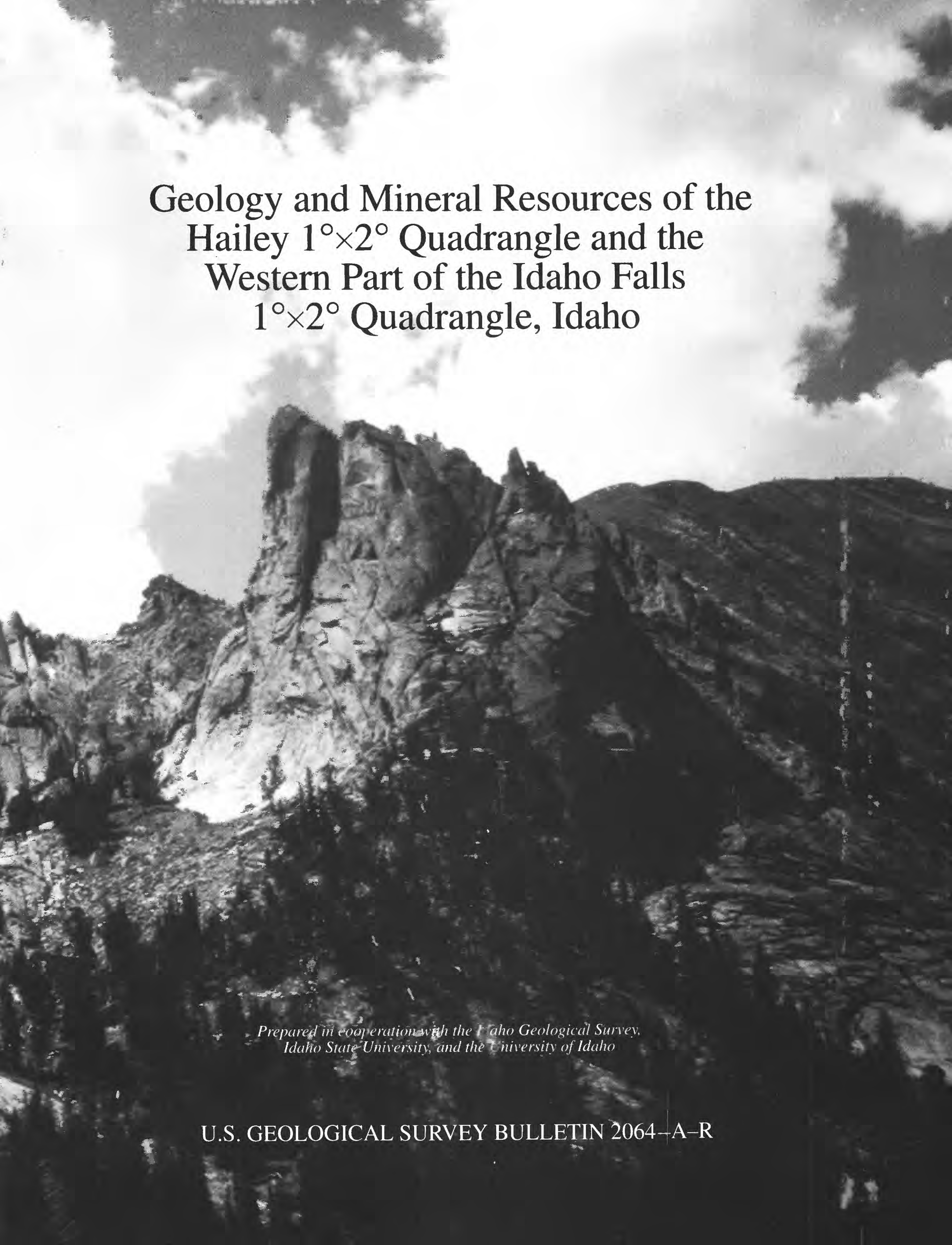


Geology and Mineral Resources of the Hailey 1°×2° Quadrangle and the Western Part of the Idaho Falls 1°×2° Quadrangle, Idaho

Prepared in cooperation with the Idaho Geological Survey, Idaho State University, and the University of Idaho

U.S. GEOLOGICAL SURVEY BULLETIN 2064–A–R

Cover. Looking south at the Mackay stock, southwest of Mackay, Idaho. The Eocene Mackay stock (center of photograph) is made up of granite, quartz monzonite, leucogranite porphyry, and felsic dike rocks. It is flanked by Paleozoic sedimentary rocks, including the Lower Mississippian McGowan Creek Formation and the Upper Mississippian White Knob Limestone. The Mackay mineralized area includes polymetallic skarn deposits. Photograph by Ronald G. Worl, 1987.

Geology and Mineral Resources of the Hailey 1°×2° Quadrangle and the Western Part of the Idaho Falls 1°×2° Quadrangle, Idaho

Edited by Ronald G. Worl, Paul K. Link, Gary R. Winkler, *and* Kathleen M. Johnson

U.S. GEOLOGICAL SURVEY BULLETIN 2064

Prepared in cooperation with the Idaho Geological Survey, Idaho State University, and the University of Idaho

Volume 1 (chapters A–R) contains regional information and a discussion of mineral deposit types hosted by sedimentary rocks. These chapters are issued together and are not available separately

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

U.S. DEPARTMENT OF THE INTERIOR

BRUCE BABBITT, Secretary

U.S. GEOLOGICAL SURVEY

Gordon P. Eaton, Director

Published in the Central Region, Denver, Colorado
Manuscript approved for publication November 4, 1993
Edited by Judith Stoeser
Graphic design by Patricia L. Wilber
Cartography by David M. Caster, Joseph A. Romero, and JRG Computer Graphics
Tables prepared by Judith Stoeser
Photocomposition by Carol Quesenberry
Cover design by Art Isom

For sale by U.S. Geological Survey, Information Services
Box 25286, Federal Center
Denver, CO 80225

Any use of trade, product, or firm names in this publication is for descriptive purposes only and does not imply endorsement by the U.S. Government

Library of Congress Cataloging in Publication Data

Geology and mineral resources of the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, Idaho / edited by Ronald G. Worl . . . [et al.].
p. cm. — (U.S. Geological Survey bulletin ; 2064)
"Prepared in cooperation with the Idaho Geological Survey, Idaho State University, and the University of Idaho."
Includes bibliographical references.
Supt. of Docs. no.: I 19.3:2064 A–R
1. Geology—Idaho. 2. Mines and mineral resources—Idaho.
I. Worl, Ronald G. II. Idaho Geological Survey. III. Idaho State University.
IV. University of Idaho. V. Series.
QE75.B9 no. 2064
[QE103]
557.3 s—dc20
[557.96] 93–44574
CIP

PREFACE

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

This CUSMAP project was a cooperative venture with the Idaho Geological Survey and the Departments of Geology at Idaho State University and the University of Idaho. Project personnel interacted closely with concurrent research investigations in the area supported by National Science Foundation Grant RUIEAR 86–18629 and an Idaho State Board of Education Economic Incentive Grant to Idaho State University. Project participants came from several branches of the U.S. Geological Survey, Idaho Geological Survey, Idaho State University, University of Idaho, Boise State University, Bureau of Land Management, U.S. Forest Service, U.S. Bureau of Mines, Rice University, and Western Washington University.

Wayne Hall and Thor Kiilsgaard proposed the Hailey 1°×2° quadrangle as a CUSMAP project and as the founding fathers were sources of guidance and inspiration for all who served on the project. Wayne spent many years prior to this CUSMAP project developing an understanding of the black shale terrane and associated mineral deposits (Hall, 1985). His work laid the foundation for studies in the black shale terrane, and his concepts on the genesis of the mineral deposits provided the guidance for many of the studies reported in volume 1 of this bulletin. Thor has actively studied the mineral deposits of central Idaho for many years starting during a period when many of the mines were still operational (Anderson and others, 1950). His first-hand knowledge and understanding of the important ore deposits in the region proved invaluable to the success of this project. In recent years, Thor has been recognized as an expert on the Trans-Challis fault system and on precious-metal deposits hosted by rocks of the Idaho batholith (Kiilsgaard and others, 1986). His work has provided new insight into this important class of mineral deposits and provided the foundation for many of the studies that will be reported in volume 2 of this bulletin.

Those who served include:

Ron Worl	Earl Bennett	Thor Kiilsgaard
Paul Link	Reed Lewis	Brian Mahoney
Sandy Soulliere	Betty Skipp	Kate Johnson
Wayne Hall	Gary Winkler	Falma Moye
Bill Hackett	Cole Smith	Anna Wilson
Jim Erdman	Dean Kleinkopf	Keegan Schmidt
Rick Sanford	Nancy Milton	Scott Southworth
Larry Snee	Brad Burton	Larry Snider
Barbara Eiswerth	Dave Stewart	Shelly Whitman
Larry Dee	Joe Wooden	Tony Gibbons

Dan Bruner	Ed Ratchford	Craig Wavra
Ryan McDermott	J.P. O'Brien	John Montgomery
Micheline Doyle	Norio Honjo	Bill Leeman
Chris Clark	Jeff Jones	Bruce Doe
Mark Stowman	Betty Bailey	Bob Criss
Flint Hall	Ann McCafferty	Spencer Wood
Bob Darling	John Blakley	Helen Whitney
Marty Power	Brenan Jordan	Victoria Mitchell
Walt Baweic	Dick Hardyman	Kevin Kunkel
John Finnegsmeir	Will Park	Darlene Batatian
Audrey Huerta		

And companions:

Blue	Minos	Nooky
Earl's Yorkies	Joe	Gretchen
Annie	Brittany	Mollie
Zonker	Minnie Moore Mutts	Basho Kitty

Investigations of the Hailey CUSMAP project expanded on previous mineral resource assessments of nearby areas in Idaho and Montana. Studies in the Hailey CUSMAP project began in 1986 and fieldwork was completed in 1989. Preliminary results were presented in May 1988 during oral and poster sessions at the 41st Annual Meeting of the Rocky Mountain Section, Geological Society of America, Sun Valley, Idaho (Link and Hackett, 1988), and on December 7, 1989, at the Northwest Mining Association 95th Annual Convention and Trade Show in Spokane, Washington (Winkler and others, 1989). A preliminary geologic map of the Hailey quadrangle was jointly prepared and released by the U.S. Geological Survey and the Idaho Geological Survey (Worl and others, 1991), and an aeromagnetic anomaly map of the Hailey quadrangle and the western part of the Idaho Falls quadrangle was released by the U.S. Geological Survey (McCafferty and Abrams, 1991).

REFERENCES CITED

Anderson, A.L., Kiilsgaard, T.H., and Fryklund, V.C., Jr., 1950, Detailed geology of certain areas in the Mineral Hill and Warm Springs mining districts, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 90, 73 p.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

Kiilsgaard, T.H., Fisher, F.S., and Bennett, E.H., 1986, The trans-Challis fault system and associated precious metal deposits, Idaho: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 81, no. 3, p. 721–724.

Link, P.K., and Hackett, W.R., eds., 1988, Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, 319 p.

McCafferty, A.E., and Abrams, G.A., 1991, Aeromagnetic anomaly map of the Hailey and western part of the Idaho Falls 1°×2° quadrangle, Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF–2160–A, scale 1:250,000.

Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., 1989, Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, 140 p.

CONTENTS

CONTENTS

METRIC CONVERSION FACTORS

Multiply	By	To obtain
Miles	1.609	Kilometers
Feet	0.3048	Meters
Inches	2.54	Centimeters
Tons	1.016	Metric tons
Short tons	0.907	Metric tons
Troy ounces	31.103	Grams
Ounces	28.35	Grams

Geology and Mineral Deposits of the Hailey 1°×2° Quadrangle and the Western Part of the Idaho Falls 1°×2° Quadrangle, South-Central Idaho—An Overview

By Ronald G. Worl *and* Kathleen M. Johnson

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–A

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

PLATE

[Plate is in pocket]

FIGURE

TABLES

Geology and Mineral Deposits of the Hailey 1°×2° Quadrangle and the Western Part of the Idaho Falls 1°×2° Quadrangle, South-Central Idaho—An Overview

By Ronald G. Worl[1] *and* Kathleen M. Johnson[1]

SUMMARY

The area of the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, Idaho, is characterized by a wide variety of geologic terranes and mineral deposit types. The eastern part of the area is a region of folded and faulted Paleozoic sedimentary rocks, covered in part by rocks of the Eocene Challis volcanic field. Much of the western part is underlain by the southern part of the Atlanta lobe of the Cretaceous Idaho batholith. Hypabyssal to epizonal bodies of Eocene intrusive rocks crop out throughout the area. Miocene and younger volcanic and sedimentary rocks cover much of the southern part of the area, and large areas of Quaternary sedimentary rocks are common.

The Paleozoic sedimentary rocks are complexly folded and faulted along major low-angle thrust faults, detachment faults, and high-angle normal faults. Wide shear zones are common in Cretaceous and Eocene plutonic rocks.

Mineral deposits of numerous types are hosted in the diverse geologic terranes of the area. Important deposit types include polymetallic quartz veins and lodes, polymetallic veins, polymetallic replacements, epithermal precious-metal veins, copper skarn, tungsten skarn, and gold placers. Significant amounts of silver, gold, lead, copper, zinc, tungsten, and antimony have been produced from these deposits. Other deposit types are prospective for a variety of elements.

Forty-six distinct mineralized areas have been defined in the Hailey quadrangle and the western part of the Idaho Falls quadrangle. These areas are based on geographic groupings of mineral deposits of similar geologic characteristics, origin, or age. They include large areas of past mineral production and areas prospective for a variety of deposit types.

At least five metallogenic events are represented in the mineral deposits of the Hailey quadrangle and the western part of the Idaho Falls quadrangle. The earliest event was syngenetic deposition of metals in oxygen-depleted basins during much of Paleozoic time, especially during the Middle to Late Devonian. Igneous and hydrothermal activity related to formation of the Cretaceous and Paleocene intrusive rocks of the Idaho batholith and its satellite plutons was the driving force for the second metallogenic event. The third event was associated with regional extensional tectonism and widespread plutonic and volcanic activity that started about 50 Ma. Miocene to Recent hydrothermal activity and the movement of groundwater constituted the fourth event. The most recent metallogenic event was the mechanical concentration of placer deposits during Quaternary time.

[1]U.S. Geological Survey, W. 904 Riverside Avenue, Spokane, Washington 99201.

INTRODUCTION

Mineral deposits are common in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle (lat 43°–44° N., long 113°15′–116° W.) and were a major factor in the settlement and development of the area. Deposits containing base-, precious-, and ferrous-metals are hosted in a variety of geologic terranes and were formed during a series of metallogenic events. This chapter summarizes the geologic terranes, mineralized areas, mineral deposit types, and metallogenic history of the quadrangles and is meant to provide background information for the more detailed reports that follow.

GEOLOGIC TERRANES

Mineral resource information in this chapter, as well as in those that follow, includes descriptions of geologic settings of mineral deposit types in terms of geologic terranes, interaction of geologic terranes, and major structural features. For this project, a geologic terrane was defined as the area in which a particular assemblage of rock types crops out.

The study area is underlain by six major terranes: Middle Proterozoic metamorphic rocks, Paleozoic sedimentary rocks, Cretaceous intrusive rocks, Tertiary intrusive rocks, Tertiary volcanic rocks, and Quaternary deposits (pl. 1). These terranes were originally defined by Worl and others (1991).

Middle Proterozoic metamorphic rocks are present in the Pioneer Mountains core complex and are mainly layered quartzo-feldspathic gneiss, quartzitic gneiss, mafic gneiss, and calc-silicate marble. Paleozoic sedimentary rocks, which range in age from Late Cambrian to Early Permian, constitute four terranes, each characterized by a particular rock assemblage: quartzite, carbonate rock, flysch, or black shale. Rocks of the quartzite terrane are of limited extent, present mainly as narrow lenses in other terranes. Rocks of the carbonate, flysch, and black shale terranes underlie extensive areas in the eastern part of the study area, in roughly parallel north-northwest-trending belts (pl. 1).

Cretaceous intrusive rocks constitute three terranes, each characterized by a particular rock type: granite, granodiorite, or tonalite. Rocks in these terranes are part of the Idaho batholith and satellitic plutons. The Idaho batholith underlies much of the western part of the area, and its eastern edge intrudes Paleozoic black shale terrane. Tertiary intrusive rocks constitute three terranes: rhyolite, granite, and diorite complex. Tertiary intrusive rocks are present throughout the area but are most prevalent in the western part. They intrude all older rock types, either as large plutons and batholiths, where they intrude Cretaceous intrusive rocks, or as smaller plutonic or hypabyssal bodies, where they intrude Paleozoic sedimentary rocks.

Tertiary volcanic rocks constitute two major terranes: Eocene and Miocene-Pliocene. The Eocene volcanic rock terrane consists of rocks of the Challis Volcanic Group in large tracts in the eastern part of the area, where rocks of the terrane overlie rocks of the Paleozoic sedimentary rock terranes. The Challis Volcanic Group is principally intermediate to mafic potassium-rich lava flow rocks and lesser amounts of volcaniclastic material. Locally the rocks are predominantly rhyolitic lava flow rocks and ash-flow tuff or volcaniclastic sedimentary rocks.

The Miocene-Pliocene volcanic rock terrane consists of the Idavada Volcanics, Banbury Basalt, and rocks of the Magic Mountain eruptive center and underlies much of the southern part of the area along the north edge of the Snake River Plain. The terrane is predominantly basaltic to rhyolitic lava flow rocks and tuff, rhyolite domes, and intercalated sedimentary rocks. The Quaternary deposits terrane includes lava flow rocks and intercalated sedimentary rocks, terrace gravels, glacial deposits, and stream gravels.

STRUCTURAL FEATURES

The structural setting of the Hailey quadrangle and the western part of the Idaho Falls quadrangle was developed during several events including the late Paleozoic Antler orogeny, the Mesozoic Cordilleran orogeny, Paleogene extensional tectonism, and Neogene basin and range extension and development of the Snake River Plain (Rodgers and others, this volume). The Antler orogeny is recorded in this area only in the stratigraphic record. Features of the Cordilleran orogeny include a major fold and thrust belt in the Paleozoic sedimentary rocks in the eastern part of the area. Several thrust plates formed from east-directed compression during the Cordilleran orogeny (Skipp, 1987). In the black shale belt these thrust plates were the loci of Cretaceous-Paleocene mineralization. Ductile deformation and metamorphism accompanied this event and are best recorded in rocks of the Pioneer Mountains core complex.

Paleogene extension was accompanied by the Eocene Challis magmatic episode (Moye, in press). The northeast-trending Trans-Challis fault system, exposed in the northwestern part of the study area, is a prominent regional feature that localized Eocene plutonic and volcanic activity. Numerous northeast-, north-, and northwest-trending brittle fracture systems developed during this episode and helped to localize Tertiary mineralization. Detachments in the Pioneer Mountains core complex and low-angle normal faults in the black shale terrane developed during Paleogene extension. Paleogene structures in some areas represent reactivation of structures formed during the Cordilleran orogeny (Rodgers and others, this volume).

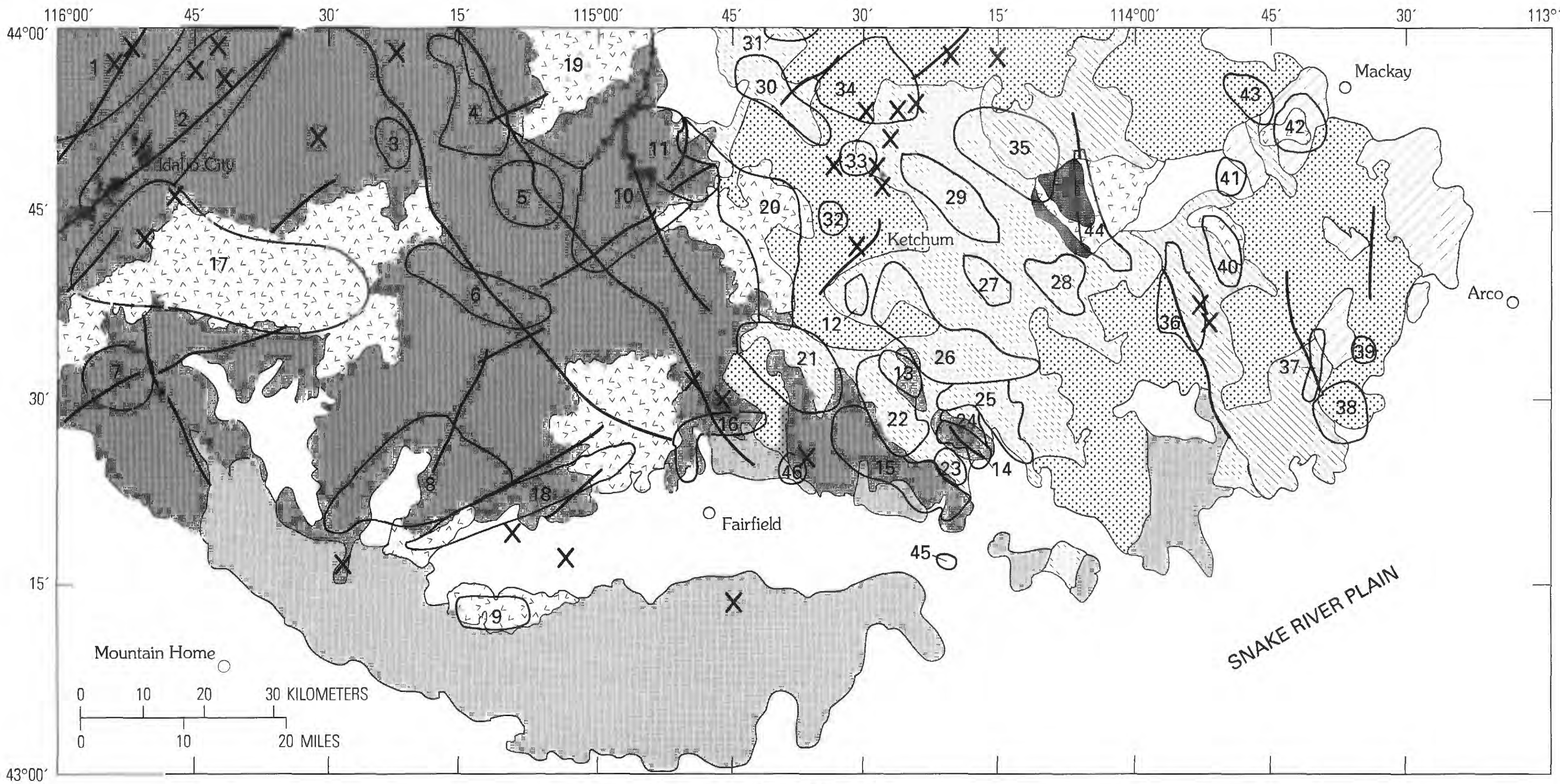
Neogene basin and range extension coincided with volcanism on the Snake River Plain. Many Neogene faults are present in the study area, and, although their trends are diverse, the most obvious are of northwest orientation. Several of the major faults host hot springs, and in some areas they host precious-metal-bearing hot-springs deposits. Faulting continues to the present along some of the range-front faults in the region, most notably the range-bounding fault on the west side of the Lost River Range.

MINERALIZED AREAS

Mineralized areas in the Hailey quadangle and the western part of the Idaho Falls quadrangle have been defined on the basis of geologic characteristics (fig. 1, table 1). These mineralized areas are geographic groupings of mineral deposits or terranes of similar character and do not represent mining districts or other political entities. The mineralized areas were defined primarily for descriptive purposes and to assist in the integration of resource, geochemical, and geophysical data for assessment of potential for undiscovered resources. Table 1 contains a summary of the geologic terranes, major structural features, and the descriptive mineral deposit types thought to be present in each mineralized area. An interpretation of the regional geochemistry of the mineralized areas in the Hailey quadrangle is provided in Smith (this volume).

MINERAL DEPOSIT TYPES

Mineral deposit types described in this section are based on characteristics of known and inferred deposits within or close to the study area. The descriptions of the



EXPLANATION

Quaternary terrane

Tertiary volcanic rock terranes
- Miocene-Pliocene terrane
- Eocene terrane

Tertiary intrusive rock terrane

Cretaceous intrusive rock terrane

Paleozoic sedimentary rock terranes
- Black shale terrane
- Flysch terrane
- Carbonate terrane

Middle Proterozoic metamorphic rock terrane

Contact

Major fault

3 Mineralized Area

MINERALIZED AREAS

1. Quartzburg
2. Idaho City
3. Swanholm
4. Black Warrior
5. Atlanta
6. Rocky Bar
7. Neal
8. Pine
9. Volcano
10. Marshall Peak
11. Vienna
12. Rooks Creek stock
13. Deer Creek stock
14. Croesus stock
15. Hailey gold belt
16. Soldier
17. Middle Fork Boise River
18. Corral Creek
19. Sawtooth
20. Smoky Mountains
21. Carrietown
22. Bullion
23. Bunker Hill
24. Minnie Moore
25. Bellevue
26. Wood River
27. Triumph
28. East Fork Wood River
29. Lake Creek
30. Galena
31. Washington Basin
32. Baker Creek
33. Boulder Basin
34. East Fork Salmon
35. Summit
36. Muldoon
37. Lead Belt
38. Lava Creek
39. Champagne Creek
40. Lake Creek stock
41. Copper Basin
42. Mackay
43. Burma Road
44. Pioneer
45. Magic
46. Elk Creek

Figure 1. Map showing generalized geologic terranes and location of mineralized areas in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho.

Table 1. Mineralized areas in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho.
[Shown by number in figure 1]

No.	Mineralized area	Geologic terrane(s)	Major structural features	Mineral deposit types having significant production	Other mineral deposit types	Comments	References
1	Quartzburg	Cretaceous granite, Tertiary diorite complex, Quaternary cover rock	Trans-Challis fault system	Tertiary epithermal precious-metal veins, Quaternary gold placers	Tertiary polymetallic veins, polymetallic quartz veins and lodes	None	Kiilsgaard, Scanlan, and Stewart (in press).
2	Idaho City	Cretaceous granodiorite, Tertiary diorite complex, Miocene-Pliocene volcanic rock, Quaternary cover rock	Trans-Challis fault system	Tertiary epithermal precious-metal veins, polymetallic quartz veins and lodes, Quaternary gold placers	None	None	Kiilsgaard, Scanlan, and Stewart (in press).
3	Swanholm	Cretaceous granite, Cretaceous granodiorite	Northwest-trending fracture systems	None	Polymetallic quartz veins and lodes (stibnite rich)	None	Popoff (1953).
4	Black Warrior	Cretaceous granite, Cretaceous granodiorite, Tertiary granite	North- to northwest-trending fracture systems	Polymetallic quartz veins and lodes (gold and base-metal varieties)	Tertiary epithermal precious-metal veins, Tertiary polymetallic veins	None	Kiilsgaard and others (1970), Kiilsgaard and others (1983).
5	Atlanta	Cretaceous granodiorite, Cretaceous granite	Intersection of regional northwest- and northeast-trending fracture systems	Polymetallic quartz veins and lodes (gold variety)	None	Polymetallic quartz veins and lodes may have important Tertiary component. These deposits are target of continuing (1991) exploration	Kiilsgaard and Bacon (in press).
6	Rocky Bar	Cretaceous granodiorite	Intersection of regional northwest- and northeast-trending fracture systems	Polymetallic quartz veins and lodes (gold variety)	Tertiary epithermal precious-metal veins, Tertiary polymetallic veins	Polymetallic quartz veins and lodes are target of continuing (1991) exploration	Bennett (in press).
7	Neal	Cretaceous granodiorite, Tertiary rhyolite	Northeast-trending fracture systems	Precious-metal veins	None	Veins are either polymetallic quartz veins and lodes (gold variety) or Tertiary epithermal precious-metal veins. Continuing exploration (1991) for precious metals	Bennett (in press).
8	Pine	Cretaceous granite, Tertiary diorite complex, Tertiary granite	Northeast- and northwest-trending fracture systems	Base- and precious-metal veins	None	Veins are either polymetallic quartz veins and lodes or Tertiary polymetallic veins	Bennett (in press).
9	Volcano	Tertiary rhyolite, Tertiary granite, Cretaceous granodiorite, Cretaceous granite	East- and northeast-trending fracture systems	Polymetallic quartz veins and lodes (gold variety)	None	Polymetallic quartz veins and lodes may have Tertiary component	Bennett (in press).

10	Marshall Peak	Cretaceous granodiorite	Regional northeast- and northwest-trending shear zones	None	Polymetallic quartz veins and lodes (gold and base-metal varieties), Tertiary epithermal precious-metal veins, Tertiary polymetallic veins	Relative importance of production from Cretaceous and Tertiary veins is unknown	Federspiel and others (1987).
11	Vienna	Cretaceous granodiorite	Northwest-trending fractures and later regional northeast-trending shear zones	Polymetallic quartz veins and lodes (silver variety)	None	Veins deposits are hosted mainly in northwest- trending fractures	Mahoney and Horn (in press).
12	Rooks Creek stock	Cretaceous granodiorite, Cretaceous tonalite, Paleozoic black shale, Tertiary rhyolite	None	Polymetallic quartz veins and lodes (gold variety)	Cretaceous polymetallic replacements	None	Park and Link (in press).
13	Deer Creek stock	Cretaceous tonalite, Paleozoic black shale	None	None	Polymetallic quartz veins and lodes (gold and base-metal varieties), Cretaceous polymetallic replacements	None	Park and Link (in press).
14	Croesus stock	Cretaceous tonalite, Paleozoic black shale	West-trending low-angle and northwest-trending high- angle fracture systems	Polymetallic quartz veins and lodes (gold and base-metal varieties)	Polymetallic replacements	Important replacement deposits close to this are in Minnie Moore mineralized area (area 24)	Worl and Lewis (in press).
15	Hailey gold belt	Cretaceous tonalite, Miocene-Pliocene volcanic rock	West- and northwest-trending low-angle fracture systems	Polymetallic quartz veins and lodes (gold and base-metal varieties)	None	Small hot springs deposits containing traces of gold and uranium may be present	Worl and Lewis (in press).
16	Soldier	Cretaceous granodiorite, Cretaceous tonalite, Tertiary diorite complex, Eocene volcanic rock, Miocene-Pliocene volcanic rock	Northwest-trending fracture systems	Polymetallic quartz veins and lodes (gold and base-metal varieties)	None	None	Worl and Lewis (in press), Lewis (in press).
17	Middle Fork Boise River	Tertiary granite, Tertiary diorite complex, Cretaceous granodiorite	None	None	None	Prospects and geochemical anomalies suggest Tertiary quartz stockworks containing molybdenum, tin, or tungsten	Kiilsgaard and Smith (in press).
18	Corral Creek	Tertiary rhyolite, Tertiary granite, Tertiary diorite complex, Cretaceous granite, Cretaceous granodiorite	Major northeast-trending fracture systems	None	Tungsten stockworks and veins	None	Lewis (in press).
19	Sawtooth	Tertiary granite, Cretaceous granite, Cretaceous granodiorite	None	None	Quartz stockworks containing Be, Sn, Mo, Ag, or Bi	None	Kiilsgaard and others (1970).

Table 1. Mineralized areas in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho—Continued.

No.	Mineralized area	Geologic terrane(s)	Major structural features	Mineral deposit types having significant production	Other mineral deposit types	Comments	References
20	Smoky Mountains	Tertiary granite, Tertiary diorite complex, Paleozoic black shale, Eocene volcanic rock	North-trending faults on east edge of Idaho batholith	None	Tertiary polymetallic replacements, Tertiary polymetallic veins	None	Link and others (this volume).
21	Carrietown	Paleozoic black shale, Cretaceous granodiorite, Cretaceous tonalite, Eocene volcanic rock	Northwest- and northeast-trending fracture systems on east edge of Idaho batholith	Cretaceous polymetallic veins, Cretaceous polymetallic replacements	None	None	Darling and others (this volume).
22	Bullion	Paleozoic black shale	Major northwest-trending fracture systems on east edge of Idaho batholith	Cretaceous polymetallic veins	None	None	Link and others (this volume).
23	Bunker Hill	Paleozoic black shale, Cretaceous tonalite	Major northwest-trending fracture systems on east edge of Idaho batholith	None	Cretaceous polymetallic veins, Cretaceous polymetallic replacements	None	Link and others (this volume).
24	Minnie Moore	Paleozoic black shale (primarily argillaceous rock), Cretaceous tonalite	Low- and high-angle, northwest-trending fracture systems in Mesozoic fold and thrust belt	Cretaceous polymetallic veins, Cretaceous polymetallic replacements	None	None	Anderson and others (1950), Link and Worl (in press).
25	Bellevue	Paleozoic black shale (primarily argillaceous rock)	Major northwest-trending fracture systems in Mesozoic fold and thrust belt	None	Cretaceous polymetallic veins	None	Link and others (this volume).
26	Wood River	Paleozoic black shale (primarily argillaceous rock), Eocene volcanic rock	Mesozoic fold and thrust belt	Stratabound barite	Cretaceous polymetallic veins	None	Link and others (this volume).
27	Triumph	Paleozoic black shale (primarily argillaceous rock)	Low- and high-angle, northwest-trending fracture systems in Mesozoic fold and thrust belt	Cretaceous polymetallic veins, Cretaceous polymetallic replacements, possible Paleozoic stratabound zinc-lead deposits	Tertiary(?) gold skarn	Some Cretaceous or Paleozoic deposits may have Tertiary overprint	Anderson and others (1950), Turner and Otto (this volume).
28	East Fork Wood River	Paleozoic black shale	Northwest-trending low-angle fracture systems in Mesozoic fold and thrust belt	None	Cretaceous polymetallic veins	None	Link and others (this volume).
29	Lake Creek	Paleozoic black shale	Major low- and high-angle northwest-trending fracture systems in Mesozoic fold and thrust belt	Tertiary polymetallic veins and replacements	None	Some Tertiary deposits may represent material remobilized from Cretaceous vein deposits	Burton and Link (this volume).

30	Galena	Paleozoic black shale (primarily argillaceous rock)	Major northeast-trending fracture systems in Mesozoic fold and thrust belt	None	Cretaceous polymetallic veins in argillaceous rocks, Tertiary(?) polymetallic veins in arenaceous rocks	None	Link and others (this volume).
31	Washington Basin	Paleozoic black shale, Cretaceous granodiorite, Cretaceous tonalite	Northeast- and northwest-trending low- and high-angle fracture systems in Mesozoic fold and thrust belt	None	Cretaceous polymetallic veins and replacements, Cretaceous molybdenum stockworks	A Cretaceous molybdenum stockwork just north of the Hailey quadrangle has known resources	Mahoney (this volume).
32	Baker Creek	Eocene volcanic rock, Eocene intrusive rock	Fracture systems that may be part of caldera complex	None	None	Geochemistry and field geology suggest undiscovered Tertiary epithermal precious-metal deposits	Erdman and others (this volume).
33	Boulder Basin	Paleozoic black shale, Tertiary granite, Tertiary diorite complex, Tertiary rhyolite, Eocene volcanic rock	Northwest- and northeast-trending low- and high-angle fracture systems	Tertiary polymetallic veins and replacements	None	None	Ratchford (in press).
34	East Fork Salmon	Tertiary granite, Tertiary diorite complex, Tertiary rhyolite, Paleozoic black shale, Eocene volcanic rock	Northeast- and northwest-trending fracture systems	None	Tertiary(?) polymetallic veins and replacements	None	Stewart and others (in press).
35	Summit	Paleozoic black shale, Paleozoic flysch, Tertiary granite	Major low-angle fracture systems and northeast- and northwest-trending high- angle fracture systems, all in Mesozoic fold and thrust belt	Tertiary polymetallic veins and replacements	Tertiary molybdenum stockworks, tungsten stockworks and veins	None	Link and others (this volume).
36	Muldoon	Paleozoic flysch, Tertiary granite, Eocene volcanic rock	North- and northwest-trending fracture systems in Mesozoic fold and thrust belt	Tertiary polymetallic veins and replacements, Tertiary epithermal precious-metal veins	Stratabound barite	The more significant Tertiary polymetallic veins and replacements are hosted in Drummond Mine Limestone Member of Copper Basin Formation. Veins and replacements are target of continuing (1991) exploration for precious metals	Winkler and others (this volume).
37	Lead Belt	Paleozoic flysch, Paleozoic carbonate, Eocene volcanic rock	Major northwest-trending low-angle fracture systems and north-trending high- angle fracture systems, all in Mesozoic fold and thrust belt	None	Tertiary polymetallic veins, Tertiary epithermal precious-metal veins	There has been some base- and precious-metal production from both known deposit types	Soulliere and others (this volume).

Table 1. Mineralized areas in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho—Continued.

