Recurrent fault movement in this zone and strong structural control over igneous intrusions suggest a fundamental tectonic feature that has influenced the tectonic development of the Idaho-Montana area from at least Middle Proterozoic time to the present.

In southwest Montana, the Great Falls tectonic zone also appears to control the location and orientation of the frontal edge of the thrust belt, and separates Early Proterozoic regional metamorphic terrane from Archean metamorphic rocks.



GREAT FALLS TECTONIC ZONE

RELATIONSHIP OF THE GOLDEN SUNLIGHT MINE TO THE GREAT FALLS TECTONIC ZONE

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Tom Chadwick
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INTRODUCTION

The Golden Sunlight Mine is situated in Jefferson County, Montana, five miles northeast of Whitehall (Fig. 1). The mine is operated by Golden Sunlight Mines, Inc., a wholly-owned subsidiary of Placer Dome U.S. Inc., of San Francisco, California. Total reserves including those mined to date are approximately 53 million short tons averaging 0.05 ounces gold per ton. Dayton (1984) provided an overview of the operations.

The mine is located within the Great Falls Tectonic Zone (GFTZ) (Fig.1). Recent studies by Golden Sunlight suggest that magmatism and associated gold mineralization at the mine were localized along a portion of this trend. These studies also suggest that the GFTZ may have been active during Proterozoic sedimentation in the mine area.

MINE AREA GEOLOGY

The Golden Sunlight Mine lies on the eastern flank of a north-trending mountain range known as Bull Mountain. Oldest exposed rocks are part of the Proterozoic Belt Supergroup, and are predominately clastic sedimentary rocks. These rocks were deposited near the southern edge of the tectonically active Helena Embayment of the Belt Basin. Recent studies by Golden Sunlight suggest that the entire Proterozoic section in the mine area represents a submarine fan, slope, and shelf complex that prograded northward over basin plain deposits.

The Mid-Proterozoic sedimentary rocks were intruded by at least three plutonic sequences that are assumed to be late Cretaceous to early Tertiary in age. Pheno-latite porphyries, commonly sill-like bodies, are cut by lamprophyre dikes. Irregular hypabyssal intermediate-mafic composition intrusions are also present. Their age relationship is unclear, but they may be comagmatic with the latite. The igneous rocks are typically altered, complicating whole rock and age date analyses.

Gold mineralization occurs within and around a pipe-shaped hydrothermal breccia that cuts the Proterozoic rocks. The breccia pipe is approximately 700 feet in diameter, and plunges 35^{0} to the west-southwest. Over 1,200 vertical feet of breccia have been drilled, and it remains open at depth.

Recent studies by Golden Sunlight revealed that brecciation and mineralization was contemporaneous with latite porphyry magmatism. Potassic alteration and molybdenite mineralization increase in the lower portions of the breccia. The breccia pipe is thought to have formed in an epithermal environment above a molybdenum porphyry system. Schmidt and others (1989) and Porter and Ripley (1985) summarized the geology of the gold system.

RELATIONSHIP TO THE GREAT FALLS TECTONIC ZONE

The Golden Sunlight Mine is located within a northeast-trending tectonic zone that extends from central Idaho to the Canadian border and beyond. Rostad (1978) summarized early work on the trend. Billingsley and Locke (1941) recognized that the northeast-trending "Tertiary dike zone" extended from Idaho into Montana.

O'Neill and Lopez (1985) named the trend the Great Falls Tectonic Zone, and summarized the structural, depositional, magmatic, and geophysical characteristics which define the zone. In central Idaho, the trend has been termed the Trans-Challis Fault System (Kiilsgaard and Lewis, 1985).

The trend appears to represent a deep seated zone of crustal weakness that localized early Tertiary magmatism in Idaho and Montana. Eocene eruptive centers that developed along the GFTZ include stratovolcanoes, cauldrons, and dome complexes. At least 3 graben-shaped cauldrons are bounded by northeast-trending faults (McIntyre and others, 1982; Foster, 1987). Numerous intrusions were emplaced along the zone, and some carry molybdenite mineralization (Rostad, 1978). The igneous rocks and molybdenite mineralization at the Golden Sunlight are likely part of this regional magmatic event.

O'Neill and Lopez (1985) documented recurrent tectonism along the GFTZ from Proterozoic to present in central Idaho, and suggested that it may form the boundary between Precambrian crustal terranes. Hughes (1983) cited evidence for Proterozoic movement along northeast-trending syndepositional structures in submarine fan facies of the Yellowjacket Basin of Idaho. Recent paleodepositional environment studies by Golden Sunlight suggest that the GFTZ may have been active during Middle Proterozoic sedimentation in southwestern Montana as well as Idaho. A 1,200' wide northeast-trending syndepositional graben has been tentatively identified in the mine area. Identification of the graben is based on limited drill data, and should be considered speculative until additional studies are completed.

The graben is defined by northeast-trending, high-angle dip-slip faults that dip towards the graben axis. The age of initial movement along these structures is unclear. However, rapid facies and thickness variations in mid fan channel and slope deposits are confined to the graben. This suggests that the graben was active during Proterozoic sedimentation.

The inferred Proterozoic graben and associated facies appear to have had a strong influence on emplacement of the Cretaceous to Tertiary gold system. The mineralized breccia developed within the graben, and the northeast-trending graben-bounding structures localized high-grade ore zones within the system. In addition, the breccia widens where it intersects submarine gravity flow deposits that accumulated in the graben. These relationships suggest that epigenetic gold mineralization was developed along, and was greatly influenced by this structural and depositional fabric.

SUMMARY

Crustal weakness associated with the GFTZ appears to have influenced development of the Golden Sunlight deposit in two ways. First, magmatism which drove the hydrothermal system was localized along the zone. Second, the northeast-trending structures provided a pathway for brecciation and gold mineralization. Initial movement along the GFTZ in Montana may have begun during Middle Proterozoic time, if not earlier.

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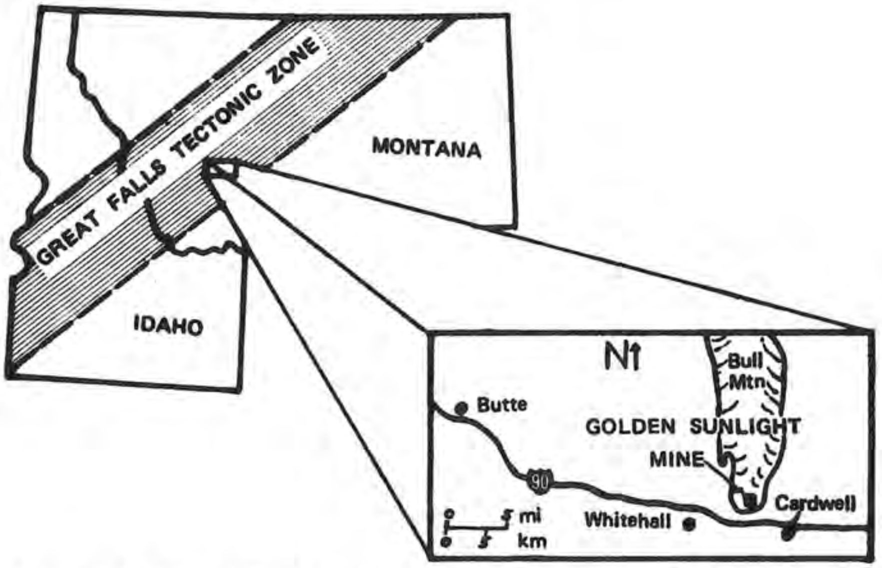

Figure 1. Location of the Golden Sunlight Mine within the Great Falls Tectonic Zone.

The tectonic disruption of a Late Cretaceous and Cenozoic drainage system in southwest Montana and adjacent Idaho in the wake of the Yellowstone hot spot

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Introduction:

Fragments of a Late Cretaceous and Cenozoic drainage system are exhumed in the fault-block mountains of southwestern Montana and adjacent Idaho. These abandoned and tectonically disrupted stream channels contain sedimentary and volcanic sequences that record the transition of the region from a stable, post-orogenic plateau to a basin-and-range province. This transition coincided with drift of the region past the Yellowstone hot spot and formation of the Snake River Plain in Late Cenozoic time (Armstrong and others, 1975).

The drainage system had its headwaters in Middle Proterozoic Belt Supergroup and associated rocks in western Montana and Idaho. We have traced a major trunk drainage across 200 km of southwestern Montana to Humphrey, Idaho, where it passes under the Snake River Plain (Figure 1). A second major drainage apparently came out of the region northwest of Salmon, but we have not yet established the linkage with the trunk drainage in Montana. We suggest that, before the Snake River Plain formed, the drainage system transported many cubic km of quartzite cobbles southeastward to a depositional basin in Jackson Hole, Wyoming (cf. Love and others, 1972).

Our preliminary work suggests that the valley system formed in Late Cretaceous time, and drained the overthrust belt of western Montana and Idaho. In Montana, some of the ancient valleys eroded Late Cretaceous rocks, and cross-cut Late Cretaceous thrust structures. The oldest deposit yet reported in the valley system is a Late Cretaceous (67-Ma) rhyolite flow in the Ruby Mountains east of Dillon (Fritz and Sears, 1989; Satterfield and others, 1989).

Some segments of the valley system were choked by Challis volcanic materials and abandoned. Some of the Challis rocks, especially the fluvial volcaniclastic facies, outline parts of the old drainage network.

Valley segments near Salmon, Idaho:

A 100 m deep channel that we think is part of the drainage system is spectacularly well-exposed 18 km north of Salmon, on the east side of Salmon River canyon along US highway 93. The channel dissects tilted beds of the Proterozoic Yellowjacket Formation, and is filled with well-indurated, cliff-forming conglomerate, part of Anderson´s (1959) Kriley Formation, or

Harrison´s (1985) Kriley Gulch Formation. Anderson (1959) thought the unit was Paleocene or Eocene; Harrison (1985) showed that it is probably Eocene, because it locally overlies Challis volcanics on an unconformity, and perhaps interfingers with other volcanic rocks elsewhere in the Salmon basin. The conglomerate contains abundant clasts of the Middle Proterozoic Yellowjacket Formation, distinctive Proterozoic granitic clasts, and grey quartzite clasts. Clast provenance suggests southeast transport.

The Leesburg basin occupies an ancient southeasterly-draining valley which transported grey quartzite cobbles and boulders, as well as Proterozoic granite, and was choked with Challis volcanics. This valley is now perched on a high plateau above the Salmon basin. Other broad, dry valleys with a veneer of rounded quartzite clasts occur on the plateau.

Valley segment in the Ruby Mountains, Montana:

The best-studied part of the drainage system is exhumed in the Ruby Range east of Dillon, Montana where an ancient valley, 200 m deep and several km wide, was eroded into Archean gniess. The oldest deposit in the valley is 67 million year old rhyolite ash-flow tuff (Satterfield and others, 1989), interlayered with quartzite-cobble conglomerate. The tuff is overlain by layers of conglomerate and fluvial volcaniclastic sandstone of Oligocene and Miocene age (Satterfield and others, 1989). The ancient valley is topped off by blankets of fluvially reworked, white rhyolite ash of Late Miocene age and interlayered quartzite cobble conglomerate (Satterfield and others, 1989). The highest deposits in the valley merge with bajadas that overlap the neighboring hills and reach far into the mountains (Sears and others, 1989). Pliocene basalt from the Snake River Plain flowed north over the gravel deposits (Fritz and Sears, 1989).

Tectonic interpretation:

Figure 2 shows our interpretation for the evolution of the drainage system. Figure 2A outlines the reconstructed drainage system, which transported material out of Idaho and Montana for about 60 million years, from Late Cretaceous to late Miocene time.