No.	Mineralized area	Geologic terrane(s)	Major structural features	Mineral deposit types having significant production	Other mineral deposit types	Comments	References
38	Lava Creek	Paleozoic flysch, Paleozoic carbonate, Tertiary granite Eocene volcanic rock	North- to northwest-trending fracture systems in Mesozoic fold and thrust belt	Tertiary polymetallic veins and replacements	Tertiary tungsten stockworks and veins, Tertiary epithermal precious-metal veins	None	Erdman and others (this volume).
39	Champagne Creek	Paleozoic flysch, Eocene volcanic rock	North-trending fracture systems	Tertiary polymetallic veins, Tertiary epithermal precious-metal veins, hot-springs precious-metal deposits	None	Hot-springs deposits are currently (1991) being mined for gold	Erdman and others (this volume), Moye and others (in press).
40	Lake Creek stock	Tertiary diorite complex, Paleozoic flysch, Eocene volcanic rock	Northeast-trending fracture systems	None	Tertiary polymetallic veins and replacements	None	Winkler and others (this volume).
41	Copper Basin	Paleozoic flysch, Tertiary rhyolite, Tertiary granite	Northeast-trending fracture systems	None	Copper skarn, Tertiary polymetallic replacements, Tertiary polymetallic veins	Area is target of recent exploration for precious-metal deposits	Wilson and others (this volume).
42	Mackay	Tertiary granite, Paleozoic flysch, Paleozoic carbonate, Tertiary rhyolite	Northeast-trending fracture systems	Copper skarn	Tertiary polymetallic replacements, Tertiary polymetallic veins, tungsten skarn	Some lead and zinc production from polymetallic veins	Wilson and others (this volume).
43	Burma Road	Eocene volcanic rock	Northeast- and northwest-trending fracture systems	None	Tertiary antimony veins, Tertiary epithermal precious-metal veins	Area is target for continuing (1990) exploration for precious-metal deposits	Moye (in press).
44	Pioneer	Proterozoic metamorphic rock, Tertiary granite, Paleozoic quartzite	Pioneer Mountains core complex	Tungsten skarn	Polymetallic replacements	None	Simons and others (1981).
45	Magic	Cretaceous granodiorite, Eocene volcanic rock, Miocene-Pliocene volcanic rock	None	None	Precious-metal-bearing hot-springs deposits	Active hot springs are in general area	Leeman (1982), Norio Honjo and W.P. Leeman (unpub. data).
46	Elk Creek	Cretaceous granodiorite, Tertiary diorite complex, Eocene volcanic rock, Miocene-Pliocene volcanic rock	Northwest- and northeast-trending fracture systems	None	Precious-metal-bearing hot-springs deposits	Active hot springs are in general area	None.

mineral deposit types are presented in the format of the mineral deposit models defined by Cox and Singer (1986), and the applicable mineral deposit model of Cox and Singer is indicated (referred to as USGS model analog). Examples of deposits listed in the Hailey descriptive models are from central Idaho; those examples within the Hailey quadrangle and the western part of the Idaho Falls quadrangles are referred to the relevant mineralized area. Table 2 lists all mineral deposit types thought to be present in the study area. Detailed descriptive models follow for deposit types for which adequate information is available.

Table 2. Mineral deposit types thought to be present in the Hailey 1°x2° quadrangle and the western part of the Idaho Falls 1°x2° quadrangle, south-central Idaho.

[Deposit types enclosed in parentheses are those for which not enough is known to develop descriptive models]

Age of mineralization	Mineral deposit type
Quaternary	(Gold placers)
	(Radioactive black-sand placers)
	(Hot springs deposits)
Tertiary	(Tungsten skarn)
	Tungsten stockworks and veins
	(Stockwork deposits containing molybdenum, tungsten, tin, or beryllium)
	Copper skarn
	Polymetallic replacements
	(Gold-bearing skarn)
	Molybdenum stockworks
	Epithermal precious-metal veins
	Polymetallic veins
	(Carbonate-hosted, jasperoid-associated, precious-metal deposits)
	(Antimony veins)
	Polymetallic veins
Cretaceous	(Tungsten skarn)
	Tungsten stockworks and veins
	Molybdenum stockworks
	Polymetallic replacements
	Polymetallic veins
	Polymetallic quartz veins and lodes
Paleozoic	(Sedimentary stratabound lead-zinc deposits)
	(Stratabound barite)

TUNGSTEN STOCKWORKS AND VEINS

Approximate Synonym

Quartz-wolframite veins

USGS Model Analog

Tungsten veins (15a) (Cox and Singer, 1986, p. 64)

Summary Description

Fissure-filled quartz veins, commonly in a stockwork pattern, are along major fractures or in and just above felsic porphyritic intrusions.

Commodities

Major.—Tungsten

Byproducts.—Antimony, gold, silver, and molybdenum

Geologic Environment

Rock types.—Deposits are in altered granitic plutons and calcareous and quartzitic sedimentary and metasedimentary rocks.

Depositional environment.—Tungsten was apparently deposited within or just above small epizonal granitic plutons that formed during the late stages of both the Cretaceous and Tertiary episodes of plutonism.

Tectonic setting.—Deposition was as fissure fillings in regional fracture systems.

Associated deposit types.—Related deposits include molybdenum stockworks and tungsten skarns. Cretaceous tungsten stockworks and veins may be part of a family of vein deposits that includes Cretaceous polymetallic veins and polymetallic quartz veins and lodes.

Deposits

Description.—Quartz veins and stockworks contain local concentrations of wolframite or scheelite. The economically important deposits are above molybdenum stockwork deposits.

Mineralogy.—Ore minerals include wolframite (ferberite and huebnerite) and some scheelite, molybdenite, stibnite, gold, and silver. The bulk of the vein material is quartz; variable amounts of fluorite, arsenopyrite, calcite, rhodochrosite, sericite, feldspar, pyrite, sphalerite, chalcopyrite, tetrahedrite, galena, and cinnabar are also present. Manganese oxide minerals are ubiquitous.

Texture/structure.—Deposits are in veins and veinlets of fissure-filling massive quartz and as lenses and pods of sulfide minerals.

Alteration.—Silicification, as shown by cavities lined with chalcedonic quartz and drusy quartz, is common near the deposits; argillic and sericitic alteration of feldspar is common in igneous host rocks.

Ore controls.—Deposits are in swarms of quartz veins in fracture zones close to granite pluton contacts; important deposits may be at intersections of major structures.

Age range.—Both Cretaceous and Eocene deposits are known.

Orebodies.—Individual orebodies are irregular in shape but tend to be elongate along fracture systems or form envelopes parallel with plutonic contacts. Orebodies range in size from less than 2 ft in width and 100 ft in length to as much as 35 ft in width and 1,700 ft in length. Grades average 0.5–5 percent WO_3, 0.05–6.0 percent antimony, 0.01–0.10 ounces of gold per ton, and 0.01–1.75 ounces of silver per ton.

Geochemical Signature

A stream-sediment sample collected below the Ima mine (Worl and others, 1989) contained anomalous concentrations of copper, lead, antimony, and tungsten. A geochemical signature commonly associated with tungsten veins is W, Mo, Sn, Bi, As, Cu, Pb, Be, and F.

Geophysical Signature

The deposits themselves have no geophysical expression, but the major fracture zones and related intrusions correspond to anomalies in regional magnetic and gravity data.

Examples

Most deposits in the Corral Creek mineralized area (fig. 1)
Big and Little Falls prospects (Summit mineralized area, fig. 1)
Blizzard Mountain prospect (Lava Creek mineralized area, fig. 1)
Ima mine (Mitchell and others, 1981)
Tungsten Jim (Mitchell and others, 1986)

Comment

These deposits probably could be separated into Cretaceous and Tertiary types.

References

Callaghan and Lemon (1941)
Cook (1956)

TERTIARY MOLYBDENUM STOCKWORKS

Approximate Synonym

Calc-alkaline stockwork

USGS Model Analog

None

Summary Description

Vein stockworks and disseminations of molybdenite are in or near hypabyssal rocks of Eocene age.

Commodities

Major.—Molybdenum
Byproducts.—Copper and tungsten
Trace.—Antimony

Geologic Environment

Rock types.—Host rocks include Paleozoic sedimentary rocks of the black shale terrane, granodiorite of the Idaho batholith, and Eocene hypabyssal intrusive rocks.
Depositional environment.—Deposits are within and above late phases of multiple-phase intrusions.
Tectonic setting.—Deposits are localized along deep-seated fracture systems, including the Trans-Challis fault system.
Associated deposit types.—Related deposits include fluorspar deposits and tungsten veins and skarns.

Deposits

Description.—Two general deposit types are recognized: deposits within and next to rhyolite dikes in shear zones and deposits within small epizonal plutons. Stockworks of molybdenite-bearing quartz veinlets and disseminations of molybdenite are characteristic of both types. Most of the molybdenite is in intrusive rock, although some extends along fractures into country rock.
Mineralogy.—Ore minerals include molybdenite and chalcopyrite and locally scheelite and stibnite. Gangue is quartz and pyrite and locally pyrrhotite and arsenopyrite.
Texture/structure.—Deposits consist of flakes and rosettes of molybdenite in veins and veinlets of quartz and as disseminations in altered rock.
Alteration.—Intense alteration consisting of a siliceous core surrounded by envelopes of argillized, sericitized, and chloritized rocks is common. Pyrite zone may overprint other alteration types.
Ore controls.—Deposits are associated with rhyolite dikes within major fractures of the Trans-Challis fault system or with later phases of multiple-phase plutons.
Age range.—Deposits are Eocene or younger.
Orebodies.—The orebodies range from small isolated pods of low-grade ore within intrusive bodies to orebodies containing 0.1 percent MoS_2 and as much as 1 billion tons of ore. Orebodies associated with rhyolite dikes are elongate in the direction of the host shear zones and dikes.
Weathering.—Oxidized pyrite gives outcrops a reddish-brown color.

Geochemical Signature

Stream sediments and soils contain anomalous amounts of molybdenum and copper, and some stream sediments also contain anomalous amounts of silver.

Geophysical Signature

The deposits themselves may be detectable using electrical methods, and buried intrusions are detectable using aeromagnetic data.

Examples

Rhyolite- and shear-zone-hosted deposits
Cumo prospect (Mitchell and others, 1986)
Little Falls prospect (Mitchell and others, 1986)
Red Mountain (Mitchell and others, 1986)
Pluton-hosted deposits
Walton-White Mountain prospect (Summit mineralized area, fig. 1)
Ima mine (Mitchell and others, 1981)

References

Rostad (1967)
Tucheck and Ridenour (1981)

TERTIARY POLYMETALLIC VEINS

Approximate Synonyms
Lead-silver-zinc-antimony-tin veins, base-metal veins

USGS Model Analog
None

Summary Description
Base- and precious-metal veins are hosted by several terranes in the vicinity of Eocene plutonic or volcanic rocks.

Commodities
Major.—Silver, lead, and zinc
Byproduct.—Gold
Trace.—Antimony, copper, tin, tungsten, and bismuth

Geologic Environment
Rock types.—Deposits are in many types of host rock including sheared Cretaceous intrusive rocks; quartzite, argillite, and carbonate beds of the black shale terrane; argillite and carbonate beds of the flysch terrane; and carbonate beds of the carbonate terrane.
Depositional environment.—Deposits formed in fracture systems that also localized Eocene intrusions.
Tectonic setting.—Many deposits are in or near major shear zones in Cretaceous intrusive rocks or regional faults in the Paleozoic sedimentary rock terranes.
Associated deposit types.—Related deposits include Tertiary polymetallic replacement deposits and epithermal precious-metal veins.

Deposits
Description.—Deposits in base- and precious-metal veins are characterized by open-space filling textures, cryptocrystalline quartz, trace to significant amounts of gold, silver, lead, and zinc, and trace amounts of antimony, copper, tungsten, bismuth, and tin.
Mineralogy.—Ore minerals include galena, sphalerite, and silver-bearing tetrahedrite in a gangue of quartz, siderite, calcite, pyrite, arsenopyrite, pyrrhotite, and cryptocrystalline quartz. Locally the ore contains minor amounts of bornite, bournonite, chalcopyrite, enargite, silver telluride (hessite?), cassiterite, zincian stannite, and silver sulfantimonide minerals.
Texture/structure.—Deposits are characterized by sheared and broken zones in which quartz bodies are broken and pulled apart. Fault gouge and highly sheared rock are common.
Alteration.—Carbonate country rocks are metamorphosed to hornfels and skarn, and most host rocks show some degree of silicification.
Ore controls.—Fissures and shear zones provided permeability for mineralizing solutions driven by Eocene volcanic and plutonic activity.
Age range.—Most deposits are Tertiary or have a Tertiary overprint.
Orebodies.—Deposits generally are small lenses, pods, or irregular bodies along tabular linear veins. Size of individual orebodies ranges from a few feet in length and depth to as much as 200 ft in length, 300 ft in depth, and 7 ft in width. Grades range from trace to 6 ounces of silver per ton, 2 to 8 percent lead, 2 to 5 percent zinc, and trace to 1 ounce of gold per ton. Known deposits of this type in shear zones in Cretaceous intrusive rocks are small but may represent higher grade zones in a large area of low-grade mineralized rock.
Weathering.—Zones that contain pyrite or pyrrhotite form conspicuous gossans, but zones that contain only silver, lead, and zinc minerals have inconspicuous outcrop. Secondary lead, zinc, and antimony minerals are common in the weathered zones.

Geochemical Signature
Rocks and stream sediments in the vicinity of deposits in the black shale belt are anomalous in Sn, As, Sb, Cd, Zn, Pb, Au, and Ag; the best geochemical indicators are probably Zn, Pb, Sb, As, Ag, and Cu.

Geophysical Signature
The deposits themselves have no direct geophysical expression, but some intrusions and major structures can be recognized by prominent anomalies and steep gradient zones in the magnetic and gravity data.

Examples
Most deposits in the Galena, Boulder Basin, Lake Creek, and Lead Belt mineralized areas and some deposits in the Marshall Peak, Mackay, Copper Basin, Lava Creek, and Muldoon mineralized areas (fig. 1)

References
Anderson (1929, 1947a)
Anderson and Wagner (1946a)
Federspiel and others (1987)
Tucheck and Ridenour (1981)

EPITHERMAL PRECIOUS-METAL VEINS

Approximate Synonym
Epithermal gold (quartz-adularia)

USGS Model Analog
Comstock epithermal veins (25c) (Cox and Singer, 1986, p. 150)

Summary Description
Epithermal gold and silver quartz veins are within or close to Eocene volcanic or hypabyssal rocks, commonly within major deep-seated regional fracture systems.

Commodities
Major.—Gold and silver
Byproducts.—Lead, zinc, and copper
Trace.—Antimony and bismuth

Geologic Environment
Rock types.—Host rocks include units of the Idaho batholith and siliceous Eocene extrusive and intrusive rocks.
Depositional environment.—Hydrothermal systems are associated with Eocene calc-alkaline peraluminous volcanism and associated intrusive activity.
Tectonic setting.—Major deposits are aligned along extensional regional fracture systems, including the Trans-Challis fault system.
Associated deposit types.—Related deposits include fluorspar veins, Tertiary antimony veins, and Tertiary polymetallic veins and replacements.

Deposits
Description.—Deposits are in banded and crustified veins and stockworks of veinlets that formed mainly by open-space filling in altered and silicified rocks.
Mineralogy.—Ore minerals include auriferous pyrite, native gold, native silver, electrum, tetrahedrite, pyrargyrite, proustite, argentite, stephanite, galena, chalcopyrite, enargite, sphalerite, and aikinite and other bismuth sulfide minerals. Gangue includes quartz (cryptocrystalline to coarsely crystalline), pyrite, calcite, adularia, siderite, barite, pyrrhotite, and arsenopyrite.
Texture/structure.—Veins are characterized by open-space-filling textures in fissures and breccias; sheared rock and fault gouge are common in and along ore shoots. Ore minerals are in clusters, thin layers, lenses, and disseminations.
Alteration.—Sericitized and silicified zones are closer to the veins, and wide propylitized zones are farther away.
Ore controls.—Most of the deposits are along high-angle fractures in the vicinity of hypabyssal dikes, stocks, and plugs.
Age range.—Deposits are Eocene or younger.
Orebodies.—Deposits are generally small lenticular pods as much as 100 ft in strike length, 100 ft in dip length, and 3 ft or more in thickness. Ore values are very irregular even within a single orebody and range from trace to 23 ounces of gold per ton and trace to 2,500 ounces of silver per ton.
Weathering.—Weathering includes bleached country rock and iron-stained outcrops.

Geochemical Signature
Bismuth is associated with gold in the Quartzburg, Idaho City, Lava Creek and Champagne Creek mineralized areas (fig. 1); heavy-mineral concentrates in some areas contain anomalous amounts of Au, Ag, Sn, Ba, Bi, Cu, Mo, and Pb.

Geophysical Signature
The deposits themselves have no geophysical expression, but the ore-controlling faults might be detected by detailed magnetic surveys and intensely altered zones might be detected by resistivity surveys. Regional magnetic data show the northeast-trending structures of the Trans-Challis fault system. Some mines are at intersections of strong east-west and northeast linear magnetic trends.

Examples
Gold Hill mine (Quartzburg mineralized area, fig. 1)
Volcanic rock-hosted deposits in the Muldoon, Lava Creek, and Champagne Creek mineralized areas (fig. 1)

Reference
Anderson (1929, 1947a, b)

TERTIARY POLYMETALLIC REPLACEMENTS

Approximate Synonyms
Distal skarn, irregular replacements of base- and precious-metals

USGS Model Analog
Zinc-lead skarns (18c) (Cox and Singer, 1986, p. 90)

Summary Description
Carbonate strata are partly to completely replaced by silicate and sulfide minerals.

Commodities
Major.—Silver, lead, and zinc
Byproducts.—Copper, gold, and antimony

Geologic Environment
Rock types.—Host rocks are calcareous or dolomitic units in flysch and carbonate terranes.
Depositional environment.—Deposits formed by replacement of carbonate-bearing units by solutions following high-angle and bedding-parallel fracture systems in the vicinity of hypabyssal intrusive bodies.
Tectonic setting.—Deposits are generally near high-angle faults that host hypabyssal bodies.
Associated deposit types.—Related deposits include Tertiary polymetallic veins and epithermal precious-metal veins.

Deposits
Description.—Deposits are tabular to elongate lenses or ovoid, pipelike, stratabound bodies of silicate and sulfide minerals.
Mineralogy.—Sulfide minerals include galena, sphalerite, pyrite, chalcopyrite, arsenopyrite, marcasite, and complex antimonide minerals. Oxidized zones contain cerussite, anglesite, scorodite, malachite, and iron oxide minerals. Gangue minerals include calcite, quartz, sericitized feldspar, fluorite, and several calc-silicate minerals.

Texture/structure.—Deposits are generally coarse grained mosaics of silicate, sulfide, and carbonate minerals.
Alteration.—Alteration effects are not extensive in carbonate rocks; they consist of some minor bleaching and silicification. Extensive silicification and sericitic alteration are present in siliceous igneous rocks.
Ore controls.—Deposits formed along permeable zones within receptive host rock close to hypabyssal intrusive bodies.
Age range.—Known deposits are probably Eocene.
Orebodies.—Numerous sulfide seams and stringers are in bodies that are as much as 1,000 ft long, 120 ft wide, and 20 ft thick. Grades range from 6 to 12 ounces of silver per ton, 5 to 10 percent lead, 3 to 7 percent zinc, 0.1 to 1.0 percent copper, and trace to 0.3 ounces of gold per ton. Some pockets of high-grade ore were mined from the oxidized zone.
Weathering.—Secondary lead and zinc minerals are common in the weathered zone.

Geochemical Signature
Stream-sediment samples contain anomalous amounts of Ag, As, Ba, Cu, Pb, Sb, and Zn. Most barren rock samples from the same stratigraphic unit that hosts the orebodies contain anomalous amounts of barium, and many contain anomalous amounts of boron and zinc (Winkler and others, this volume).

Geophysical Signature
No geophysical signatures are known, but local detailed electrical surveys might detect individual orebodies.

Examples
Eagle Bird mine area (Muldoon mineralized area, fig. 1)
Phi Kappa mine (Summit mineralized area, fig. 1)

References
Anderson and Wagner (1946a)
Tucheck and Ridenour (1981)

COPPER SKARN

Approximate Synonyms
Base-metal skarn, polymetallic skarn

USGS Model Analog
Copper skarn (18b) (Cox and Singer, 1986, p. 86)

Summary Description
Chalcopyrite-rich lenses, pods, and disseminations of sulfide minerals are in calc-silicate contact metasomatic rocks.

Commodities
Major.—Copper, lead, zinc, and silver
Byproducts.—Gold, molybdenum, and iron
Trace.—Tungsten and fluorine

Geologic Environment
Rock types.—Host rocks are skarn composed of Ca-Fe-Mg-Mn-silicate minerals within carbonate-bearing sedimentary rocks at the contact with Tertiary granite and leucogranite porphyry.
Depositional environment.—Deposits are in thermal and chemical aureoles of Tertiary plutons intruding carbonate-bearing rocks.
Tectonic setting.—Plutons are most commonly along northeast- or northwest-trending regional structures. Deposits are in faulted margins of plutons.
Associated deposit types.—Outer zones of these skarns contain polymetallic veins and local iron skarns.

Deposits
Description.—Skarn deposits contain significant copper minerals. The orebodies are pipelike, chimney-, or pod-shaped and are along the contacts of intrusive rock and skarn, along the contacts of skarn and carbonate-bearing sedimentary rock, and along joints, breccias, faults, and bedding planes within skarn. Peripheral to and, in some cases, overlapping the skarn deposits are replacement veins containing silver, lead, and zinc minerals. The two deposit types are genetically related.
Mineralogy.—Historic ore production was mainly from oxidized ore containing chrysocolla, malachite, azurite, and tenorite in a gangue of calc-silicate minerals and iron hydroxide. Primary ore contains chalcopyrite, local concentrations of molybdenite, sphalerite, and bornite, and rare fluorite and scheelite associated with pyrite, pyrrhotite, calcite, quartz, magnetite, and specularite in a matrix of calc-silicate minerals.
Texture/structure.—The ore-bearing skarn is coarse to very coarse grained and, in general, coarser grained than surrounding barren skarn. Copper sulfide minerals are massive to disseminated; the other sulfide minerals are disseminated and interstitial to the calc-silicate minerals.
Alteration.—Skarn formation.
Ore controls.—Deposits are related to receptive carbonate-bearing beds, Eocene felsic plutons, and permeable areas within the metasomatic zone.
Age range.—All known occurrences are associated with Eocene plutonic rocks.
Orebodies.—Individual orebodies are 15–200 ft long, 5–55 ft wide, and as much as 600 ft deep. Grades range from 2 to 6 percent copper in oxidized ore and are 2.5 percent or less copper in sulfide ore.
Weathering.—Exposures have been described as calc-silicate rock carrying disseminations and spongelike aggregates of metallic minerals. Secondary copper and iron minerals are common.

Geochemical Signature
Stream-sediment samples in the vicinity of the Empire mine contain anomalous amounts of copper (Worl and others, 1989).

Geophysical Signature

Magnetic anomalies may indicate buried intrusive bodies in Paleozoic carbonate rocks. Local detailed electrical surveys might detect individual orebodies.

Examples

Most deposits in the Mackay and Copper Basin mineralized areas (fig. 1)

Reference

Umpleby (1917)

CRETACEOUS MOLYBDENUM STOCKWORKS

Approximate Synonym

Calc-alkaline stockwork

USGS Model Analog

Porphyry molybdenum, low-fluorine (21b) (Cox and Singer, 1986, p. 120)

Summary Description

Vein stockworks and disseminations of molybdenite are in or near plutonic rocks of Cretaceous age.

Commodities

Major.—Molybdenum
Byproducts.—Silver and tungsten

Geologic Environment

Rock types.—The deposits are within or associated with compositionally zoned stocks of Cretaceous biotite granodiorite and porphyritic biotite granite.
Depositional environment.—The intrusions are outliers of the Idaho batholith and are within a late Mesozoic-Tertiary magmatic arc in the western North American Cordillera that hosts many economically significant molybdenum stockwork deposits associated with fluorine-deficient, I-type, compositionally zoned granitoid plutons that are probably subduction related.
Tectonic setting.—Known deposits are on the eastern edge of the Idaho batholith.
Associated deposit types.—Related deposits include tungsten veins and stockworks and tungsten skarn.

Deposits

Description.—Deposits are stockworks of veins and veinlets of quartz, biotite, potassium feldspar, and white mica in which sulfide minerals other than molybdenite are rare.
Mineralogy.—Ore minerals include molybdenite and locally scheelite, chalcopyrite, galena, and silver minerals. Gangue includes biotite, potassium feldspar, and white mica.
Texture/structure.—Molybdenite is present as rosettes, flakes, fracture fillings, intercalations with secondary micas, and selvages.
Alteration.—Central potassic zone grades outward through phyllic, argillic, and probably propylitic zones.
Ore controls.—Deposits are in quartz stockworks in and above compositionally zoned plutons.
Age range.—Cretaceous.
Orebodies.—Mineralized areas range in size from a few feet in diameter to bodies as much as 11,100 ft in length, 3,000 ft in width, and more than 2,200 ft in thickness.
Weathering.—Yellow ferrimolybdite, after molybdenite, is present locally.

Geochemical Signature

Stream-sediment and heavy-mineral-fraction samples in the vicinity of the Thompson Creek mine contain anomalous amounts of tungsten, molybdenum, bismuth, and copper; stream sediments also contain anomalous amounts silver, and the heavy-mineral fraction contains anomalous amounts of lead, antimony, boron, and thorium (Worl and others, 1989). A reconnaissance biogeochemical survey in 1979 included wood samples from two Douglas fir trees from the window overlying the orebody at Thompson Creek. These samples yielded silver concentrations of 300 and 500 ppm, as compared to a norm of about 20 ppm. Anomalous molybdenum concentrations were also measured (J. Erdman, U.S. Geological Survey, unpublished data, 1992).

Geophysical Signature

Known deposits are on the flanks of positive magnetic anomalies that probably indicate buried granitic plutons.

Examples

Thompson Creek mine (Mitchell and others, 1986)
Cabin Creek deposit (Mitchell and others, 1986)
Virginia-Beth deposit (Mitchell and others, 1986)
Little Boulder Creek (mainly a skarn deposit) (Mitchell and others, 1986)

References

Cavanaugh (1979)
Hall and others (1984)

POLYMETALLIC QUARTZ VEINS AND LODES

Approximate Synonym

Mixed base- and precious-metal veins

USGS Model Analog

Mixed base- and precious-metal veins (James D. Bliss, U.S. Geological Survey, Tucson, Arizona, written commun., 1991)

Summary Description

Quartz veins and lodes locally enriched in precious- or base-metals are within or close to the Idaho batholith.

Commodities

Major.—Gold and silver; locally antimony, copper, lead or zinc

Byproducts.—Gold and silver where other metals dominant

Trace.—Molybdenum and tungsten

Geologic Environment

Rock types.—Host rocks include all lithologic phases of the Idaho batholith and siliceous and calcareous sedimentary and metamorphic rocks near the batholith.

Depositional environment.—Mineralization occurred during and directly following formation of the Idaho batholith and prior to Eocene volcanism.

Tectonic setting.—Deposits are in fractures that formed during cooling of the Idaho batholith; regionally, this was a period of change from a dominantly compressive to a dominantly tensional structural environment.

Associated deposit types.—These veins and lodes represent the quartz-gold-rich end member of a family of vein deposits that were emplaced during and directly following batholith formation. The other end member of this family is represented by the polymetallic veins that are present in the black shale terrane next to the batholith. Cretaceous tungsten stockworks and veins may also be part of this family of vein deposits and closely related to the polymetallic quartz veins and lodes.

Deposits

Description.—The deposits are multistage quartz veins and lodes within shear zones and fault fissures. They are spatially and genetically related to the Idaho batholith and are in rocks of the batholith and in nearby siliceous and calcareous rocks. Several varieties of deposits are recognized; they reflect in part gradations between the two end members and in part mineral or metal zoning within the polymetallic quartz veins and lodes. Subtypes include base-metal veins, gold veins, antimony veins, and silver veins.

Base-metal veins are characterized by multistage quartz and siderite or calcite gangue and locally abundant galena, tetrahedrite, and sphalerite. The carbonate minerals commonly are present as fracture fillings in brecciated quartz and the sulfide minerals as replacements of carbonate minerals. Production from these veins was mainly for silver, lead, and zinc. Examples are in the Croesus mine (Croesus stock mineralized area, fig. 1).

Gold veins are characterized by several stages of quartz including early massive columnar and late vuggy cockscomb varieties and generally minor amounts of pyrite, chalcopyrite, pyrrhotite, and (or) arsenopyrite. Sericite is present in most veins and vein selvages. Production from these veins was mainly for gold and locally for copper. Examples include deposits in the Hailey gold belt, Rocky Bar, and Atlanta mineralized areas (fig. 1).

Antimony veins are characterized by vuggy cockscomb quartz and locally abundant stibnite; other sulfide minerals may be present. Production from these veins was mainly for antimony, but some veins may have significant precious-metal content as well. Examples include deposits in the Swanholm mineralized area (fig. 1).

Silver veins are characterized by vuggy banded cockscomb quartz, ruby silver minerals, and an absence of arsenopyrite, pyrrhotite, or sphalerite; stibnite may be present. Production from these veins was mainly for silver. The silver veins may grade into gold veins at depth. Examples include deposits in the Vienna mineralized area (fig. 1).

Mineralogy.—Quartz is dominant in all deposits. Other nonsulfide gangue minerals include siderite, calcite, and sparse barite, sericite, albite, adularia(?), epidote, magnetite, and fluorite. Sulfide gangue minerals are mostly insignificant and include pyrite, arsenopyrite, and pyrrhotite. Ore minerals are chalcopyrite, galena, sphalerite, tetrahedrite, pyrargyrite, stibnite, huebnerite, and scheelite. Silver is mostly associated with galena, tetrahedrite, or pyrargyrite, and gold is associated with chalcopyrite or pyrite.

Texture/structure.—The deposits include massive single veins of quartz in fissures and zones of many veins, veinlets, breccia fillings, and disseminations of quartz in large areas of broken, sheared, and crushed country rock. Both syn- and post-mineral deformation is evident in most deposits. Much of the quartz is barren or metal-deficient massive and columnar varieties. Generally, metal-enriched veins and lodes have a late quartz stage with drusy, cockscomb, coliform, and vuggy textures. Sulfide minerals are present as bands, pods, and disseminations within the quartz.

Alteration.—Alteration is mostly confined to the zone of broken, sheared, and crushed country rock. Argillic and sericitic alteration is pervasive through the sheared zones, and pyritic alteration and silicification are present next to the veins and lodes.

Ore controls.—The deposits are in shear zones and fracture systems, and groups of deposits commonly have similar orientations, but there is no apparent regional control to their location. Deposits in parts of the batholith are genetically related to the muscovite-biotite granite phase. Deposits formed at or near the edges or tops of batholith phases.

Age range.—Deposits range from Late Cretaceous to Paleocene in age.

Orebodies.—Individual veins and lodes are from tens of feet to several miles in length, a few feet to a few tens of feet in width, and hundreds to thousands of feet in depth. Individual orebodies are lenticular pipelike ore shoots within the veins or lodes and are characterized by consistency through large vertical extents. Some of the larger orebodies have strike lengths of as much as 300 ft, dip lengths of more than 700 ft, and thicknesses of a few inches to 8 ft. Grades vary considerably; recorded averages range from trace to 1 ounce of gold per ton, trace to 15 ounces of silver per ton, 0.01 to 4.5 percent copper, 2 to 8 percent lead, and 4 to 9 percent zinc.

Weathering.—Abundant quartz chips and placer gold concentrations are in soils and stream sediments; local minor lenses of gossan, manganese-oxide stain, and malachite or azurite stain are present.

Geochemical Signature

Anomalous amounts of arsenic, gold, silver, and antimony and locally tungsten, lead, copper, and zinc are in stream-sediment and heavy-mineral-fraction samples.

Geophysical Signature

High-angle shear zones and fractures, where mineralized, can be mapped using detailed magnetic and electrical surveys.

Examples

Many deposits hosted by rocks of the Idaho batholith including, in the Hailey quadrangle, most deposits from the Atlanta, Rocky Bar, Black Warrior, Swanholm, Volcano, Vienna, Hailey gold belt, and Croesus stock mineralized areas (fig. 1)

References

Allen (1952)
Anderson (1939, 1943)
Anderson and others (1950)
Anderson and Wagner (1946b)
Kiilsgaard and others (1970)
Popoff (1953)
Ross (1927)

CRETACEOUS POLYMETALLIC VEINS

Approximate Synonyms

Wood River lead-silver veins, lead-silver-zinc-antimony veins

USGS Model Analog

Polymetallic veins (22c) (Cox and Singer, 1986, p. 125)

Summary Description

Deposits are rich siderite-silver-lead veins in the Idaho black shale belt.

Commodities

Major.—Silver
Byproducts.—Lead, zinc, and gold
Trace.—Antimony and copper

Geologic Environment

Rock types.—Carbonaceous micritic limestone and black argillite are the most common hosts, but deposits are also present in siltite, siltstone, shale, quartzite, and sandy limestone.

Depositional environment.—Deposition was along major fracture systems in the black shale terrane within the thermal aureoles of Cretaceous granodioritic to granitic intrusions.

Tectonic setting.—Many deposits are in or near major regional low-angle faults.

Associated deposit type.—These veins represent the siderite-silver-lead-rich end member of a family of vein deposits emplaced during and directly following batholith formation. The other end member of this family is represented by polymetallic quartz veins and lodes hosted by rocks of the Cretaceous intrusive terrane.

Deposits

Description.—Deposits comprise multistage veins characterized by siderite and quartz gangue. Deposits formed along fracture systems in black shale terrane through a combination of open-space filling and replacement of carbonate beds. Veins have simple mineralogy and contain few trace metals; they are mainly in argillaceous rocks of the black shale terrane, grade into polymetallic replacement deposits, and are within the metamorphic and metasomatic influence of the Cretaceous intrusive rocks.

Mineralogy.—Ore minerals are argentiferous galena, sphalerite, and silver-bearing tetrahedrite in a gangue of siderite, calcite, and quartz. Variable amounts of pyrite, arsenopyrite, and chalcopyrite are present in the ore.

Texture/structure.—Coarse-grained sulfide minerals are present as bands, pods, and disseminations within siderite and crushed country rock.

Alteration.—The country rocks are metamorphosed and metasomatized to hornfels and skarn. Alteration, mainly formation of sericite, extends only a few inches from individual fractures and is not a conspicuous feature.

Ore controls.—Fissures and shear zones close to Cretaceous igneous rock intruding black shale terrane provided permeability for metal-bearing solutions.

Age range.—Most deposits are probably Cretaceous and Paleocene, although some may have an Eocene overprint.

Orebodies.—Small lenses, pods, or tabular veins and large irregular bodies. Size ranges from a few tons to 200,000 tons for a single orebody. Average grades range from 5 to 45 ounces of silver per ton, 5 to 20 percent lead, and 5 to 13 percent zinc; trace amounts of gold are present. The very rich ores resulted from secondary enrichment.

Weathering.—Zones that contain pyrite or pyrrhotite form conspicuous gossans, but zones that contain only silver, lead, and zinc minerals have inconspicuous outcrop. Secondary lead, zinc, and antimony minerals are common in the weathered zones. Reported gold production was from gossans derived from pyrite-rich veins.

Geochemical Signature

Best geochemical indicators are probably Zn, Pb, Sb, As, Ag, Cu, and organic carbon.

Geophysical Signature

The deposits themselves have no direct geophysical expression, but some intrusions and major structures exhibit prominent anomalies and steep gradient zones in the magnetic and gravity data.

Examples

Livingston mine (Mitchell and others, 1986)
Hoodoo mine (Mitchell and others, 1986)
Triumph mine (Triumph mineralized area, fig. 1)
Minnie Moore mine (Minnie Moore mineralized area, fig. 1)
Most deposits in the Bullion mineralized area (fig. 1)

References

Anderson and others (1950)
Umpleby and others (1930)
Van Noy and others (1986)

CRETACEOUS POLYMETALLIC REPLACEMENTS

Approximate Synonym

Irregular replacements of base- and precious-metals

USGS Model Analog

Polymetallic replacement deposits (19a) (Cox and Singer, 1986, p. 99)

Summary Description

Deposits formed by selective replacement by silicate and sulfide minerals of carbonate strata in Paleozoic black shale terrane.

Commodities

Major.—Silver, lead, and zinc
Byproducts.—Copper and gold

Geologic Environment

Rock types.—Host rocks are generally sandy or silty calcareous or dolomitic rock interbedded with black carbonaceous argillite of the black shale terrane.
Depositional environment.—Solutions emanating from Cretaceous intrusive rocks and moving along fracture systems replaced receptive host rocks.
Tectonic setting.—Most of the richer deposits are at the intersections of the host rocks and steep faults. Regional fold structures and low-angle faults may have helped localize ore.
Associated deposit type.—Cretaceous polymetallic veins commonly are in the same location, but they are thought to be slightly younger.

Deposits

Description.—Deposits of tabular to irregular-shaped bodies are in part stratabound but locally crosscut strata. In some areas the deposits may represent local reconcentration of metals.
Mineralogy.—Ore minerals include galena, sphalerite, tetrahedrite, and chalcopyrite; gangue includes siderite, quartz, pyrite, and arsenopyrite. Deposits are characterized by the presence of siderite gangue and jamesonite.
Texture/structure.—Deposits are generally coarse grained; ore minerals are present as pods, veinlets, irregular masses, and disseminations.
Alteration.—Dolomitization(?), silicification, and sericitization occurred around some of the deposits.
Ore controls.—Mineralization was controlled by location of major fracture systems cutting receptive carbonate rocks within the influence of Cretaceous igneous intrusions.
Age range.—Deposits are mostly Cretaceous in age; they may be Paleocene in part.
Orebodies.—Orebodies are generally irregular, discontinuous, elongate lenses or ovoid pipelike bodies that follow bedding in the host rock. Orebodies range from a few tons to more than 200,000 tons. Grades range from a few percent lead and zinc and less than an ounce of silver per ton to 30–40 percent lead and 40–60 ounces of silver per ton. Some of the richer orebodies mined in the past were in the secondary enrichment zone.
Weathering.—Secondary lead and zinc minerals are common in the weathered zone.

Geochemical Signature

Stream-sediment samples from the vicinity of the Clayton Silver mine (Mitchell and others, 1986) contain anomalous amounts of B, Ag, Cu, Zn, Pb, and As, and heavy-mineral concentrates contain anomalous amounts of B, Ag, Cu, Zn, Pb, Ba, Mo, Cd, Sn, and W (Worl and others, 1989).

Geophysical Signature

No geophysical signature is known, but local detailed electrical surveys might detect individual orebodies.

Examples

Some deposits in the Minnie Moore, Triumph, Bullion, Rooks Creek stock, and Deer Creek stock mineralized areas (fig. 1)
Clayton Silver mine (Mitchell and others, 1986)

References

Anderson and others (1950)
Hall and others (1978)
Ross (1937)

METALLOGENIC EVENTS

Mineralizing events recognized in the rocks of the Hailey quadrangle and the western part of the Idaho Falls quadrangle include (1) sedimentation in marine basins that were oxygen depleted during much of Paleozoic time, (2) igneous and hydrothermal activity related to formation of the Cretaceous Idaho batholith, (3) igneous and hydrothermal activity during formation of Eocene granitic intrusions and the Eocene Challis volcanic field, (4) hydrothermal activity related to Miocene and younger volcanic activity, and (5) Pleistocene to Holocene weathering and erosion. These events are listed in table 3 and described below.

PALEOZOIC EVENTS

The earliest recorded event was stratabound concentration of metals in black shales during most of the Paleozoic, especially during the Middle Devonian. Metals concentrated include Ag, Ba, Cu, Mo, Va, Pb, Zn, and Ni. The metals entered the basin as emanations directly onto the sea floor and in detritus derived from weathering of pre-existing deposits in the sediment source area (Sanford and Wooden, this volume). Metal concentrations took the form of lenses, stringers, and disseminations of sulfide minerals and probably metal-bearing nonsulfide minerals. The metal-enriched rocks are mainly black carbonaceous argillite, siltite, shale, and micritic limestone, which are shown as argillaceous rocks within the black shale terrane on plate 1. Known syngenetic metal deposits include stratabound zinc deposits at the Hoodoo and Livingston mines in the Boulder Creek district just north of the Hailey quadrangle (Hall, 1985) and some of the stratabound zinc-lead orebodies in the Triumph mineralized area (Turner and Otto, this volume) (fig. 1). Numerous other deposits in the black shale terrane are thought to be stratabound in part. In addition to the stratabound deposits that are known to be present or may be present, the carbonaceous rocks (black shale) constitute an enormous possible low-grade resource of metals.

Table 3. Metallogenic events in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho.

Age	Metallogenic event(s)
Pleistocene and Holocene	Mechanical concentration of gold and heavy minerals into placer deposits.
Miocene to present	Hydrothermal cells associated with volcanism developed hot springs deposits containing gold and locally containing uranium.
50–40 Ma	Mineralization associated with Eocene extensional and igneous activity included: Veins and veinlets along major deep-seated, high-angle fault systems containing Au, Ag, Bi, Sb, Pb, Zn, Cu, fluorspar, and Ba; Disseminations and stockworks containing molybdenum and tungsten within and above granitic and rhyolitic intrusions; Replacement deposits including skarns and distal skarns containing Pb, Zn, Ag, W, and Au in the vicinity of granitic-rhyolitic intrusions; Greisens and pegmatitic pods containing molybdenum, tungsten, beryllium, or tin in and just above the tops of pink granite intrusions; and Remobilization of pre-existing ores, mainly affecting vein and massive deposits of lead, silver, and zinc and massive quartz veins containing gold and silver.
78–57 Ma	Mineralization during transition from dominantly compressional to dominantly extensional tectonism developed a family of deposits characterized by several generations of quartz and containing some combination of Au, Ag, Sb, As, Pb, Cu, W, and Zn. Deposits formed in Paleozoic black shale terrane were enriched in silver, lead, and zinc and those formed in Cretaceous intrusive rocks and siliceous metamorphic rocks were enriched in gold and locally silver, antimony, or tungsten.
74(?) Ma	Intrusion of leucogranitic phase of Idaho batholith developed disseminations and stockworks containing molybdenum, tungsten, and antimony(?) within and surrounding the plutonic bodies.
90–80 Ma	Intrusion of Idaho batholith border phases formed lead-, zinc-, and silver-enriched massive replacement bodies in carbonate country rocks; metals were derived, in part, from older deposits and enrichments.
Late Pennsylvanian and Permian	Metalliferous sedimentation concentrated barium, lead, silver, and zinc in carbonaceous sediments; metals were derived, in part, from detrital material from older deposits.
Early Mississippian	Metalliferous sedimentation concentrated barium, lead, silver, and zinc in turbidite beds.
Middle and Late Devonian	Significant concentration of silver, lead, and zinc in stratiform lenses and disseminations of sulfide minerals interbedded with calcareous mudrock, chert, and limestone; the source of metals was metalliferous solutions discharging on the sea floor.
Ordovician and Devonian	Metalliferous sedimentation concentrated Pb, Ag, Zn, V, Mo, Ba, Cu, Ni, and possibly other metals in carbonaceous sediments.

CRETACEOUS EVENTS

Emplacement of the Idaho batholith during Cretaceous and Paleocene time resulted in many of the important mineral deposits in the area. Three physical types of deposits formed during this event: replacement deposits, stockworks of veinlets, and large veins and lodes. Polymetallic replacement deposits, most notable in the Minnie Moore and Carrietown mineralized areas (fig. 1), have been mined for their silver, lead, and zinc content. These are stratabound to irregular-shaped, massive deposits of galena and sphalerite that formed by replacement of carbonate-bearing members of the black shale sequence near its contacts with intrusive rocks of the Idaho batholith and satellite bodies.

Stockwork deposits formed during this event concentrated molybdenum, tungsten, and some copper. Known deposits of this type in the region include Thompson Creek, a producing world-class molybdenum deposit north of the study area in the Challis quadrangle, and the Little Boulder Creek molybdenum prospect, just north of the Hailey quadrangle.

A variety of vein deposits formed during the late stages of batholith formation (about 78–57 Ma; L.W. Snee, U.S. Geological Survey, 1992, unpublished data), during a period of transition from dominantly compressional to dominantly extensional tectonism. Two general types are recognized: polymetallic veins and polymetallic quartz veins and lodes. Cretaceous polymetallic veins, characterized by quartz-siderite gangue, are mainly within the black shale terrane and have produced silver, lead, and zinc. Notable examples are in the Minnie Moore, Triumph, and Bullion mineralized areas (fig. 1). Polymetallic quartz veins and lodes are characterized by several generations of quartz and limited sulfide minerals. They are mainly within rocks of the Idaho batholith or in siliceous and calcareous sedimentary rocks within or close to the batholith and have produced principally gold and silver. Most deposits in the Atlanta, Rocky Bar, and Hailey gold belt mineralized areas (fig. 1) are of this type.

EOCENE EVENTS

Tertiary igneous and hydrothermal activity and attendant metallization started about 50 Ma during regional extension characterized by deep-seated high-angle fractures and rift structures. Metals introduced during this event include Au, Ag, Mo, W, Be, Sn, Sb, Bi, F, Pb, Cu, and Zn. Solutions transporting these metals also may have mobilized and reconcentrated metals from earlier deposits. Tertiary metal concentrations are of three general types: replacement of sedimentary carbonate rocks, stockworks of veins and veinlets, and vein systems.

Tertiary replacement deposits, including skarn, concentrated copper, lead, zinc, silver, and gold, formed in carbonate-bearing sedimentary rocks around Tertiary intrusive bodies. Replacement deposits in the Mackay mineralized area (fig. 1) are typical copper skarns mined mainly for copper. Replacement deposits in the Summit and Muldoon mineralized areas (fig. 1), hosted by the Drummond Mine Limestone Member of the Copper Basin Formation (pl. 1), have produced some lead and silver and may be distal zinc-lead skarns.

Stockwork deposits of Eocene age concentrated molybdenum and tungsten within or close to plutonic bodies. Examples are in the Summit mineralized area (fig. 1). Eocene granite batholiths in this area are enriched in beryllium, molybdenum, tungsten, and tin; it is possible that greisens formed during some of the later crystallization stages (Smith, this volume).

Hydrothermal activity, starting about 50 Ma and continuing intermittently to the present, produced a variety of vein deposits. Polymetallic veins that have produced some silver, lead, and zinc formed in open fractures and breccia zones in the vicinity of Tertiary plutons in all terranes. Examples in the Mackay mineralized area (fig. 1) are in carbonate terrane; those in the Lead Belt, Lava Creek, and Muldoon mineralized areas (fig. 1) are in flysch terrane; and those in the Summit mineralized area (fig. 1) are in black shale terrane.

Epithermal precious-metal veins, some of which have been major gold producers, formed from hydrothermal cells in areas of volcanism. Most of these deposits are hosted by, or are closely associated with, Eocene volcanic and hypabyssal rocks. Examples are in the Quartzburg, Lava Creek, and Burma Road mineralized areas (fig. 1).

MIOCENE AND YOUNGER EVENTS

A few gold- and uranium-bearing deposits that are related to hot-springs systems formed during Miocene to Holocene time. Examples are in the Magic and Elk Creek mineralized areas (fig. 1). Hydrothermal solutions coursing along major high-angle fracture systems formed large bodies of jasperoid in Paleozoic carbonate, flysch, and black shale terranes and in altered rock in Tertiary volcanic terranes. Some of these solutions may have been metal bearing, in which case large low-grade sediment- and volcanic-hosted epithermal precious-metal deposits may have formed (Soulliere and others, this volume).

PLEISTOCENE TO HOLOCENE EVENTS

During Pleistocene and Holocene time, radioactive black-sand and gold placers formed. Minerals concentrated in the late-stage crystallization products of the Cretaceous Idaho batholith and Eocene granitic plutons were the sources for the minerals in the radioactive black-sand placer deposits. Lode deposits were sources for the placer gold.

REFERENCES CITED

Allen, R.M., Jr., 1952, Geology and mineralization of the Volcano district, Elmore County, Idaho: Moscow, University of Idaho, M.S. thesis, 55 p.

Anderson, A.L., 1929, Geology and ore deposits of the Lava Creek district, Idaho: Idaho Bureau of Mines and Geology Pamphlet 32, 70 p.

———1939, Geology and ore deposits of the Atlanta district, Elmore County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 49, 71 p.

———1943, Geology of the gold-bearing lodes of the Rocky Bar district, Elmore County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 65, 39 p.

———1947a, Epithermal mineralization at the Last Chance and Horn Silver Mines, Lava Creek District, Butte County, Idaho: Geological Society of America Bulletin, v. 56, p. 451–482.

———1947b, Geology and ore deposits of Boise Basin, Idaho: U.S. Geological Survey Bulletin 944–C, 319 p.

Anderson, A.L., Kiilsgaard, T.H., and Fryklund, V.C., Jr., 1950, Detailed geology of certain areas in the Mineral Hill and Warm Springs mining districts, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 90, 73 p.

Anderson, A.L., and Wagner, W.R., 1946a, A geological reconnaissance in the Little Wood River (Muldoon) district, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 75, 22 p.

———1946b, A geologic reconnaissance of the Hailey gold belt (Camas district), Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 76, 26 p.

Bennett, E.H., in press, Geology and mineral deposits of part of the western half of the Hailey 1°×2° quadrangle, Idaho, *with a section on the* Neal mining district *by* Thor H. Kiilsgaard and Earl H. Bennett *and a section on the* Dixie mining district *by* Thomas M. Jacob: U.S. Geological Survey Bulletin 2064–W.

Callaghan, Eugene, and Lemmon, D.M., 1941, Tungsten resources of the Blue Wing district, Lemhi County, Idaho: U.S. Geological Survey Bulletin 931–A, 21 p.

Cavanaugh, P.C., 1979, The geology of the Little Boulder Creek molybdenum deposit, Custer County, Idaho: Missoula, University of Montana, M.S. thesis, 100 p.

Cook, E.F., 1956, Tungsten deposits of south-central Idaho: Idaho Bureau of Mines and Geology Pamphlet 108, 40 p.

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

Federspiel, F.E., Gabby, P.N., Neumann, T.N., and Leszcykowski, A.M., 1987, Mineral resources of the South Boise Yuba Study Area and additions, Blaine, Camas, and Elmore Counties, Idaho: U.S. Bureau of Mines Open-File Report MLA 64–87, 117 p.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658, p. 117–132.

Hall, W.E., Rye, R.O., and Doe, B.R., 1978, Wood River mining district, Idaho—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 579–592.

Hall, W.E., Schmidt, E.A., Howe, S.S., and Broch, M.J., 1984, The Thompson Creek, Idaho, porphyry molybdenum deposit—An example of a fluorine-deficient molybdenum deposit: Proceedings of the Sixth Quadrennial IAGOD Symposium, v. 6, 1, p. 349–357.

Kiilsgaard, T.H., and Bacon, L.D., in press, Geology and mineral deposits of the Atlanta Hill area, Idaho: U.S. Geological Survey Bulletin 2064–Z.

Kiilsgaard, T.H., Benham, J.R. and Avery, D.W., 1983, Mineral resource potential of the Ten Mile West Roadless Area, Boise and Elmore Counties, Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF–1500–C, scale 1:62,500.

Kiilsgaard, T.H., Freeman, V.L., and Coffman, J.S., 1970, Mineral resources of the Sawtooth Primitive Area, Idaho: U.S. Geological Survey Bulletin 1319–D, 174 p.

Kiilsgaard, T.H., Lewis, R.S., and Bennett, E.H., in press, Plutonic and hypabyssal rocks of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 2064–U.

Kiilsgaard, T.H., Scanlan, T.M. and Stewart, D.E., in press, Geology and mineral deposits of the Boise Basin area, Boise and Ada Counties, Idaho: U.S. Geological Survey Bulletin 2064–BB.

Kiilsgaard, T.H., and Smith, C.L., in press, Geology, geochemistry, and mineral resources of the lower part of the Middle Fork Boise River drainage basin, Boise and Elmore Counties, Idaho: U.S. Geological Survey Bulletin 2064–AA.

Leeman, W.P., 1982, Geology of the Magic Reservoir area, Snake River Plain, Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 369–376.

Lewis, R.S., in press, Alteration and mineralization in the eastern part of the Soldier Mountains, Camas County, Idaho: U.S. Geological Survey Bulletin 2064–V.

Link P.K., and Worl, R.G., in press, Geology and mineral deposits of the Minnie Moore and Bullion mineralized areas, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–Y.

Mahoney, J.B., and Horn, M.C., in press, Geology of the Vienna mineralized area, Blaine and Camas Counties, Idaho: U.S. Geological Survey Bulletin 2064–DD.

Mitchell, V.E., Strowd, W.B., Hustedde, G.S., and Bennett, E.H., 1981, Mines and prospects of the Dubois quadrangle, Idaho: Idaho Bureau of Mines and Geology, Mines and Prospects Map Series, scale 1:250,000, 22 p.

———1986, Mines and prospects of the Challis quadrangle, Idaho (2nd ed.): Idaho Bureau of Mines and Geology, Mines and Prospects Map Series, scale 1:250,000, 56 p.

Moye, F.J., in press, The southeastern Challis volcanic field, central Idaho—Mineral deposit models: U.S. Geological Survey Bulletin 2064–FF.

Park, W.A., in press, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–CC.

Popoff, C.C., 1953, Hermada antimony deposit, Elmore County, Idaho: U.S. Bureau of Mines Report of Investigations 4950, 28 p.

Ratchford, M.E., in press, Epigenetic lead-tin-silver-antimony-gold veins in Boulder Basin, Blaine and Custer Counties, Idaho—Potential for economic tin mineralization: U.S. Geological Survey Bulletin 2064–JJ.

Ross, C.P., 1927, Ore deposits in Tertiary lava in the Salmon River Mountains, Idaho: Idaho Bureau of Mines and Geology Pamphlet 25, 20 p.

Rostad, O.H., 1967, Geochemical case history at the Little Falls molybdenite prospect, Boise County, Idaho, *in* Symposium on geochemical prospecting: Canadian Geological Survey Paper 66–54, p. 249–252.

Skipp, Betty, 1987, Basement thrust sheet in the Clearwater orogenic zone, central Idaho and western Montana: Geology, v. 15, no. 3, p. 220–224.

Stewart, D.E., Schmidt, K.L., Moye, F.J., Mahoney, J.B., and Link, P.K., in press, Eocene intrusive complexes in the Smoky and Boulder Mountains, south-central Idaho: U.S. Geological Survey Bulletin 2064–HH.

Tuchek, E.T., and Ridenour, James, 1981, Economic appraisal of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497, p. 181–292.

Umpleby, J.B., 1917, Geology and ore deposits of the Mackay region, Idaho: U.S. Geological Survey Professional Paper 97, scale 1:375,000, 129 p.

Umpleby, J.B., Westgate, L.G., and Ross, C.P., 1930, Geology and ore deposits of the Wood River region, Idaho: U.S. Geological Survey Bulletin 814, 250 p.

Van Noy, R.M., Ridenour, James, Zilka, N.T., Federspiel, F.E., Evans, R.K., Tuchek, E.T., and McMahan, A.B., 1986, Economic appraisal of the eastern part of the Sawtooth National Recreation Area, Idaho: U.S. Geological Survey Bulletin 1545–E, p. 232–472.

Worl, R.G., Kiilsgaard, T.H., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Worl, R.G., and Lewis, R.S., in press, Vein deposits hosted by plutonic rocks in the Croesus stock and Hailey gold belt mineralized areas, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–X.

Worl, R.G., Wilson, A.B., Smith, C.L., Kleinkopf, M.D., and Sykes, R.C., 1989, Mineral resource potential and geology of the Challis National Forest, Idaho: U.S. Geological Survey Bulletin 1873, 101 p.

Structural Framework of Mineral Deposits Hosted by Paleozoic Rocks in the Northeastern Part of the Hailey 1°×2° Quadrangle, South-Central Idaho

By David W. Rodgers, Paul Karl Link, *and* Audrey D. Huerta

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–B

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

PLATE

[Plate is in pocket]

FIGURES

Structural Framework of Mineral Deposits Hosted by Paleozoic Rocks in the Northeastern Part of the Hailey 1°×2° Quadrangle, South-Central Idaho

By David W. Rodgers,[1] Paul Karl Link,[1] *and* Audrey D. Huerta[1]

ABSTRACT

The style, geometry, and timing of structures in the northeastern part of the Hailey 1°×2° quadrangle, south-central Idaho, has had a strong influence on mineralization of Paleozoic rocks. Polyphase deformation has produced a myriad of faults and folds and resulted in significant shortening, extension, and differential uplift. Devonian rift-related faults may have influenced syngenetic silver-lead-zinc mineralization in the Milligen Formation. The Late Devonian and Early Mississippian Antler orogeny was characterized by folding and thrusting of lower Paleozoic siliceous deep-water strata over coeval carbonate platform rocks. Crustal warping probably related to the Middle Pennsylvanian Ancestral Rockies orogeny is manifested by uplift of the Copper Basin highland relative to the subsiding Wood River Basin to the west.

During the Late Jurassic to Cretaceous Sevier orogeny, folds and thrusts formed and the Atlanta lobe of the Idaho batholith was emplaced. Open, upright, north-trending folds are present in the Smoky Mountains, whereas tight, overturned, northwest-trending folds are present to the east in the Pioneer and Boulder Mountains. Map-scale folds accommodated at least 25 percent shortening, whereas pressure solution, small-scale folding, and shearing resulted in additional unmeasured shortening. Of the many thrust faults previously identified, only the Pioneer thrust fault and possibly the Trail Creek fault are interpreted as major; these have inferred top-to-the-east-northeast slip of several tens of kilometers. Silver-lead-zinc veins in the Triumph mine formed in minor thrust faults above inferred Late Cretaceous intrusive bodies. Vein deposits of the Minnie Moore area are hosted by northwest-striking, steeply dipping shear zones that formed shortly after intrusion of plutons about 90 Ma. In the Carrietown mineralized area, smaller shear-zone-hosted silver-lead-zinc veins formed about 80 Ma adjacent to the Idaho batholith.

During the Late Cretaceous to Eocene, the southeastern Idaho batholith and adjacent country rock were exhumed 8–12 km, after which occurred an extended period of early Eocene(?) to middle Oligocene deformation. In the early stages of this tectonism top-to-the-northwest slip occurred along low-angle faults including the Lake Creek and Trail Creek faults in the Pioneer and Boulder Mountains. The Lake Creek fault slipped about 18 km in a dextral-normal sense; most slip occurred before 51–48 Ma. Shear zones along the Lake Creek fault contain silver-lead veins, and in the Triumph and Minnie Moore mineralized areas low-angle faults offset Cretaceous veins. During and after eruption of the main body of the Challis Volcanic Group (48–47 Ma), northeast-striking high-angle faults and dikes formed; northwest-southeast extension of perhaps 6–25 percent probably was accommodated by faulting. Some fault zones near volcanic vents show silicification and alteration. In the middle Eocene and middle Oligocene, the Pioneer Mountains metamorphic core complex was uplifted. Uplift was probably associated with 17–23 km of top-to-the-northwest slip along the Wildhorse detachment fault.

Neogene basin and range faulting produced northwest-striking northeast-tilted fault blocks. The amount of southwest-northeast extension is probably about 10–15 percent. Existing mineral deposits were offset, but no new deposits formed near the surface.

[1]Department of Geology, Idaho State University, Pocatello, Idaho 83209.

INTRODUCTION

The location of mineral deposits hosted by Paleozoic rocks in south-central Idaho is influenced by structures that formed during several orogenic events. Ore deposits are concentrated in joints, shear zones, fault zones, and the cores of folds and are offset by high- and low-angle faults that have meters to tens of kilometers of slip. The area is part of the "central Idaho black-shale mineral belt," as defined by Hall (1985) to include carbonaceous mudrock of various ages along the southeastern edge of the Idaho batholith.

We focus herein on the structural framework of mineralized Paleozoic strata in the northeastern part of the Hailey 1°×2° quadrangle in south-central Idaho. The framework of mineralization and the mineralized areas are described by Worl and Johnson (this volume). Details of Paleozoic stratigraphy are described by Link and others (this volume), and more detailed discussions of mineral genesis and remobilization are presented by Burton and Link (this volume), Darling and others (this volume), Mahoney (this volume), Turner and Otto (this volume), Winkler and others (this volume), Link and Worl (in press), Mahoney and Horn (in press), and Park (in press).

The study area is bisected by the northwest-trending Wood River valley, along which are the towns of Ketchum, Hailey, and Bellevue. To the west of the Wood River valley are the Smoky Mountains, to the north are the Boulder Mountains, and to the northeast are the Pioneer Mountains. The headwaters of the Salmon River are in the northern Smoky Mountains, the headwaters of the East Fork Salmon River are in the northern Boulder Mountains, and the Big Lost River flows through the eastern Pioneer Mountains and Boulder Mountains. Major mineralized areas include the Triumph area in the western Pioneer Mountains near Hailey and Ketchum, the Minnie Moore area west of Bellevue, the Bullion area west of Hailey, and the Carrietown area in the southern Smoky Mountains. The location of individual 7.5-minute quadrangles in the study area is shown in figure 1.

The distribution of rocks in the study area is complex, but a general northwest-trending outcrop pattern is shown by most major rock units (plate 1). From southwest to northeast, the major rock units are the Cretaceous Idaho batholith, the upper Paleozoic Sun Valley Group (Dollarhide, Grand Prize, and Wood River Formations), lower Paleozoic strata (Milligen Formation and unnamed Ordovician, Silurian, and Devonian strata), and the Mississippian Copper Basin Formation. Overlying all these units is the Eocene Challis Volcanic Group. Proterozoic to Paleozoic metasedimentary rocks and Eocene plutons crop out in the Pioneer Mountains metamorphic core complex in the central Pioneer Mountains. Structurally, Paleozoic strata throughout the area are involved in map-scale folds that have north- to northwest-trending hinges. Four types of map-scale faults are present: (1) thrust faults such as the north-northwest-striking Pioneer thrust fault in the eastern Boulder Mountains and northern and eastern Pioneer Mountains; (2) dextral-normal faults including the northwest-striking Lake Creek and Trail Creek faults in the western Boulder and Pioneer Mountains and the arched Wildhorse detachment fault of the Pioneer Mountains core complex; (3) northeast-striking normal faults exposed throughout the map area, most notably the White Mountain fault within the south-central Pioneer Mountains; and (4) north-northwest striking normal faults exposed throughout the area.

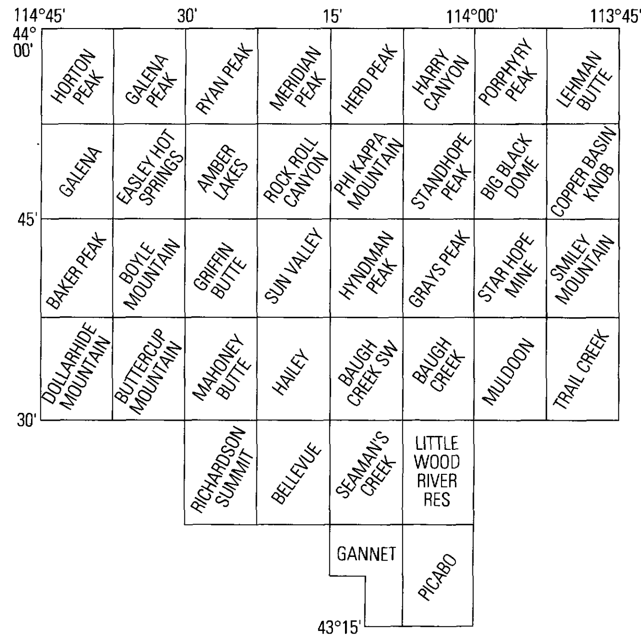


Figure 1. Index map showing location of 7.5-minute quadrangles for the northeastern part of the Hailey 1°×2° quadrangle, south-central Idaho.

Paleozoic strata of south-central Idaho record a protracted orogenic history that has been unraveled in large part by previous workers (Lindgren, 1900; Umpleby and others, 1930; Anderson and others, 1950; Nilsen, 1977; Hall, Rye, and Doe, 1978; Skipp and others, 1979; Dover, 1981, 1983; Hall, 1985; Link and others, 1988). We briefly review this history before describing the structural evolution of the area in more detail.

Subsidence during the latest Proterozoic of Early Proterozoic crystalline rocks and overlying Middle Proterozoic strata (exposed in the core of the Pioneer Mountains) produced a west-facing passive continental margin that persisted until the end of the Silurian. Isolated exposures of lower Paleozoic shelf strata of this passive margin crop out east of the Cretaceous Pioneer thrust fault, and coeval deep-water (siliceous) strata are exposed west of the Pioneer thrust fault from Trail Creek north to the North Fork of the Big Lost River. A pulse of Devonian rifting, subsidence, and stratiform syngenetic silver-lead-zinc mineralization is recorded within the Milligen Formation, which is extensively exposed east of the Wood River valley near Hailey and Ketchum and in the Boulder Mountains. The Lower Mississippian part of the Copper Basin Formation accumulated during and after the Late Devonian–Early Mississippian Antler orogeny in a foreland basin-flysch trough to the east of a thrusted Antler highland, and the upper part of the formation is interpreted to be a marginal-marine facies that represents filling of the foreland basin. Pennsylvanian and Permian crustal warping, possibly associated with formation of the Ancestral Rockies, is reflected by relative uplift of the area now east of the Pioneer thrust fault to form the Copper Basin highland and relative subsidence of the area now west of the Pioneer thrust fault to form the epicratonic Wood River Basin in which the Sun Valley Group was deposited.

During the Late Jurassic to Cretaceous Sevier orogeny, Paleozoic and older strata were metamorphosed, folded, thrust eastward, and intruded by granitoid rocks of the Atlanta lobe of the Idaho batholith. North- to northwest-trending, east-vergent folds accommodated most of the shortening. A smaller, but unresolved amount of shortening was accommodated along thrust faults, primarily the Pioneer thrust. Late Cretaceous granitoid rocks are exposed in the Smoky Mountains and include the Atlanta lobe and three satellite plutons (Rooks Creek, Deer Creek, and Croesus stocks) near Hailey and Bellevue. The historically productive Minnie Moore and Triumph mines near Hailey contain shear-zone-hosted silver-lead-zinc veins that formed during late stages of Cretaceous intrusive activity.

Significant exhumation of the Atlanta lobe and adjacent Paleozoic strata occurred during the Late Cretaceous and early Tertiary. By the middle Eocene, rocks formerly at 8–12 km depth were at the surface.

Paleozoic rocks were extensively deformed in early Eocene(?) to middle Oligocene time, before, during, and after eruption of the middle Eocene Challis Volcanic Group. Slippage occurred on several low-angle faults including the Lake Creek and Trail Creek faults. Northeast-striking high-angle normal faults and dikes throughout the area are structurally parallel with the Trans-Challis fault system. The Pioneer Mountains metamorphic core complex formed during this time and was episodically uplifted along the Wildhorse detachment fault. Remobilization and introduction of metals accompanied these deformational events.

Neogene development of the Basin and Range province is manifested by differential uplift and northeast tilting along north- to northwest-striking normal faults. The study area is characterized by rugged topography that formed as a result of erosion by Quaternary glaciers and streams and has a present-day maximum elevation greater than 3,500 m.

Acknowledgments.—This synthesis reflects knowledge obtained by many geologists during the course of the U.S. Geological Survey Conterminous United States Mineral Assessment Program (CUSMAP) in the Hailey quadrangle and the Idaho Initiative project of the Idaho Geological Survey. In particular, we would like to thank Idaho State University graduate students who have attacked the area with inquisitive minds and youthful vigor: L.D. Batatian, B.R. Burton, R.S. Darling, J.K. Geslin, B.T. Jordan, J.B. Mahoney, J.P. O'Brien, D.E. Stewart, S.K. Whitman, and Eric Wilson. Their mapping, stratigraphic, structural, and petrologic studies have been invaluable contributions to our understanding of the geologic history of the area. Other assistance was provided by Kevin Greer, who first compiled and interpreted folded bedding attitudes. B.R. Burton, J.B. Mahoney, and S.E. Box are thanked for excellent reviews of the manuscript and figures. Errors of omission or fact may remain, however, and we take full responsibility for them.

PALEOZOIC SYNGENETIC STRATIFORM MINERALIZATION AND THE ANTLER OROGENY

Fine-grained carbonaceous Devonian strata including the Milligen Formation and correlative units (parts of the Salmon River assemblage and unnamed Silurian and Devonian units) were deposited over much of the eastern part of the Hailey 1°×2° quadrangle and the eastern part of the Challis 1°×2° quadrangle to the north. Turner and Otto (this volume) and Link and others (this volume) describe the stratigraphy of these Devonian rocks. At the Triumph mine east of Ketchum (and at the Livingston and HooDoo mines in the Challis quadrangle), these strata contain stratiform silver-lead-zinc mineral deposits that have demonstrable syngenetic texture (Hall, 1985). By comparison with other large Devonian mineral deposits in Canada, these deposits are thought to have formed in a transtensional or extensional strain regime (Turner and Otto, this volume). In the Milligen Formation, the stratiform ore deposits may have been concentrated in small fault-bounded deep-water basins that were subject to exhalation of hydrothermal fluids enriched in metals (Hall, 1985; Turner and Otto, 1988, this volume). Turner and Otto (1988, this volume) identified possible tuffs and mafic sills in the Milligen Formation in the Triumph mineralized area; however, the absence of faults and Devonian radiometric ages has so far prevented a direct tie between these metal concentrations and sedimentary exhalative processes. These stratiform deposits are broadly constrained as Middle Devonian (Eifelian) in age (Link and others, this volume). Devonian ore deposits may be a significant source of metals that were remobilized during Cretaceous and Eocene magmatic activity and are now in Paleozoic rocks of various ages in the black shale mineral belt (Hall, Rye, and Doe, 1978; Hall, 1985, 1987; Sanford and Wooden, this volume).

Based on recognition of soft-sediment folds of inferred tectonic origin in the Milligen Formation (Turner and Otto, 1988, this volume), Milligen deposition may have been concurrent with, or closely followed by, early compressional phases of the Late Devonian to Early Mississippian Antler orogeny. The transition from basin formation and associated rifting to compressional deformation and, indeed, the details of Antler deformation in south-central Idaho are not well understood.

The later phases of the Antler orogeny produced axial planar cleavage and tight to isoclinal folds in the Milligen Formation in the Triumph area and in the Boulder Mountains (Davis, 1984; Turner and Otto, 1988, this volume), complex folds in the Milligen Formation in the Lake Creek area (Burton and Link, this volume), and regional tilting of the Milligen Formation below a Middle Pennsylvanian unconformity in the northern Boulder Mountains (Batatian, 1991; Burton and Link, this volume). These structures are attributed to the Antler orogeny because they have only been found in pre-Mississippian rocks and were overprinted by northeast-vergent folds during the Mesozoic Sevier orogeny. Although some outcrop belts of Milligen Formation west of the Trail Creek fault show clear polydeformational structures, only one fabric can be demonstrated in other areas. Polydeformational fabrics have not been identified east of the Trail Creek fault in the Lower Ordovician to Middle Silurian Phi Kappa Formation, Middle Silurian Trail Creek Formation, Devonian Milligen Formation, and unnamed Silurian and Devonian units. Our observation is that an Antler fabric is present in many, but not all, outcrops of Devonian and older strata west of the Trail Creek fault but is not present east of the fault. The presence of an Antler fabric in central Idaho has been long debated (Dover, 1980), and the nature of such a fabric requires further study.

Based on the presence of a Mississippian-age flysch trough (Copper Basin Formation) containing clasts derived from the Milligen Formation, Skipp and Hall (1975) and Nilsen (1977) suggested that the Milligen Formation was thrust eastward to form a highland during the Late Devonian to Early Mississippian Antler orogeny. The thrust fault that accommodated this shortening cannot, however, be conclusively located in south-central Idaho (Dover, 1980). If it is present in the Hailey 1°×2° quadrangle, it may be concealed beneath the outcrop belt of upper Paleozoic Wood River Formation in the Boulder Mountains (Roberts and Thomasson, 1964), or it may have been reactivated to become the Cretaceous Pioneer thrust fault (Wust, 1986) or the early Eocene(?) Trail Creek fault. The Pioneer thrust fault is an unlikely candidate because it contains Mississippian flysch in its hanging wall (Link and others, this volume). We observe that the Trail Creek fault forms the eastern boundary of a demonstrable Antler fabric in the Milligen Formation and that the fault zone shows several stages of deformation, especially in its footwall where the Milligen Formation is tightly folded and northwest of Trail Creek where the fault zone is a mylonitic zone tens of meters wide (Dover, 1983). Similar stratigraphic members of the Milligen Formation are present in the footwall and hanging wall of the Trail Creek fault (Bruner, 1991; Link and others, this volume), and thus any Antler-age slip along the Trail Creek fault would be less than the distance over which facies changes occurred in the original Milligen basin.

Latest Devonian to Early Mississippian uplift of the Antler highland in Nevada and Idaho produced a source for clastics that formed a wedge of chert-rich flysch and molasse which becomes finer grained and more carbonate-rich eastward (Poole and Sandberg, 1991). Within central Idaho these strata comprise the eastward-fining Mississippian Copper Basin Formation, which was deposited in the foreland basin east of the Antler highland. With the exception of a few isolated exposures in the hanging wall of the Pioneer thrust fault (Link and others, this volume), the Copper Basin Formation now crops out east of the Pioneer thrust fault (Nilsen, 1977; Skipp and others, 1979; Skipp and Hall, 1980; Dover, 1983; Link and others, this volume). Black and gray chert and cleaved argillite clasts in the Copper Basin Formation that were derived from the Devonian Milligen Formation suggest that the Milligen was deformed prior to Early Mississippian time and was part of the western Antler highland (Davis, 1984).

PENNSYLVANIAN-PERMIAN WOOD RIVER BASIN

Middle Pennsylvanian to Early Permian subsidence of the Wood River Basin produced a depositional site for thick sequences (more than 3,000 m) of fine-grained mixed carbonate-siliciclastic strata of the Sun Valley Group (Mahoney and others, 1991). The Sun Valley Group contains the Wood River, Grand Prize, and Dollarhide Formations, each of which on plate 1 is divided into lower and upper units.

The Sun Valley Group contains two upward-fining stratigraphic cycles interpreted to represent two periods of tectonic subsidence (Mahoney and others, 1991). Subsidence during the lower cycle (Desmoinesian to Missourian, Middle to Late Pennsylvanian) provided accommodation space for the Hailey Member and correlative strata that make up the basal part of the Sun Valley Group. These units were deposited in a southwest-flowing braid-delta system and adjacent deep-water slope. The braid delta drained the Copper Basin highland (now the eastern Pioneer Mountains) where the Copper Basin Formation cropped out (Winsor, 1981). The Copper Basin highland may have been uplifted along an east-dipping reverse fault, as suggested by Mahoney and others (1991), but the location of this fault has not been identified. Uplift of the Copper Basin highland and coeval subsidence of the Wood River Basin to the southwest was probably related to the Ancestral Rockies orogeny (Skipp and Hall, 1980; Kluth, 1986; Mahoney and others, 1991; Link and others, this volume).