Independent evidence from Wyoming shows that between Late Cretaceous and late Miocene time, rivers flowing from the northwest layed down thick deposits of quartzite-cobble conglomerate in the Harebell, Pinyon, and Colter formations of Jackson Hole (Love and others, 1972). Many of the cobbles in Jackson Hole are red, feldspathic quartzites which closely match parts of the Belt Supergroup in western Montana; others are grey quartzites which match Proterozoic rocks near Salmon.

In Late Miocene time (Figure 2B), the paleo-valleys of southwest Montana became choked with rhyolite ash and conglomerate. Four separate layers of fluvially reworked ash are interlayered with conglomerate in the Ruby Mountains. We suggest that the ash erupted from the Miocene Yellowstone hot spot. The Blue Creek and Rexburg calderas along the Snake River Plain erupted at least four distinct ash units in late Miocene time (McBroome and others, 1981).

We suggest that when the ancient drainage system crossed the path of the hot spot uplift, southwest Montana became a closed basin which began to fill with gravel and volcanic materials from the hot spot. At the same time, Jackson Hole became isolated from its external source of quartzite cobbles (Love and others, 1972). Pliocene basalt from the Snake River Plain flowed north, down the reversed stream gradient, and mixed with gravel in southwest Montana.

After reversal of the drainage, southwest Montana broke into the modern basins and ranges, and uplifted segments of the old valley system were exhumed (Figure 2C). The modern drainage of southwest Montana was probably established in Pleistocene time (Sears and others, 1989).

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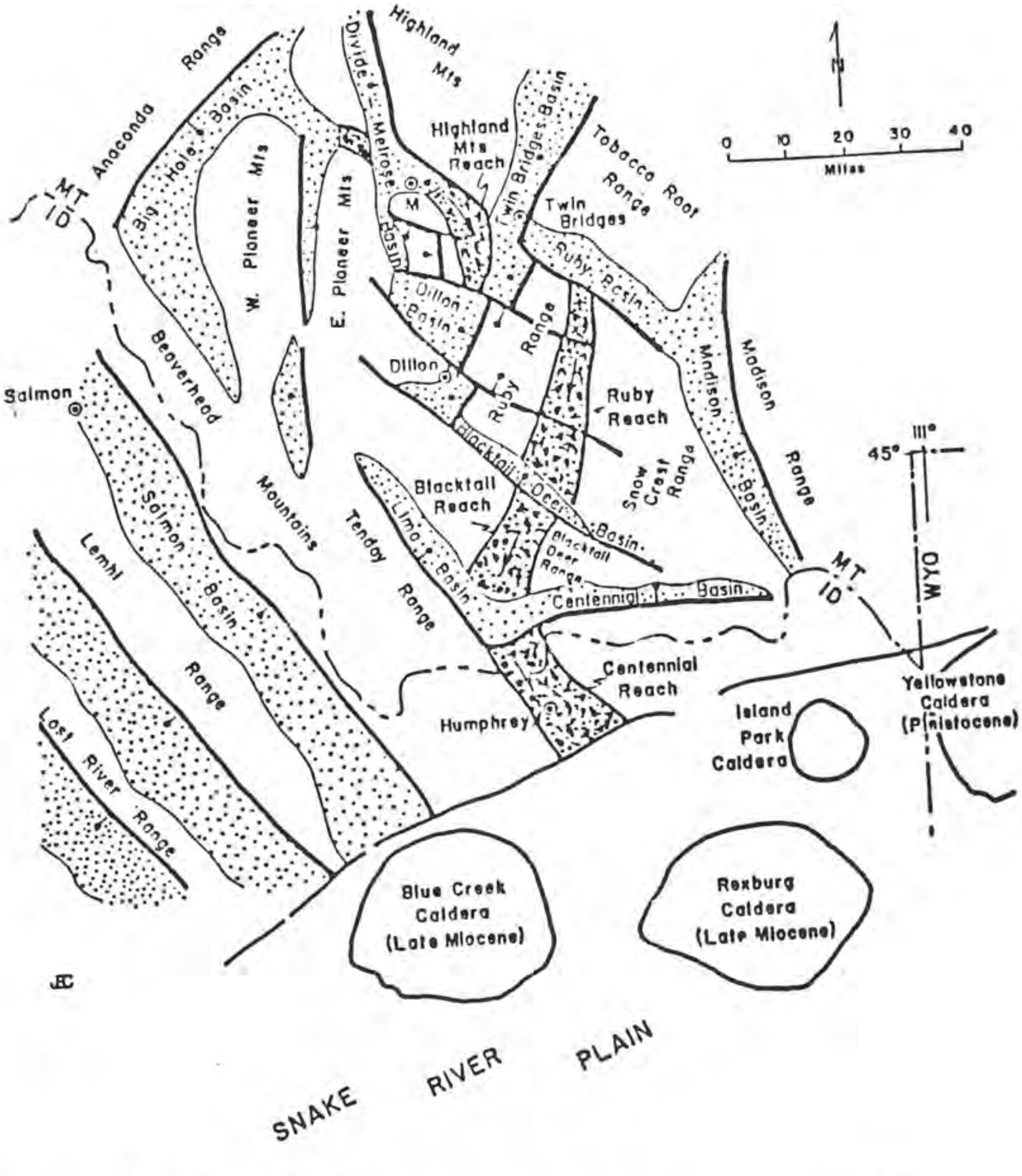

Figure 1. Trunk drainage of the Cretaceous-Tertiary valley system, (heavy stipple) and present basins (light stipple) and ranges of southwestern Montana and adjacent Idaho.

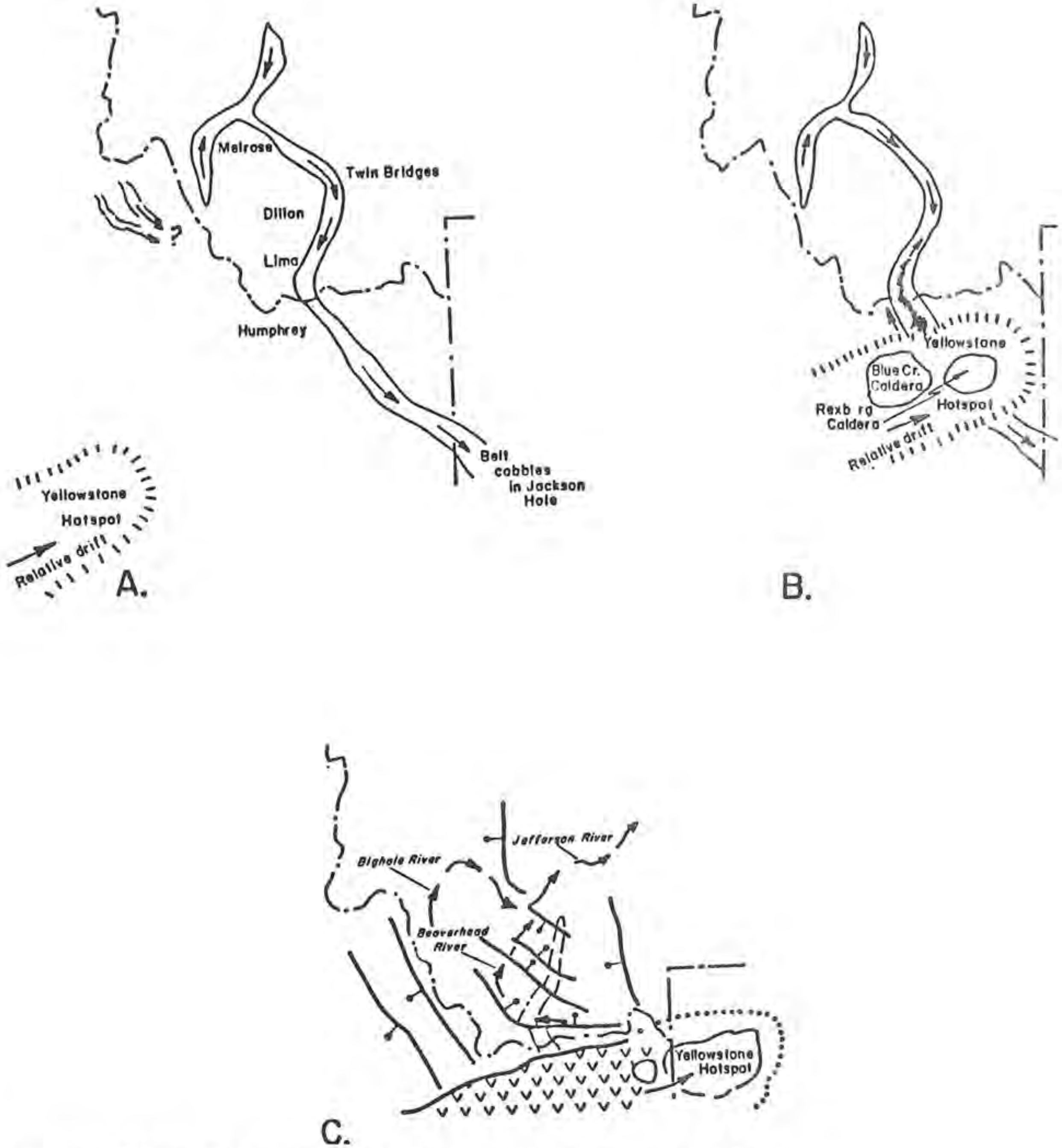

Figure 2. Generalized history of the Cretaceous -Tertiary drainage as the region drifted past the Yellowstone hot spot.

A. Quartzite cobbles from Montana and Idaho are transported to Jackson Hole, Wyoming.

B. Drainage reverses as it crosses Yellowstone hot spot, basalt (black) flows north.

C. Region breaks up into modern basins and ranges, Cretaceous-Tertiary drainage is abandoned, modern drainage forms.

ACCOMMODATION OF EN ECHELON EXTENSION BY CLOCKWISE ROTATION OF THE SAPPHIRE TECTONIC BLOCK, WESTERN MONTANA AND IDAHO.

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INTRODUCTION

In western Montana and central Idaho, the Bitterroot dome (a metamorphic core complex) is a major feature of middle Eocene crustal extension (Figure 1). The genetic relationship between the mid-crustal Bitterroot dome and its flanking crustal detachment, the Bitterroot mylonite, has been known for some time [Chase and Talbot, 1973]. The mylonite separates the Bitterroot dome from the upper-crustal Sapphire tectonic block to the east [Hyndman and others, 1975; Hyndman, 1980]. Hyndman [1980] and Hyndman et al. [1975, 1988] propose sliding the Sapphire tectonic block off the Bitterroot dome in the late Cretaceous. However, recent radiometric age dates indicate an Eocene age of formation and uplift [Chase et al., 1983; Garmezy and Sutter, 1983]. As a result, Chase et al. [1983] suggest pulling the Bitterroot dome out from under the Sapphire tectonic block during the Late Cretaceous and Eocene. Because of the possible tectonic rotation of either the Bitterroot dome or Sapphire tectonic block, these hypotheses are ideal for paleomagnetic testing.

Conveniently, a widespread middle Eocene magmatic event emplaced hypabyssal rhyolitic dikes, volcanics, and plutons in both the Bitterroot dome and Sapphire tectonic block. In the Sapphire tectonic block, the dike swarm trends about N46°E. The swarm is part of the similarly oriented Idaho porphyry belt in central Idaho [Badley, 1977]. To the west, in the Bitterroot dome, the dike swarm trends about N21°E and is perpendicular to the stretching lineation (110°) in the Bitterroot mylonite. The difference (25°) in the trend of the dike swarms supports the separation of the Bitterroot dome from the Sapphire tectonic block since the middle Eocene, (e.g., Chase and others; 1983), but does not conclusively validate either of the tectonic models.

In order to test the models for relative displacement between the Bitterroot dome and Sapphire tectonic block, we present the results of a paleomagnetic study of 19 dikes from three localities in this area.

PALEOMAGNETIC METHODS AND RESULTS

Six to eight oriented cores or two handsamples (Selway locality only) were examined from each dike. Each sample underwent stepwise thermal or alternating field demagnetization until only a few percent of the magnetization remained. Samples typically lose up to 70% of their magnetization by 50mT or 300°C; thereafter, the intensity of magnetization changes little until the Curie temperature of magnetite (580°C), when the intensity drops to the level of measurement error. Generally, the low coercivity component appears to be a modern day viscous remanent overprint. After the removal of this secondary component, most samples have one remaining component of magnetization. This characteristic component of high stability is inferred to be the primary thermal remanent magnetization. Both normal and reversed polarity dikes occur at the Selway and Sula localities and suggest adequate sampling of paleosecular variation. The Lolo locality only contains normal polarity dikes and may not adequately sample paleosecular variation.

Each sample's characteristic remanent magnetization was analyzed using standard principal component analysis; maximum angular deviations of linear segments were generally below 5°. At each site, the principal component directions from each sample were averaged together to produce a site mean direction by the method of Fisher [1953]. The same methods were used to calculate locality mean directions and their associated 95% confidence intervals (α_{95}). Further paleomagnetic detail and results may be found in Doughty [1990].

Diehl and others [1983] published a North American paleomagnetic pole, calculated from alkalic intrusives in eastern Montana, for the period from 47 to 54 mybp. Because this pole may contain some regional non-dipole components of the Eocene field also present at our three localities, it provides the most reliable reference pole for our data. The reference pole, located at 82.0°N and 170.2°E, yields an expected declination of 348.5° and inclination of 65.5° at our localities. Some of our results differ from the expected direction and when analyzed by standard techniques (Beck, 1980; Demarest, 1983) suggest relative rotations among the three localities (Table 1).

Table 1. Paleomagnetic results from the Bitterroot dome/Sapphire tectonic block, Idaho and Montana.

Locality	Location	N	α_{95}	k	Dec.	Inc.	R	± R	F	± F
Lolo	N. B'dome	5	7.2	114.1	331.6	59.5	-16.8	12.2	6.4	6.6
Selway	W. B'dome	6	9.5	51.1	336.8	66.5	-11.7	19.7	-1.0	7.8
Sula	S. S.T.B.	8	12.9	19.3	44.4	69.0	55.9	30.6	-3.8	10.4

Note: N, number of sites; k, Fisher (1953) precision parameter; α_{95}, radius of 95% confidence cone in degrees; Inc. and Dec., inclination and declination of locality-mean in degrees; R and ± R, Rotation ($D_{observed}$ - $D_{expected}$) and 95% confidence interval in degrees; F and ± F, Flattening (I_{ex} - I_{ob}) and 95% confidence interval in degrees.

The Sula locality, in the southern Sapphire tectonic block, displays a rotation (R) of 55.9° (± 30.6°) and 25.3° of significant clockwise rotation at the 95% confidence level. The Selway and Lolo localities, both within or near the Bitterroot dome, exhibit 0° and -4.6° of rotation at the 95% confidence level. No locality shows significant flattening. The Lolo locality may not sufficently sample paleosecular variation and may not provide reliable rotation or flattening results. Assuming that the Selway and Sula localities adequately average paleosecular variation, there is no evidence for post-Eocene rotation of the Bitterroot dome. However, our data indicate 25.3° of clockwise rotation of the Sapphire tectonic block about a nearby vertical axis.

TECTONIC IMPLICATIONS

Although there is no paleomagnetic evidence that rocks within the Bitterroot dome have rotated, the geologic evidence argues for translation since the Eocene. The Bitterroot dome appears to be one of several en echelon metamorphic core complexes linked together along the right-lateral Lewis and Clark fault system [Ewing, 1980; Rehrig and Reynolds, 1981; Sheriff and others, 1984; Rehrig and others, 1987]. Although the system probably had a left-lateral component during the late Cretaceous [Hyndman et al., 1988], post-Eocene right-lateral displacement has been estimated at approximately 68km [cf., Harrison et al., 1974]. The amount of apparent crustal thinning south of the fault system suggests 40km of right-lateral offset [Sheriff et al., 1984]. Given the orientation of the Lewis and Clark line and the uncertainty in the data, we cannot resolve the extent of westward translation of rocks within the Bitterroot dome. However, with their location south of the Lewis and Clark line, these rocks probably lie some 40km or more west of their original position.

The eastern end of the Lewis and Clark line bounds the northern edge of the

Sapphire tectonic block and displays a minor amount of right-lateral displacement in this area. Much of this Eocene displacement may result from minor extension east of the Sapphire tectonic block. This relation prohibits significant clockwise rotation of the Sapphire tectonic block about a vertical axis located in the southern portion of the study area. Rather, the structural continuity across the Lewis and Clark line suggests rotation about a vertical axis in the northern half of the Sapphire tectonic block. Thus the southern portion of the Sapphire tectonic block moved west, most likely in conjunction with movement of the Bitterroot dome, while the northern portion remained relatively fixed.

The geologic and paleomagnetic data are consistent with a clockwise rotation of the Sapphire tectonic block as the Bitterroot dome moved westward out from beneath it (Figure 1). These data also support the assertion that displacement on the Bitterroot mylonite reaches a maximum at the northern end and decreases toward the south [Hyndman et al., 1988]. This produced a narrowing of the Bitterroot Valley toward the south. Because the Bitterroot mylonite effectively detached the two tectonic domains at the northern end, the northern margin of the Sapphire tectonic block remained stationary. At the southern end, where the tectonic domains were not separated, they both moved some 40km toward the west.

At the southern margin of the Sapphire tectonic block, extension was transferred eastward to the Big Hole basin. The Big Hole basin narrows toward the north, opposite from the taper of the Bitterroot Valley. The geometry of these basins suggest an en echelon extensional system with zones of high east-west extension underneath the northern Bitterroot Valley and the central Big Hole basin. Perhaps the clockwise rotation of the Sapphire tectonic block results from an upper crustal block which spans the transition zone between these two en echelon decollements. Rather than fail along a dextral strike-slip fault, the Sapphire tectonic block rotated clockwise above a mid-crustal detachment and around a northerly axis.

This behavior is consistent with the regional geology. West of the Bitterroot mylonite, middle Eocene extension proceeded along a related set of metamorphic core complexes connected by dextral strike-slip faults. Many such faults of the Lewis and Clark line parallel the stretching lineation within the Bitterroot mylonite and intersect extensional basins at right angles. For example, extension within the Priest River complex and Bitterroot dome is coupled together along the St. Joe and Kelly Creek faults. The Boehls Butte area (Figure 1) may be another core complex related to this extensional system. Presumably, this type of ductile extension stepped south of the Bitterroot dome into the Salmon River arch or to the east in the Pioneer block. East and north of this system, in the brittle carapace, extension becomes less organized and large crustal rotations are expected.

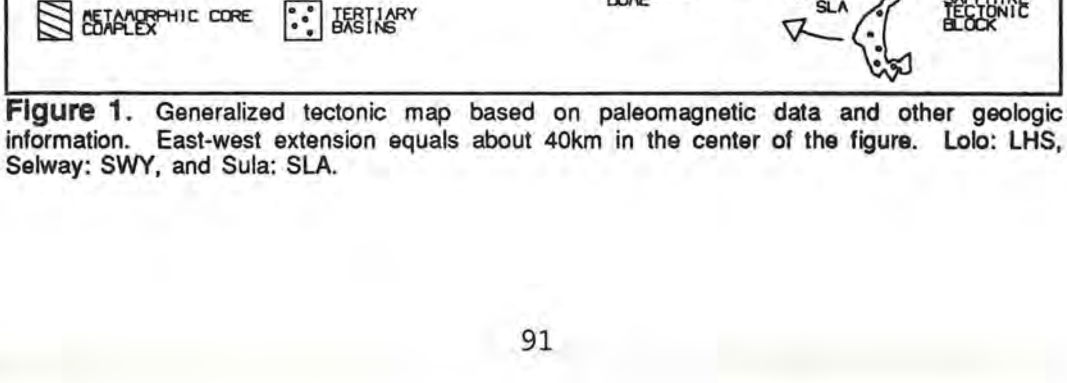

Figure 1. Generalized tectonic map based on paleomagnetic data and other geologic information. East-west extension equals about 40km in the center of the figure. Lolo: LHS, Selway: SWY, and Sula: SLA.

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COBALT MINERALIZATION IN THE NORTHERN LEMHI RANGE, LEMHI COUNTY, IDAHO

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The northern end of the Lemhi Range is littered with abandoned mines and prospects (fig. 1). Many of the mines and prospects were developed in dark-colored, impure lithic arkose (quartzite) of the Middle Proterozoic Yellowjacket Formation, a known cobalt producer to the west. Less important hosts include an overlying sequence of lighter colored Middle Proterozoic arenites (Lemhi Group?), a still younger sequence of Lower Paleozoic sedimentary rocks, and small Tertiary(?) granodioritic plugs.

Historic production in the area has been small and has consisted entirely(?) of copper and byproduct silver and gold (Ross, 1925, p. 31). Production appears to have been largely confined to discordant shear zones and quartz veins. Primary mineralization was dominantly chalcopyrite, although pyrite was locally important. Surface exposures are strongly weathered and signs of sulfide mineralization away from the mines and prospects include extensive gossanization and small but widespread shows of malachite and azurite.

A suite of 108 rock samples was collected from outcrop, from float, and from mine dumps and prospects during a reconnaissance examination of the area during the summer of 1989. The rocks contain anomalously high amounts of many metals other than copper and silver. Copper-rich samples also contain high amounts of antimony, bismuth, cobalt, gold, lead, molybdenum, nickel, yttrium, and zinc. The highest observed concentrations of these metals (in parts per million) were as follows:

Metal	Concentration	Source, location
Ag	200 ppm	Quartz vein, Copper Creek prospect
Au	.2 ppm	Quartzite breccia, Porterfield prospect, and quartz vein, 12-mile Canyon prospects
Bi	1000 ppm	Quartz vein, 12-mile Canyon prospects
Co	2000 ppm	Quartz vein, Harmony mine
Mo	30 ppm	Quartzite breccia, Pope-Shenon mine
Ni	150 ppm	Chalcopyrite, Harmony mine
Pb	1000 ppm	Quartz vein, Copper Creek prospect
Sb	1000 ppm	Quartzite breccia, Pope-Shenon mine
Y	2000 ppm	Quartzite breccia, K Mountain quarry
Zn	3000 ppm	Chalcopyrite?, prospect A

(Analyst: R. Hopkins, U.S. Geological Survey)

These metals, excluding yttrium and perhaps gold were originally carried in sulfide minerals, either in their own phases or as impurities. The copper-yttrium association is an odd one, inasmuch as yttrium is not a chalcophile element, but it might be an artifact of the weathering process.