During the second cycle of subsidence (Virgilian to Wolfcampian, Late Pennsylvanian to Early Permian), the Copper Basin highland ceased to be a significant source of sediment. Fine-grained carbonate detritus derived from the

Snaky Canyon Formation (now exposed 100 km to the east) may have washed over the site of the former highland on its way to the Wood River Basin. This second phase of subsidence may be related to crustal loading by thrust plates to the west (Geslin, 1993) or, possibly, to accretion of terranes along the western margin of the continent. Such loading may have produced the Dry Mountain orogenic phase in east-central Nevada (Trexler and others, 1991) that was synchronous with the Virgilian-Wolfcampian time of most rapid subsidence in the Wood River Basin (Mahoney and others, 1991).

MESOZOIC SHORTENING

Shortening of sedimentary rocks during the Late Jurassic to Cretaceous Sevier orogeny was accommodated by folding and to a lesser extent by thrusting and cleavage development. North-trending folds are upright to east-overturned, and slip along thrusts was generally top-to-the-east. Shortening of rocks now exposed in the core of the Pioneer Mountains metamorphic core complex is manifested by ductile shear zones and folds of all scales. The geology of the core of the Pioneer Mountains has been described by Dover (1981, 1983), O'Neill and Pavlis (1988), and Silverberg (1988, 1990a, b).

STRUCTURAL STYLE AND GEOMETRY

The presence of far-traveled Mesozoic thrust plates in the study area has long been the accepted tectonic model (Umpleby and others, 1930; Hal, Rye, and Doe, 1978; Hall, 1985; Link and others, 1988). Numerous faults have been previously identified as thrusts based on the criteria of brittle shear zones, gentle fault dips, and stratigraphic juxtaposition. Slip along the thrust faults was inferred to be perpendicular to fold hinges, generally top-to-the-east-northeast. Recent mapping (Mahoney and others, 1991; Worl and others, 1991; Mahoney, 1992; Mahoney and Link, 1992; Stewart and others, 1992) and reconstruction of sedimentary facies of the Sun Valley Group (Mahoney and others, 1991; Link and others, this volume) indicate, however, that many of these faults are not thrust faults but instead are (1) shear zones with little slip, (2) gently dipping faults characterized by normal and strike-slip displacement to the northwest, or (3) facies transitions between formations of the Sun Valley Group. With our improved understanding of the regional geology, we now consider a fault a thrust only if it places older rocks on top of younger rocks and (or) preserves evidence of west-southwest–east-northeast slip. All thrust faults shown on plate 1 show one or both of these criteria. Following Silverberg (1990b), we have modified Dover's (1981, 1983) mapping of the lower plate of the Pioneer Mountains core complex to show only older-over-younger thrust faults. The paucity of thrust faults shown on plate 1 represents a significant reinterpretation of the structural geology.

The Pioneer thrust fault places the Ordovician and Silurian Phi Kappa and Silurian Trail Creek Formations or the Devonian Milligen Formation and correlative strata over the Mississippian Copper Basin Formation (Skipp and Hall, 1975; Skipp and Hait, 1977; Dover, 1981, 1983; Link and others, 1988). The fault strikes north, dips west, and is somewhat discontinuously exposed in the northeastern part of the map area (plate 1). Along most of its exposed trace, the thrust fault shows a hanging wall that is flat through the Phi Kappa Formation; a buried footwall-flat most likely is present in the same unit. Several faults immediately west of the trace of the Pioneer thrust fault repeat lower Paleozoic strata and are interpreted to be imbricate splays that merge into the Pioneer thrust fault at depth (Dover, 1981, 1983). This imbricate system of thrusts was shown by Dover as placing both older-on-younger and younger-on-older strata, but based on the structural style to the west (see Wood River thrust discussion below) we interpret younger-on-older juxtapositions near the Pioneer thrust as locally sheared depositional contacts and show a simpler imbricate system in cross section *C–C′* (plate 1). The direction of slip along the Pioneer thrust fault is assumed to be east-northeast, perpendicular to numerous fold hinges in its hanging wall and footwall.

The Glide Mountain thrust of Dover (1981, 1983) places coarse-grained siliciclastic facies of the Copper Basin Formation (unit Mcu, plate 1) over approximately coeval, finer grained, more carbonate rich facies of the Copper Basin Formation (unit Mcl, plate 1). Dover showed the Glide Mountain thrust fault as a continuous structure to the north, east, and south of the Pioneer Mountains core complex, but the fault as mapped is unusual because it cuts both up- and down-section in all directions and eliminates and repeats section and because the fault surface displays irregular changes in orientation. To explain these features, Dover invoked map-scale folding before and after thrusting. An alternative interpretation prepared by Wilson (1992, 1994) and Wilson and Rodgers (1993), based on new mapping, is that the mapped Glide Mountain thrust of Dover is not a regionally extensive fault. It is, instead, a depositional contact in Muldoon Canyon (just east of the map area) and Phi Kappa Creek, a thrust fault having about 200 m of east-northeast slip along the East Fork of the Big Lost River, and a gently northwest-dipping normal fault having about 800 m of top-to-the-northwest slip in Big Fall Creek. In Big Fall Creek the Glide Mountain fault decapitates the 48-Ma Summit Creek stock, indicating a 48-Ma or younger age of movement. Plate 1 shows the Glide Mountain fault as mapped and interpreted by Wilson (1994).

The Wood River thrust fault of Hall, Rye, and Doe (1978) and Dover (1983) was placed by them at the contact between the Devonian Milligen Formation and overlying Pennsylvanian and Permian strata of the Wood River

Formation, about 5–15 km west of the Pioneer thrust fault, but we do not show it on plate 1. Reinvestigation of the contact (Skipp and others, 1986; Link and others, 1988; Burton and others, 1989; Mahoney and others, 1991; Worl and others, 1991; Burton and Link, this volume) suggests instead that at least three geologic relations are present across it (see cross sections, plate 1): (1) an undisturbed unconformity with the Milligen Formation depositionally overlain by the Hailey Member or the Eagle Creek Member; (2) an unconformity across which the Hailey Member has been sheared, boudinaged, and locally eliminated, with east-northeast-trending striations and mineralized quartz veins locally preserved along the shear zones; and (3) a dextral-normal fault characterized by stratigraphic elimination, typically with Eagle Creek and Wilson Creek Members of the Wood River Formation placed over Milligen Formation (Trail Creek and Lake Creek faults, cross sections *C–C′*, *D–D′*, plate 1). Thus, only where the Wood River-Milligen contact is sheared to the east-northeast could it be a thrust fault (having a hanging wall-flat over footwall-flat geometry), but even this type of contact could be the result of flexural slip during folding. We do not recognize the Wood River thrust fault.

The Trail Creek fault (plate 1) is present from the North Fork of the Big Lost River southeastward to the Pioneer Mountains core complex. In the eastern Boulder Mountains, the fault places little-deformed Wilson Creek Member of the Wood River Formation on complexly folded Devonian and older strata. Dover (1983) mapped this younger-on-older fault relation as part of the "Wood River thrust system." At least two interpretations are possible for this fault. First, it may be a thrust fault of Sevier age that was reactivated as an Eocene dextral-normal fault (whose structural style is described later). This possibility is suggested by the intensity of folding in the footwall (interpreted to reflect unusually high strain near the thrust fault), as well as by the present relation of younger rocks on older. Alternatively, it may be only an Eocene dextral-normal fault, and the intense folding of footwall rocks may be unrelated to slip along the fault. Detailed study of the fault is needed to clarify its kinematic history.

Several thrust faults having older-on-younger stratigraphic relations are present west of the Pioneer thrust fault in its hanging wall, but they have relatively short fault traces and minor amounts of slip. The Deer Creek and Murdock Creek thrust faults are in the cores of map-scale folds; if this structural style is typical, then unrecognized blind thrust faults may be associated with other east-vergent folds. Identified minor thrust faults include:

1. The Deer Creek thrust fault west of Hailey, which places the lower member of the Dollarhide Formation on the Eagle Creek Member of the Wood River Formation (Skipp and others, 1994).
2. The Boulder Peak thrust fault in the southern Boulder Mountains, which places the Milligen Formation over the Eagle Creek Member of the Wood River Formation (Ratchford, 1989).
3. The Washington Basin thrust fault, near the northwestern corner of the map area (plate 1), which places the Eagle Creek Member of the Wood River Formation over member 2 of the Grand Prize Formation (Mahoney, this volume).
4. The Murdock Creek thrust fault, exposed in the southern Boulder Mountains between the East Fork of the North Fork of the Wood River and Eagle Creek, which places the Milligen Formation above the Eagle Creek Member of the Wood River Formation, and passes southeastward into the core of a northeast-overturned anticline (Batatian, 1991).
5. Four thrust faults that repeat sections of the Milligen Formation in the Triumph area (Turner and Otto, this volume).

Shortening during the Sevier orogeny was accommodated by northwest- to north-northeast-trending map-scale folding. Folds are open to tight; anticlines are tighter and more angular than synclines, and fold wavelengths are commonly several kilometers. Folds are generally symmetric and upright west of the Wood River valley but asymmetric and east-vergent east of the Wood River valley (cross sections *C–C′*, *D–D′*, plate 1). Mesoscopic folds are uncommon in the coarser grained Grand Prize and Wood River Formations, more common in the fine-grained Dollarhide and Milligen Formations, and quite common in lower Paleozoic strata near the Pioneer thrust fault. These differences suggest that folding was facilitated by high strain, deep burial depths, and large variations in lithology and hence rheology.

Map- and outcrop-scale fold hinges trend north-northeast to northwest and plunge gently to subhorizontally north or south (plate 1, fig. 2). Three general variations in fold orientation were recognized within the Pennsylvanian and Permian Sun Valley Group. First, within the Carrietown mineralized area (Dollarhide Mountain quadrangle, fig. 2) isoclinal folds and cleavage are present in the lower member of the Dollarhide Formation adjacent to the southeastern edge of the Idaho batholith (Whitman, 1990, this volume). The folds are tight to isoclinal, and their hinges plunge moderately to the east-northeast, orthogonal to the regional trend of fold hinges. A first cleavage is sparsely preserved, whereas a second cleavage is well preserved and axial planar to numerous mesoscopic folds. The anomalous trend of fold hinges is interpreted to be the result of high shear strain, in which fold hinges having an initial northwest trend were passively rotated to an east-northeast trend, parallel with the overall shear strain direction (Whitman, 1990, this volume). Formation of the isoclinal folds only adjacent to the batholith suggests that the high strain resulted from lower rock strength during contact metamorphism.

A second variation in fold-hinge trend is a counterclockwise rotation from west to east, from S. 4° E. in the Boyle

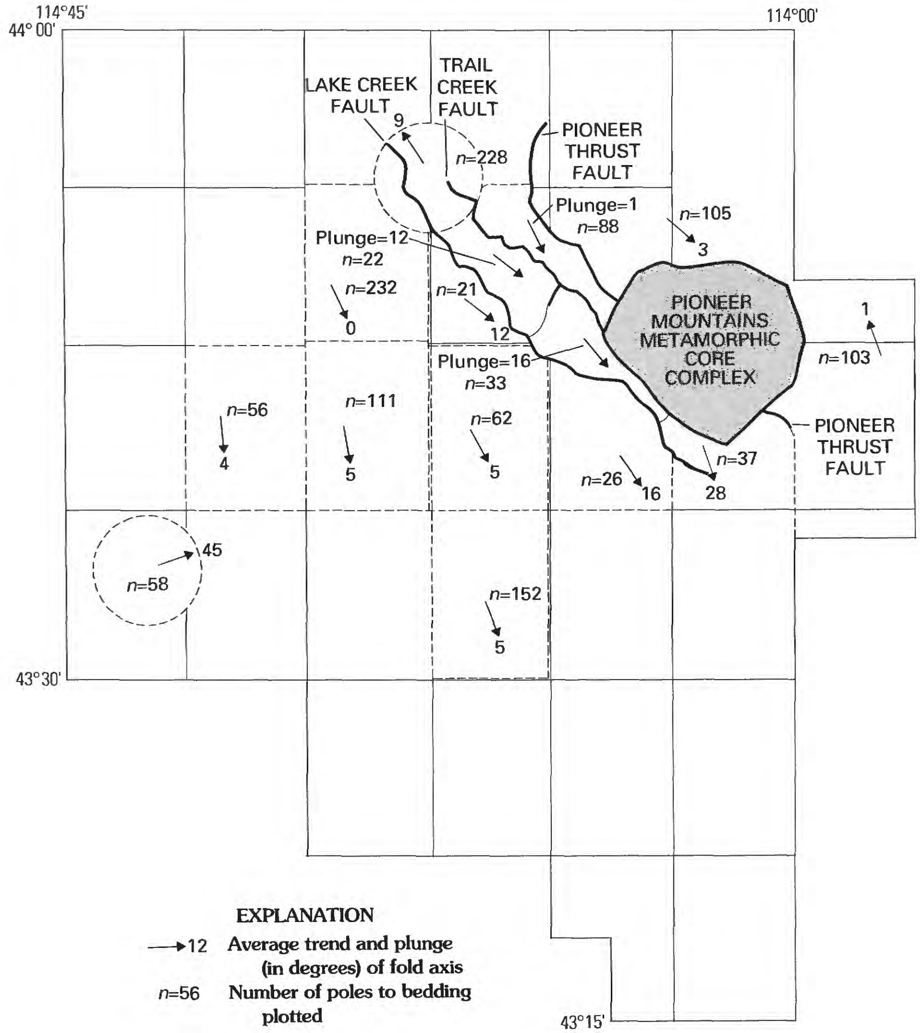


Figure 2. Cretaceous(?) fold-axis orientations in Paleozoic strata across the central part of the study area, Hailey 1°×2° quadrangle. See text for discussion of data. Folds west of the Trail Creek fault involve the Middle Pennsylvanian to Lower Permian Sun Valley Group, folds between the Trail Creek fault and Pioneer thrust fault involve Ordovician to Devonian strata, and folds east of the Pioneer thrust fault involve the Mississippian Copper Basin Formation. To construct the figure, (1) structural domains (dashed lines) were selected using 7.5-minute quadrangles (fig. 1), major faults, or local map areas as boundaries, (2) all Paleozoic bedding attitudes within each domain were plotted on a lower hemisphere equal-area projection, and (3) an average fold axis was measured according to the great-circle distribution of poles to bedding. The derived fold-axis attitude and observed outcrop and map-scale fold-hinge attitudes are consistent within each domain (see also plate 1). Sources of data include Batatian (1991), Batchelder and Hall (1978), Burton (1988), Hall, Batchelder, and Tschanz (1978), Whitman (1990), O'Brien and others (1993), P.K. Link (unpublished mapping, Boyle Mountain, Griffin Butte, and Seamans Creek quadrangles, 1986–1989), and W. Hall and C. Wavra (unpublished mapping, Amber Lakes quadrangle, 1980–1985).

Mountain and Griffin Butte quadrangles to S. 22°–29° E. in the Amber Lakes, Sun Valley, and Hailey quadrangles, to an extreme rotation of S. 53° E. in the Rock Roll Canyon and Phi Kappa quadrangles (fig. 2). We are unsure of the reason for this pattern, but it may be related to (1) an increase in strain caused by greater heating and (or) proximity to the Pioneer thrust fault resulting in the rotation of fold axes toward the east-northeast transport direction, (2) buttressing against the northeastern edge of the Wood River Basin as the hanging wall of the Pioneer thrust fault moved eastward, or (3) buttressing against concealed footwall ramps in the Pioneer thrust fault system. The rotation cannot be attributed to superimposed rotation accommodated by younger faults because fold-hinge trends do not vary across faults such as the Lake Creek and Trail Creek (fig. 2). The first two hypotheses do not satisfactorily explain the observed variations because strata directly against the Pioneer thrust fault (which also demarcates the edge of the Sun Valley Group) have fold-hinge trends of S. 26° E. to S. 20° E., significantly less rotation than fold hinges west of the thrust fault and the edge of the Sun Valley Group outcrop belt. Thus, fold-hinge trends may reflect patterns of footwall ramps in the Pioneer thrust fault or some other, as yet unrecognized, phenomena.

The third trend in fold-hinge orientations is a consistent increase in southeasterly plunge from north to south, from the Ryan Peak quadrangle in the northwestern corner of the study area to the Grays Peak quadrangle in the southeastern corner. In the Ryan Peak quadrangle fold hinges plunge 9° NE. and in the Rock Roll Canyon and Hyndman Peak quadrangles fold hinges plunge 14°–16° SE., whereas in the Grays Peak quadrangle fold hinges plunge 28° SE. The southward tilting of fold hinges may reflect (1) tilting of strata along a broad, south-facing lateral ramp beneath the Pioneer thrust fault, (2) doming of the Pioneer Mountains core complex, or (3) southward tilting of strata due to Neogene downwarping of the Snake River Plain. The first option is the preferred hypothesis because doming would cause radial tilting of fold hinges away from the core, a feature that is not recognized (fig. 2), and the Snake River Plain downwarp extends only a few tens of kilometers north of the plain (Zentner, 1989).

AMOUNT OF SHORTENING

The total amount of shortening by folding and thrusting is difficult to estimate because of the absence of pinpoints across the Pioneer thrust fault, the possibility that flexural flow was an important process of fold formation, and the variable but unknown extent of pressure dissolution during cleavage formation. Skipp and Hait (1977) proposed that 130–150 km of slip occurred along the Pioneer thrust fault because the fault places lower Paleozoic continental slope strata over coeval cratonal strata. This amount of slip is greater than need be, especially if facies juxtaposition occurred by slip along a concealed Late Devonian Antler thrust fault rather than along the Cretaceous Pioneer thrust fault. If the Pioneer thrust fault is similar to better understood Cretaceous thrust faults in the Sevier foreland, only a few tens of kilometers of slip may have occurred along it. In the hanging wall of the Pioneer thrust fault, a minimum of 10 km, or 25 percent, of east-northeast-directed shortening is evident when folds and thrusts in cross section *C–C′* (plate 1) are restored, but this is only an estimate because several folds in this cross section are schematic. Significantly more shortening occurred as a result of small-scale folding, cleavage development, and internal shearing. East of the Pioneer thrust fault, about 25 percent shortening by folding is evident in the Copper Basin Formation (using the cross sections of Dover, 1983).

TIMING OF DEFORMATION

Folds and thrust faults formed by the end of the Cretaceous. According to Skipp and others (1994), the Deer Creek thrust fault in the Mahoney Butte quadrangle (fig. 1) cut the core of an anticline prior to emplacement of the Deer Creek stock, dated at 94.4±0.3 Ma using the $^{40}Ar/^{39}Ar$ technique on hornblende (L.W. Snee, U.S. Geological Survey, unpublished data, 1992). Whitman (1990, this volume) identified coeval folding, cleavage development, and metamorphism in the lower part of the Dollarhide Formation near the Buttercup mine (plate 1) and obtained a whole-rock K-Ar age of 83.9±3.4 Ma from a carbonaceous argillite (R.L. Armstrong, written commun., 1990). Just north of the study area, the 83.0±2.8-Ma (K-Ar, biotite) White Cloud stock intrudes north-trending folds, as does an unnamed and undated stock in Washington Basin (Mahoney, 1992). A foliated gneissose granodiorite in the core of the Pioneer Mountains core complex yielded a hornblende K-Ar age of 67.6±1.6 Ma (Dover, 1983), and hornblende from a schist in the complex yielded a $^{40}Ar/^{39}Ar$ plateau age of 79.3±1.1 Ma (Silverberg, 1990b). All of the ages are best interpreted as ages of cooling below the argon retention temperature of the mineral analyzed and, as such, are minimum ages of metamorphism and plutonism. Thus, available age constraints indicate that deformation had ceased by about 79 Ma, but the actual ages of shortening remain unconstrained. Regionally, the Atlanta lobe of the Idaho batholith was emplaced between about 95 and 75 Ma (Johnson and others, 1988); if shortening was approximately synchronous with magmatism, then it was a Late Cretaceous event.

CRETACEOUS STRUCTURES AND MINERALIZATION

Intrusion of the Idaho batholith and satellite plutons produced the most historically productive mineral deposits in the Wood River region. Shear-zone-hosted silver-lead-zinc veins having a gangue of quartz or siderite are present in several mineralized areas. These mesothermal veins developed during faulting that was synchronous with or followed intrusive activity and were filled by mineral deposition from hydrothermal systems driven by the Cretaceous intrusive activity. East of the Wood River these veins occupy thrust faults and sheared stratigraphic contacts. In the southern Smoky Mountains, the dominant strike of these veins is parallel with Mesozoic fold hinges and sheared axial zones. The mineralized areas (plate 1) (Worl and Johnson, this volume; Link and others, this volume) include Triumph, Minnie Moore, Bullion, Bellevue, Hailey gold belt, Bunker Hill, Deer Creek stock, Rooks Creek stock, Croesus stock, Carrietown, Vienna, and Washington Basin.

Host structures for the Cretaceous vein deposits include:

1. Thrust faults and sheared areas along the unconformity between the Milligen Formation and the Hailey Member of the Wood River Formation. In the Triumph area, much of the lead-silver-zinc ore was produced from shear-zone-hosted veins along thrust faults. The mineralized Fissure fault at the Triumph mine and the ore zone at the North Star mine are interpreted to be the same structure offset by a normal fault (Turner and Otto, this volume). In the Triumph mineralized area the mineralized thrusts are intruded by granitic (quartz porphyry) and andesitic dikes (Kiilsgaard, *in* Anderson and others, 1950) that are displaced by low-angle normal faults, including the Triumph Shaft flat fault (Turner and Otto, this volume). The sheared unconformity between the Milligen and Wood River Formations hosts quartz veins that contain small lead-zinc deposits in the Wood River and Lake Creek mineralized areas (Burton and Link, this volume).

2. Northwest-striking high-angle normal faults. These control locations of veins in the Minnie Moore and Bullion mineralized areas, west of Bellevue and Hailey (Link and Worl, in press). These faults, though generally parallel with earlier fold axes, cut the Croesus and Deer Creek stocks (intruded about 90 Ma) and may represent a period of extension during initial uplift of the stocks. The shear-zone-hosted veins are interpreted as Cretaceous in age because they contain mesothermal mineral assemblages and fluid inclusions and because there are no Eocene intrusive rocks exposed nearby.

3. High-angle faults and minor fractures parallel with Mesozoic fold hinges. These are especially important in the Rooks Creek stock mineralized area (Park, 1990, in press) and the Washington Basin area (Mahoney, this volume). In the Rooks Creek mineralized area, the mineralized veins strike northeast and contain sericite dated by $^{40}Ar/^{39}Ar$ at about 91 Ma (Park, in press).

Mineralization of the Carrietown, Vienna, and Marshall Peak mineralized areas is associated with intrusion of the main-phase batholith about 80 Ma (Darling and others, this volume; Mahoney and Horn, in press). The Marshall Peak and Vienna areas are not shown on plate 1 but are directly west and north of Galena Summit, in the northwestern part of the map area. The host structures are northwest- and northeast-striking shear zones that developed after intrusion of the batholith in the Vienna area but may have been inherited from the pre-intrusion compressional structures in the Carrietown area.

CRETACEOUS TO EOCENE EXHUMATION

Exhumation of the Atlanta lobe and adjacent country rock occurred during Late Cretaceous to Eocene time, during and (or) after folding, thrusting, and batholith emplacement. Evidence for exhumation includes isotopic and geochronologic data from the batholith (Jordan, 1994) and the unconformable contact of the Challis volcanic group on batholith rocks. Distinct lateral and vertical trends have been observed in both the alteration and the apparent K-Ar ages of biotite from Cretaceous plutons in the southern Atlanta lobe, mostly just west of the study area (Criss and others, 1982). These trends were attributed to postemplacement exhumation and reheating of the batholith, including exhumation during the Late Cretaceous to early Paleogene at a rate of about 120 m/m.y., arching during the middle Eocene (and perhaps earlier) about a north-trending axis in the east-central part of the batholith, and reheating during the middle Eocene, with the intensity of reheating increasing toward the east-central part of the core, with depth, and near individual Eocene plutons. Second, in the Smoky Mountains the Idaho batholith was emplaced at a depth of 8–12 km (Whitman, 1990), but basal lavas of the Eocene Challis Volcanic Group unconformably overlie the batholith and its country rock. Thus a cumulative uplift of 8–12 km occurred between 84 Ma, the minimum age of metamorphism and mineralization (Darling, 1987; Whitman, 1990, this volume), and 51 Ma, the approximate age of initial eruption of Challis volcanic rocks (Moye and others, in press).

Exhumation is indirectly indicated by Upper Cretaceous to Eocene sedimentary deposits. In the Boulder and Pioneer Mountains, the conglomerate of Smiley Creek (Paull, 1974) is sparsely preserved beneath the basal Challis lavas (on plate 1 the Smiley Creek is included with the Challis Volcanic Group). The conglomerate is an alluvial fan deposit containing clasts of subjacent Paleozoic formations that suggests well-developed topographic relief within the area. Paull (1974) and Dover (1981, 1983) suggested that the conglomerate could be as old as Cretaceous or as young as Eocene. Burton and Blakley (1988) documented a conformable transition from conglomerate to tuffaceous sandstone and lahar deposits of the Eocene Challis Volcanic Group that suggests deposition occurred just prior to Eocene volcanism, about 51 Ma. The lack of a sedimentologic tie between the Idaho batholith and the conglomerate of Smiley Creek makes it unclear, however, if this deposit reflects uplift of the batholith. In Oregon and California, Eocene arkosic sandstone is interpreted to reflect exhumation of the Idaho batholith. The Idaho batholith and Eocene Tyee Formation of western Oregon show similar Nd-Sm, Rb-Sr, K-Ar, ^{18}O-^{16}O, and D/H isotopic values (Heller and others, 1985), evidence that the batholith was the major source of sediment for the Tyee Formation (although the isotopic variation of the batholith is not sufficient to uniquely identify which lobe provided sediment). Similarly, the Upper Cretaceous to lower middle Eocene Montgomery Creek Formation of northern California contains detrital muscovite that has Cretaceous to Eocene $^{40}Ar/^{39}Ar$ ages unlike those from nearby crystalline rocks but similar to those in the Idaho batholith (Renne and others, 1990).

Based on the relatively long duration and slow, steady rate of Late Cretaceous to Paleogene exhumation, Criss and others (1982) proposed that exhumation was related to regional isostatic uplift. Late Cretaceous to Paleogene drainage patterns and pollen indicate that mountains once capped the Idaho batholith (Axelrod, 1968; Heller and others, 1985; Renne and others, 1990), mountains that probably formed in response to crustal heating and thickening. Erosion of the mountains and consequent isostatic uplift would have caused exhumation of the deep-seated batholith (England and Molnar, 1990; Jordan, 1994). In contrast, Eocene exhumation may reflect more localized isostatic uplift that was a response to more localized Challis magmatism and extension (Criss and others, 1982).

EARLY EOCENE(?) TO OLIGOCENE DEFORMATION

Most faults in the study area are normal and strike-slip faults that accommodated northwest-southeast slip during the Paleogene. Major fault sets and associated structures include low-angle faults, high-angle normal faults, dikes, and the Pioneer Mountains metamorphic core complex. Recent studies by Kim (1986), Wust (1986), O'Neill and Pavlis (1988), Silverberg (1988, 1990a, b), Huerta (1992), and Wilson (1994) significantly improve our kinematic understanding of these structures and reveal that tectonic activity occurred episodically for more than 15 m.y., before, during, and after eruption of the middle Eocene Challis Volcanic Group. In the following section we discuss low-angle faults first because available age constraints indicate that movement on most low-angle faults occurred prior to formation of the other structures.

EARLY EOCENE(?) DEXTRAL-NORMAL FAULTING

Several northwest-striking, gently southwest dipping faults that accommodated dextral-normal slip formed at some time after Cretaceous folding and prior to or during the early stages of Eocene Challis volcanism. We infer that most slip occurred in the early Eocene, but more data are needed to document the precise age(s) of faulting.

STYLE AND GEOMETRY OF FAULTS

Several low-angle faults are exposed in the Boulder, Pioneer, and Smoky Mountains (Kim, 1986; Burton, 1988; Batatian, 1991; Worl and others, 1991; Burton and Link, this volume). The faults generally strike northwest and dip 4°–45° SW. A concentration of fault striations that trend and plunge gently west-northwest to northwest indicate predominantly strike slip displacement, and along at least one fault, the Lake Creek fault, the sense of slip is dextral (Huerta, 1992). The faults have strike lengths of from 1 to more than 40 km and are typically characterized by stratigraphic elimination. The faults are expressed as sharp contacts or brecciated zones several centimeters to meters thick, and in some places, notably at the head of Lake Creek and Trail Creek in the Boulder Mountains, the fault zones are filled with 1–2 m of quartz veins.

Two major low-angle, top-to-the-northwest faults, the Trail Creek and Lake Creek faults, in the northern Boulder Mountains (plate 1) (Batatian, 1991; Burton and Link, this volume) were previously mapped as Mesozoic thrust faults by Umpleby and others (1930) and Dover (1981, 1983) but have been reinterpreted by us as dextral-normal faults because they cut across folds, generally eliminate stratigraphic section, and locally have northwest-trending striations. The trace of the better studied Lake Creek fault extends for more than 40 km from the central Boulder Mountains to the western Pioneer Mountains (plate 1). The fault strikes N. 45°–52° W. and dips 20°–30° SW. (Huerta, 1992). In most places the fault places younger rocks on older rocks, but in a few places the reverse is true because the rocks were folded prior to faulting.

Numerous other low-angle faults having northwest slip have been identified within the study area. Kim (1986) measured northwest-trending striations on the Pioneer thrust fault system, the Glide Mountain thrust of Dover (1983), and several unnamed low-angle faults north and west of the core. The Glide Mountain fault north of Summit Creek clearly displays a top-to-the-northwest fabric (Wilson, 1992, 1994). Turner and Otto (this volume) mapped four subhorizontal faults having northwest striations that cut Challis volcanic rocks in the Triumph mineralized area just west of the exposed core complex. In the central and southern Smoky Mountains low-angle normal faults have been mapped in several places, including in the Mahoney Butte quadrangle (Skipp and others, 1994), in Colorado Gulch southwest of Hailey (Link and others, 1988; Link and Worl, in press), and in the central Smoky Mountains near Baker Peak (Stewart, 1987; Stewart and others, 1992). The low-angle Rockwell fault displaces the Minnie vein in the Minnie Moore mine (Anderson and others, 1950; Link and Worl, this volume), and a low-angle fault places Wood River Formation above Dollarhide Formation in the Bullion mineralized area (Skipp and others, 1994; Link and Worl, in press).

In order to determine if the gentle dips of low-angle faults are due to superimposed tilting, the attitudes of other structures and contacts were compared to their expected initial attitudes. Middle Eocene volcaniclastic rocks are variably tilted as much as 20°, more consistently northeast than other directions (Worl and others, 1991), and the regional outcrop patterns of subplanar volcanic and hypabyssal contacts (for example, Mahoney, 1987) suggest that 10°–15° of tilting to the northeast has occurred, probably in association with basin and range faulting. North-northwest-trending Cretaceous fold hinges consistently plunge less than 16° (fig. 2), except near the Idaho batholith and just south of the Pioneer Mountains core complex, and northeast-striking middle Eocene dikes throughout the region and rare northwest-striking dikes are subvertical. These attitudes are typical of those for untilted hinges and dikes. Finally, the deformational style along major low-angle faults does not vary along strike. For instance, along its 40-km length (subparallel with its slip direction) the Lake Creek fault shows no evidence of ductile shear, though ductile fabrics would be expected if the fault had been tilted southeast by more than about 15°. Taken together, the data suggest that low-angle faults have been tilted less than 15°, making them originally low-angle faults.

AMOUNT OF OFFSET

The offset on most low-angle faults has not been determined. One exception is the Lake Creek fault, the offset of which was determined by restoring piercing points defined by the intersection of a map-scale synclinal trough and the fault surface (Huerta, 1992). Devonian and Pennsylvanian-Permian strata within both the hanging wall and footwall of the fault are folded into northwest-trending, map-scale, overturned synclines (plate 1). The synclines extend along the entire length of the fault, but the fault cuts obliquely across them. Two piercing points are located where the planar contact of the Wilson Creek and Eagle Creek Members of the Wood River Formation is folded into a linear trough and intersects the Lake Creek fault surface (fig. 3). The vector between these piercing points trends N. 50° W., plunges 2° NW., and is 18 km long. This slip vector is corroborated by the unique geometries of the synclines. To the northwest both synclines are overturned, whereas to the southeast both synclines are upright, and restoration of 15 km slip juxtaposes overturned parts and upright parts of the folds (fig. 3). The calculated slip vector indicates that displacement along the Lake Creek fault involved significant dextral slip and minor normal slip.

Turner and Otto (this volume) estimate a minimum of several hundred meters of northwest-directed slip along low-angle faults near the Triumph mine southeast of Ketchum.

TIMING OF FAULTING

The age of dextral-normal faulting is bracketed by crosscutting relations with the Challis Volcanic Group. Regionally, intermediate volcanism occurred from 51 to 48 Ma and silicic volcanism occurred from 48 to 44 Ma (Moye and others, in press; L.W. Snee, U.S. Geological Survey, unpublished data, 1992).

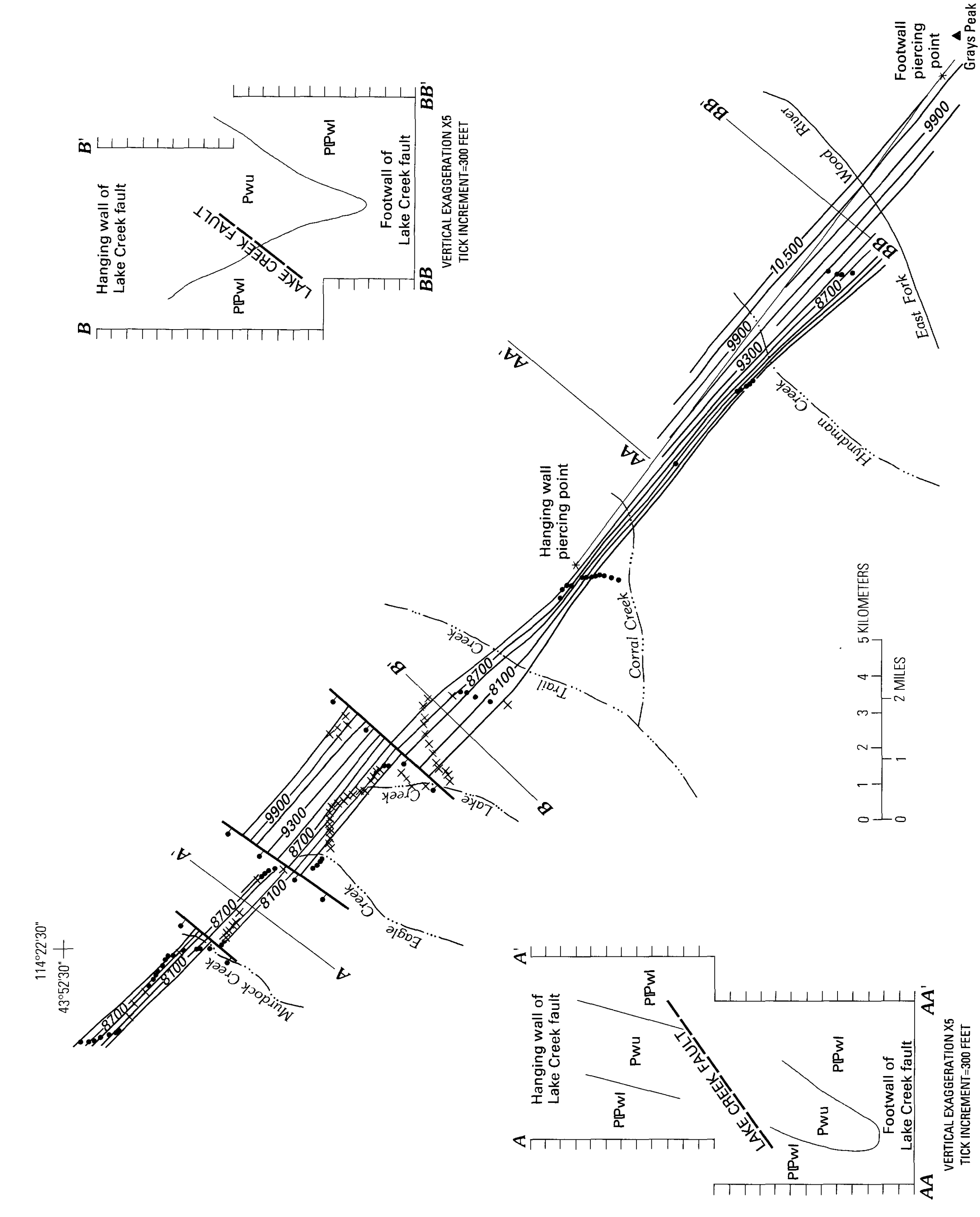
1. The low-angle normal fault near Big Fall Creek previously identified as the Glide Mountain thrust fault by Dover (1981, 1983) was thought by Dover to be pinned by the Summit Creek stock. New mapping by Wilson (1994) indicates that the fault decapitates the stock, which was emplaced 48.5±2.0 Ma (K-Ar, hornblende) (Zartman, unpublished data *in* Silverberg, 1990b). Porphyroblasts in the contact aureole of the stock are sheared in the normal fault zone, further evidence that some slip occurred after stock emplacement (Wilson, 1992).