The most interesting geochemical feature of the northern Lemhi Range is the presence of a strong cobalt anomaly. Eight samples contained 100 ppm or more cobalt; five of these were collected in or around the Harmony mine, one was collected from a prospect in Mormon Canyon, and one each were collected from prospects A and B (fig. 1). The Yellowjacket Formation in the Blackbird Mining District 35 km to the west is host to large stratabound cobalt deposits. A principal ore host there is an iron-rich metasedimentary rock consisting mostly of biotite (mafic tuff of Hughes, 1983). Similar biotite-rich rocks containing over 20 percent total iron oxide were widely observed in the northern Lemhi Range and strongly suggest that the Yellowjacket there includes strata equivalent to the ore zone of the Blackbird Mining District. If these strata reflect mineralizing processes genetically related to those at Blackbird, then these processes in the northern Lemhi Range resulted in major copper-silver deposition with only minor cobalt, as opposed to major copper-cobalt deposition at Blackbird.

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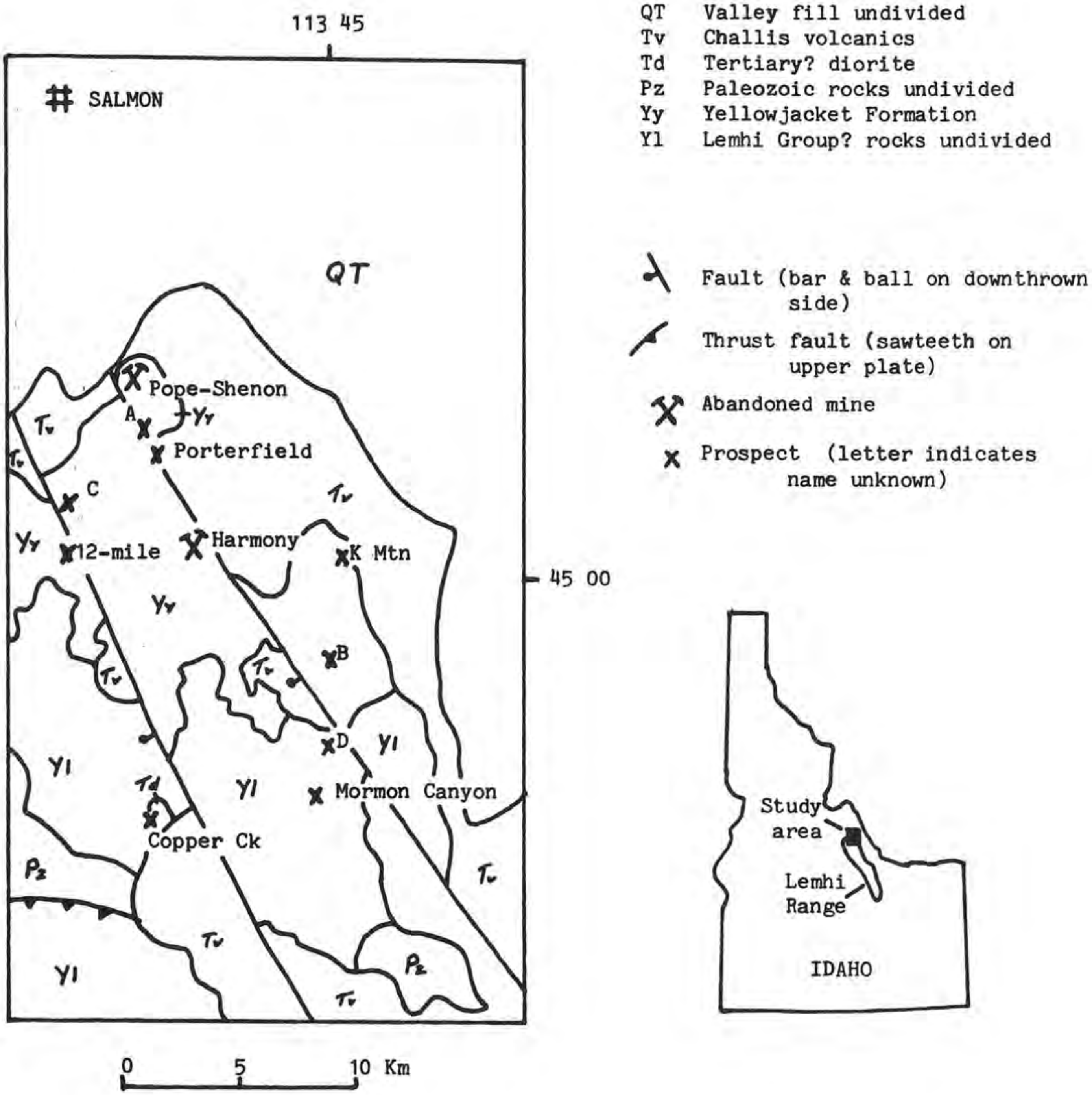

Figure 1. Index map and geologic map showing mines and prospects in the northern Lemhi Range, Lemhi County, Idaho.

LATE PROTEROZOIC(?) TUFF NEAR CHALLIS, IDAHO

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Abstract

Drill exploration near Challis, Idaho has delineated a new unit, of volcanic nature, located beneath the lower Paleozoic Bayhorse anticline sequence. The unit is a silicic, possibly rhyolitic, lithic tuff, probably of subaerial origin. Source, correlation, and regional significance is uncertain.

No tectonism, and little apparent hiatus in deposition is evident at the contact with overlying dolomite; the basal contact was not penetrated.

Inclusion within the Bayhorse sequence and low-grade regional metamorphic fabric preclude correlation with the voluminous Eocene volcanic deposits of central Idaho. Possible temporal association with lowermost Bayhorse rocks implies a late Proterozoic or early Cambrian age; however, the tuff could be much older.

Introduction

Diamond-drill exploration conducted in 1987 by Shama Minerals Corporation near Challis, Idaho penetrated known strata of the Bayhorse anticline prior to intersection of lithic tuff not previously discovered in central Idaho. The hole is located approximately 1/2 mile northwest of Daugherty Spring in T13N R18E, sec 4 of the Boise Principal Meridian (Figure 1).

The Bayhorse Anticline is a broad north-south trending open antiform exposing a generally concordant sequence of at least 5000 feet of weakly meta-

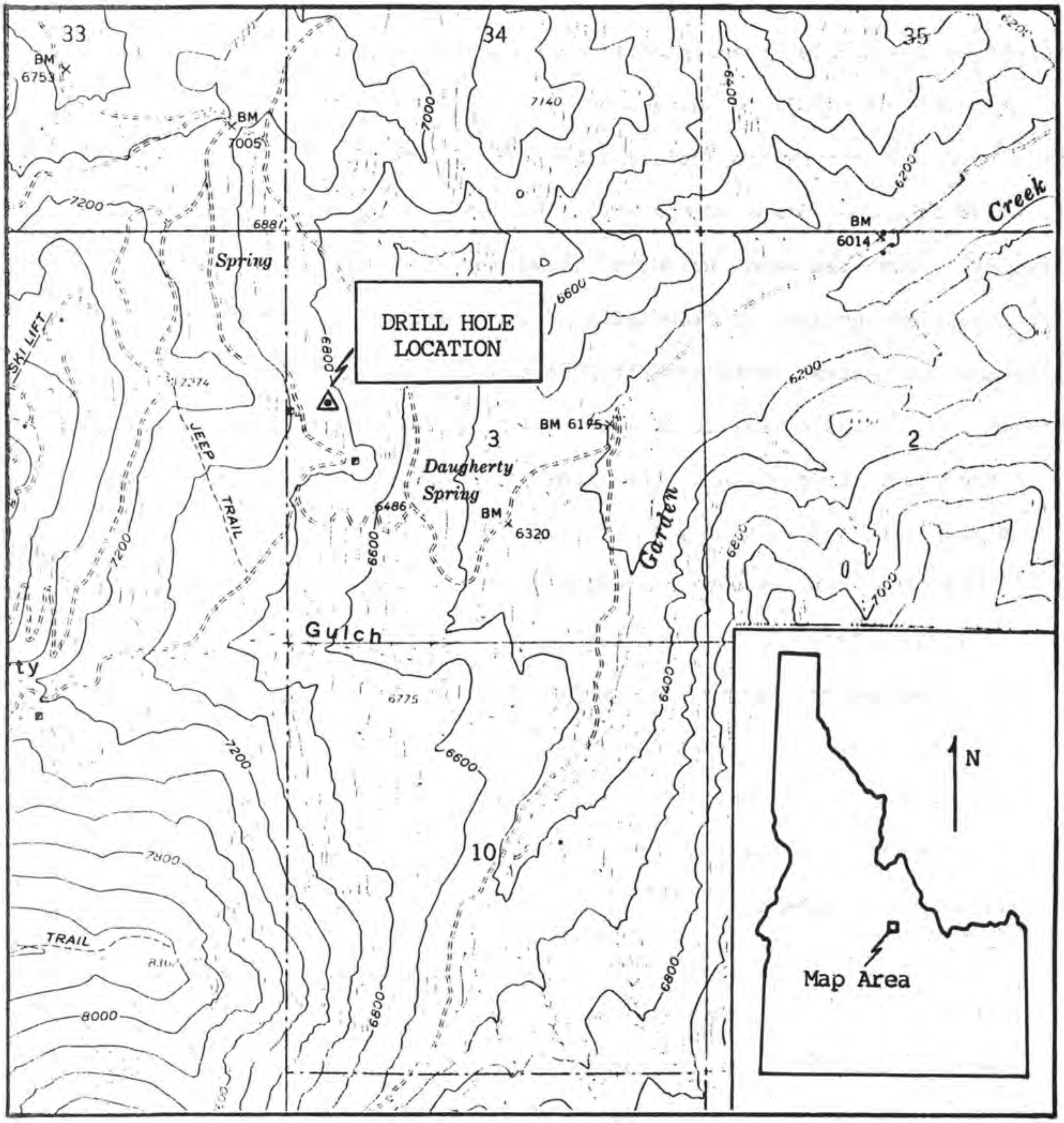


Figure 1. Map showing drill site approximately five miles west of Challis, Idaho. Base from U.S. Geological Survey Bayhorse, Idaho quadrangle. Scale 1:24,000.

morphosed (to middle greenschist facies) marine sediment, and has been described by Ross (1937), Hobbs and others (1975), and Hobbs (1985). The four main units comprising the anticline are, from top to bottom, Ramshorn Slate, Bayhorse Dolomite, Garden Creek Phyllite, and informally named (Hobbs and others, 1975) Bayhorse Creek Dolomite. These workers have tentatively designated the strata as Cambrian to Ordovician in age. Definitive correlation with rocks elsewhere in central Idaho has not been possible due to regional structural complexity (including thrust faulting) and widespread Eocene volcanic cover.

Megascopic inspection of the lithic tuff does not readily disclose a volcanic affinity, and initially led to interpretation of the unit as diamictite, likely late Precambrian glaciomarine tillite, following the descriptions of Ludlum (1942), Aalto (1971), and Condie (1969). However, recent collaboration with U.S. Geological Survey personnel (S.W. Hobbs, B.F. Leonard, D.H. McIntyre) in identification of these rocks, including petrographic work by Leonard and McIntyre (unpublished), has disclosed the volcanic nature of matrix components and hence tentative identification as silicic, lithic tuff, possibly ash-flow tuff or lahar.

Resolution of this metavolcanic into the scheme of central Idaho geology is problematical as no rocks of similar description, particularly with regard to age, are known to exist in Idaho geology. Therefore the purpose of this paper is to present descriptive material, rather than interpretation

Drill Hole Stratigraphy/Structure

The 3749 foot vertical hole collared adjacent to an outcrop of Bayhorse Dolomite on the east limb of Bayhorse anticline, near the northernmost exposure of that structure. Intersected were 11 feet of overburden, 488 feet of Bayhorse Dolomite, 2866 feet of Garden Creek Phyllite, 196 feet of Bayhorse Creek Dolomite, and 189 feet of lithic tuff (in which the hole terminated). Estimated true thickness is 471 feet, 2482 feet, 170 feet, and 164 feet for the Bayhorse Dolomite, Garden Creek Phyllite, Bayhorse Creek Dolomite, and tuff sections, respectively, assuming an average 30° dip for all but the Bayhorse Dolomite in which bedding dips near 15° (bedding/foliation steepens and remains near 30 degrees at depth).

Contacts between units are conformable, excepting the Bayhorse Creek Dolomite/lithic tuff contact which is, in the strictest sense, a nonconformity; however, the top of the tuff has been weakly sorted by sedimentary processes and rests conformably beneath the dolomite. The Bayhorse Dolomite/Garden Creek Phyllite contact is gradational over approximately 5 feet, the Garden Creek Phyllite/Bayhorse Creek Dolomite contact over 2 feet.

Strong internal deformation throughout the Garden Creek Phyllite section is evident from wide variation in dip of lamination and schistosity (which are locally parallel but commonly intersect at angles between 20° and 50°), and the local occurrence of small-scale folds and crenulation.

Handspecimen Descriptions

BAYHORSE DOLOMITE

Bayhorse Dolomite is light to medium grey to brown in color with local sandy, silty, and phyllitic horizons, the latter generally of very light grey color. One such phyllitic horizon at the base of the unit grades, over a 5 foot interval, into the Garden Creek Phyllite. Although most of the unit is massive, some horizons display fine to coarse lamination. Beds range generally from 3 to 12 feet in massive dolomite, 1 to 5 feet where phyllitic. Massive beds are fine- to medium-grained crystalline.

GARDEN CREEK PHYLLITE

Garden Creek Phyllite is dark grey to black carbonaceous, but locally imbued with white, calcite-rich laminae (the overall color is nevertheless dark grey). Disseminated fine- to medium-grained, euhedral pyrite (authigenic?) is widespread in concentrations to 2%, and appears laminae-controlled. Stringers of fracture-fill pyrite occur locally.

Horizons rich in dark grey, silty, carbonaceous limestone and dolomite are common within the lower 1200 feet of the unit, constituting as much as 50 percent of the section in beds ranging up to approximately 4 inches in thickness. Dolomite predominates in the lower 500 feet. Schistosity and lamination are less pronounced within phyllite of the carbonate zone; slaty cleavage is locally well developed.

BAYHORSE CREEK DOLOMITE

This white to light grey, massive, finely crystalline dolomite locally contains sandy horizons with poorly defined bedding. Some fractures are coated with white sericite. A greenish cast near the base of the unit results from incorporation of small amounts of sericite-chlorite within the groundmass.

LITHIC TUFF

The top of the tuff unit is marked by approximately 6 inches of weakly sorted (clast-free), weakly bedded quartzite with local sericite-chlorite laminae which fines upward into 2 inches of light green, sericite-chlorite phyllite. The overlying dolomite rests with apparent conformity upon the phyllite, although the contact is fractured (parallel to bedding).

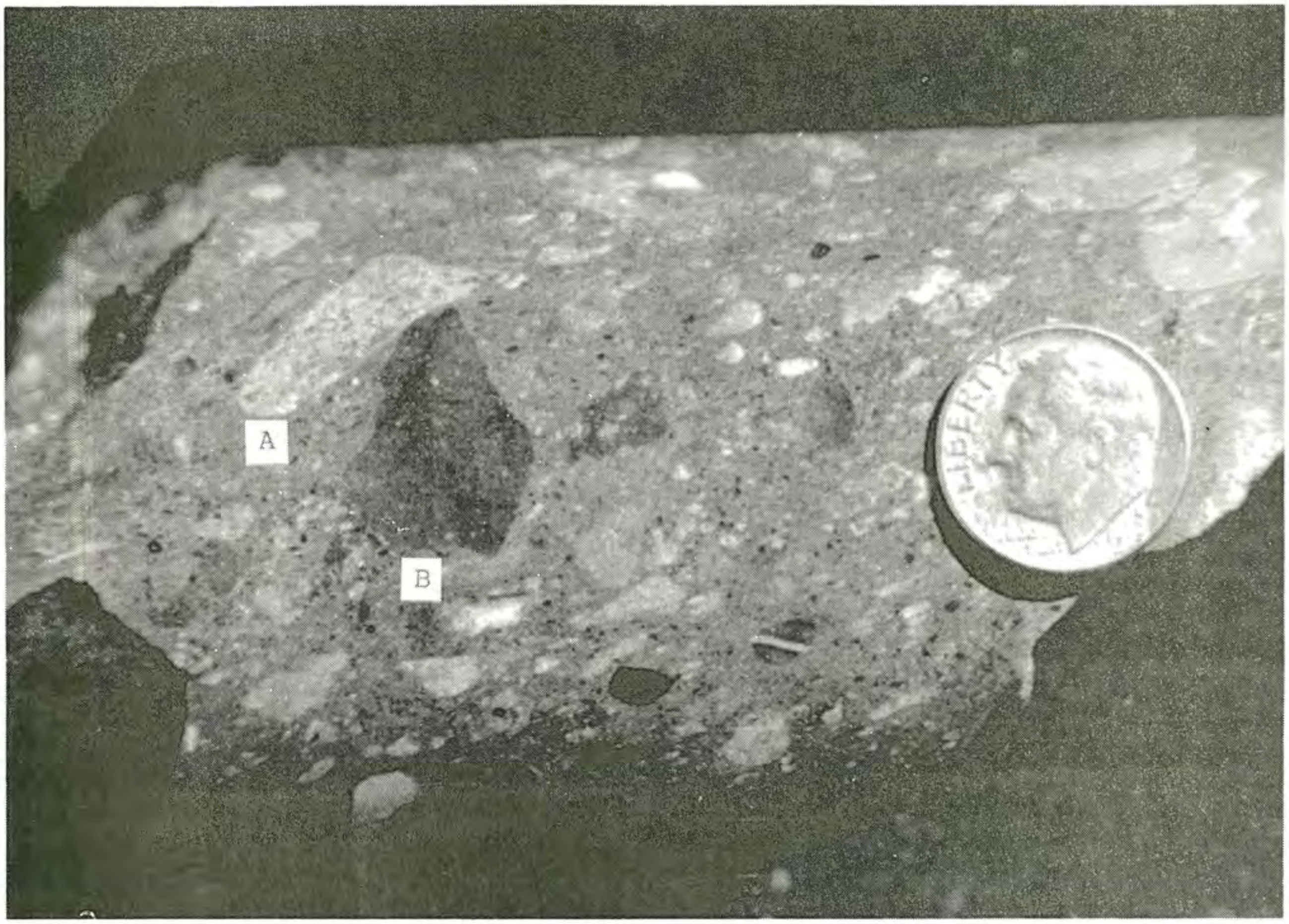
The body of the lithic tuff comprises light to medium grey/green to grey, sericite-rich, weak to moderately phyllitic rock (Figure 2). Weak schistosity, imparted by the platy sericite and chlorite, is pervasive within the matrix but affects few clasts. Although metamorphic index minerals are lacking, the presence of schistosity, and inclusion within a sequence affected by low-grade regional metamorphism, implies lower greenschist facies metamorphism within the tuff unit.

The clast/matrix ratio, megascopically, ranges from approximately 25/75 to 35/65 (except at the top as described above). Clasts are of quartzite (maroon to pink to purple), argillite/phyllite (light green to dark green), dolomite and marble (tan to green), felsic and intermediate volcanic porphyry (tan to grey to green), welded tuff, and rare pumice. Clasts are subangular to moderately rounded, and mainly between 1/8 inch and 3 inches in diameter; megaclasts of quartzite to 10 inches occur rarely. Quartzite clasts decrease in abundance downsection. Size sorting among clasts or matrix is absent.

Tabular clasts are crudely aligned within the plane of schistosity, possibly defining crude primary foliation or bedding; some, excepting quartzite, are slightly flattened. Schistosity is locally deformed about relatively large or competent clasts.

Supporting the extraordinarily diverse assortment of lithic fragments is a moderately crystal-rich matrix of silicic, possibly rhyolitic composition; some felsic volcanic clasts may(?) be genetically related to the groundmass. In thin-section the matrix displays faint to distinct wavy foliation which can be interpreted to mimic primary welding, suggesting an ash-flow tuff in which foliation is indented by lithic fragments.

Original matrix constituents have been altered to a very-fine-grained ag-



Tom Jacob

Figure 2. HQ core of lithic tuff collected at a depth of 3576 feet. Polylithic assortment of clasts is supported in a matrix of weakly phyllitic, tuffaceous material. Note slightly flattened felsic volcanic clast (A) and adjacent angular quartzite clast (B) left of center. Schistosity/bedding crosses diagonally from lower left to upper right.

gregate of quartz, sericite, unidentifiable clay minerals, opaque specks, and locally chlorite, calcite, zoisite, and K-feldspar.

Quartz grains constitute 5 to 10 percent of the matrix, and are quartz phenocrysts rather than detrital grains (rounded grains were not recognized). Some are identical to those of felsic lithic fragments: blunted, dipyrimidal, near-equant, high-quartz form, resorbed edges, local crackling, rhyolitic inclusions (B. F. Leonard, written communication, 1990). Broken phenocrysts are common, but do not display these features. Feldspar phenocrysts (uncommon) are altered to carbonate, rare mafic minerals to a mixture of quartz mosiac, opaque oxides, and minor carbonate.

Conclusions

The lithic tuff unit described herein is a new addition to the puzzle of late Precambrian, early Paleozoic stratigraphy in central Idaho. Recognition within a single, relatively deep drill hole limits the scope of definitive evaluation.

Tectonic setting of the tuff is unknown. The diverse collection of lithic fragments suggests close proximity to an area of broad uplift (the craton/continent margin?), rather than an off-shore volcanic island source.

Although the tuff underlies marine strata with apparent structural conformity (no angular discordance), fabric evidence suggestive of subaqueous deposition was not recognized. Tentative identification as ash-flow tuff implies subaerial origin, the unit then inundated by marine transgression. Deposition as lahar is a possibility.

No correlative volcanic rocks are known to exist. The only recognized pre-Eocene volcanic rocks in central Idaho are marine-deposited mafic tuffs within the Yellowjacket Formation (Hughes, 1982), of middle Proterozoic age (Ruppel, 1975). Position within the Bayhorse sequence constrains the tuff as no younger than late Proterozoic or possibly early (?) Cambrian in age.

Lithic fragments, excepting some of volcanic origin, are extraformational and of varied provenance. A possible source are late Precambrian formations to the east, such as the quartzite-dolomite-argillite of Leaton and Pennal Gulch, Lawson Creek Formation, or Swauger Quartzite mapped by McIntyre and Hobbs (1987) in the Challis, Idaho 15 minute quadrangle. This suggests deposition near the Proterozoic/Paleozoic boundary.

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Acknowledgements

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STRUCTURES IN THE ELK CITY, IDAHO REGION AND THEIR RELATIONSHIP TO THE TRANS-CHALLIS FAULT SYSTEM

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Two predominant sets of steeply-dipping faults were recognized during recent geologic mapping by the Idaho Geological Survey in the Elk City region. The first set of faults has a northerly trend (north-northeast to north-northwest) and the second a northeast trend.

Most of the northerly-trending faults are present west of Elk City. The Orogrande shear zone, which cross-cuts both Precambrian metasedimentary rocks and the Cretaceous Idaho batholith, is an example of one of these structures. Gold mineralization along the northerly-trending structures is spatially associated with the roof/batholith contact and is probably Late Cretaceous in age. Recurrent motion along some of these faults has down-dropped unconsolidated sediments of Miocene(?) age.