2. Near their northern terminations the Lake Creek and Trail Creek faults encounter a dacite-rhyolite hypabyssal center that was uplifted 450 m along a steeply south dipping normal fault (plate 1, cross section *A–A′*) (Batatian, 1991). The hypabyssal center extends across the low-angle faults, indicating that emplacement the center broadly postdates slip along the faults. Within the center all rhyolite dikes pin the faults, whereas many dacite dikes are offset a few to at least several tens of meters (Batatian, 1991; oral commun., 1992). Dacite was dated at 50.2±1.8 Ma using the K-Ar technique (Fisher and others, 1983), and hornblende from a rhyolite porphyry 2 km north of the center was dated at 47.2 Ma using the K-Ar technique (Dover, 1981). This, and the fact that the Lake Creek fault to the south is overlain by undated Eocene volcanic breccia, is interpreted to indicate that most slip along the Lake Creek and Trail Creek faults occurred prior to Challis volcanism but that minor slip was coeval with early Challis volcanism.

3. Near Baker Peak a low-angle fault is intruded by dacite porphyry dikes 48 m.y. old (K-Ar age, whole rock) (Stewart and others, 1992, in press).

4. Low-angle normal faults cut dacite lava in the Triumph area southeast of Ketchum (Turner and Otto, this volume) and at several other locations, and they cut dacite *and* rhyolite volcanic rocks in the Easley Hot Springs quadrangle (Ratchford, 1989).

In summary, it is likely that the majority of movement on the Lake Creek and Trail Creek faults was prior to intermediate volcanism (51–48 Ma), that movement on other faults was at some time prior to silicic volcanism (48–44 Ma), and that movement on a few faults was after silicic volcanism. We theorize that the pre-Challis dextral-normal faults formed just prior to Challis magmatism, not well before it, because in the Challis 1°×2° quadrangle northwest-striking dextral strike-slip faults are associated with a northwest alignment of early Challis volcanic vents (McIntyre and others, 1982).



114°22'30"
43°52'30"
8700
8100
9300
9900
10,500
Murdock Creek
Eagle Creek
Lake Creek
Trail Creek
Corral Creek
Hyndman Creek
East Fork
Wood River
Hanging wall piercing point
Footwall piercing point
Grays Peak
A
A'
B
B'
AA
AA'
BB
BB'
0 1 2 3 4 5 KILOMETERS
0 1 2 MILES
Hanging wall of Lake Creek fault
Footwall of Lake Creek fault
LAKE CREEK FAULT
Pwu
PIPwl
VERTICAL EXAGGERATION X5
TICK INCREMENT=300 FEET

MIDDLE EOCENE TO OLIGOCENE EXTENSION

Widespread middle Eocene to middle Oligocene extension is manifested by high-angle faults, ductile shear zones, boudinage, and dike swarms. Extension was coeval with and postdated the main phase of eruption of the Challis Volcanic Group and resulted in uplift of the Pioneer Mountains metamorphic core complex.

STRUCTURAL STYLE AND GEOMETRY

Dike swarms associated with Challis volcanism have been mapped in several places, notably in the northern Boulder Mountains (Tschanz and others, 1986; Batatian, 1991; Schmidt, 1994; Moye and others, in press) and in the Smoky Mountains west of the area of plate 1 (Worl and others, 1991; Stewart and others, in press). Individual dikes are present throughout the study area but are too small to show on plate 1. Dike swarms are more common to the west and east of the study area, where substantial exposures of plutonic and volcanic rocks are present. All dike swarms and most individual dikes strike northeast and dip almost vertically.

Normal faults that strike northeast, dip moderately to steeply southeast or northwest, and extend for several hundreds to several thousands of meters are present throughout the study area (plate 1). Stratigraphic juxtapositions indicate dip-slip displacements of a few hundred to rarely a few thousand meters. Most of these faults are unnamed, but one, the White Mountains fault, forms the southeast boundary of the Pioneer Mountain core complex (Dover, 1981, 1983) (plate 1, cross section *A–A′*). Because the high-angle fault set is distributed throughout the study area and is parallel with the Trans-Challis fault system in central Idaho (Bennett, 1986), it is inferred to be part of that system.

Figure 3 (facing page). Map view showing structure contours and slip vector of the Lake Creek fault. Contour interval 300 ft (91 m); contours are dashed where extrapolated. Slip vector of N. 50° W., plunge 2°, and 18 km long is based on displacement of piercing points where trough of a Mesozoic syncline pierces the fault surface. Solid circles indicate well-constrained locations; x's indicate poorly constrained locations. Bar and ball on fault line indicates minor normal fault. Cross sections show geometry of Lake Creek fault and of contact between the Wilson Creek Member of the Wood River Formation (Pwu) and the Eagle Creek and Hailey Members of the Wood River Formation (PPwl) (not shown in map view) in the hanging wall (*A–A′*, *B–B′*) and the footwall (*AA–AA′*, *BB–BB′*) of the fault. Stacking of cross sections shows the inferred fold geometry after restoration of 18 km of slip along fault. Modified from Huerta (1992).

PIONEER MOUNTAINS METAMORPHIC CORE COMPLEX

The Pioneer Mountains metamorphic core complex includes a core of Early Proterozoic gneiss separated by the Hyndman ductile shear zone from Middle Proterozoic and lower Paleozoic metasedimentary rocks and intruded by Cretaceous(?) and Tertiary plutons. The core is separated by the Wildhorse detachment fault from overlying, internally faulted Paleozoic and Eocene rocks (plate 1). Dover (1981, 1983) mapped the core and surrounding cover rocks, and Silverberg (1988, 1990a, b) studied the metamorphic, structural, and geochronologic history of the core. Wust (1986) studied the Wildhorse detachment fault, and O'Neill and Pavlis (1988) studied the detachment kinematics in some detail. Kim (1986), Burton (1988), Burton and others (1989), Huerta (1992), Wilson (1994), and Burton and Link (this volume) described the geometry and kinematics of faults in the upper plate. We refer the reader to these studies for detailed discussions of the core and here only briefly review the structural geology of the detachment and upper plate.

The Wildhorse detachment fault (plate 1) is a brittle fault zone several meters thick that flanks the core on three sides (Dover, 1981, 1983; Wust, 1986). The fault surface is not planar but takes the shape of a map-scale kink fold that plunges northwest. The kink axis is along the northwest corner of the core and plunges about 25° NW., whereas the sides of the fault dip 45°–70° to the northeast and southwest, away from the core (Wust, 1986). In all places, the detachment fault shows west-northwest- to northwest-trending striations (Wust, 1986), making it a dextral-normal or sinistral-normal fault, except along the kink axis where it is a dip-slip normal fault. The fault overlies and (or) cuts across a 5–100-m-thick ductile shear zone, characterized by mylonitic gneiss, marble, and quartzite, that has kinematic indicators also indicating top-to-the-northwest shear strain (Wust, 1986; O'Neill and Pavlis, 1988; Silverberg, 1990b). $^{40}Ar/^{39}Ar$ analysis of minerals in the shear zone indicates that the zone cooled rapidly during the early Oligocene (36-33 Ma); Silverberg (1990a, b) interpreted the cooling to reflect tectonic unroofing due to slip along the shear zone.

The upper plate of the Pioneer Mountains core complex extends in all directions away from the core and has no definite outer boundary. The upper plate contains two main sets of Eocene(?) to Oligocene faults having northwest-southeast slip. Previously described low-angle faults (Trail Creek, and Lake Creek faults) accommodated top-to-the-northwest slip after Cretaceous folding and before the main phase of Challis volcanic activity. The previously described northeast-striking high-angle normal fault set was active during and (or) after the middle Eocene. The role of these faults in core complex formation and uplift is not clear. The Pioneer and Trail Creek faults are cut by the Wildhorse detachment fault (Dover, 1981, 1983) (plate 1), whereas the crosscutting relation between the detachment fault and the Lake Creek fault is not known. High-angle normal faults may be temporally related to core complex development, but the small offset accommodated by them contrasts with upper plate fault patterns typical of core complexes. Upper plate rocks in the study area are not significantly tilted by faulting, in contrast to many other core complexes where upper plate strata are strongly tilted by domino-style faulting. At this point in our studies we have not documented significant upper plate extension that is temporally related to uplift of the core complex.

AMOUNT OF EXTENSION

Slip along the Wildhorse detachment fault was estimated by Wust (1986) as at least 17 km based on two lines of evidence: first, upper plate rocks in the Boulder Mountains northwest of the core originated southeast of the core, and second, the Summit Creek stock north of the core might be the beheaded equivalent of an Eocene pluton in the eastern half of the core. There is no stratigraphic evidence, however, that strata northwest and southeast of the core were originally contiguous, and a gravity study of the Summit Creek stock was unable to resolve whether the stock is rooted (Wust, 1986). Silverberg (1990a, b) presented evidence that the core rose as much as 8.4 km during the middle Eocene and proposed 23 km of horizontal displacement along the Wildhorse detachment fault, if it accommodated the 8.4 km of uplift and dipped 20° NW.

Slip along most high-angle faults has not been measured, but the steep dips and relatively small displacements of the faults suggest that the faults individually accommodated small amounts of extension as compared to that along more gently dipping faults. However, because of the large number of high-angle faults, the total northwest-southeast extension accomplished along them may be significant. As an estimate of the approximate cumulative extension, each fault is assumed to dip 70° and show dip-slip displacement of 100 m. If the faults are spaced at a 250-m interval across a fault zone 100 km wide (measured northwest-southeast), the total northwest-southeast extension is about 14 km. Slightly modifying these assumptions (dips of 60°–80°, slips of 50–200 m, fault spacing of 200–400 m) yields estimates of extension ranging from 5 to 20 km, a cumulative extension of from 6 to 25 percent across the fault zone.

TIMING OF MAGMATISM AND EXTENSION

The ages of most structures are incompletely known at this time, but age constraints indicated by crosscutting textures and radiometric dating include the following:

1. Northeast-trending Challis dikes are dated, or served as feeders to volcanic flows that are dated, between 51–48 Ma for intermediate compositions and 48–44 Ma for silicic compositions (Moye and others, in press).

2. High-angle normal faults associated with the Trans-Challis system cut most dikes and rocks of the Challis Volcanic Group and cut low-angle faults having dextral-normal or indeterminate slip directions (Kim, 1986). The high-angle White Mountains fault cuts the Wildhorse detachment fault.

3. Silverberg (1988, 1990a, b) used $^{40}Ar/^{39}Ar$ geochronology on the lower plate of the Pioneer Mountains metamorphic core complex to document two phases of rapid cooling, at 48–45 Ma and at 36–33 Ma. The first phase was associated with 8.4 km of uplift, but the fault(s) that accommodated uplift is not known. The second phase reflects 4 km of uplift along the Wildhorse detachment fault.

According to these data, northwest-southeast extension occurred in the middle Eocene and the middle Oligocene. In the middle Eocene, dikes formed and one or more low-angle faults (the Wildhorse detachment?) must have slipped to produce 8 km of lower plate uplift. The Pioneer Mountains core complex was uplifted in two stages; the Wildhorse detachment fault was active during at least the second stage. The age(s) of high-angle faulting is not certain: the structurally parallel Trans-Challis fault zone formed in the middle Eocene (Bennett, 1986), but at least one high-angle fault in the study area cuts the middle Oligocene Wildhorse detachment fault.

EARLY EOCENE(?)–OLIGOCENE MINERALIZATION

Mineralized areas affected by hydrothermal cells generated during Challis magmatism and associated faulting include much of the Galena, Boulder Basin, Lake Creek, Smoky Mountains, Summit, and East Fork Salmon River areas. Regionally in south-central Idaho, Eocene gold-bearing epithermal ore deposits are associated with silicification along Trans-Challis fault zones formed during northwest-directed crustal extension. Generally the deposits are associated with extensive argillic and sericitic alteration of wallrock.

Most Eocene mineral deposits in Paleozoic rocks of the northeastern Hailey 1°×2° quadrangle are in the Boulder and Smoky Mountains. The deposits are associated with hypabyssal intrusive rocks, especially granite and rhyolite, that make up the final Eocene intrusive phase, dated about 48–44 Ma (L.W. Snee, U.S. Geological Survey, unpublished data, 1992; Stewart and others, in press).

Deposits in the Summit mineralized area of the northern Pioneer Mountains are associated with Eocene granodiorite of the Summit Creek stock and include skarns and veins in the Phi Kappa mine, molybdenum stockwork deposits in the Summit Creek stock in Fall Creek, and epithermal veins in unnamed Silurian and Devonian units in the East Fork of Trail Creek and Bear Canyon (Bruner, 1991).

The Lake Creek mineralized area in the Boulder Mountains (Burton and Link, this volume) contains silver-lead-zinc veins at the Homestake and Long Grade prospects. These veins are in northwest-striking normal faults interpreted to be splays of the underlying Lake Creek dextral-normal fault. Intrusive rocks in the area include Eocene dacite porphyry dikes; larger stocks may be present at depth.

A variety of mineral deposits in the Boulder Basin, Galena, and Smoky Mountains areas are associated with Eocene silicic intrusive complexes. Deposits contain polymetallic veins and replacements in calcareous wallrock of the Milligen Formation and Sun Valley Group (Ratchford, in press).

The main period of mineralization in the Triumph mineralized area was interpreted by Kiilsgaard (*in* Anderson and others, 1950) to have occurred in the early Tertiary. The mineralization was interpreted to have been postbatholith and prevolcanic and associated with postbatholith andesite dikes. In contrast, Umpleby and others (1930), Hall, Rye, and Doe (1978), and Turner and Otto (this volume) interpreted the main period of mineralization to have been associated with Cretaceous intrusive activity and Eocene intrusive rocks in the Triumph mineralized area to postdate mineralization.

NEOGENE BASIN AND RANGE EXTENSION

Neogene tectonic activity in the Wood River area is manifested by the development of northwest-striking basin and range normal faults. No near-surface remobilization of metals can be attributed to these faults, although they influence the locations of active hot springs and possible metal remobilization and deposition at depth. The young faults also cut mineral deposits, locally terminating minable ore shoots.

STRUCTURAL STYLE AND GEOMETRY

In the Smoky and Boulder Mountains, Neogene basin and range faults have broken the crust into gently northeast tilted blocks (plate 1, cross section *B–B′*) (Mahoney, 1987; Stewart and others, in press). The faults have north to northwest strikes, steep to moderate dips, and relatively long continuous map traces and cut across almost all other faults, folds, and pre-Quaternary rocks. Major faults include the Sun Valley fault zone, which strikes north along the east side of the Wood River valley (plate 1, cross sections *B–B′*, *C–C′*, *D–D′*) and cuts Quaternary sediments near Sun Valley (Hall, Batchelder, and Tschanz, 1978), the Boulder front fault, which strikes northwest along the front of the Boulder Mountains (Tschanz and others, 1986; Mahoney and Link, 1992), and the Big Smoky fault, which strikes north through much of the Smoky Mountains along Big Smoky Creek (plate 1, cross section *B–B′*) (Mahoney, 1992; Stewart and others, 1992). Offset along these faults was accompanied by 10°–15° of tilting to the northeast, as shown by regional outcrop patterns of subplanar volcanic and hypabyssal contacts (for example, Mahoney, 1987) and a few measurements of the attitudes of Challis Group volcaniclastic rocks.

Field observation (Umpleby and others, 1930) and analysis of aerial photographs and satellite imagery (Southworth, 1988) demonstrate two prominent sets of lineaments, trending northeast and north-northwest, that control topography and cut rocks of all ages in the study area. One prominent northeast lineament is aligned with the East Fork of the Wood River. Turner and Otto (this volume) describe northeast-striking, steeply dipping faults, having tens of meters of apparent right-lateral displacement, that may be related to the formation of northeast-trending lineaments. Although the northeast-striking faults and lineaments are oriented perpendicular to basin and range faults, crosscutting relations suggest at least some postdate basin and range faults (Turner and Otto, this volume).

AMOUNT AND TIMING OF EXTENSION

The amount of northeast-southwest extension accommodated by basin and range faults is about 10 percent, if basin and range faults in cross section *B–B′* or *C–C′* (plate 1) are restored. Other unmapped faults having lesser offsets are present, making this estimate a minimum, but the true amount of extension probably is not substantially larger.

Little is known about the age of basin and range extension in the study area. The age of initial basin and range extension is not known because the oldest sediments in the Wood River valley are concealed beneath younger basin fill. Some recent slip is indicated by a fault scarp in Quaternary sediments near Sun Valley (Hall, Batchelder, and Tschanz, 1978). In southern Idaho, basin and range faulting probably began in the early Miocene and continued sporadically to the present (Allmendinger, 1982; Rodgers and others, 1990), and a similar age span is likely for the northeastern part of the Hailey quadrangle.

REFERENCES CITED

Allmendinger, R.W., 1982, Sequence of late Cenozoic deformation in the Blackfoot Mountains, southeastern Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 505–516.

Anderson, A.L., Kiilsgaard, T.H., and Fryklund, V.C., Jr., 1950, Detailed geology of certain areas in the Mineral Hill and Warm Springs mining districts, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 90, 73 p.

Axelrod, D.I., 1968, Tertiary floras and topographic history of the Snake River Basin, Idaho: Geological Society of America Bulletin, v. 79, p. 713–734.

Batatian, L.D., 1991, Paleozoic stratigraphy and Cenozoic structure, central Boulder Mountains, Blaine and Custer Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 85 p.

Batchelder, J.N., and Hall, W.E., 1978, Preliminary geologic map of the Hailey 7.5 minute quadrangle, Idaho: U.S. Geological Survey Open-File Report 78–545, scale 1:24,000.

Bennett, E.H., 1986, Relationship of the trans-Challis fault system in central Idaho to Eocene and Basin and Range extension: Geology, v. 14, p. 481–484.

Bruner, D.J., 1991, Origina of mineralizationwithin Devonian-Silurian sedimentary breccias, northeastern Boulder Mountains, Custer County, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 118 p.

Burton, B.R., 1988, Stratigraphy of the Wood River Formation in the eastern Boulder Mountains, Blaine and Custer Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 165 p.

Burton, B.R., and Blakley, J.B., 1988, The basal Challis conglomerate, Custer County, south-central Idaho—Implications for the initiation of Challis volcanism: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 408–409.

Burton, B.R., Link, P.K., and Rodgers, D.W., 1989, Death of the Wood River thrust—Structural relations in the Pioneer and Boulder Mountains, south-central Idaho: Geological Society of America Abstracts with Programs, v. 21, no. 5, p. 62.

Criss, R.E., Lanphere, M.A., and Taylor, H.P., Jr., 1982, Effects of regional uplift, deformation and meteoric-hydrothermal metamorphism on K-Ar ages of biotites in the southern half of the Idaho batholith: Journal of Geophysical Research, v. 87, p. 7029–7046.

Darling, R.S., 1987, The geology and ore deposits of the Carrietown silver-lead-zinc district, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 168 p.

———1989, Geologic map and cross-sections of the northern half of the Dollarhide Mountain quadrangle: Idaho Geological Survey Technical Report 89–1, 1 plate.

Davis, S., 1984, A microstructural perspective of orogenesis in the Pioneer Mountains, central Idaho: Missoula, University of Montana, M.S. thesis, 55 p.

Dover, J.H., 1980, Status of Antler orogeny in central Idaho, clarification and constraints from the Pioneer Mountains, *in* Fouch, T.D., and Magathan, E.R., eds., Paleozoic paleogeography of the West-Central United States: Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, West-Central United States Paleogeography Symposium 1, p. 371–381.

———1981, Geology of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–A, p. 1–75.

———1983, Geologic map and sections of the central Pioneer Mountains, Blaine and Custer Counties, central Idaho: U.S. Geological Survey Miscellaneous Investigations Series Map I–1319, scale 1:48,000.

England, P., and Molnar, P., 1990, Surface uplift, uplift, and exhumation of rocks: Geology, v. 18, p. 1173–1177.

Fisher, F.S., May, G.C., and Johnson, F.L., 1983, Mineral resource potential, geology, and geochemistry of part of the White Cloud-Boulder Roadless Area, Custer County, Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF–1580, scale 1:62,500.

Gehlen, W.T., 1983, The geology and mineralization of the eastern part of the Little Smoky Creek mining district, Camas County, Idaho: Moscow, University of Idaho, M.S. thesis, 116 p.

Geslin, J.K., 1993, Evolution of the Pennsylvanian-Permian Oquirrh–Wood River Basin, southern Idaho: Los Angeles, University of California, Los Angeles, Ph.D. dissertation, 272 p.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

———1987, Stratiform syngenetic deposits of precious and base metals in argillic rocks and micritic limestone, *in* Fisher, F.S., and Johnson, K.M., eds., 1987, Preliminary manuscript for "Mineral-resource assessment and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 205–208.

Hall, W.E., and Batchelder, J.N., 1979, Preliminary geologic map of the Baugh Creek Southwest 7½ minute quadrangle, Idaho: U.S. Geological Survey Open-File Report 79–1617, scale 1:24,000.

Hall, W.E., Batchelder, J.N., and Tschanz, C.M., 1978, Geologic map of the Sun Valley 7.5-minute quadrangle, Blaine County, Idaho: U.S. Geological Survey Open-File Report 78–1058, scale 1:24,000.

Hall, W.E., Rye, R.O, and Doe, B.R., 1978, Wood River mining district—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 579–592.

Heller, P.L., Peterman, Z.E., O'Neill, J.R., Shafiqullah, M., 1985, Isotopic provenance of sandstone from the Eocene Tyee Formation, Oregon Coast Range: Geological Society of America Bulletin, v. 96, p. 770–780.

Huerta, A.D., 1992, Lake Creek fault—Evidence of pre-Challis shear within south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 57 p.

Johnson, K.M., Lewis, R.S., Bennett, E.H., and Kiilsgaard, T.H., 1988, Cretaceous and Tertiary intrusive rocks of south-central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 55–86.

Jordan, B.T., 1994, Emplacement and exhumation of the southeastern Atlanta lobe of the Idaho batholith and outlying stocks, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 110 p.

Kim, A.Y.H., 1986, The kinematics of brittle polyphase deformation within the Pioneer metamorphic core complex, Pioneer Mountains, Idaho: Bethlehem, Pa., Lehigh University, M.S. thesis, 130 p.

Kluth, C.F., 1986, Plate tectonics of the ancestral Rocky Mountains, *in* Peterson, J.A., ed., Paleotectonics and sedimentation in the Rocky Mountain region, United States: American Association of Petroleum Geologists Memoir 41, p. 353– 369.

Lewis, R.S., 1991, Geologic map and geochemical data of the eastern part of the Soldier Mountains, Camas County, Idaho: Idaho Geological Survey Technical Report 91–1, scale 1:24,000.

Lindgren, W., 1900, The gold and silver veins of Silver City, DeLamar, and other mining districts in Idaho: U.S. Geological Survey 20th Annual Report, pt. 3, p. 65–256.

Link, P.K., Skipp, Betty, Hait, M.H., Jr., Janecke, Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 5–42.

Link, P.K., and Worl, R.G., in press, Geology and mineral deposits of the Minnie Moore and Bullion mineralized areas, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–Y.

Mahoney, J.B., 1987, The geology of the northern Smoky Mountains and stratigraphy of a portion of the Lower Permian Grand Prize Formation, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 143 p.

———1992, Geologic map of the Frenchman Peak quadrangle, Blaine and Camas Counties, Idaho: Idaho Geological Survey Technical Report 92–2, scale 1:24,000.

Mahoney, J.B., and Horn, M.C, in press, Geology and mineral deposits of the Vienna mineralized area, Blaine and Camas Counties, Idaho: U.S. Geological Survey Bulletin 2064–DD.

Mahoney, J.B., and Link, P.K., 1992, Geologic map of the Galena quadrangle, Blaine and Camas Counties, Idaho: Idaho Geological Survey Technical Report 92–3, scale 1:24,000.

Mahoney, J.B., Link, P.K., Burton, B.R., Geslin, J.K., and O'Brien, J.P., 1991, Pennsylvanian and Permian Sun Valley Group, Wood River Basin, south-central Idaho, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 2, p. 551–580.

McIntyre, D.H., Ekren, E.B., and Hardyman, R.F., 1982, Stratigraphic and structural framework of the Challis volcanics in the eastern half of the Challis 1°×2° quadrangle, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 3–22.

Moye, F.J., Hackett, W.R., Snider, L.G., and Mertzman, S.A., in press, Stratigraphy, chronology, and petrology of the southeastern Challis volcanic field, central Idaho: U.S. Geological Survey Bulletin 2064–EE.

Nilsen, T.H., 1977, Paleogeography of Mississippian turbidites in south-central Idaho, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 1, p. 275–299.

O'Brien, J.P., Lewis, R.S., Link, P.K., Geslin, J.K., Whitman, S.K., and Schmidt, K.L., 1993, Geologic map of the Buttercup Mountain quadrangle, Blaine and Camas Counties, Idaho: Idaho Geological Survey Technical Report 93–1, scale 1:24,000.

O'Neill, R.L., and Pavlis, T.L., 1988, Superposition of Cenozoic extension on Mesozoic compressional structures in the Pioneer Mountains metamorphic core complex, central Idaho: Geological Society of America Bulletin, v. 100, p. 1833–1845.

Park, W.A., 1990, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: Pocatello, Idaho State University, M.S. thesis, 135 p.

———in press, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–CC.

Paull, R.A., 1974, Upper Cretaceous(?) to lower Eocene post-orogenic conglomerate, south-central Idaho: Geological Society of America Abstracts with Programs, v. 6, no. 5, p. 466.

Poole, F.G., and Sandberg, C.A., 1991, Mississippian paleogeography and conodont biostratigraphy of the Western United States, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 1, p. 107–136.

Ratchford, M.E., 1989, Geology of the Boulder Basin, Blaine and Custer Counties, Idaho: Moscow, University of Idaho, M.S. thesis, 154 p.

———in press, Epigenetic lead-tin-silver-antimony-gold veins in Boulder Basin, Blaine and Custer Counties, Idaho—Potential for economic tin mineralization: U.S. Geological Survey Bulletin 2064–JJ.

Renne, P.R., Becker, T.A., and Swapp, S.M., 1990, $^{40}Ar/^{39}Ar$ laser-probe dating of detrital micas from the Montgomery Creek Formation, northern California—Clues to provenance, tectonics, and weathering processes: Geology, v. 18, p. 563–566.

Roberts, R.J., and Thomasson, M.R., 1964, Comparison of late Paleozoic depositional history of northern Nevada and central Idaho: U.S. Geological Survey Professional Paper 475–D, p. D1–D6.

Rodgers, D.W., Hackett, W.R., and Ore, H.T., 1990, Extension of the Owyhee Plateau, eastern Snake River Plain, and Yellowstone Plateau: Geology, v. 18, p. 1138–1141.

Sanford, R.F., Whitney, H.A., McCafferty, A.E., and Gese, D.D., 1989, Mineral resources of the Little Wood River Wilderness Study Area, Blaine County, Idaho: U.S. Geological Survey Bulletin 1721–D, 15 p.

Schmidt, D.L., 1962, Quaternary geology of the Bellevue area in Blaine and Camas Counties, Idaho: U.S. Geological Survey Open-File Report 62–0120, 134 p.

Schmidt, K.L., 1994, Geology and petrology of the Ibex Canyon igneous complex, Blaine and Custer Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 177 p.

Silverberg, D.S., 1988, Petrologic, $^{40}Ar/^{39}Ar$ isotopic and structural constraints on the Tertiary tectonic evolution of the Pioneer core complex, south-central Idaho: Geological Society of America Abstracts with Programs, v. 20, no. 7, p. A18.

———1990a, Denudation rates and timing of diachronous Paleogene extension in south-central Idaho: Geological Society of America Abstracts with Programs, v. 22, p. 330.

———1990b, The tectonic evolution of the Pioneer metamorphic core complex, south-central Idaho: Cambridge, Massachusetts Institute of Technology, Ph.D. dissertation, 280 p.

Skipp, Betty, and Hait, M.H., Jr., 1977, Allochthons along the northeast margin of the Snake River Plain, Idaho: Wyoming Geological Association Guidebook, Twenty-Ninth Annual Field Conference, p. 499–515.

Skipp, Betty, and Hall, W.E., 1975, Structure and Paleozoic stratigraphy of a complex of thrust plates in the Fish Creek Reservoir area, south-central Idaho: U.S. Geological Survey Journal of Research, v. 3, no. 6, p. 671–689.

———1980, Upper Paleozoic paleotectonics and paleogeography of Idaho, *in* Fouch, T.D., and Magathan, E.R., eds., Paleozoic paleogeography of the West-Central United States: Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, West-Central United States Paleogeography Symposium 1, p. 387–422.

Skipp, Betty, Hall, W.E., and Link, P.K., 1986, Pre-middle Eocene extension in south-central Idaho—The Wood River allochthon: Geological Society of America Abstracts with Programs, v. 18, p. 753.

———1994, Geologic map of the Mahoney Butte quadrangle, Blaine County, Idaho: U.S. Geological Survey Geologic Quadrangle Map GQ–1742, scale 1:24,000.

Skipp, Betty, Sando, W.J., and Hall, W.E., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States–Idaho: U.S. Geological Survey Professional Paper 1110–AA, p. AA1–AA43.

Southworth, C.S., 1988, Preliminary lineament map of the Hailey, Idaho, and west half of the Idaho Falls, Idaho, 1°×2° quadrangles, with inset of regional image mosaic and geologic structures: U.S. Geological Survey Open-File Report 88–640, scale 1:250,000.

Stewart, D.E., 1987, Geology of the northern half of the Baker Peak quadrangle, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 113 p.

Stewart, D.E., Mahoney, J.B., Geslin, J.G., Darling, R.S., and Link, P.K., 1992, Geologic map of the Baker Peak quadrangle, Blaine and Camas Counties, Idaho: Idaho Geological Survey Technical Report 92–1, scale 1:24,000.

Stewart, D.E., Schmidt, K.L., Moye, F.J., Mahoney, J.B., and Link, P.K., in press, Eocene intrusive complexes in the Smoky and Boulder Mountains, south-central Idaho: U.S. Geological Survey Bulletin 2064–HH.

Trexler, J.H., Jr., Snyder, W.S., Cashman, P.H., Gallegos, D.M., and Spinosa, C., 1991, Mississippian through Permian orogenesis in eastern Nevada: Post-Antler, pre-Sonoma tectonics of the western Cordillera, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 1. p. 317–330, v. 2, p. 551–579.

Tschanz, C.M., Kiilsgaard, T.H., and Seeland, D.A., 1986, Geology of the eastern part of the Sawtooth National Recreation Area, Idaho, *in* Mineral resources of the eastern part of the Sawtooth National Recreation Area, Custer and Blaine Counties, Idaho: U.S. Geological Survey Bulletin 1545–A, p. 17–43.

Turner, R.J.W., and Otto, B.R., 1988, Stratigraphy and structure of the Milligen Formation, Sun Valley area, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 153–167.

Umpleby, J.B., Westgate, L.G., and Ross, C.P., 1930, Geology and ore deposits of the Wood River region, Idaho: U.S. Geological Survey Bulletin 814, 250 p.

Whitman, S.K., 1990, Metamorphic petrology and structural geology of the Pennsylvanian-Permian Dollarhide Formation, Blaine and Camas Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 108 p.

Wilson, Eric, 1992, Mesozoic compression and Cenozoic extension along the Glide Mountain fault, south-central Idaho: Geological Society of America Abstracts with Programs, v. 24, no. 6, p. 68.

———1994, Demise of the Glide Mountain thrust fault, Pioneer Mountains, Blaine and Custer Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 188 p.

Wilson, Eric, and Rodgers, D.W., 1993, Demise of the Glide Mountain thrust, Pioneer Mountains, south-central Idaho: Geological Society of America, Abstracts with Programs, v. 25, no. 4, p. 164.

Winsor, H.C., 1981, The paleogeography and paleoenvironments of the Middle Pennsylvanian part of the Wood River Formation in south-central Idaho: San Jose, Calif., San Jose State University, M.S. thesis, 81 p.

Worl, R.G., Kiilsgaard, T.H., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Wust, S.L., 1986, Extensional deformation with northwest vergence, Pioneer core complex, central Idaho: Geology, v. 14, p. 712–714.

Zentner, N.C., 1989, Neogene normal faults related to the structural origin of the eastern Snake River Plain, Idaho: Pocatello, Idaho State University, M.S. thesis, 48 p.

Stratigraphic Setting of Sediment-Hosted Mineral Deposits in the Eastern Part of the Hailey 1°×2° Quadrangle and Part of the Southern Part of the Challis 1°×2° Quadrangle, South-Central Idaho

By Paul Karl Link, J. Brian Mahoney, Daniel J. Bruner, L. Darlene Batatian, Eric Wilson, *and* Felicie J.C. Williams

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–C

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

CONTENTS

PLATE

[Plate is in pocket]

FIGURES

TABLES

Stratigraphic Setting of Sediment-Hosted Mineral Deposits in the Eastern Part of the Hailey 1°×2° Quadrangle and Part of the Southern Part of the Challis 1°×2° Quadrangle, South-Central Idaho

By Paul Karl Link,[1] J. Brian Mahoney,[2] Daniel J. Bruner,[3] L. Darlene Batatian,[3] Eric Wilson,[3] *and* Felicie J.C. Williams[4]

ABSTRACT

The central Idaho black-shale mineral belt includes most of the outcrop area of Paleozoic strata in the eastern part of the Hailey 1°×2° quadrangle and part of the southern part of the Challis 1°×2° quadrangle. Syngenetic deposits of silver, lead, and zinc are present in the Devonian Milligen Formation, unnamed Silurian and Devonian strata, and the Devonian part of the Salmon River assemblage. Remobilization of these metals during Cretaceous and Eocene deformation and magmatism produced a variety of epigenetic mineral deposits including shear-zone-hosted veins, replacement deposits, and skarns. The most important time of mineralization probably was the Late Cretaceous (about 90–80 Ma), when folding and thrust faulting within the Sevier orogenic belt temporally and spatially overlapped intrusion of the Atlanta lobe of the Idaho batholith and satellite plutons to the east. Structures that host silver-lead-zinc veins include sheared unconformities, northwest- and northeast-striking Cretaceous high-angle faults, northwest-striking low-angle oblique-slip Paleogene faults, and northeast-striking high-angle normal Eocene faults associated with the Trans-Challis fault system. Replacement deposits are present in calcareous rocks that are adjacent to mineralized structures and along silicification fronts produced during fluid migration. Skarn deposits are hosted by calcareous rocks adjacent to plutons.

Mapping of informal members within the Devonian Milligen Formation (lower argillite, quartzite of Cait, argillite of Triumph, limestone of Lucky Coin, and sandstone of Independence) suggests that the formation is present in the Minnie Moore, Bellevue, Triumph, Lake Creek stock, and Summit mineralized areas where the Milligen Formation is interpreted to pass by facies change into unnamed Silurian and Devonian strata. Because the same stratigraphic units are present from west to east within the Mesozoic Pioneer thrust plate, it is doubtful that the plate contains an older (Antler) thrust fault having significant (tens of kilometers) displacement.

The Mississippian Copper Basin Formation contains west-derived siliciclastic turbidite and east-derived calciclastic turbidite strata. The formation is present east of the Pioneer thrust fault and also in apparent stratigraphic continuity with rocks of the Pioneer thrust plate. Calciclastic silty turbidites of the medial Drummond Mine Limestone Member are host to skarn and vein deposits in the Phi Kappa mine area south of Summit Creek.

The Pennsylvanian and Permian Sun Valley Group (Wood River, Dollarhide, and Grand Prize Formations) includes strata previously thought to be present in different thrust plates. The contacts between these formations are generally facies changes rather than structural contacts. The basal Hailey Member of the Wood River Formation hosts vein deposits along its sheared unconformity with the underlying Devonian Milligen Formation. This relationship was previously mapped as the "Wood River thrust." The medial Eagle Creek and the upper Wilson Creek Members of the Wood River Formation host vein, replacement, and skarn deposits adjacent to the Rooks Creek stock and in several areas east of the Wood River valley. Carbonaceous micritic siltstone of the lower and upper members of the Dollarhide Formation contains shear-zone-hosted veins in the Minnie Moore, Bullion, Deer Creek stock, Bunker Hill, and Smoky Mountains mineralized areas. In the Deer Creek

[1]U.S. Geological Survey and Department of Geology, Idaho State University, Pocatello, Idaho 83209.

[2]U.S. Geological Survey and University of British Columbia, Vancouver, British Columbia V6T 1Z4.

[3]Department of Geology, Idaho State University, Pocatello, Idaho 83209.

[4]Westmont Mining Inc., N. 10220 Nevada Street, Suite 140, Spokane, Washington 99218.

stock area, the lower member of the Dollarhide Formation contains a laminated barite deposit. In the Carrietown area, adjacent to the Atlanta lobe of the Idaho batholith, mineralized veins are near the contact between foliated banded quartzite and carbonaceous marble within the lower member of the Dollarhide Formation. The four members of the Grand Prize Formation are host to skarn, replacement, and vein deposits in the Washington Basin, Smiley Creek, and Galena mineralized areas in the White Cloud Peaks, Smoky Mountains, and northern Boulder Mountains.

INTRODUCTION

This report provides a synthesis of the stratigraphic setting of sediment-hosted mineral deposits in the eastern half of the Hailey 1°×2° quadrangle and part of the southern part of the Challis 1°×2° quadrangle (fig. 1, plate 1) and a summary of the present understanding of the Paleozoic stratigraphy of the area. The sediment-hosted mineral deposits are in the "central Idaho black-shale mineral belt" of Hall (1985) and have produced primarily silver, lead, and zinc and small amounts of gold. The deposits are included within mineralized areas defined by Worl and Johnson (this volume) (fig. 1).

Research summarized in this report has resolved some long-standing problems of stratigraphy but only scratched the surface of others. In particular, we understand the Pennsylvanian and Permian part of the stratigraphic section far better than we understand the lower Paleozoic units, especially the Silurian, Devonian, and Mississippian units. We therefore present this summary as a progress report.