The northeast-trending faults cross-cut rocks of Precambrian to Eocene age southeast of Elk City. The Blanco Creek and Bargamin Creek shear zones are the two largest structures with this orientation. The extent of pre-Tertiary motion along these northeast-trending faults is unknown. Pink granite of Eocene age has intruded along, and been cut by, the Bargamin Creek shear zone. The area with northeast-trending faults also contains numerous northeast-trending Eocene dike swarms. Mylonitized Eocene dikes within the Bargamin Creek shear zone have s-c fabrics which indicate down to the west normal motion. Hydrothermally altered Eocene dikes are present within the Blanco Creek shear zone, but mineral production from this zone has been negligible. The majority of the other northeast-trending faults are not mineralized.

The northeast-trending fault set in the Elk City region parallels the Trans-Challis fault system, which is present 80 km to the southeast. Both are interpreted as coeval structures formed as a result of northwest-southeast extension during Eocene time. The spatial relationship between Eocene magmatism and northeast-trending faults, found in both the Elk City region and in the trans-Challis system, implies a link between extension and magmatism during the Eocene in central Idaho.

TECTONO-METALLOGENIC MAP OF THE DILLON 1° x 2° QUADRANGLE, MONTANA AND IDAHO

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This 1:250,000-scale map shows the relationship between structures, rock units, and mineral deposits in such a way that inferences can be made about controls of mineralization and exploration targets can be readily defined. The map covers 46 mining districts along with numerous mineral occurrences outside of formal mining districts. Principal tectonic elements are the Medicine Lodge thrust plate, Grasshopper thrust plate, the frontal thrust and fold zone, Ruby and Tobacco Root uplifts of the Rocky Mountain foreland, the Boulder, Pioneer, and Idaho batholiths, and part of the northeast-trending Trans-Challis-Great Falls tectonic zone or lineament (Fig. 1) of O'Neill and Lopez (1985) and Kiilsgaard and others (1986). Each thrust plate has a distinctive stratigraphic sequence and structural style (Ruppel and others, 1981), whereas the foreland uplifts have similar stratigraphic units. Major structures include thrust faults, reverse faults, high-angle faults with dip slip and(or) strike slip, folds, shear zones, and a caldera.

Rock units shown on the map were chosen so as to emphasize favorable hosts for mineralization. Proterozoic units that contain syngenetic and(or) remobilized metals include the Helena-Wallace interval (middle carbonate of the Belt Supergroup), Chamberlain Shale-Newland Limestone undivided, Yellowjacket Formation, and Lemhi-Swauger formations undivided. Felsic intrusions of Proterozoic age are also shown separately.

Mississippian through Cambrian strata are mostly carbonates and commonly host replacement deposits adjacent to intrusions. Tertiary and(or) Cretaceous intrusive rocks and Tertiary volcanic rocks are shown as felsic, intermediate, or mafic. Quaternary and Tertiary strata and sediments undivided constitute widespread basin fill and surficial deposits with associated gold placers which have dominated mineral production in this quadrangle. The Virginia City district (Fig. 1, location 3) has estimated production of 3,000,000 ounces of placer gold in that part of the district covered by the Dillon quadrangle (Loen and Pearson, 1984).

In addition to the mineral deposit types discussed above, there are veins hosted by nearly all of the rock units noted, disseminated deposits (mostly molybdenum-bearing in felsic intrusions), syngenetic-stratiform massive to semi-massive sulfide bodies in Proterozoic strata, podiform chromite in crystalline basement rocks, and breccia- or stockwork-hosted deposits (production dominated by the Golden Sunshine mine in the Cardwell district).

The total value at the time of production for the Dillon 1° x 2° quadrangle was estimated to be slightly over $191 million by Loen and Pearson (1984), with most of the value from gold. Minimum

production of gold is estimated to be about 5,000,000 ounces, not **including the recently opened Beal deposit in the Siberia** (German **Gulch) district. Gold reserves at the Golden Sunlight** mine (Fig. **1, location 1) and the Beal deposit (Fig. 1, location 2)** are about **2,450,000 and 440,000 ounces, respectively (Foster, 1989;** Hastings **and Harrold, 1988). Exploration potential for precious** metals is **deemed to be excellent in the Dillon quadrangle,** judging from the **presence of favorable host rocks and structures** and the styles of **mineralization.**

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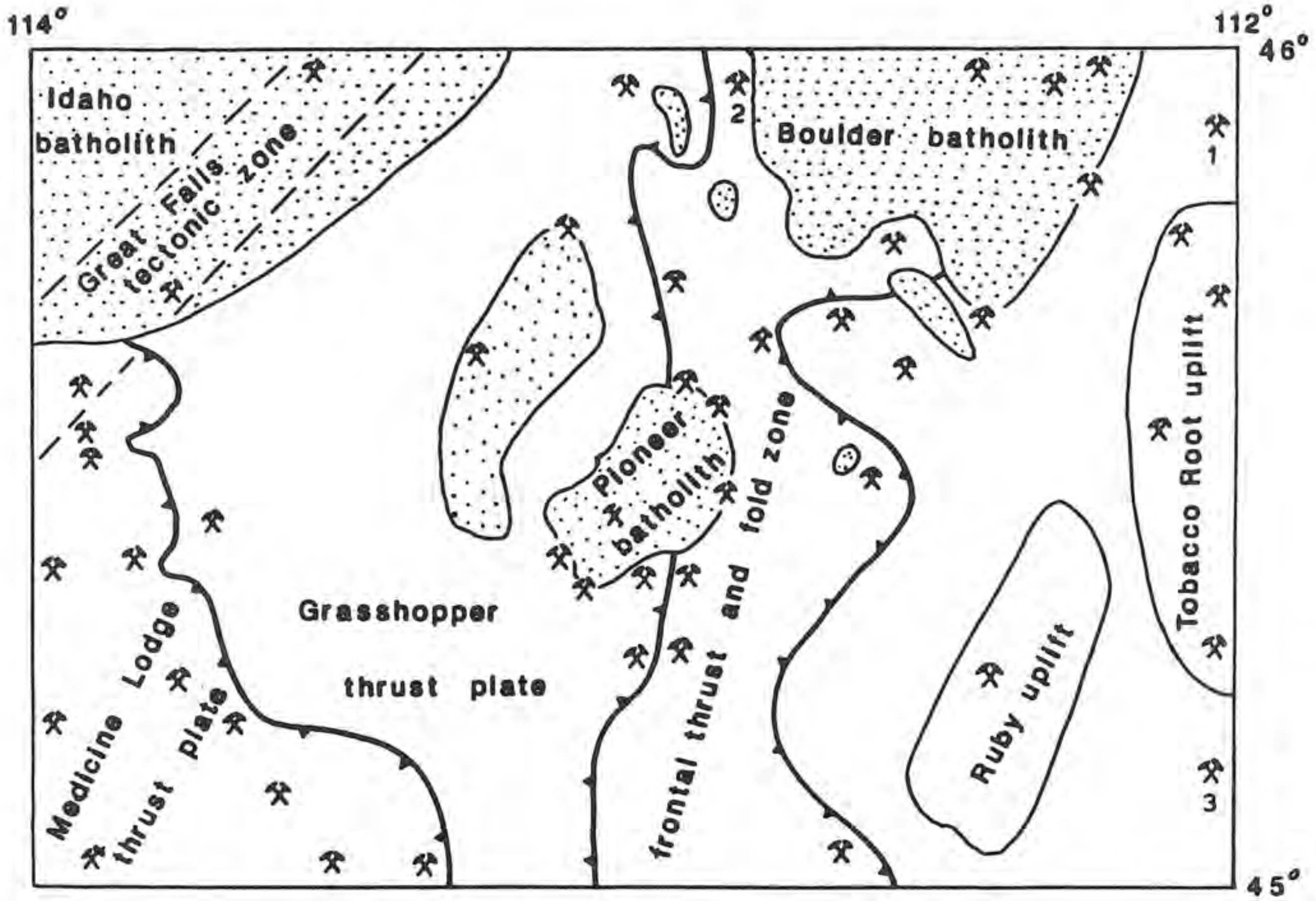
Figure 1. Generalized tectonic map of Dillon 1° x 2° quadrangle showing mining districts (⚒). Tertiary-Cretaceous intrusions stippled. Numbered localities noted in text.



VLF RESISTIVITY SIGNATURES OF SOME MINERALIZED STRUCTURES IN EAST-CENTRAL IDAHO

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INTRODUCTION

Resistivity surveys can help delineate subsurface structural features because resistivity of earth materials varies with mineral composition, degree of fracturing and moisture content. In most geologic environments, rocks which are highly fractured and have high clay and moisture contents have relatively low resistivity. Conversely, unaltered and unfractured crystalline rocks and rocks with high silica and low moisture contents have relatively high resistivity.

Measurements of magnetotelluric-type resistivity and phase at Very Low Frequencies (VLF-R) provide detailed information on geologic structure between the surface and depths of approximately 50 to 300 feet. Inversion modeling of the data allows apparent resistivity to be calculated at various depths, providing output which can be displayed as multi-layer contour maps, surface nets or as pseudosections (Grissemann and Reitmayr, 1972; Van der Poel, 1988, 1989). VLF-R signatures of ore-bearing structures vary with geologic setting, but often provide indications of continuity, geometry and genetic processes.

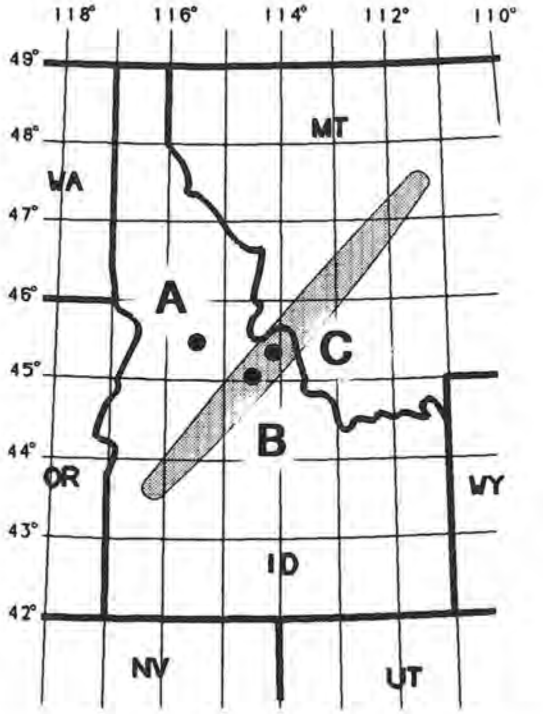

Figure 1. Location of deposits studied with VLF-R. A = Robinson Dike, B = Trail's End, C = Beartrack. Patterned area indicates approximate boundary of the trans-Challis fault system and possible extension into Montana as the Great Falls tectonic zone (Kiilsgaard and others, 1986; O'Neill and Lopez, 1985).

The Robinson Dike deposit (Figure 1, A) lies north of the trans-Challis fault system of Kiilsgaard, Fisher and Bennett (1986) and partially occurs in a zone adjacent to a strong resistivity low. The signature has similarities to those of other deposits in the region where clay development is the most geophysically conspicuous alteration feature in or adjacent to ore zones. These signatures often have a close spacial association with Tertiary intrusives. The Trail's End and Beartrack deposits (Figure 1, B and C) locally host mineralization within resistivity highs caused by silica and/or carbonate concentrations. The high-resistivity zones are wholly or partially surrounded by resistivity lows caused by varying degrees of clay development in the alteration envelopes. Variants of this resistivity signature are common where steep faults or fractures served as plumbing for hydrothermal activity, a setting shared by many ore deposits within the trans-Challis fault system.

Pseudosection data are plotted beneath surface topographic profiles, with no vertical exaggeration, to depths where the VLF signal is attenuated to approximately one third of its value at the surface (skin depth). As with all resistivity methods, this method allows a greater depth of investigation through high-resistivity material than through low-resistivity material. Small dots on the pseudosections represent points at which apparent resistivitv was calculated.

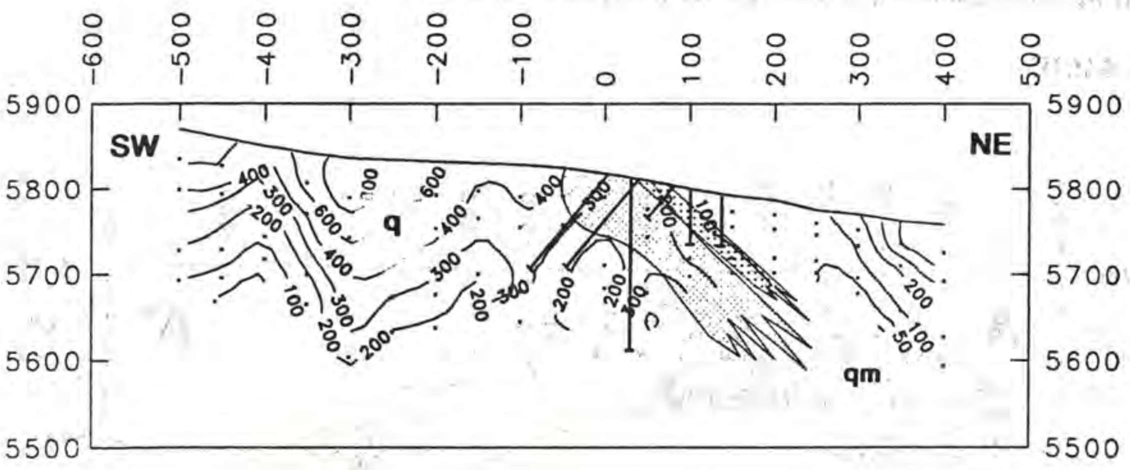

Figure 2. VLF-R pseudosection across a portion of the Robinson Dike property. Light overlay pattern shows the ore zone; heavier dots show the mafic dike; bolder lines are drill holes. qm = quartz monzonite; q = quartzite. Elevations and grid units are feet; apparent resistivity in ohm-meters.

ROBINSON DIKE

The Robinson Dike property is located approximately three miles south of Dixie, between Elk City and the Salmon River (Figure 1, A). Mineralization occurs in a zone described by Lorain (1938, p.54) as "granite or granodiorite that had been highly shattered, silicified and mineralized with fine seams and disseminated pyrite". The ore zone was reported to have an average width of 100 feet, measured horizontally, and a strike length of over 4,500 feet. A mafic dike forms the hanging wall of part of the zone and dips northeast at a steep to intermediate angle. Lorain also reported grades from exploration in the early 1900s ranging from approximately 0.1 to rare high grades of 0.3 ounces per ton gold. Recent drilling indicates disseminated grades closer to 0.06 ounces per ton for portions of the deposit. The property is currently controlled by Canyon Resources Corporation.

Figure 2 is a VLF-R pseudosection across the ore zone near some of the early workings, along with simplified geologic data taken from drill logs. The ore zone and its mafic dike hanging wall are both seen to be in the footwall of a low-resistivity slab which dips northeast at

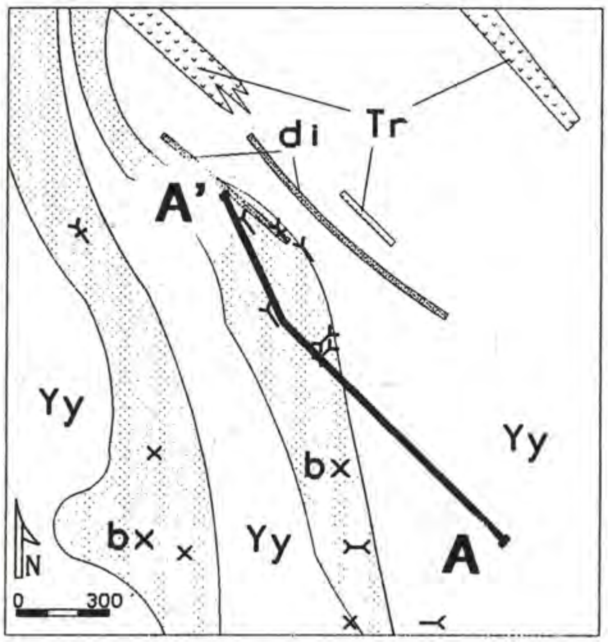

Figure 3. Generalized geologic map of the Trail's End property. bx = fault breccia; Yy = Yellowjacket formation; di = diorite; Tr = Tertiary rhyolite dikes.

an intermediate angle. The ore is localized in the area where steep and southwest-dipping contours intersect the northeast-dipping contours, a pattern which recurs at or near the ore zone intermittently along strike for at least a thousand feet. It is speculated that the clay-rich zone restricted permeability for ascending hydrothermal fluids, facilitating metal precipitation in its footwall.

TRAIL'S END

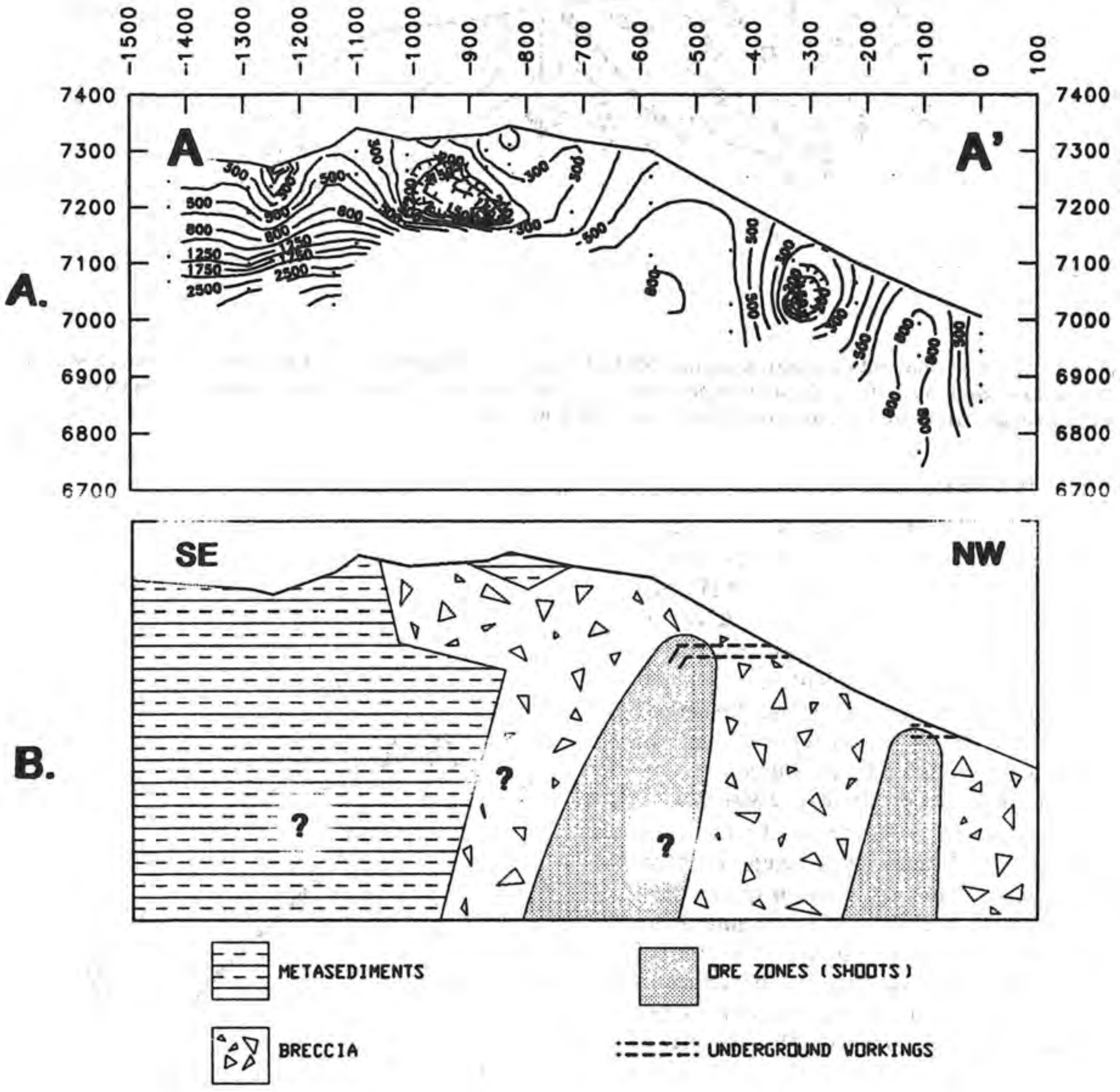

Figure 4. Sections obliquely across the brecciated eastern fault zone at the Trail's End property which hosts precious metals in quartz-carbonate shoots. Elevation and grid units are feet; apparent resistivity in ohm-meters. A. Pseudosection produced by inversion modeling. B. Interpretive geologic cross section.

The Trail's End property is south of Yellowjacket Creek, approximately 1.5 miles southeast of the Yellowjacket mine (Figure 1, B). Mineralization occurs as irregular masses of sulfide minerals in pods or shoots of quartz-carbonate gangue in two northwest-striking fault zones in Yellowjacket formation metasediments (Figure 3). Figure 4 shows sections obliquely crossing the east fault zone which pinches and swells along strike and is filled with brecciated countryrock which has locally been altered and recemented. Kidneys of ore from this property reportedly range from 1 pound to 1/2 ton and though overall production has been relatively small, shipments of hand-picked ore have averaged near 0.76 ounces per ton gold, 56 ounces silver, 15% copper and 11% lead (Anderson, 1953; Earthworks Inc., 1986).

The northern halves of the sections in Figure 4 (approximately 0 to 900S) show the ore-bearing shoots, portions of which were mined underground near 100S and 550S. These appear as resistivity highs, flanked by resistivity lows which represent the less-cemented breccia. Sulfide concentrations within the shoots are not conductive enough to lower overall resistivity in the gangue. Clay alteration is only weakly developed and is partially overprinted by silica and carbonate cement or replacement, as evidenced by "low" apparent resistivities in the 150-300 ohm-meter range. It is noteworthy that the resistivity contours within the fault zone dip steeply, in contrast to the contours within the metasedimentary countryrock which are more horizontal. Horizontal contours with increasing resistivity at depth are typical of rock units which are weathered near the surface and are otherwise unaltered. The data suggest that additional exploration for the known type of ore occurrence might focus on the downward projection of steeply dipping resistivity highs within the fault zone.

BEARTRACK

The Beartrack deposit is located along a northeast-trending segment of the Panther Creek fault, approximately 10 miles northwest of Salmon, Idaho (Figure 1, C). Figure 5 (A and B) shows apparent resistivity pseudosurfaces of a portion of the Beartrack deposit, near the discovery on "Gold Ridge". The deposit partially occurs in a quartz-adularia stockwork in quartz monzonite which is in fault contact with Precambrian Yellowjacket metasediments. Sericitic alteration and silicification characterize the ore zone where gold was precipitated in disseminated pyrite, now partially oxidized within about 200 feet of the surface. Hydrothermal circulation inferably occurred along the fault and in adjacent fractured rock. The deposit is currently controlled by Meridian Gold Company and is in the pre-production phase of development.