Acknowledgments.—Work on the Paleozoic rocks of south-central Idaho was supported by the Hailey project of the U.S. Geological Survey Conterminous United States Mineral Assessment Program (CUSMAP), Idaho State Board of Education Grant 89–56 to P.K. Link, and the Idaho Initiative project of the Idaho Geological Survey. This manuscript benefitted from review by S.J. Soulliere, Betty Skipp, David Seeland, and B.R. Burton.

PREVIOUS WORK

The mineral deposits of the Wood River area were first documented by Lindgren (1900) and have been more recently described by W.E. Hall and colleagues of the U.S. Geological Survey (Hall and Czamanske, 1972; Hall and others, 1978; Hall, 1985; Howe and Hall, 1985; Hall, 1987a, b; Hall and Hobbs, 1987). Hall concluded that many of the deposits were formed by hydrothermal circulation systems developed during both Cretaceous and Tertiary magmatism. These hydrothermal systems were thought to have derived metals from a country rock source (the lower Paleozoic black shales of the Milligen Formation and Salmon River assemblage). In the Wood River area, ore mineral deposition was thought to be localized near intrusive bodies and below regional thrust faults that acted as a permeability barrier to mineralizing solutions. Hall (1985) also suggested that the Paleozoic black shales hosted syngenetic stratabound mineral deposits, particularly in the Paleozoic Salmon River assemblage at the Hoodoo mine (plate 1), the Devonian Milligen Formation in the Triumph mineralized area, and the Middle Pennsylvanian to Lower Permian Dollarhide Formation at the Deer Creek barite deposit, 12 km west of Hailey.

Recent work has modified some facets of these models. Syngenetic stratabound mineral deposits are believed to be present mainly in lower Paleozoic, primarily Devonian, strata but not in upper Paleozoic rocks. Lead isotope data suggest that the source of the metals in polymetallic vein deposits is remobilized Devonian syngenetic sulfide deposits and Precambrian continental crust (Sanford and Wooden, this volume). The concept of ore concentration below regional thrust faults is rejected. Many contacts previously mapped as thrust faults are now mapped as stratigraphic contacts, sheared unconformities, or low-angle oblique-slip faults (Skipp and others, 1986; Burton and others, 1989; Mahoney and others, 1991; Burton and Link, this volume; Rodgers and others, this volume).

This report is a synthesis of the stratigraphic setting of mineral deposits in the eastern part of the Hailey 1°×2° quadrangle. For studies of individual mines and mineral districts that include maps, the reader is referred to Lindgren (1900), Umpleby (1915), Umpleby and others (1930), Anderson and others (1950), Tuchek and Ridenour (1981), and Van Noy and others (1986) and to reports in this volume and the second volume of this bulletin (in press).

STRATIGRAPHY

This report contains a summary of Paleozoic stratigraphy of the eastern part of the Hailey and part of the southern part of the Challis 1°×2° quadrangles and emphasizes new developments. We discuss only the Paleozoic rocks structurally above the Wildhorse detachment fault, which forms the upper boundary of metamorphic and intrusive rocks of the Pioneer Mountains core complex (fig. 1). Proterozoic and Paleozoic metasedimentary strata of the core are discussed by Dover (1969, 1981, 1983). Our work is grounded in, but differs in detail from, the reports of Hall and others (1974), Skipp and Hall (1975, 1980), Sandberg and others (1975), Skipp, Sando, and Hall, (1979), Dover (1981, 1983), Hall (1985), Hall and Hobbs (1987), Mahoney and Sengebush (1988), and Link and others (1988).

Allochthonous Ordovician through Lower Permian strata in the eastern part of the Hailey 1°×2° quadrangle are generally of three lithologic types: dark-colored carbonaceous argillite, light-colored calcareous sandstone, siltstone, and limestone, and pebble and cobble conglomerate. Because the rocks are mostly unfossiliferous and lithologically similar, formations of different ages have been difficult to differentiate with confidence. Age assignments have been, and remain, contentious.

Although we have made significant progress in understanding the Pennsylvanian and Permian parts of the section, the Devonian rocks are less well understood. They are structurally complex and have only recently been subdivided stratigraphically (Turner and Otto, 1988, this volume; Bruner, 1991). Biostratigraphic studies in the Milligen Formation and correlative lower Paleozoic units are especially needed.

ORDOVICIAN, SILURIAN, AND DEVONIAN SHELF-FACIES STRATA

Calcareous and quartzitic shelf-facies strata of early Paleozoic age are exposed in two structural inliers in Dry Canyon and Wildhorse Creek in the northeastern part of the Hailey 1°×2° quadrangle (Dover, 1981, 1983) (figs. 1, 2, plate 1). Kim (1986) mapped the upper boundaries of these inliers as west-dipping low-angle normal faults. Dover (1981, 1983) showed these contacts as thrust faults. Wilson (1994) interpreted these contacts as unconformable.

Formations mapped within the Wildhorse inlier include the Middle and Upper Ordovician Hanson Creek Formation (cherty dolomite, more than 64 m thick), an unnamed Middle Silurian limestone unit (40–120 m thick) containing a prominent bed of black dolomite and black chert, the Silurian Roberts Mountains Formation (calcareous to dolomitic siltstone, about 600 m thick), and an unnamed carbonate and conglomerate and breccia unit of Devonian age (more than 10 m thick). The lower contact of this sequence is not exposed. The upper contact is a fault or an unconformable contact with the Lower Mississippian lower part of the Copper Basin Formation.

The Dry Canyon inlier contains a partial section of the Lower and Middle Devonian Carey Dolomite (thickness unknown), the Upper Devonian Jefferson Formation (cliff-forming, medium-bedded to massive, finely crystalline dolomitic limestone, dolomite, and bioclastic limestone and local intraformational conglomerate, about 350 m thick) and the Upper Devonian Picabo Formation (sandstone, at least 6 m thick). The lower contact of this sequence is not exposed. The upper contact is an unconformity with the Lower Mississippian part of the Copper Basin Formation. Near Fish Creek Reservoir in the Idaho Falls 1°×2° quadrangle, to the east of the study area, the Picabo Formation lies unconformably below the Lower Mississippian part of the Copper Basin Formation (Link and others, 1988).

No mineral deposits are hosted by the carbonate rocks of these structural inliers.

ORDOVICIAN AND SILURIAN PHI KAPPA AND SILURIAN TRAIL CREEK FORMATIONS

The Lower Ordovician to Middle Silurian Phi Kappa and Middle Silurian Trail Creek Formations are imbricated in a series of thrust-bounded slices in the hanging wall of the Pioneer thrust fault in the headwaters of Trail Creek, east of Ketchum (Dover and others, 1980) (figs. 1, 2, plate 1). The Pioneer thrust fault places lower Paleozoic rocks above the Mississippian Copper Basin Formation.

The Phi Kappa and Trail Creek Formations were defined by Umpleby and others (1930). Their structural repetition was demonstrated by Churkin (1963) and Carter and Churkin (1977). The units were revised by Dover and others (1980). The Phi Kappa Formation is 240 m thick and consists of the basal massive, gray, fine-grained Basin Gulch Quartzite Member (55 m thick) overlain by 165 m of dark-gray, red-weathering, carbonaceous, locally silicified shale and argillite of the "main body" of the formation (Dover, 1983). Thin limestone beds are present in the middle of the formation. A rich graptolite fauna reveals that the Phi Kappa Formation contains Lower to Upper Ordovician strata gradationally overlain by 17 m of Lower to Middle Silurian strata of the upper member of the formation (Dover and others, 1980; Dover, 1981, 1983) (table 1).

The Trail Creek Formation (revised by Dover and others, 1980) is at least 100 m thick and contains buff-weathering, banded tan and white siliceous metasiltstone and very fine grained quartzite. The base of the Trail Creek Formation is gradational over a few meters with the underlying argillaceous Middle Silurian beds of the upper part of the Phi Kappa Formation, but biostratigraphic control is lacking in the Trail Creek Formation (Dover and others, 1980; Dover, 1983). As mapped by Dover (1983), the top of the Trail Creek Formation is everywhere a thrust fault.

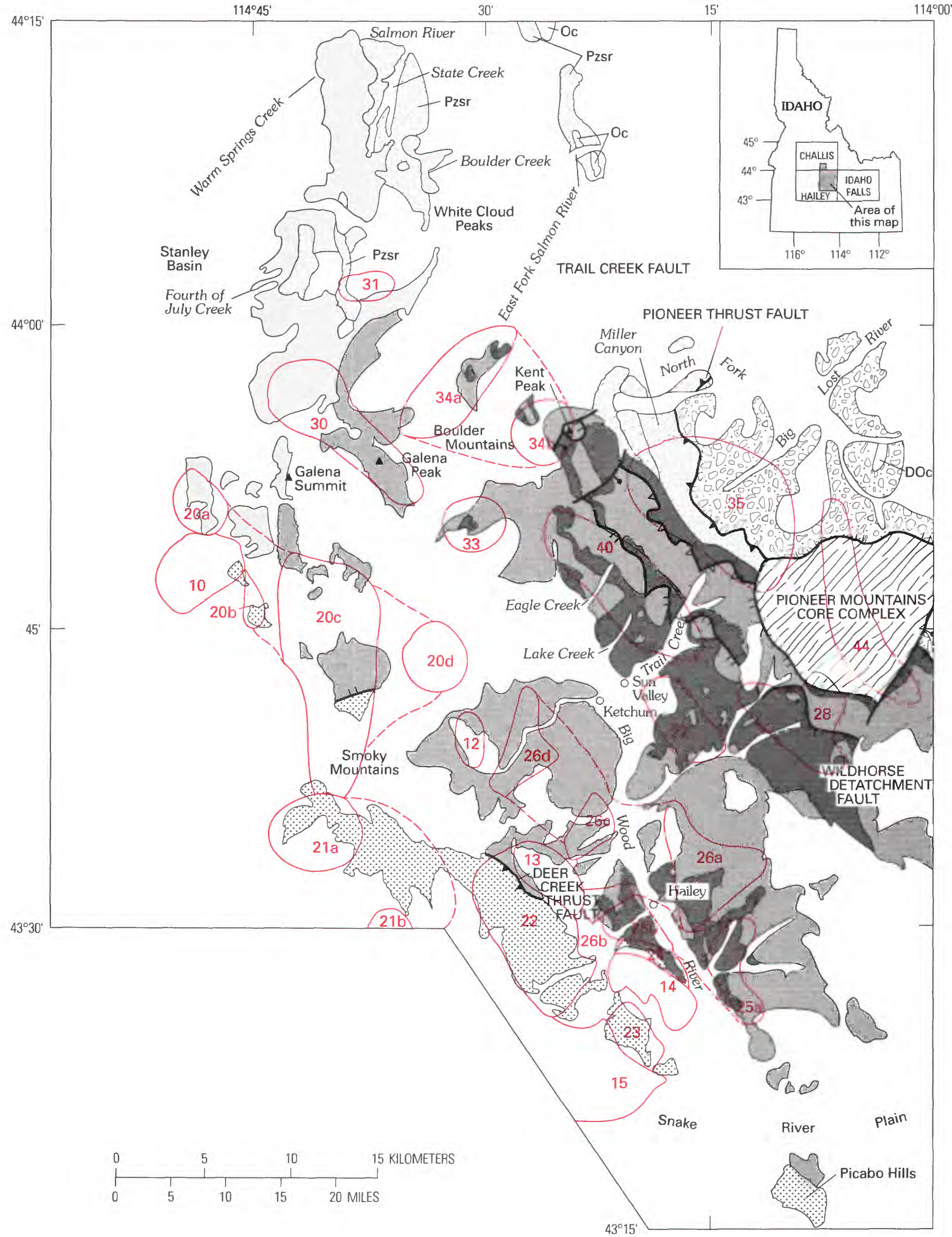
The map of Dover (1983) shows the contacts between the Phi Kappa and Trail Creek Formations and unnamed Silurian and Devonian units to be thrust faults. Our work suggests that some of these contacts may be sheared stratigraphic contacts, implying stratigraphic continuity between these units (discussed following).

No significant mineral deposits are hosted by the Phi Kappa or Trail Creek Formations. The Basin Gulch Member of the Phi Kappa Formation is quarried for building stone in the headwaters of Summit Creek.

SILURIAN AND DEVONIAN ARGILLACEOUS STRATA

DEVONIAN MILLIGEN FORMATION

The Lower to Upper Devonian Milligen Formation is the host for rich silver-lead-zinc ores in the historically productive Minnie Moore and Triumph mineralized areas near Hailey and Bellevue (fig. 1). The Milligen Formation was named by Umpleby and others (1930) for carbonaceous and argillaceous rocks originally included in the lower part of the Wood River Formation by Lindgren (1900). As described by Sandberg and others (1975), Dover (1981), and Turner and Otto (1988, this volume), the Devonian Milligen Formation


114°45'
30'
15'
114°00'
44°15'
44°00'
45'
43°30'
43°15'
Salmon River
State Creek
Pzsr
Oc
Warm Springs Creek
Boulder Creek
White Cloud Peaks
East Fork Salmon River
Stanley Basin
Fourth of July Creek
TRAIL CREEK FAULT
PIONEER THRUST FAULT
Miller Canyon
North Fork
Lost River
Big
Kent Peak
Boulder Mountains
Galena Peak
Galena Summit
DOc
Eagle Creek
Lake Creek
Trail Creek
Sun Valley
Ketchum
Big Wood River
PIONEER MOUNTAINS CORE COMPLEX
WILDHORSE DETATCHMENT FAULT
Smoky Mountains
DEER CREEK THRUST FAULT
Hailey
Snake River Plain
Picabo Hills
IDAHO
CHALLIS
HAILEY
IDAHO FALLS
Area of this map
45°
44°
43°
116°
114°
112°
0 5 10 15 KILOMETERS
0 5 10 15 20 MILES
10
12
13
14
15
20a
20b
20c
20d
21a
21b
22
23
25
26a
26b
26c
26d
28
30
31
33
34a
34b
35
40
44

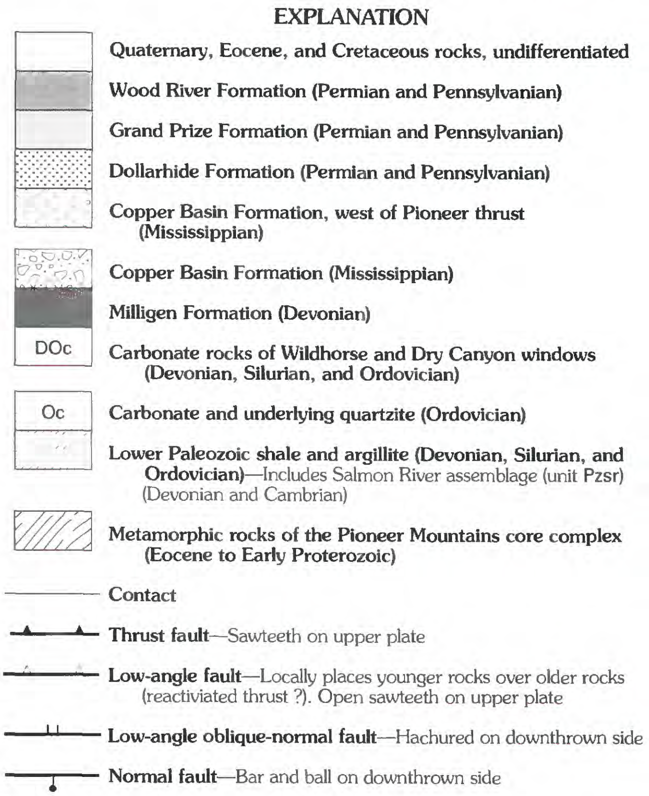


Figure 1 (above and facing page). Generalized geologic map of the eastern part of the Hailey 1°×2° quadrangle and part of the southern part of the Challis 1°×2° quadrangle, south-central Idaho (modified from Mahoney and others, 1991, and Batatian, 1991). Mineralized areas (Worl and Johnson, this volume), referred to by number, are shown in red: 10, Marshall Peak; 12, Rooks Creek stock; 13, Deer Creek stock; 14, Croesus stock; 15, Hailey gold belt; 20, Smoky Mountains—(a) Smiley Creek, (b) West Fork Big Smoky Creek, (c) Norton-Baker Peaks, (d) Baker Creek; 21, Carrietown—(a) Carrietown, (b) Buttercup mine; 22, Bullion; 23, Bunker Hill; 24, Minnie Moore; 25, Bellevue—(a) Slaughterhouse Gulch, (b) Vorberg Gulch; 26, Wood River—(a) Quigley Gulch, (b) Carbonate Mountain, (c) Greenhorn Gulch, (d) Warm Springs Creek; 27, Triumph; 28, East Fork Wood River; 30, Galena; 31, Washington Basin; 33, Boulder Basin; 34, East Fork Salmon—(a) East Fork Salmon River, (b) Ryan Peak; 35, Summit; 40, Lake Creek stock; 44, Pioneer.

contains more than 1,200 m of unfossiliferous, fine-grained, locally carbonaceous strata that are poorly exposed on sagebrush- and grass-covered slopes in the Boulder and Pioneer Mountains east of the Wood River. The base of the Milligen Formation has not been observed in detailed mapping of the Triumph mine area or on the western side of the Boulder Mountains (Dover, 1983; Turner and Otto, this volume). The upper contact varies along strike and is locally an unconformity below the Middle Pennsylvanian Hailey Member of the Wood River Formation, a possibly conformable contact with strata mapped as Mississippian Copper Basin Formation, or a fault against Pennsylvanian and Permian strata of the Sun Valley Group.

Though Umpleby and others (1930) thought that the bulk of the Milligen Formation was Mississippian in age, Sandberg and others (1975) restricted the age of the formation in the type area of Milligen Gulch southeast of Ketchum (plate 1) to Devonian, based on the presence of sparse Eifelian, Frasnian, and Famennian conodonts (fig. 3). Dark-colored Lower Mississippian rocks in the Idaho Falls 1°×2° quadrangle to the east, originally included in the Milligen Formation by Umpleby and others (1930) and many subsequent workers, were renamed the McGowan Creek Formation by Sandberg (1975). Dover (1969, p. 29; 1981, p. 34) and Sandberg (1975) cautioned against using the name Milligen Formation beyond the type area originally defined in the Boulder and Pioneer Mountains east of the Wood River.

A stratigraphic division of the Milligen Formation in the area near the Triumph mine (Turner and Otto, 1988, this volume) (fig. 3) includes, as informal members, (1) a lower argillite member (130+ m thick), which locally contains diamictite or is chert- or sandstone-rich, (2) the limestone member of Lucky Coin (50–250 m thick), which contains limestone turbidite, black argillite, and diamictite, (3) the quartzite member of Cait (lenses of variable thickness, usually less than 10 m), which contains carbonaceous coarse-grained sandstone, diamictite, and black argillite, (4) the argillite member of Triumph (0–150 m thick), which contains black argillite, locally cherty, and interbedded sandstone, and (5) the sandstone member of Independence (150+ m thick), which contains interbedded sandstone and limestone turbidite. Diamictite in the lower argillite member, coarse-grained sandstone in the quartzite of Cait, and calciclastic turbidite of the limestone of Lucky Coin represent distinctive markers that allow correlation of the Milligen Formation across structurally complex areas.

In the areas of the Triumph and Minnie Moore mines, the Milligen Formation is present in northeast-vergent overturned folds that have southwest-dipping limbs. Inferred syngenetic stratabound mineral deposits in these areas are in distinctive stratigraphic units in the middle part of the formation (limestone member of Lucky Coin and quartzite member of Cait).

Our recent work suggests several changes in understanding of the Milligen Formation. These revisions are grounded in geologic mapping and stratigraphic studies (Batatian, 1991; Bruner, 1991; Worl and others, 1991). Several of the revisions await confirmation from further biostratigraphic studies. These revisions include the following:

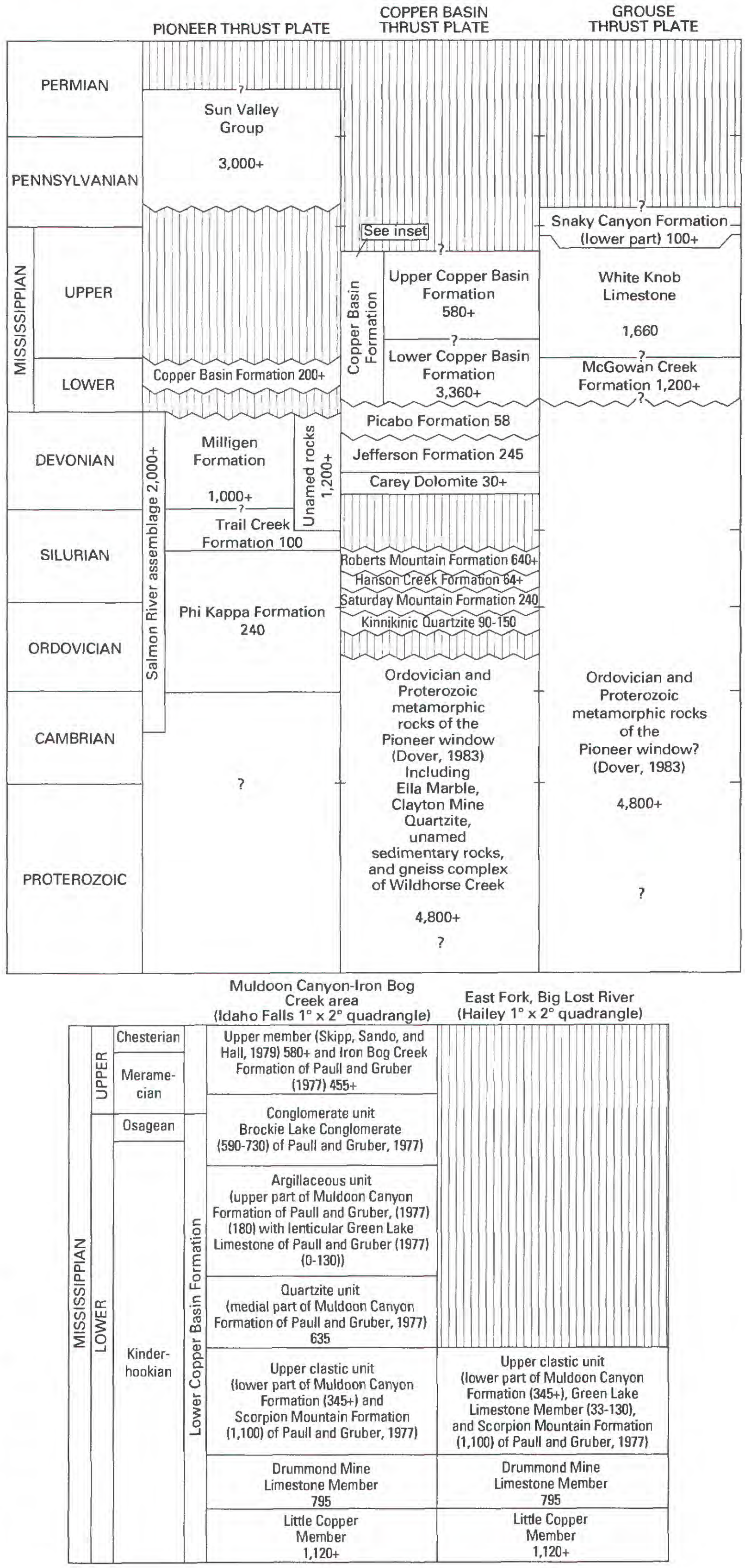


Figure 2. Correlation chart for Paleozoic rocks of the eastern part of the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho. Modified from Skipp, Sando, and Hall (1979), Dover (1983), Link and others (1988), and Batatian (1991). All thicknesses are in meters. Inset shows detailed Lower Mississippian stratigraphic correlation as proposed by Wilson (1994) and modified from Paull and others (1972), Paull and Gruber (1976), Nilsen (1977), Dover (1981), Skipp (1989), and Skipp and others (1990).

Table 1. Graptolites from Phi Kappa Formation, south-central Idaho.
[Collected by D.J. Bruner and colleagues, identified by Claire Carter, U.S. Geological Survey, 1990 and 1991]

82PL89A	Probably Early or Middle Ordovician but could be Silurian. Right fork of Trail Creek, elevation 8,200 ft, west of creek, unsurveyed, T. 7 N., R. 18 E., Rock Roll Canyon quadrangle, Blaine County, Idaho.
82PL89B	*Orthograptus* sp; probably Middle or Late Ordovician. West side of Squib Canyon, elevation 8,700 ft, northwest of peak 10,284, unsurveyed T. 7 N., R. 18 E., Meridian Peak quadrangle, Custer County, Idaho.
82PL89C	Diplograptids and *Dicranograptus* sp.; Middle Ordovician. In right fork of right fork of Trail Creek, elevation 9,000 ft, unsurveyed T. 7 N. R. 18 E., Meridian Peak quadrangle, Blaine County, Idaho.
DJB–G2–90	*Climacograptus spiniferus* Ruedemann, *Orthograptus*?; Middle Ordovician zone of *C. spiniferus*. West side of Miller Canyon, northeast of 9413, elevation 8,400 ft, T. 7 N., R. 18 E., Meridian Peak quadrangle, Custer County, Idaho.
DJB–G3–90	*Climacograptus caudatus* Lapworth, *Amplexograptus*?, *Orthograptus*?; Middle Ordovician zone of *C. spiniferus*. East fork of Squib Canyon, elevation 8,600 ft, T. 7 N., R. 18 E.; Meridian Peak quadrangle, Custer County, Idaho.
DJB–G4–90	*Climacograptus bicornis* (Hall), *Glossograptus ciliatus* Emmons, unidentifiable diplograptid; Middle Ordovician zone of *C. bicornis*. Head of Miller Canyon, elevation 9,800 ft, northwest of peak 10,356, sec. 35, T. 7 N., R. 18 E., Meridian Peak quadrangle, Custer County, Idaho.
DJB–25A–90	*Climacograptus* cf. *C. tubuliferus* Lapworth, *C.* sp. (small), *C.* cf *C. spiniferus* Ruedemann, *Orthograptus*?, *Glyptograptus*?, *Dicellograptus* sp.; Middle Ordovician zone of *C. spiniferus*. Upper Bear Creek, elevation 8,600 ft, sec. 34, T. 7 N., R. 18 E., Meridian Peak quadrangle, Custer County, Idaho.

1. The contact of the Milligen Formation with the overlying Pennsylvanian and Permian Wood River Formation is not a regional thrust fault but rather is a locally sheared unconformity or a low-angle normal fault (Burton and others, 1989; Batatian, 1991; Mahoney and others, 1991; Burton and Link, this volume; Rodgers and others, this volume).

2. Pending biostratigraphic verification, the informal members of the Milligen Formation from the type area can be recognized east of Kent Peak and in the Miller Canyon area on the east slope of the Boulder Mountains near the head of the North Fork of the Big Lost River (Batatian, 1991; Bruner, 1991; B. Otto, written commun., 1991).

3. The base of the Milligen Formation in the eastern Boulder Mountains may be gradational with underlying calcareous siltstone that contains Silurian (Ludlovian), as well as Early Devonian (Lochkovian and Pragian) conodonts (Brennan, 1987; Batatian, 1991) (fig. 3).

4. In the Boulder Mountains, Devonian argillite, apparently stratigraphically above the upper part of the Milligen Formation exposed in the Triumph mine area, appears to be locally conformably overlain by argillite and lithic wacke mapped as the upper part of the lower Copper Basin Formation (Lower Mississippian) (Bruner, 1991; Huerta and others, 1991) (fig. 2).

UNNAMED SILURIAN AND DEVONIAN ROCKS

Unnamed Silurian and Devonian rocks including black, siliceous to carbonaceous argillite (map unit DSa of Dover, 1983), tan, calcareous, dolomitic, or siliceous siltstone (map unit DSs of Dover, 1983), and bioclastic limestone (map unit DSl of Dover, 1983) are tectonically imbricated with the Phi Kappa and Trail Creek Formations in steep grassy slopes at the head of Trail Creek and in the North Fork of the Big Lost River northeast of Ketchum (fig. 1, plate 1). The rocks were originally included in the Phi Kappa and Trail Creek Formations by Umpleby and others (1930) but were redefined based on biostratigraphy by Dover (1981, 1983). Structural complexity and poor exposure prevented Dover (1981, 1983) and previous workers from dividing this sequence into formations. Limestone (mapped as unit DSl by Dover, 1983) in the unnamed Silurian and Devonian rocks contains Middle Devonian (Eifelian) conodonts (Dover and others, 1980; Dover, 1981, 1983; Brennan, 1987) (fig. 3). Limestone beds in a siltstone unit (unit DSs of Dover, 1983) exposed in the headwaters of the North Fork of the Big Lost River contain Upper Silurian (Ludlovian) and Lower Devonian (Lochkovian and Pragian) conodonts (Dover, 1983; Brennan, 1987) (fig. 3).

We show the contact between undifferentiated Ordovician, Silurian, and Devonian rocks and the Milligen

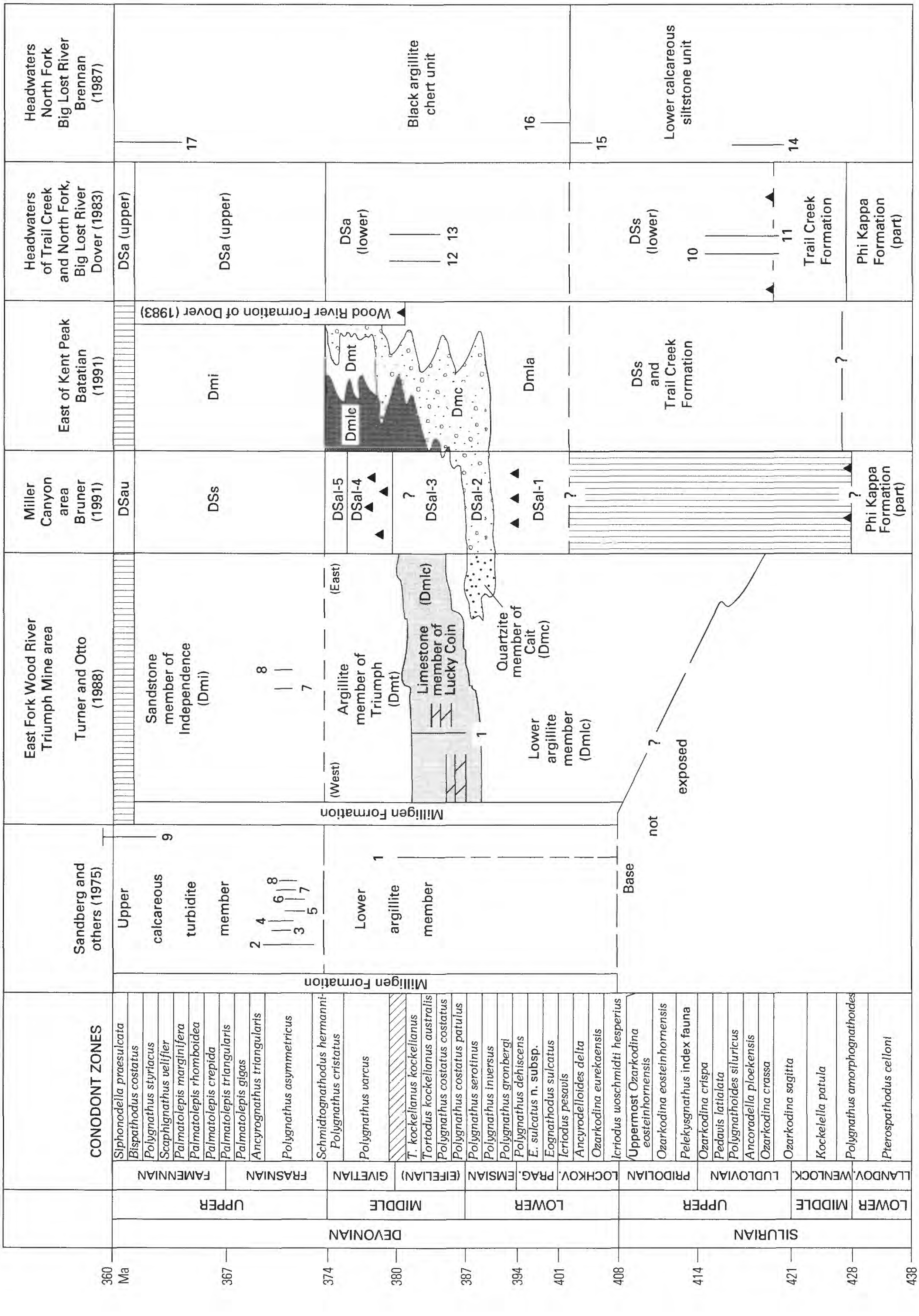

CONODONT ZONES
Sandberg and others (1975)
East Fork Wood River Triumph Mine area
Turner and Otto (1988)
Miller Canyon area Bruner (1991)
East of Kent Peak Batatian (1991)
Headwaters of Trail Creek and North Fork, Big Lost River Dover (1983)
Headwaters North Fork Big Lost River Brennan (1987)
DEVONIAN
SILURIAN
UPPER
MIDDLE
LOWER
FAMENNIAN
FRASNIAN
GIVETIAN
(EIFELIAN)
EMSIAN
PRAG.
LOCHKOV.
PRIDOLIAN
LUDLOVIAN
WENLOCK.
LLANDOV.
Siphonodella praesulcata
Bispathodus costatus
Polygnathus styriacus
Scaphignathus velifier
Palmatolepis marginifera
Palmatolepis rhomboidea
Palmatolepis crepida
Palmatolepis triangularis
Palmatolepis gigas
Ancyrognathus triangularis
Polygnathus asymmetricus
Schmidtognathodus hermanni-Polygnathus cristatus
Polygnathus varcus
T. kockelianus kockelianus
Tortodus kockelianus australis
Polygnathus costatus costatus
Polygnathus costatus patulus
Polygnathus serotinus
Polygnathus inversus
Polygnathus gronbergi
Polygnathus dehiscens
E. sulcatus n. subsp.
Eognathodus sulcatus
Icriodus pesavis
Ancyrodelloides delta
Ozarkodina eurekaensis
Icriodus woschmidti hesperius
Uppermost Ozarkodina eosteinhornensis
Ozarkodina eosteinhornensis
Pelekysgnathus index fauna
Ozarkodina crispa
Pedavis latialata
Polygnathoides siluricus
Ancoradella ploekensis
Ozarkodina crassa
Ozarkodina sagitta
Kockelella patula
Polygnathus amorphognathoides
Pterospathodus celloni
Milligen Formation
Upper calcareous turbidite member
Lower argillite member
Base
not exposed
Sandstone member of Independence (Dmi)
Argillite member of Triumph (Dmt)
Limestone member of Lucky Coin (Dmlc)
Quartzite member of Cait (Dmc)
Lower argillite member (Dmlc)
(West)
(East)
DSau
DSs
DSal-5
DSal-4
DSal-3
DSal-2
DSal-1
Phi Kappa Formation (part)
Wood River Formation of Dover (1983)
Dmi
Dmt
Dmlc
Dmc
Dmla
DSs and Trail Creek Formation
DSa (upper)
DSa (lower)
DSs (lower)
Trail Creek Formation
Black argillite chert unit
Lower calcareous siltstone unit
360 Ma
367
374
380
387
394
401
408
414
421
428
438

Figure 3 (facing page). Correlation chart for Silurian through Devonian rocks in the eastern Boulder Mountains, south-central Idaho. Modified from Batatian (1991) and Bruner (1991). Solid triangles indicate chert, and open irregular rectangles indicate intrabasinal clasts. Conodont zonation for Devonian System is as presented in Brennan (1987) based on sections in the Roberts Mountains, Nevada, and at Cellon, Carnic Alps, Austria (after Walliser, 1964; Klapper and Murphy, 1974; and Sandberg, 1979). Complete data for collections 1 through 9 are given in Sandberg and others (1975): 1, Sawmill Gulch, SAW–1; 2, Trail Creek, USGS 8928–SD; 3, Slaughterhouse Creek, AXE–510; 4, Slaughterhouse Creek, BLV–3; 5, Fisher Canyon, FIC–1; 6, Fisher Canyon, FIC–2; 7, Milligen Gulch (principal reference section), C–189; 8, Milligen Gulch (principal reference section), C–58; 9, North Fork Big Lost River, location uncertain, collected by R. Paull. 10, Bioclastic limestone in unit DSs of Dover (1983), no specific location given; Ludlovian conodonts (A.G. Harris, written commun., 1975, to J.H. Dover). 11, Silty limestone, possibly equivalent to number 10 above; unit DSs?, tributary North Fork Big Lost River (C. Sandberg, oral commun., *in* Dover, 1983). 12, Dover 74–5209 (USGS loc. 9496–SD), middle to late Eifelian, unit DSa of Dover (1983), bioclastic phosphatic limestone. In stream cut draining small cirque east of Kent Peak in headwaters of North Fork of Big Lost River. In intensely sheared dominantly argillite section immediately below Wood River thrust fault (J.H. Dover, written commun., 1990). P elements *Polygnathus linguiformis linguiformis*, P elements *P. costatus costatus* Klapper, I element *Icriodus* cf. *I. stelcki* Chatterton, I element I. *nodosus* (Huddle), I element *I. espansus* Branson and Mehl, *Pryantodus* elements, *Hindeodella* element, *Angulodus* elements, *Spathodnathodus* cf. *S. bipennatus* Bischoff and Ziegler; conodonts identified by A.G. Epstein (Harris), 1974. 13, Dover 74–5226 (USGS loc. 9497–SD), Eifelian, unit DSl of Dover (1983), bioclastic phosphatic limestone. East side of south fork of North Fork of Big Lost River in chiefly argillite section (J.H. Dover, written commun., 1990). P elements of *Polygnathus linguiformis linguiformis*, P. elements of *P. costatus costatus* Klapper, I elements of *Icriodus nodosus* (Huddle), I elements of *I. expansus* Branson and Mehl transitional to *I. stelcki* Chatterton, *Angulodus* elements, *Spathognathodus* element, *Ligonodina* elements, *Lonchodina* element, *Neoprioniodus* element. Identified by A.G. Epstein (Harris), 1974. 14, Samples NF–8, NF–12, NF–13, NF–27, NF–29 of Brennan (1987), Ludlovian, lower part of calcareous siltstone unit (lower DSs unit of Dover, 1983, correlated with Trail Creek Formation by Spoelhof, 1972). Meadow in east fork of headwaters of the North Fork of the Big Lost River. *Polygnathus siluricus*, *P. latialata*, *Anacoradella ploekensis*. Identified by B.D. Brennan. 15, Samples NF–10, NF–11 of Brennan (1987), Lochkovian, upper part of calcareous siltstone unit (upper part of lower unit DSs of Dover, 1983). *Ancyrodelloides delta*. Identified by B.D. Brennan. 16, Sample NF–14 identified by Brennan (1987), Pragian, lower part of cherty argillite unit with thin limestone beds (unit DSa of Dover, 1983; lower argillite member of fig. 6, this paper (Bruner, 1991), equivalent to lower argillite member of Milligen Formation of Turner and Otto (this volume). *Eognathodus sulcatus*. Identified by B.D. Brennan. 17, Sample NF–37 of Brennan (1987), Famennian. East side, North Fork Big Lost River, limestone (unit DSl of Dover, 1983). *Polygnathus styriacus*(?). Identified by B.D. Brennan.

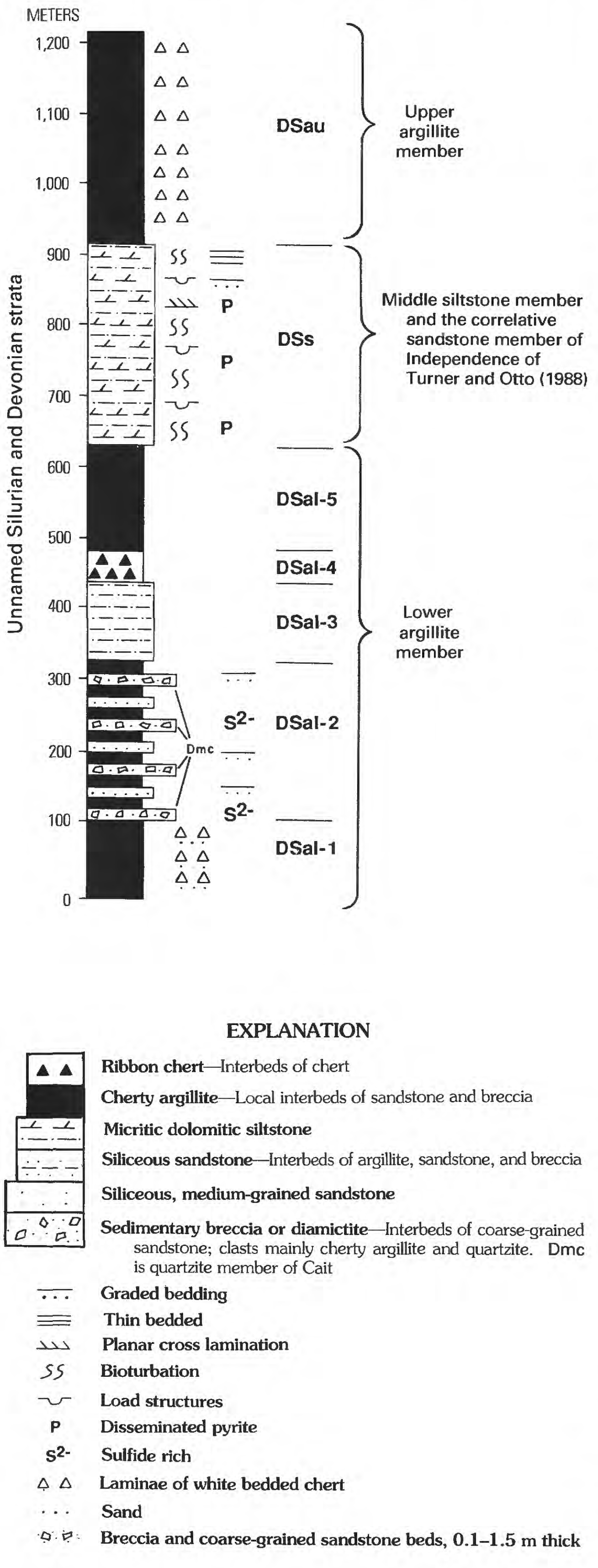
Formation as a gradational line of facies change in the area east of the Trail Creek fault from east of Kent Peak south to the Pioneer Mountains core complex (plate 1). In discussion of the unnamed Silurian and Devonian rocks, Dover (1981, p. 35) suggested that "at least some of these rocks are equivalent in lithology, conodont fauna, and age with the Milligen Formation [although] other parts of the sequence seem to have closer affinity in lithology and age to the Silurian Trail Creek Formation than to the Milligen." Furthermore, he suggested that "structural continuity in the subsurface [between the unnamed Silurian and Devonian rocks and the main Milligen outcrop belt] is probable."

MILLER CANYON AREA

Bruner (1991) described more than 1,200 m of unnamed Silurian and Devonian rocks from measured stratigraphic sections and drill core in the northeastern Boulder Mountains west of Miller Canyon (fig. 4, plate 1).

In the Miller Canyon area, graptolitic shale of the Lower Ordovician to Middle Silurian Phi Kappa Formation (table 1) forms the lowest exposed stratigraphic interval. The Phi Kappa Formation is exposed in imbricate fault slices (with apparent younger-over-older relations) below a folded thrust plate of tan dolomitic siltstone of the unnamed Silurian and Devonian rocks (unit DSs, fig. 4). The Middle Silurian Trail Creek Formation was not observed in the area and is apparently sheared out along the younger-over-older faults.

The unnamed Silurian and Devonian rocks structurally overlying the Phi Kappa Formation west of Miller Canyon are subdivided into three informal members: lower argillite, middle siltstone, and upper argillite (Bruner, 1991) (fig. 4). The lower argillite (unit DSal, at least 630 m thick) contains five subunits (DSal1–DSal5, figs. 3, 4). A basal contact of the lower argillite member has not been observed. The lowest exposed subunit (104+ m thick) is black cherty argillite containing laminae of white chert. The second subunit (224 m thick) contains black cherty argillite and interbeds of limestone, distinctive coarse-grained quartz sandstone, and pebble to boulder diamictite or sedimentary breccia (fig. 5). Diamictite beds are host to stratabound, inferred syngenetic, sulfide mineral deposits that are described later. The third subunit of the lower argillite unit (106 m thick) is siliceous siltstone, the fourth (45 m thick) is a black ribbon chert, and the fifth (151 m thick) is black cherty argillite. Gradationally overlying the lower argillite member is 234 m of orange-weathering bioturbated siltstone of the middle siltstone member (unit DSs, figs. 3, 4). The upper argillite member (unit Dsau, figs. 3, 4) conformably overlies the middle siltstone member and contains at least 300 m of phyllitic cherty argillite. The upper argillite member may be stratigraphically overlain by Mississippian Copper Basin Formation



conglomerate west of the East Fork of Trail Creek (plate 1) (Huerta and others, 1991).

Strata in the Miller Canyon area are correlative with strata in the headwaters of the North Fork of the Big Lost River (fig. 1), where conodonts of Silurian and Devonian ages have been recovered (Dover, 1983; Brennan, 1987) (fig. 3).

The unnamed strata exposed near Miller Canyon may be lithologically correlative with the Devonian Milligen Formation as shown in figure 3 (Bruner, 1991). The lower argillite in Miller Canyon is interpreted to correlate with the lower argillite member and the argillite member of Triumph of the Milligen Formation because subunit 2 of the lower argillite in Miller Canyon (unit DSal2, fig. 4) contains distinctive diamictite and coarse-grained sandstone beds characteristic of the quartzite member of Cait. Thin limestone beds recognized in drill core from the lower argillite member may represent the limestone member of Lucky Coin.

The middle siltstone member (unit DSs, fig. 4) may correlate with the sandstone member of Independence. Note that this siltstone member is the stratigraphically higher of two siltstone units mapped as unit DSs by Dover (1983). The lower siltstone unit of Dover (1983), exposed in the headwaters of the Big Lost River east of Kent Peak, may correlate with the Trail Creek Formation but is not exposed in the Miller Canyon area (fig. 3).

The upper argillite member of the unnamed Silurian and Devonian rocks (unit DSau, fig. 4) may contain strata younger than any exposed in the Milligen Formation in the Triumph area.

KENT PEAK AREA

On the east side of Kent Peak, in the headwaters of the North Fork of the Big Lost River, Batatian (1991) reinterpreted strata mapped by Dover (1983) as "unnamed Devonian and Silurian" rocks and as Pennsylvanian and Permian Wood River Formation to be correlative with several of the informal members of the Milligen Formation (Turner and Otto, 1988, this volume). These correlations are based on lithologic similarity and require biostratigraphic confirmation.

Figure 4. Composite stratigraphic column of unnamed Silurian and Devonian rocks exposed in the area west of Miller Canyon, eastern Boulder Mountains, Custer County, Idaho, and their correlation with the Devonian Milligen Formation. Modified from Bruner (1991). Unit Dmc is quartzite member of Cait of Turner and Otto (1988).

Batatian (1991) mapped the cirque east of Kent Peak (plate 1) as containing a folded but continuous stratigraphic section of Silurian and Devonian rocks. Diamictite identical with that in the lower argillite member of the unnamed Silurian and Devonian strata in the Miller Canyon area, and containing very coarse quartz sand grains similar to the quartzite of Cait of the Milligen Formation, is present southeast of Kent Peak (plate 1). Batatian (1991) suggested that the strata are correlative with the Milligen Formation and that the lower argillite member, limestone of Lucky Coin, and sandstone of Independence are present (fig. 4).

Dover (1983) mapped some of the rocks east of Kent Peak as the upper part of the Wood River Formation and others as unnamed Silurian and Devonian argillite and siltstone. He mapped a low-angle fault (the "Wood River thrust") between Silurian and Devonian strata and the questionable strata in the cirque east of Kent Peak. He noted that argillite in the unit mapped as Wood River Formation contains "thick layers and lenses of dark chert pebble conglomerate [and] includes unit 7 of Hall and others (1974) and probably younger beds as well." Our observations suggest that these mixed carbonate-siliciclastic strata contain trace fossils (*Taenidium* and *Zoophycos*; Burton and Link, 1991) identical to those in the Wilson Creek Member of the Wood River Formation, supporting Dover's (1983) interpretation.

East of the problematic rocks in the cirque east of Kent Peak, and apparently downsection from them, is light-colored silicified banded siltstone (mapped as a lower interval of unit DSs by Dover, 1983) (fig. 3) that has yielded Silurian conodonts (Brennan, 1987) (fig. 5). These siltstone strata resemble the Silurian Trail Creek Formation, which is exposed along strike to the southwest in Park Creek and on Trail Creek Summit (Dover, 1983).

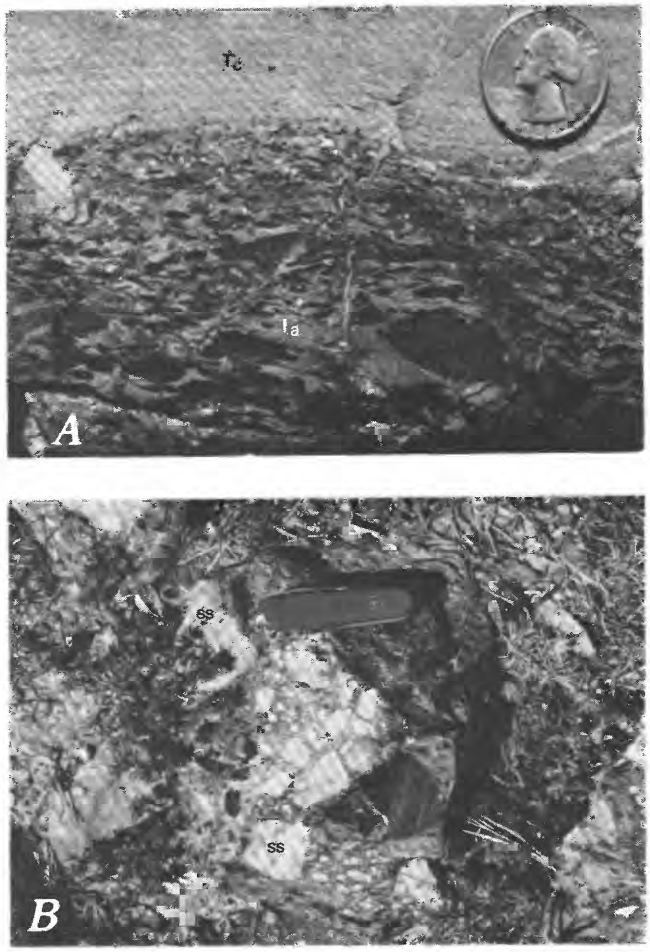


Figure 5. Unnamed Devonian strata east of Kent Peak. *A*, Graded bed of pebble breccia (T_a turbidite interval). Intrabasinal dark chert clasts float in matrix of calcareous silt at the base. Top of pebble bed contains extrabasinal quartz pebbles and granules. Overlying medium-grained calcareous sandstone is wavy parallel-laminated (T_c) interval. Coin shown for scale. *B*, Ungraded bed of pebble breccia containing intrabasinal dark cherty argillite clasts and coarse-grained sandstone clasts (ss). Knife shown for scale.

SUMMARY OF PROPOSED CORRELATIONS OF SILURIAN AND DEVONIAN STRATA

We suggest that some of the unnamed Silurian and Devonian strata in the North Fork of the Big Lost River correlate with the Trail Creek Formation and the informal members of the Milligen Formation as shown in figure 3. In particular, the lower interval of unit DSs (Dover, 1983; Batatian, 1991) may correlate with the Trail Creek Formation. The lower argillite member of the Milligen Formation (unit DMla of Batatian, 1991) may be present both in the Miller Canyon area and east of Kent Peak. This unit contains layers of diamictite (sedimentary breccia) and coarse-grained sandstone that correlate with the quartzite member of Cait, which is host to syngenetic stratabound mineral deposits. The limestone member of Lucky Coin may be present as thin interbeds in the lower argillite member of the unnamed Silurian and Devonian rocks in the Miller Canyon area (Bruner, 1991) and may be correlative with an approximately 70-m-thick limestone interval east of Kent Peak (Batatian, 1991). If these correlations are correct, the argillite member of Triumph, which is present above the diamictite and sandstone beds of the quartzite member of Cait in the Triumph mine area (Turner and Otto, this volume), may be present as unit DSla5 in the Miller Canyon area but thin or missing in the Kent Peak area. Strata correlative with the sandstone of Independence are present in both areas. The upper argillite member in the Miller Canyon area (unit DSau) may represent strata younger than any exposed in the Milligen Formation in the Triumph area.

UPPER CONTACT OF UNNAMED SILURIAN AND DEVONIAN ROCKS

The upper contact of the unnamed Silurian and Devonian rocks may be an unconformity with the Mississippian Copper Basin Formation. In the East Fork of Trail Creek, rocks mapped as unit DSa by Dover (1983) contain dark-gray and brown carbonaceous argillite and dark-gray to black, coarse-grained quartzite. West of the East Fork of Trail Creek (plate 1), these rocks are apparently conformably overlain by sandy and conglomeratic argillite and granule to boulder conglomerate shown as the upper part of the lower Copper Basin Formation (Lower Mississippian) (Huerta and others, 1991; Rodgers and others, this volume). Biostratigraphic studies are required to confirm this interpretation.

PALEOZOIC SALMON RIVER ASSEMBLAGE

The Paleozoic Salmon River assemblage (Hobbs, 1985) is exposed in the southern part of the Challis 1°×2° quadrangle directly to the north of the Hailey 1°×2° quadrangle. The Salmon River assemblage contains isoclinally folded slices of dark-colored argillite, lighter colored siltstone, limestone, and quartzite. The assemblage may be about 2,000 m thick, but isoclinal folds and faults make this estimate tenuous (Hall and Hobbs, 1987). Samples collected in situ from the Salmon River assemblage contain Late Cambrian and Late Devonian conodonts; however, the Late Cambrian fauna were derived from carbonaceous limestone within what is interpreted to be a tectonic slice at the base of the assemblage (Hall, 1985). Fossils in two blocks of limestone float found north of the Salmon River contain Late Mississippian fossils (Nilsen, 1977; Hobbs, 1985). The Salmon River assemblage is described as Late Mississippian by Nilsen (1977) and Tschanz and others (1986) and as simply Paleozoic by Fisher and others (1983). The Mississippian age assignment for the assemblage is doubtful because bioclastic limestone beds similar to those that yielded the Late Mississippian fossils are unrecognized within Salmon River assemblage stratigraphy. It is our opinion that the majority of the Salmon River assemblage is Devonian in age.

The Salmon River assemblage occupies a structural position similar to the entire Ordovician through Devonian section in the Boulder Mountains west of the Pioneer thrust fault. On the west side of the White Cloud Peaks it underlies the Pennsylvanian and Permian Grand Prize Formation along a sheared unconformity that had formerly been interpreted as a regional thrust fault similar to the "Wood River thrust" (Sengebush, 1984; Hall, 1985; Hall and Hobbs, 1987; Mahoney and Sengebush, 1988; Mahoney and others, 1991; Mahoney, this volume). On the east side of the outcrop belt the Salmon River assemblage is thrust over Ordovician carbonate and quartzite strata including the Clayton Mine Quartzite, Ella Marble, Kinnikinic Quartzite, and Saturday Mountain Formation (Hobbs, 1985; Hall and Hobbs, 1987) (fig. 1, plate 1).

In the Washington Basin mineralized area (Mahoney, this volume) (fig. 1) of the White Cloud Peaks, the Salmon River assemblage underlies sheared conglomerate boudins of the basal Grand Prize Formation. These boudins are similar to those of the Hailey Member of the Wood River Formation along the sheared contact between the Wood River and Milligen Formations. The map of Tschanz and others (1986) shows this unconformable relationship, though it includes rocks now recognized as Grand Prize Formation within the Wood River Formation and assigns a Mississippian age to the Salmon River assemblage.

The bulk of the Salmon River assemblage south of the Salmon River in the White Cloud Peaks (fig. 1, plate 1) contains dark-colored argillite, siltite, and sandstone that yield Late Devonian fossils. The Paleozoic Salmon River assemblage and the Devonian Milligen Formation are similar lithologically and stratigraphically in that both contain thick sections of homogeneous black argillite, abundant thin-bedded fine-grained turbidite sequences, and intercalated blue-gray, locally tremolitic, limestone. The Salmon River assemblage and the Milligen Formation both host inferred syngenetic sulfide deposits that have identical lead-isotope signatures (Sanford and Wooden, this volume). We suggest that the Salmon River assemblage in the White Cloud Peaks area mostly correlates with the Milligen Formation. Hall (1985, p. 121) noted that the Salmon River assemblage contains a fracture cleavage almost parallel with bedding and differs from that of the Milligen, "which has a penetrative shear cleavage that is at a large angle to bedding, has a much higher metamorphic grade, and has a phyllitic sheen." Our mapping has not revealed a systematic difference in degree of deformation between the two units.

MINERAL DEPOSITS IN SILURIAN AND DEVONIAN ROCKS

Correlation of the Milligen Formation with the unnamed Silurian and Devonian strata of the North Fork of the Big Lost River and the Salmon River assemblage to the north in the White Cloud Peaks area suggests stratigraphic and structural continuity between these areas. This continuity has important implications for mineral resource potential. The entire area underlain by the Milligen Formation, the unnamed Silurian and Devonian units, and the Salmon River assemblage has potential for the occurrence of syngenetic stratabound silver-lead-zinc deposits. Syngenetic stratabound zinc and barite deposits within the Salmon River assemblage at the Hoodoo mine in Slate Creek have textures and lead-isotope signatures similar to inferred syngenetic silver-lead-zinc deposits in the Milligen Formation in parts of the Triumph mine (Hall, 1985; Hall and Hobbs, 1987; Sanford and Wooden, this volume) (plate 1). Syngenetic textures are also present in lead-zinc sulfide minerals in the lower argillite member of the unnamed Silurian and Devonian rocks in the Miller Canyon area, as described later (Bruner, 1991). In addition, epigenetic polymetallic veins are present in the Milligen Formation and correlative units and in younger formations that are in stratigraphic or structural contact with lower Paleozoic strata.

Mineral exploration activity has been sporadic, and continues, in the Slate Creek, Washington Basin, and North Fork of the Big Lost River drainages (fig. 1). Unnamed Silurian and Devonian strata in the Summit mineralized area near Miller Canyon (fig. 1) host sulfide-mineralized rocks. Structurally complex areas near Miller Canyon have been explored since the mid-1980's for syngenetic stratabound silver-lead-zinc mineral deposits. Bruner (1991) documented textures in polished sections and in hand samples indicative of both syngenetic sedimentary exhalative and epigenetic sulfide mineralization. These textures are described in the section on mineral deposit models.

MISSISSIPPIAN COPPER BASIN FORMATION

The Lower and Upper Mississippian Copper Basin Formation crops out extensively in the northeastern part of the Hailey 1°×2° quadrangle, in the headwaters of the East and North Forks of the Big Lost River (fig. 1). The Copper Basin Formation was named by Ross (1962) for a thick sequence of coarse-grained siliciclastic strata in the Pioneer Mountains. Paull and others (1972) raised the formation to group rank and defined six formations within the Copper Basin Group. Paull and Gruber (1977) revised the definition of the Copper Basin Group. Nilsen (1977) and most subsequent workers (Skipp, Sando, and Hall, 1979; Dover, 1981, 1983) recognized stratigraphic and structural complications within these Mississippian strata, including structural repetition by the Glide Mountain thrust fault, and retained the name Copper Basin Formation after Ross (1962) (fig. 2). Wilson (1994) questioned the premise that the Copper Basin Formation occupies two thrust sheets and reaffirmed that, in the type area, the stratigraphy is essentially as described by Paull and others (1972) and Paull and Gruber (1977). Wilson and othes (1994) reviewed the stratigraphy and facies relations of the Copper Basin Formation.

The Copper Basin Formation contains graded beds of cobble and pebble conglomerate, as well as sand- and silt-size siliciclastic turbidite and argillite. Clasts in conglomerate include dark-colored chert and argillite similar to rocks of the Devonian Milligen Formation, as well as a large proportion of light-colored quartzite clasts of uncertain provenance.

The lower part of the Copper Basin Formation is of Early Mississippian age (Kinderhookian and Osagean) and is at least 3,000 m thick. As mapped by Nilsen (1977) and Dover (1981, 1983), these Lower Mississippian strata are exposed in two thrust sheets separated by the Glide Mountain thrust fault. Detailed examination of the Glide Mountain thrust fault by Wilson (1994; Wilson and others, 1994) suggests that the mapped structure is variously a stratigraphic contact (locally sheared), a normal fault, and thrust fault. For further discussion see Rodgers and others (this volume).

According to Wilson (1994), in the Hailey 1°×2° quadrangle the exposed Copper Basin Formation is solely of Early Mississippian age. Because the formation is internally sheared and locally tightly folded, thickness estimates are structural. The basal contact drapes older rocks in the Wildhorse inlier. As divided on plate 1, the formation consists of a lower interval (unit Mcl) that contains a basal of dark-gray argillite, siltite, and granule conglomerate turbidite (the Little Copper Member, from 0 to more than 660 m thick) overlain by a discontinuous, but generally eastward thickening wedge of fine-grained mixed carbonate-siliciclastic turbidite (Drummond Mine Limestone Member, 0–910 m thick). An upper unit (unit Mcu) of pebble and cobble conglomerate, sandstone, and argillite (more than 1,150 m thick) is mapped on plate 1. This upper unit includes the Scorpion Mountain Formation of Paull and Gruber (1977), the lower part of the Muldoon Canyon Formation of Paull and Gruber (1977), and the beds mapped as Green Lake Limestone Member of Copper Basin Formation north of Dry Canyon along the East Fork of the Big Lost River by Dover (1981).

The upper part of the Copper Basin Formation (Upper Mississippian, Meramecian and Chesterian) is not exposed in the map area of plate I. In figure 2 it is shown as the upper Copper Basin Formation and includes at least 580 m of shallow-water sandstone and mudstone exposed in the western part of the Idaho Falls 1°×2° quadrangle (Skipp, Sando, and Hall, 1979; Skipp, 1989) and including the Iron Bog Creek Formation of Paull and Gruber (1977) (fig. 2).

The siliciclastic parts of the Copper Basin Formation contain rare marine fossils. The primary age control on the Copper Basin Formation is from the intervals of limestone turbidite (the Drummond Mine Limestone Member and the lenticular Green Lake Limestone Member), which have yielded several identical assemblages of upper Kinderhookian conodonts (Sandberg, 1975; Skipp, Sando, and Hall, 1979; Dover, 1981; Wilson, 1993). Deep-water ichnofossils in these mixed carbonate-siliciclastic turbidites include *Taenidium*, *Phycosiphon*, *Phyllodocites*, *Cosmorhaphe*, and *Chondrites*.

The upper part of the Copper Basin Formation along Iron Bog Creek in the Idaho Falls 1°×2° quadrangle, 17 km east of the east edge of the Hailey quadrangle, is dated as Chesterian in age by several macrofossil collections (Skipp, Sando, and Hall, 1979; Skipp, 1989).

The Copper Basin Formation is thought to represent Lower Mississippian flysch and overlying Upper Mississippian molasse that filled a foreland basin east of an emergent highland of the Antler orogenic belt (Poole, 1974; Nilsen, 1977; Poole and Sandberg, 1977, 1991). Paleogeographic reconstructions (Poole and Sandberg, 1991) show that the Copper Basin Formation mainly represents deposits of a westward-derived, siliciclastic Scorpion Mountain–Brockie Lake submarine fan system, although recent work (Wilson and others, 1994) suggests that a fault-bounded southern source area provided the bulk of the coarse detritus to unit Mcu. During a relative sea-level highstand in late Kinderhookian time the eastward-derived calciclastic Drummond Mine fan or ramp system prograded westward into the basin. The Upper Mississippian part of the formation is interpreted to be a marginal-marine facies that represents filling of the Antler foreland basin (Nilsen, 1977; Skipp, Sando, and Hall, 1979).

The majority of the Copper Basin Formation is exposed east of the Pioneer thrust fault, though there are small outcrops west of the Pioneer thrust fault (Rodgers and others, this volume). The Pioneer thrust fault places Lower Ordovician to Middle Silurian Phi Kappa and Middle Silurian Trail Creek Formations and an unnamed Silurian and Devonian argillite unit over the upper part of the lower Copper Basin Formation. The Pioneer thrust fault is the westernmost major Mesozoic thrust fault in south-central Idaho (Dover, 1981, 1983; Rodgers and others, this volume) (fig. 1).

West of the Pioneer thrust fault (from the East Fork of Trail Creek north across the North Fork of the Big Lost River) are several outcrops of undated conglomeratic rocks mapped as Copper Basin Formation (Dover, 1983). A poorly exposed contact west of the East Fork of Trail Creek separates undated argillite and quartzite mapped as undifferentiated Silurian and Devonian argillite (unit Dsa of Dover, 1983) from overlying undated argillite, siltstone, fine-grained lithic wacke, quartzite, and channel-fill conglomerate mapped as Copper Basin Formation (plate 1). Although Dover (1983) mapped this contact as a thrust fault, the contact probably is gradational because bedding attitudes are consistent across it and stratigraphic tops are consistently to the west (Huerta and others, 1991). This relation implies that in Early Mississippian Osagean (?) time the western part of the Copper Basin flysch trough may have onlapped Devonian argillaceous rocks in stratigraphic continuity with Milligen Formation strata thought to have composed the Antler highland. The relation is consistent with the identification of terrigenous and shallow-water facies in western outcrops of the Copper Basin Formation (Dover, 1981). Below this contact, approximately 300 m to the east, is the Lower Ordovician Basin Gulch Quartzite Member, the basal unit of the Phi Kappa Formation. This relation indicates that several hundred meters of Devonian and (or) Silurian strata has been faulted out, but the stratigraphic location of this fault is unclear. More detailed geologic mapping and biostratigraphic study is needed to adequately document this relation.

East of the Pioneer thrust fault, in the Dry Canyon and Wildhorse inliers of the northeastern part of the Hailey 1°×2° quadrangle, lower Paleozoic shelf carbonate rocks are exposed beneath the Copper Basin Formation. The contacts were interpreted to be thrust faults by Dover (1981, 1983), low-angle normal faults by Kim (1986), and locally sheared angular unconformities by Wilson (1994). To the southeast, in the Idaho Falls 1°×2° quadrangle, the Middle Devonian Carey Dolomite unconformably underlies the Copper Basin Formation near Garfield Canyon (Kunkel, 1989) and the Upper Devonian Picabo Formation unconformably underlies the Copper Basin Formation near Fish Creek Reservoir (Link and others, 1988). Farther to the east, in the White Knob Mountains, the Copper Basin Formation of the Copper Basin plate is thrust over finer grained coeval strata (Lower Mississippian McGowan Creek Formation) of the Grouse thrust plate (Link and others, 1988; Skipp and others, 1990) (fig. 2).

In at least two locations, strata of the Copper Basin Formation have been transported along Eocene low-angle normal faults. In the headwaters of Summit Creek, a block of Copper Basin Formation conglomerate lies in low-angle normal fault contact above Wood River Formation and unnamed Devonian strata (Huerta and others, 1991). This contact was shown as a thrust fault by Dover (1983). Tectonic slices of Mississippian-age siltstone are also present between the Middle Pennsylvanian and Lower Permian Wood River Formation and the Devonian Milligen Formation in the Rock Roll Canyon quadrangle (C.M. Tschanz, U.S. Geological Survey, unpublished data, 1987).

MINERAL DEPOSITS IN THE COPPER BASIN FORMATION

Lead-silver skarn, replacement, and vein deposits are present in laminated calc-turbidites of the Drummond Mine Limestone Member in the Summit mineralized area (fig. 1) in the Phi Kappa mine (Winkler and others, this volume) and at several locations in the Idaho Falls 1°×2° quadrangle to the east. Characteristics of these deposits are summarized in the section on mineral deposit models.

PENNSYLVANIAN AND PERMIAN SUN VALLEY GROUP

Hall (1985) recognized three lithologically similar upper Paleozoic formations in south-central Idaho: the Wood River, Dollarhide, and Grand Prize Formations. Each of these consists of generally fine grained, locally carbonaceous, mixed carbonate-siliciclastic strata. Link and others (1988) proposed that these strata were deposited in the epicratonic Wood River Basin during Pennsylvanian and Permian time. Mahoney and others (1991) formally defined the Sun Valley Group to include these formations and, using new paleontological collections, established that all three formations are of Middle Pennsylvanian to Early Permian age (figs. 6, 7).

Hall (1985) proposed that the formations now included in the Sun Valley Group belonged to distinct thrust complexes or tectonic stacks and that their mutual boundaries were everywhere thrust faults of significant lateral displacement. Recent geologic mapping and measurement of stratigraphic sections (Mahoney and others, 1991; O'Brien, 1991; Worl and others, 1991) demonstrate that boundaries between formations of the Sun Valley Group in most places represent facies changes rather than structural contacts (fig. 8). Faulted contacts are also present (plate 1). The three formations of the Sun Valley Group are mappable lateral variations of originally continuous strata deposited in the Wood River Basin.

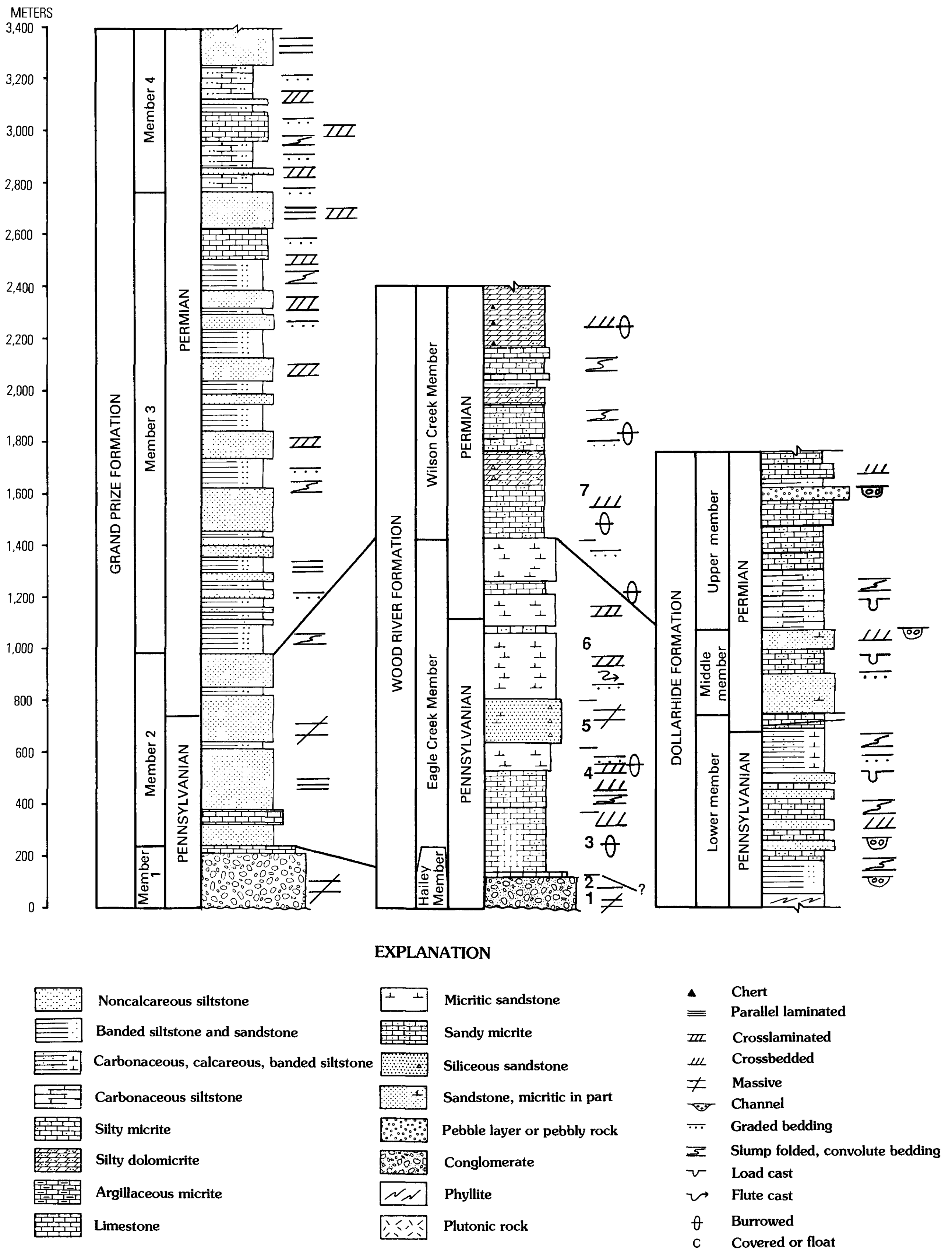
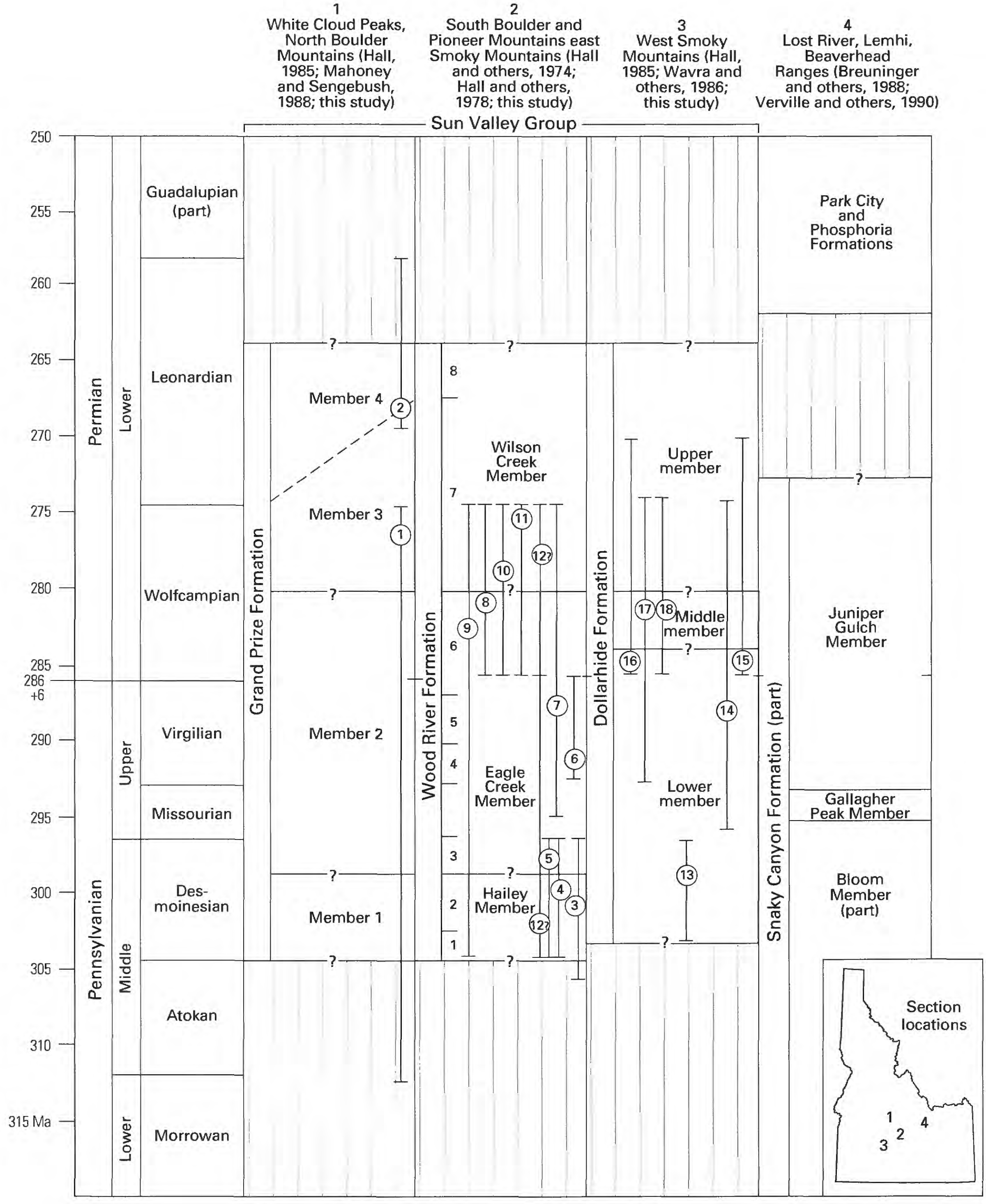


Figure 6. Generalized stratigraphic columns of the Sun Valley Group, south-central Idaho. Numbers next to Wood River Formation column indicate informal units of Hall and others (1974). Modified from Mahoney and others (1991, fig. 3).