Raw resistivity data in Figure 5B show a complex pattern of highs and lows in the area but do not identify the ore zone well. Figure 5A shows the same area after apparent resistivity in the upper 65 feet has been filtered from the raw data by inversion modeling. This image shows part of the ore-bearing structure as a distinct resistivity high because it is more silicified, oxidized and leached than surrounding rock at comparable depth.

The pseudosection in Figure 6A shows that the mineralized zone roughly coincides with a complex resistivity high bounded to the northwest by the fault contact, shown by steeply dipping, closely spaced contours. To the southeast, the mineralized zone has a more-gradational boundary, shown by a change in the density and complexity of the contours. An irregular, arcuate band of low-resistivity material appears to partially surround the core of the ore zone to depths of 100-200 feet. This low-resistivity zone may represent a combination of higher clay content in the alteration envelope, the local water table and the limit of surface oxidation.

ACKNOWLEDGMENTS

Thanks to Canyon Resources Corporation, Trail's End Mining and Meridian Gold Company for permission to use the data.

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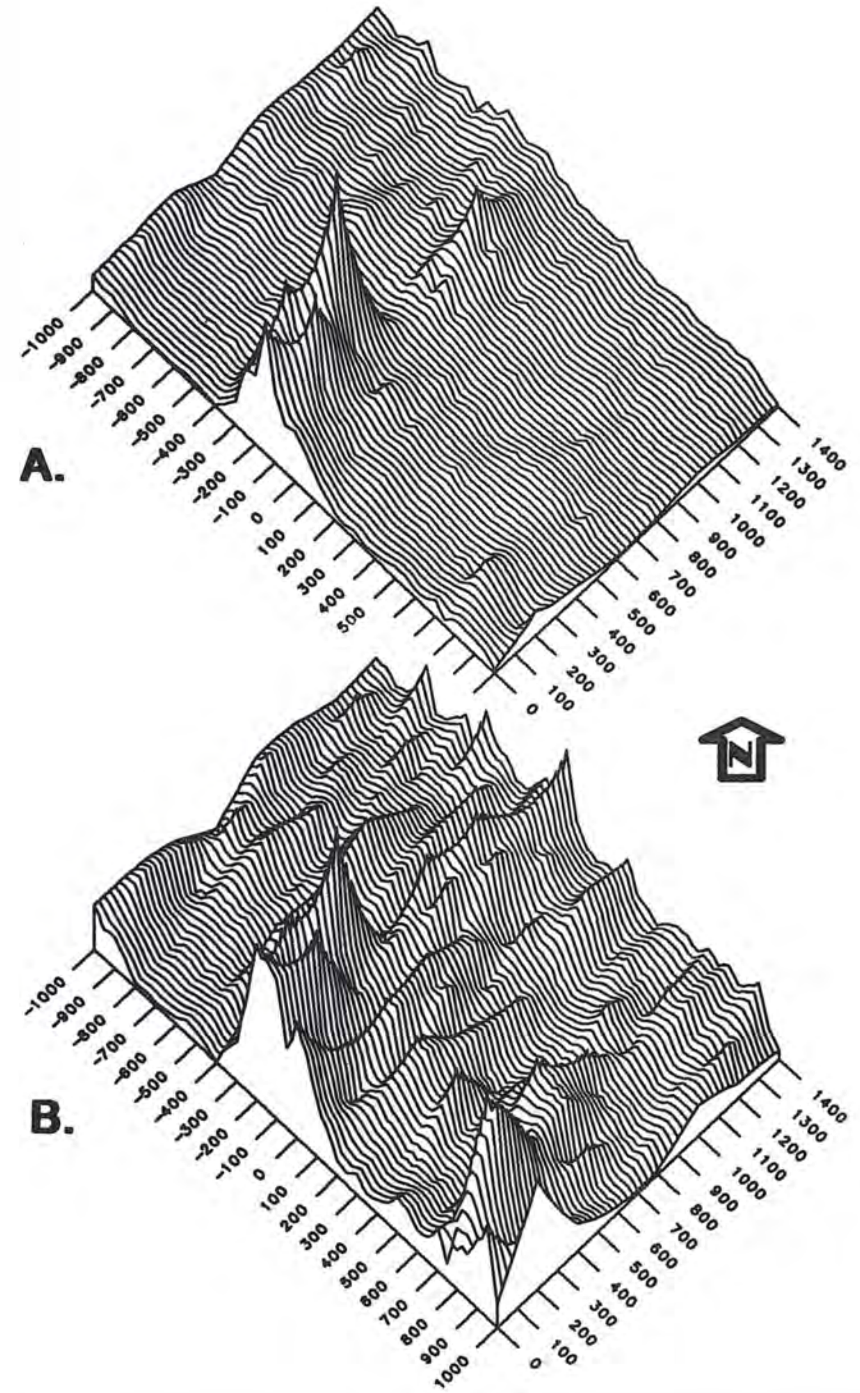

Figure 5. Apparent resistivity of a portion of the Beartrack deposit near "Gold Ridge", where gold is disseminated in a resistive stockwork. Elevated portions of the pseudosurfaces represent higher resistivities. A. Apparent resistivity from 0-65 foot depth, filtered from total apparent resistivity by inversion modeling. B. Total apparent resistivity of the same area, unfiltered.

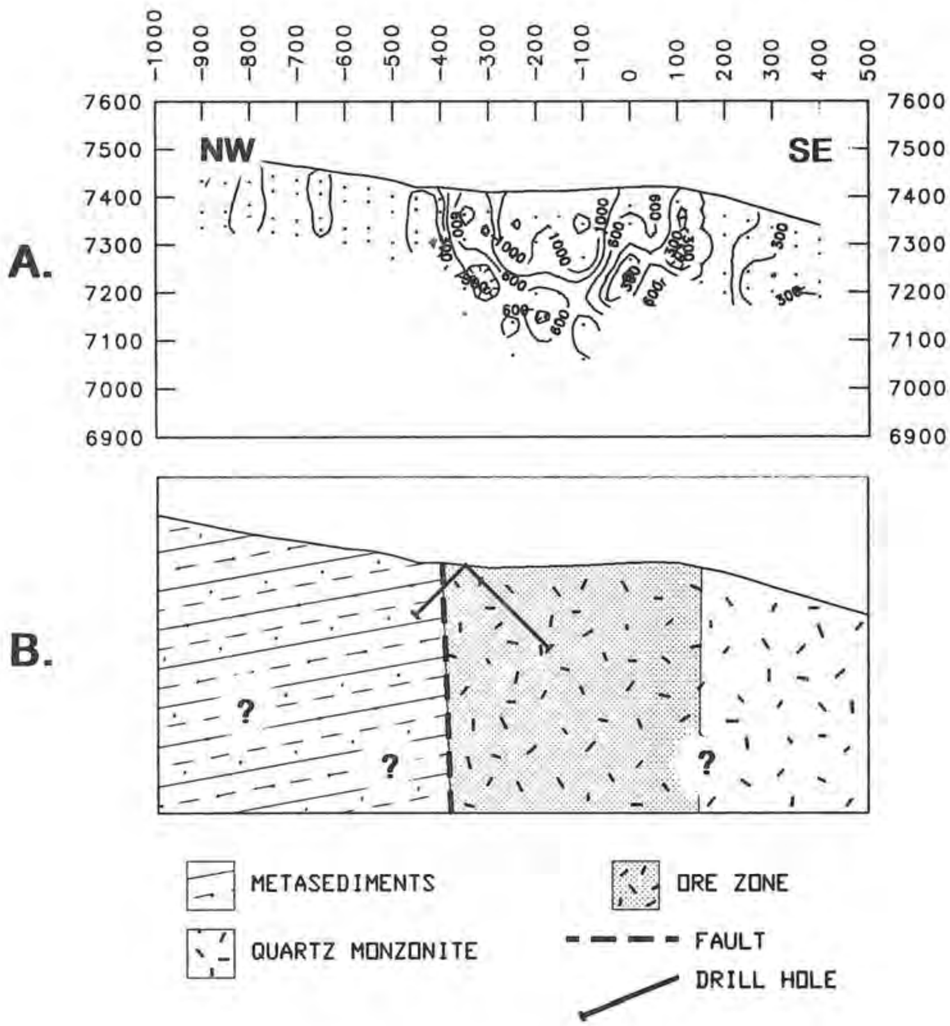

Figure 6. Sections along line 200NE of the area shown in Figure 5. Elevation and grid units are feet; apparent resistivity in ohm-meters. A. Pseudosection produced by inversion modeling. B. Interpretive geologic cross section.

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Little Belt Mountains, Montana: comparison of regional tectonics with plate tectonic model

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ABSTRACT

The structural dome which forms the Little Belt Mountains in Central Montana is one of the Laramide ranges—a series of basement cored uplifts which formed isolated ranges on the east side of the Rocky Mountains in the Early Tertiary. Thrusts and folds on the SW flank of the Little Belts are part of a SE salient of the Disturbed Belt. The salient followed the Helena embayment of the Proterozoic Belt basin. The Willow thrust postdates one igneous stock and predates another in the Castle Mountains on the S flank of the dome. The 19 laccoliths and laccolithic domes, 1 bysmalith, several stocks and many dikes and sills are alkali-calcic with most radiometric age dates falling between 54 and 48 m.y. To account for the volume of quartz latite intrusives, assimilation of granitic melt from the deep crust is required. A diatreme is located at the west end of the sapphire bearing Yogo dike. Uplift and erosion of the dome during the Eocene exposed Belt sediments on the SW flank of the dome on which Oligocene (?) basalt and rhyolite flows were extruded. Miocene ash and tuff filled the Smith River Valley and nearly buried the Little Belt range. Continued uplift and erosion is exhuming previous topography.

Dickinson's (1979) plate tectonic model of a fast moving North American plate overriding the subducted Farallon plate so that it extended as far east as the Black Hills of S.D. explains the eastward sweep of sparse volcanism across Montana (70–65 m.y.) and the transition from thin-skinned tectonics, which formed the fold and thrust belt in the sedimentary cover of the Northern Rockies, to thick-skinned tectonics involving deformation and uplift of brittle basement. Deep thrusts (in the case of the Wind River range extending through half the thickness of the crust) uplifted a number of Laramide ranges as large blocks of basement. In the Black Hills differential movement of a number of smaller blocks accounts for the domal shape of the uplift. The Little Belts probably have a similar structure, as is indicated by normal faults in the eastern part of the range and shear surfaces in the exposed Archean basement. As the North American plate slowed, the Farallon plate sank and hot asthenosphere rushed in to fill the void, thus explaining the widespread Eocene volcanism which became increasingly alkalic with time, the regional uplift and deep erosion of Laramide ranges in the Eocene and formation of diatremes at the deep end of the sinking plate.

A tear in the subducted plate extending from Glacier Park to south of the Williston basin, separating a shallow Farallon plate under Montana and a steeply dipping subducted plate under Alberta, explains the absence of Laramide ranges and arc volcanics north of the border. The tear also explains the left lateral shear of the Cat Creek, Lake Basin and Nye-Bowler fault zones and apparent counterclockwise rotation of domal axes of the Little Belts and adjacent Laramide ranges in terms of differential movement within the North American plate north and south of the border which was distributed though Montana and Wyoming.

The motion of the subducted Farallon plate as modeled by Engebretsen *et al* (1981) shows the effect of the disappearance of the Kula plate at about 55 m.y. by a counterclockwise rotation of the velocity vector: due E at 65 m.y., N45°E at 50 m.y. and N20°E at 44 m.y. The change in direction with time explains why major uplift in N–S Laramide ranges (such as the Laramie range) occurred earlier than in E–W ranges (such as the Uinta range).

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