Figure 7 (facing page). Correlation chart for Middle Pennsylvanian to Lower Permian rocks of the Sun Valley Group, south-central Idaho. Modified from Mahoney and others (1991, fig. 2). Bars with numbers indicate ranges of previously published and new biostratigraphic collections. Ranges for fusulinids are based on Loeblich and Tappan (1988). Circled numbers indicate stratigraphic position of the collection. Identified taxa and name of biostratigrapher are as follows. Grand Prize Formation: 1, Conodont elements of *Adetognathus* sp., elements of *Hindeodus* cf. *H. Ninutus* (Ellison), elements of *Idiognathosus* sp., (late Morrowan to Wolfcampian), Member 2? or 3?, Peach Creek, identified by A.G. Harris (Hall, 1985, p. 125); 2, Conodont *Neogondelella idahoensis* (middle and late Leonardian), member 3 or 4, Pole Creek, identified by B.R. Wardlaw (Hall, 1985, p. 125). Wood River Formation: 3, *Wedekindellina* (late Atokan(?) to Desmoinesian), Hailey Member limestone, north of Seamans Creek in principal reference section, (Bostwick, 1955) and *Pseudozaphrentoides* (Middle Pennsylvanian brachiopod), Eagle Creek Member (unit 3), Wilson Creek section (Thomasson, 1959); 4, *Beedeina* and numerous brachiopods (Desmoinesian), Hailey Member limestone, north of Seamans Creek in principal reference section (Bostwick, 1955; Hall and others, 1974, p. 91); 5, *Beedeina* (Desmoinesian), Eagle Creek Member (unit 3), north of Seamans Creek in principal reference section (Hall and others (1974, p. 91); 6, *Triticites*, *Pseudofusulinella* sp. and *Triticites* sp. aff *cullomensis* Dunbar and Condra (Virgilian), Eagle Creek Member (unit 4), Seamans Creek in principal reference section (Hall and others, 1974, p. 92); 7, *Triticites* sp. (Missourian to Wolfcampian), Eagle Creek Member (unit 5), north of Seamans Creek in principal reference section (Hall and others, 1974, p. 93.); 8, *Schubertella* and *Staffella*, unit 6 north of Seamans Creek in principal reference section (Hall and others, 1974, p. 94), as well as *Triticites cullomensis* (Wolfcampian), Eagle Creek Member, mid-upper unit 6, Wilson Creek Ridge type section (Burton, 1988), identified by C.A. Ross (1988); 9, *Pseudofusulinella* (Desmoinesian to Wolfcampian), Eagle Creek Member, limestone at top of unit 6, Wilson Creek Ridge type section (Burton, 1988), identified by C.A. Ross (1988); 10, *Triticites confertus*, *T. pinguis*, *T. meeki*, *T. cellamagnus*, *Pseudofusulinella utahensis*, *P. grandensis*, *P. elkoensis*, (Wolfcampian), Wilson Creek Member, lower unit 7, Basin Gulch type section (Burton, 1988), identified by C.A. Ross (1988); 11, *Pseudofusulina grandensis*, *P. elkoensis*, *P. wellsensis*, *Schwagerina*, *Paraschwagerina* (Wolfcampian), Wilson Creek Member, upper unit 7 or 8, Basin Gulch type section (Burton, 1988), identified by C.A. Ross (1988); this collection is thought to be from the same upper limestone bed tentatively assigned to Leonardian and Guadalupian(?) by Hall and others (1978, p. 581); 12, Bioclastic limestone east of summit of Kent Peak (Batatian, 1991). The unit yielded conodont collection *Neogondolella* sp. (not *N. idahoensis*), *Drepanodus*-like sp., and *Sweetognathus* sp. indicating a Late Pennsylvanian to Early Permian age (D. Van Hofwegan, C. Spinsoa, written commun., 1991) and macrofossils including *Chaetetes* sp. indicating a Desmoinesian (Middle Pennsylvanian) age (D.E. Fortsch, oral commun., 1991). Elevation 10,800 ft, ridge northeast of the summit of Kent Peak, Ryan Peak quadrangle, lat 45°53'35" N., long 114°23'32" W. Dollarhide Formation (14 through 18 are reported in Mahoney and others (1991) and O'Brien (1991): 13, Conodonts *Idiognathodus delicatus* (Gunnell) and *Neognathodus dilatus* (Stauffer and Plummer), Desmoinesian (Middle Pennsylvanian), lower member Dollarhide Formation, NW¼SW¼ sec. 25, T. 3 N., R. 17 E., Mahoney Butte 7.5-minute quadrangle, Blaine County, Idaho, lat 43°33.75' N., long 114°24.00' W., collected by Betty Skipp, identified by R.G. Stamm and B.R. Wardlaw (written commun., 1990, *in* Skipp and others, in press). USGS collection number 31400–PC; 14, Fusulinids, possibly *Triticites* (Missourian to Wolfcampian), lower member?, Sky Ranch flat, west of Bellevue, collected by P.K. Link, identified by R.C. Douglass (written commun., 1988); 15, Fusulinids *Schwagerina sp.* (Wolfcampian to mid-Leonardian), lower member, Sky Ranch flat, west of Bellevue, two collections, by P.K. Link and R.S. Lewis, identified by C.A. Ross (written commun., 1988) and D.A. Myers (written commun., 1989); 16, Fusulinids, possibly *Pseudofusulina* (Wolfcampian to Leonardian), lower member, Wolf Tone Creek, R.C. Douglass (written commun., 1978, *in* Hall, 1985, p. 124); 17, Fusulinids suggestive of *Schwagerina* (Virgilian and Wolfcampian), elevation 8,640 ft on ridge at northern headwaters of Deer Creek, SE¼ sec. 29, T. 3 N., R. 16 E., Buttercup Mountain 7.5-minute quadrangle, collected by J.P. O'Brien, identified by D.A. Myers (written commun., 1989); 18, Fusulinid *Bartramella* sp. Eelevation 8,640 ft on ridge at northern headwaters of Deer Creek, SE¼ sec. 29, T. 3 N., R. 16 E., Buttercup Mountain 7.5-minute quadrangle, collected by R.S. Lewis, identified by D.A. Myers (written commun., 1989). Because this locality yielded both *Bartramella* and *Schwagerina* (collection 15 above), the collection may be *B. heglarensis*, which is associated with *Schwagerina sublettensis* (Wolfcampian) in the Sublett Range of southern Idaho (Thompson and others, 1958).


1
White Cloud Peaks, North Boulder Mountains (Hall, 1985; Mahoney and Sengebush, 1988; this study)
2
South Boulder and Pioneer Mountains east Smoky Mountains (Hall and others, 1974; Hall and others, 1978; this study)
3
West Smoky Mountains (Hall, 1985; Wavra and others, 1986; this study)
4
Lost River, Lemhi, Beaverhead Ranges (Breuninger and others, 1988; Verville and others, 1990)
Sun Valley Group
250
255
260
265
270
275
280
285
286 +6
290
295
300
305
310
315 Ma
Permian
Pennsylvanian
Lower
Upper
Middle
Lower
Guadalupian (part)
Leonardian
Wolfcampian
Virgilian
Missourian
Des-moinesian
Atokan
Morrowan
Grand Prize Formation
Member 4
Member 3
Member 2
Member 1
Wood River Formation
Wilson Creek Member
Eagle Creek Member
Hailey Member
Dollarhide Formation
Upper member
Middle member
Lower member
Snaky Canyon Formation (part)
Park City and Phosphoria Formations
Juniper Gulch Member
Gallagher Peak Member
Bloom Member (part)
Section locations

The Sun Valley Group contains eight lithofacies—conglomerate, bioclastic limestone, micritic sandstone, banded micritic sandstone-siltstone, graded silty micritic limestone, silty micritic limestone, sandy micritic limestone, and carbonaceous siltstone—described by Mahoney and others (1991). The Middle Pennsylvanian part of the Sun Valley Group (Hailey Member of the Wood River Formation and correlative Grand Prize Formation, member 1) consists mostly of proximally derived chert- and quartzite-pebble conglomerate and interbedded biostromal limestone deposited on a braid delta and adjacent slope. The Middle Pennsylvanian to Lower Permian part of the group (Eagle Creek and Wilson Creek Members of the Wood River Formation, all of the Dollarhide Formation, and the upper three members of the Grand Prize Formation) comprises fine-grained mixed carbonate and siliciclastic rocks deposited in a south-sloping ramp-apron system with distal (carbonate bank and cratonal) provenance.

Hall (1985) defined the Grand Prize and Dollarhide Formations as distinct from the long-recognized (Lindgren, 1900) Wood River Formation on the basis of lithologic variations, carbon content (as demonstrated by color), and metamorphic overprint. The Grand Prize Formation, in addition to being sandier and coarser grained than the other formations, has undergone metamorphic silicification. The Dollarhide Formation is characteristically darker colored, more carbonaceous, and finer grained than the Wood River or Grand Prize Formations. Specific relations between the formations are described after the following stratigraphic discussion, modified from Mahoney and others (1991).

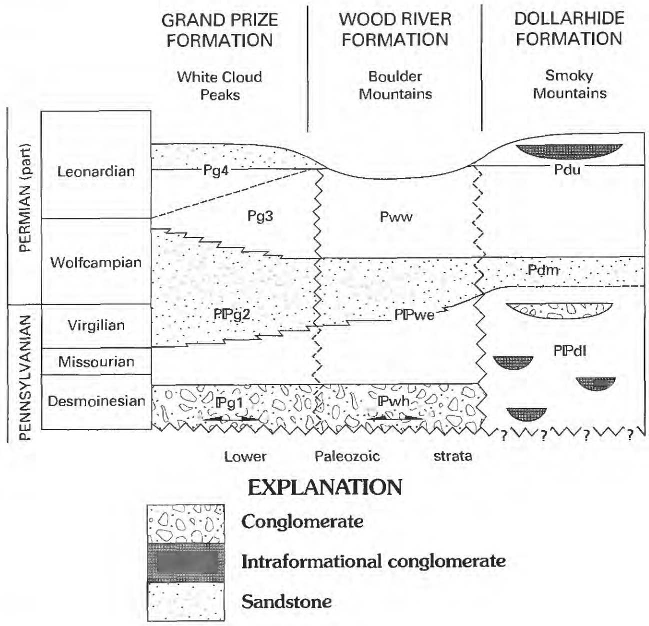


Figure 8. Schematic diagram showing facies changes within the Sun Valley Group across the Wood River Basin, south-central Idaho. Unpatterned areas represent mixed siliciclastic and carbonate rocks, and light-shaded areas indicate prograding wedges of medium- and fine-grained sand. Solid sawtooth line indicates observed transition between formations; dashed sawtooth line indicates inferred transition; double-ended arrow indicates sheared unconformity at base of Sun Valley Group; and dashed line between units Pg3 and Pg4 indicates that age of boundary is uncertain. Grand Prize Formation: ℙg1, member 1, Pℙg2, member 2, Pg3, member 3, Pg4, member 4. Wood River Formation: ℙwh, Hailey Member, Pℙwe, Eagle Creek Member, Pww, Wilson Creek Member. Dollarhide Formation: PℙPdl, lower member, Pdm, middle member, Pdu, upper member.

WOOD RIVER FORMATION

The Wood River Formation, as named by Lindgren (1900), included all Paleozoic rocks exposed near the Wood River Valley. Umpleby and others (1930) restricted the formation to light-colored rocks of Pennsylvanian and Permian age. The formation is now known to contain rocks of Desmoinesian to Leonardian age (Bostwick, 1955; Ross, 1960; Hall and others, 1974; Mahoney and others, 1991) (fig. 6) but may include rocks as young as Guadalupian (Hall and others, 1978; Skipp and Hall, 1980).

The Wood River Formation is divided into three formal members (Mahoney and others, 1991) (figs. 6, 7), the Hailey Member (basal 200 m), the Eagle Creek Member (middle 880+ m), and the Wilson Creek Member (upper 800+ m). The principal reference section of the Wood River Formation, east of Bellevue (section W5, plate 1), which includes the type section of the Hailey Member at its base, was measured by Hall and others (1974). Mahoney and others (1991) designated several supplemental reference sections. The Wilson Creek section (section W4, fig. 9*B*, plate 1), exposed on the ridge crest north of Wilson Creek, contains the type section of the Eagle Creek Member. The Basin Gulch section (section W3, fig. 9*A*, plate 1), exposed on steep slopes northeast of the head of Lake Creek and on the ridge crest to the northeast, contains the type section of the Wilson Creek Member. Descriptions of these stratigraphic sections are contained in table 2. Two other reference sections that contain exposures of the entire formation are on the ridge north of Lake Creek (section W2, fig. 9*C*, plate 1) and on the ridge north of Murdock Creek (section W1, plate 1). Precise locations of these sections are described in table 3.

The basal Hailey Member of the Wood River Formation, as defined by Mahoney and others (1991), consists of the Hailey Conglomerate Member and the bioclastic limestone of unit 2 of the Wood River Formation, as described

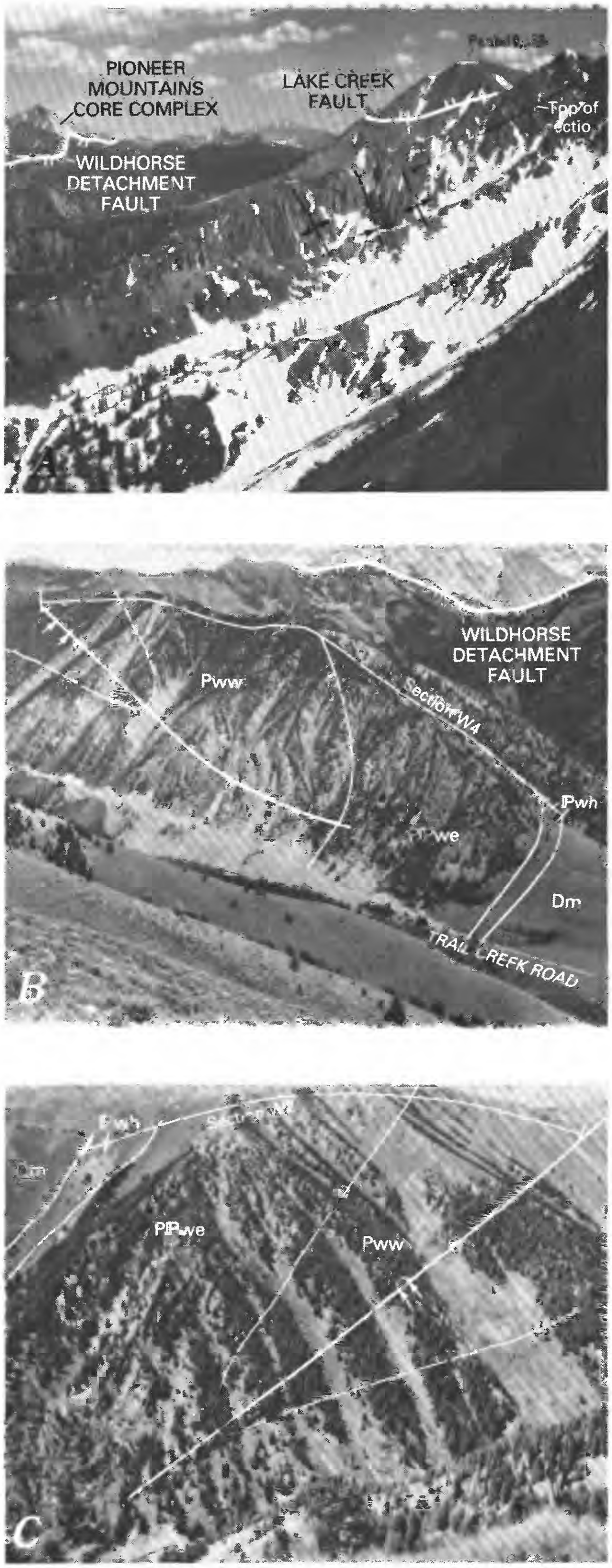


Figure 9. Reference sections of the Wood River Formation northeast of Ketchum, Idaho. Locations of sections shown on plate 1 and described in table 3. *A*, Basin Gulch section (W3), upper part within Wilson Creek Member, on the near ridge. Section was measured in southwest-dipping beds up to axis of eastward-overturned anticline. Upper part of the section is repeated in folded rocks to the northeast. View is looking southeast from west of the left fork of upper Trail Creek, toward the Pioneer Mountains core complex on horizon. Traces of Wildhorse detachment fault and Lake Creek fault on peak 10,458 are shown. *B*, Wilson Creek ridge section (W4), view looking southeast from ridge west of Trail Creek. Complete section of Wood River Formation (Pwh, Hailey Member; PPwe, Eagle Creek Member; Pww, Wilson Creek Member) below the Lake Creek fault is present in overturned southwest-dipping beds from sheared unconformity with Devonian Milligen Formation (Dm) to axis of northeast-overturned syncline. In distance is the lower plate of the Wildhorse detachment fault, the core of the Pioneer Mountains core complex. *C*, Lake Creek ridge section (W2), view looking northwest from ridge west of Trail Creek. Complete section of Wood River Formation (Pwh, Hailey Member; PPwe, Eagle Creek Member; Pww, Wilson Creek Member) above the Lake Creek fault is present in overturned southwest-dipping beds from sheared unconformity with Devonian Milligen Formation (Dm) to axis of northeast-overturned syncline.

by Hall and others (1974). Unit 2 is vertically and laterally gradational with the conglomerate (Burton, 1988; Mahoney and others, 1991). The Eagle Creek Member consists of units 3 through 6 of Hall and others (1974), and the Wilson Creek Member consists of units 7 and 8 of Hall and others (1974, 1978).

The Hailey Member is 0–200 m thick and consists of 0–180 m of light-brown to light-gray conglomerate gradationally overlain by 15–30 m of bluish-gray bioclastic limestone (fig. 6). The member is primarily exposed east of the Wood River in the Pioneer and Boulder Mountains, although it is exposed in a few areas west of the Wood River near Hailey (plate 1). To the east, near Fish Creek Reservoir in the Idaho Falls 1°×2° quadrangle, the Hailey Member contains 180 m of calcareous sandstone and dispersed chert-pebble conglomerate and bioclastic beds (Link and others, 1988). The Hailey Member may have been deposited on an irregular topographic surface and is locally depositionally absent (Winsor, 1981).

The basal contact of the Hailey Member is a locally sheared unconformity with the underlying Devonian Milligen Formation (Burton, 1988; Burton and others, 1989; Ratchford, 1989, in press; Mahoney and others, 1991; Burton and Link, this volume). In many areas interstratal slip along the unconformity has produced a shear zone in which the Hailey Member is present as kilometer-scale boudins, or is attenuated, particularly in fold limbs, or is thickened in

Table 2. Descriptions of type sections for three members of the Wood River Formation, south-central Idaho.
[Modified from Mahoney and others (1991, table 2)]

TYPE SECTION OF THE HAILEY MEMBER OF THE WOOD RIVER FORMATION

Measured by Hall and others (1974) east of Bellevue on the ridge north of Seamans Creek in northeast-dipping beds from 6,390 to 6,600 ft elevation, SW¼ sec. 28, T. 2 N., R. 19 E., Seamans Creek quadrangle (section W5, fig. 2).

	Thickness (meters)
Conformable contact to Eagle Creek Member (unit 3)	
Top of Hailey Member	
Unit 2 of Hall and others (1974)	
Limestone, bluish-gray, medium- to thick-bedded, fine-grained, locally abundant crinoidal debris, bryozoa, and brachiopod fragments. Contains 5–10 percent detrital quartz grains	15.0
Conformable contact to unit 2	
Unit 1	
Chert-pebble conglomerate, light-gray, siliceous cement	1.5
Limestone, brown, fine-grained, algal; chert-pebble conglomerate, beds 0.3–0.6 m thick	12.5
Chert-pebble conglomerate, as above	0.3
Limestone, brown, fine-grained, algal; 10 percent fine-grained quartzite beds 1.3–2.5 cm thick	16.2
Chert-pebble conglomerate, light-gray and brownish-gray, as above	64.0
Limestone, micritic, brown, silty, in part silicified	0.6
Chert-pebble conglomerate, light-gray, chalcedonic matrix	3.7
Limestone, brown, fine-grained, sandy	1.5
Chert-pebble conglomerate, siliceous matrix, chert and quartzite clasts; quartzite, light-green, fine-grained, thick-bedded	16.8
Limestone, sandy, brown	0.3
Chert-pebble conglomerate, light-gray, thick-bedded, siliceous cement, a few well-rounded light-brown limestone and white quartzite clasts, subrounded chert pebbles 1–2.5 cm long	31.0
Total thickness Hailey Member	135.5

TYPE SECTION OF THE EAGLE CREEK MEMBER OF THE WOOD RIVER FORMATION

Measured on ridge north of Wilson Creek, in northeast-dipping beds from 6,600 to 8,800 ft elevation, NE¼ sec. 14 and SW¼ sec. 12, T. 5 N., R. 18 E., Rock Roll Canyon quadrangle (Burton, 1988) (section W4, fig. 2).

	Thickness (meters)
Wilson Creek Member (gradational contact)	
Top of Eagle Creek Member	
Unit 6 of Hall and others (1974)	
Micritic sandstone, medium-gray, weathering brown, fine-grained trough cross-laminated, syndepositional convolute laminae, thin-bedded, interbedded with silty micrite	20
Micritic sandstone (quartzarenite), gray, weathering brown, thick-bedded to massive, fine- to coarse-grained, fractured; sparse trough cross lamination, load casts and flute casts; *Arenicolites*. Section is cut by fault at 470 m	483
Unit 5 of Hall and others (1974)	
Fine sandstone (quartzarenite), siliceous or partly calcareous, light-brown, well-indurated, thickly bedded to massive; crude parallel laminae, dish structures, intensely fractured	160
Unit 4 of Hall and others (1974)	
Silty micrite to fine micritic sandstone, brown, thin- to medium-bedded, trough cross- and convolute-laminated, weakly graded beds	160
Silty micrite, dark-gray, weathering gray-brown, wavy parallel laminae; interbeds of dark-gray micritic mudshale	40
Very fine micritic sandstone, dark-brown, thin-bedded, convolute laminae, load casts, micritic mudstone partings; *Scalarituba*	20
Unit 3 of Hall and others (1974)	
Silty micrite and mudshale, dark-brown to purple-gray, pink-gray weathering, moderately bioturbated, crudely fissile; *Neonereites, Scalarituba, Phycosiphon, Zoophycos*	175
Silty micrite, gray-brown and dark-brown, thin-bedded, wavy parallel laminae, burrowed to intensely bioturbated; contains interbeds as much as 1.5 m thick of silty allochem limestone containing crinoid, brachiopod, and rugose coral bioclasts, *Neonereites, Scalarituba, Muensteria?, Spirophycos?*	85
Top of Hailey Member (bioclastic packstone of unit 2 of Hall and others, 1974)	
Total thickness Eagle Creek Member	1,143

Table 2. Descriptions of type sections for three members of the Wood River Formation, south-central Idaho—Continued.
[Modified from Mahoney and others (1991, table 2)]

TYPE SECTION OF THE WILSON CREEK MEMBER OF THE WOOD RIVER FORMATION

Measured southwest to northeast across ridge northeast of the head of Lake Creek; from 9,200 ft elevation across 10,200-foot ridge crest to 9,200-foot knob on ridge southwest of 9,677 ft, north of Basin Gulch; starts in NW¼ sec. 3, T. 5 N., R. 18 E. and continues into NE¼ sec. 34, T. 6 N., R. 18 E., Rock Roll Canyon quadrangle (Burton, 1988) (section W3, fig. 2).

	Thickness (meters)
Top of measured section: hinge of tight syncline	
Unit 7 of Hall and others (1974)	
Fine sandy micrite, medium-brown, weathering dark-brown, thin-bedded, partly silicified and dolomitic; arranged in fining-upward cyclic packets 8–17 m thick of massive siliceous micrite in lower 2–3 m, overlain by thin-bedded dolomicrite containing trough cross and convolute laminae, overlain by intensely bioturbated carbonaceous silty micrite	175
Sandy and coarse silty micrite, thin-bedded, siliceous, trough cross- and convolute-laminated; mottled gray orange to brown black; *Neonereites*	180
Allochemic sandy micrite, medium- to dark-gray; bioclasts include crinoid columnals, scaphopods?, cephalopods, bryozoa, fusulinids	30
Micritic mudshale, very dark brown, silty, bioturbated; *Neonereites, Phycosiphon, Paleophycus?*	20
Fine micritic sandstone (quartzarenite), medium-gray, weathering brown, calcareous, thick-bedded to massive	20
Interbedded coarse silty micrite and fine micritic sandstone, medium-gray to brown, silty micrite parts bioturbated	67
Micritic mudshale, very dark brown, silty, bioturbated; *Neonereites, Phycosiphon, Paleophycus?*	8
Silty micrite and very fine sandy micrite, medium-gray, weathering light-gray to yellow-brown; arranged in thinning- and fining-upward sequences (15–25 m thick) containing complete turbidites at the base passing upward to partial (base-cutout) T_{ace} and T_{cde} turbidites and to thin, very dark brown, intensely silicified T_{def} turbidites at the top; basal sequences are intensely bioturbated	252
Silty micrite and micritic sandstone, medium-gray to light-brown; medium beds contain T_{acde} silt turbidite sequences; *Neonereites, Scalarituba, Phycosiphon, Zoophycos, Planolites*	35
Fine sandy micrite, medium-gray, weathering medium-brown, thin- to medium-bedded, brown micritic mudstone partings	8
Fine micritic sandstone, medium-gray, thin-bedded, micritic mudstone partings, load casts, *Scalarituba*	5
Conformable contact, top of Eagle Creek Member (micritic sandstone, unit 6 of Hall and others, 1974)	
Total thickness Wilson Creek Member	800+

fold hinges. Silver-lead-zinc vein deposits are present locally along this sheared unconformity (Burton and Link, this volume). The upper contact of the Hailey Member is placed at the first appearance of the distinctive light-purple silty micrite of the Eagle Creek Member (table 2, fig. 6). The age of the Hailey Member is Desmoinesian, based on the biostratigraphy of coral, fusulinid, and phylloid green algae occurrences (Bostwick, 1955; Hall and others, 1974) (fig. 9).

The Eagle Creek Member of the Wood River Formation (upper Desmoinesian to Wolfcampian) is 880–1,300 m thick and consists of 260 m of light-purple silty micritic limestone overlain by 620–1,140 m of light-brown micritic sandstone, light-gray sandy micritic limestone, and subordinate quartzarenite (fig. 6). The Eagle Creek Member forms the bulk of the outcrop area of the Wood River Formation both east and west of the Wood River. It is best exposed on the high ridges of the Boulder Mountains (fig. 9).

About 250 m of anomalous carbonate-rich facies assigned to the Eagle Creek Member by Batatian (1991) crops out just east of the summit of Kent Peak. The rock is coarse-grained carbonaceous bioclastic limestone and resembles bioclastic beds in the Wilson Creek Member (David Seeland, U.S. Geological Survey, written commun., 1992). The unit yielded conodonts (collection 12, fig. 7) that indicate a Late Pennsylvanian to Early Permian age and macrofossils, including *Chaetetes* sp., that indicate a (Middle Pennsylvanian) Desmoinesian age (D.E. Fortsch, Idaho State University, oral commun., 1991). If the Desmoinesian age is correct, the beds may be the carbonaceous equivalent of limestone in the upper part of the Hailey Member. The facies resembles bioclastic parts of the Bloom Member of the Snaky Canyon Formation (Chesterian to Missourian) exposed 55 km to the east across several thrust faults in the Lost River Range (Skipp, Kuntz, and Morgan, 1979) (fig. 7).

The Wilson Creek Member (Wolfcampian to Leonardian) consists of more than 800 m of dark-gray carbonaceous siltstone and sandstone, thin-bedded light-brown graded silty micritic limestone, light-brown silty micritic limestone, light- to dark-gray sandy micritic limestone, and subordinate light-brown medium-bedded micritic sandstone

Table 3. Measured stratigraphic reference sections of the Pennsylvanian and Permian Sun Valley Group, south-central Idaho.
[Diagrammatic stratigraphic columns for these sections are shown in Mahoney and others (1991, figs. 4–6). Locations of sections listed below are shown in fig. 2)]

	Wood River Formation
Section W1	Murdock Creek (unconformable base of Hailey Member to hinge of syncline in upper part of Wilson Creek Member, described by Burton, 1988). Measured in northeast-dipping beds on the ridge between Murdock Creek and the East Fork of the North Fork of the Wood River, Amber Lakes 7.5-minute quadrangle, starting just southwest of peak 8,635 and proceeding northeastward along long ridge line to peak 9,783. NE¼ sec. 27, through sec. 23, to SW¼ sec. 13, T. 6 N. R. 17 E., Blaine County, Idaho. Base of section lat 43°40'23" N., long 114°24'22" W.
Section W2	Lake Creek section (unconformable base of Hailey Member to hinge of fold in lower part of Wilson Creek Member, described by Burton, 1988). Measured southwest to northeast along ridge between Lake and Eagle Creeks in northeast-dipping beds, Rock Roll Canyon 7.5-minute quadrangle; from 9,040 to 9,000 ft elevation, across peak 9,675; NW¼ sec. 8 to NE¼ sec. 5, T. 5 N., R. 18 E., Blaine County, Idaho. Base of section lat 43°46'56" N., long 114°20'34" W;
Section W3	Basin Gulch section; includes type section of Wilson Creek Member (unconformable base of Hailey Member to hinge of syncline in upper part of Wilson Creek Member, described by Burton, 1988). Measured southwest to northeast along ridge northeast of the head of Lake Creek in northeast-dipping beds, Rock Roll Canyon 7.5-minute quadrangle; measured from 7,960 ft elevation to peak 9,677, across 10,200 ft ridge crest; SW¼ sec. 3, T. 6 N., R. 18 E. and NE¼ sec. 34, T. 6 N., R. 18 E., Blaine County and Custer Counties, Idaho. Base of section lat 43°47'58" N., long 114°18'45" W.
Section W4	Wilson Creek Ridge; includes type section of Eagle Creek Member (unconformable base of Hailey Member to hinge of syncline in upper part of Eagle Creek Member, described by Burton, 1988). Measured from southwest to northeast on the ridge between Wilson Creek and Trail Creek in northeast-dipping beds, Phi Kappa Mountain and Rock Roll Canyon 7.5-minute quadrangles. Hailey Member measured about 500 m east of the base of the ridge on the west bank of Wilson Creek, NW¼ sec. 13, T. 5 N., R. 19 E. Eagle Creek and Wilson Creek Members measured along the ridge from 6,600 ft elevation to swale at 9,000 ft elevation in NE¼ sec. 14 and SW¼ sec. 12, T. 5 N., R. 18 E. Blaine County, Idaho. Base of section lat 43°46'13" N., long 114°16'18" W.
Section W5	Bellevue (principal reference section of the Wood River Formation and type section of Hailey Member), unconformable base of Hailey Member to topographic surface in upper part of Wilson Creek Member, described by Hall and others (1974). Composite section; units 1–6 (Hailey and Eagle Creek Members) measured southwest to northeast along ridge north of Seamans Creek, Seamans Creek 7.5-minute quadrangle, from 6,390 ft elevation to top of ridge; base of section lat 43°29'30" N., long 114°14'00" W.; unit 7 (Wilson Creek Member) measured on Quigley Creek in sec. 21, T. 3 N., R. 19 E., starting 2,300 ft north of the southwest corner of section 21 and measured toward the east up the ridge from 6,640 to 7,300 ft elevation, base of section lat 43°34'30" N., long 114°12'30" W. Blaine County, Idaho.
	Grand Prize Formation
Section G1	Hoodoo Mine section (lower part of member 2 to lower part of member 3). Measured on east flank of peak 10,050, west of Hoodoo Lake in west-dipping beds; section measured from 8,820 to 10,050 ft, Robinson Bar 7.5-minute quadrangle, Custer County, Idaho. Base of section lat 44°10'35" N., long 114°38'15" W.
Section G2	Fourth of July Creek section (composite reference section, unconformable base of member 1 to topographic surface in upper part of member 4; also shown in Mahoney and Sengebush, 1988). Composite section. Interval from 0 to 500 m measured in east-dipping beds, Strawberry Basin, north of Blackman Peak, from 9,800 to 10,111 ft elevation, Washington Peak 7.5-minute quadrangle; base of section lat 44°04'28"N, long 114°39'32" W. Interval from 550 to 2,500 m measured from west to east in steeply west dipping overturned beds on ridge north of Fourth of July Creek, from 7,800 to 9,200 ft elevation, Obsidian and Washington Peak 7.5-minute quadrangles, Custer County, Idaho. Base of section lat 44°03'18"N, long 114°43'48" W.
Section G3	Washington Basin section (sheared base of member 1 to topographic surface in lower part of member 3). Measured from east to west in steeply west dipping beds on east flank of peak 10,519, section starts at 10,200 ft and goes along sawtooth ridge to top of peak 10,519, Washington Peak quadrangle, Custer County, Idaho. Base of section lat 44°15'00", long 114°39'38".

Table 3. Measured stratigraphic reference sections of the Pennsylvanian and Permian Sun Valley Group, south-central Idaho—Continued.

	Grand Prize Formation—Continued
Section G4	Champion Lakes section (lower part of member 2 to topographic surface in upper part of member 4). Measured on ridge south of Champion Lakes Basin, measured from east to west in steeply west dipping beds; section starts at 9,920 ft, southeast of peak 10,167, and continues along ridge to west side of Champion Lakes Basin, to north of peak 10,081, Horton Peak and Washington Peak 7.5-minute quadrangles, Custer County, Idaho. Base of section lat 43°58'53"N, long 114°40'00" W.
Section G5	Contains two sections (G5a and G5b of Mahoney and others, 1991) repeated across normal fault and containing parts of members 3 and 4. Comprises the type section of Grand Prize Formation of Hall (1985). Measured in northwest-dipping beds on north side of Pole Creek, near its confluence with Grand Prize Gulch, on south flank of peak 10,166, from 7,970 to 9,560 ft elevation, Horton Peak 7.5-minute quadrangle, Custer County, Idaho. Base of section lat 43°56'27" N., long 114°41'00" W.
Section G5b	Pole Creek section 2, separated from Pole Creek section 1 by low-angle fault. Type section of Hall (1985) crosses this fault. Section continues from 9,560 to 10,166 ft elevation above section G5a, Horton Peak 7.5-minute quadrangle, Custer County, Idaho. Base of section lat 43°56'28" N., long 114°41'00" W.
Section G6	Salmon River Headwaters section (partial sections of members 2 and 3, described in Mahoney, 1987). Measured on west side of Salmon River, on east flank of peak 9,423 in northwest-dipping beds, from 7,800 to 9,423 ft elevation, Frenchman Creek 7.5-minute quadrangle, Camas County, Idaho. Base of section lat 43°48'45" N., long 114°46'33" W.
	Dollarhide Formation
Section D1	Willow Creek section (principal reference section, intrusive contact at exposed base of lower member to topographic surface in upper part of upper member) described by Geslin (1986) and O'Brien (1991). Measured from southwest to northeast in northeast-dipping beds along ridge due west of Cup knob (benchmark 9,147); lower part measured from 7,245 ft (contact with Idaho batholith) to fault at knob 7,856, S½ sec. 36, T. 3 N., R. 15 E.; base of section lat 43°32'49" N., long 114°37'00" W. Upper part of section measured 0.6 mi to the north of peak 7,856, from elevation 7,000 ft in creek bottom to knob 7,852, NE¼ sec. 36 to SE¼ sec. 25, T. 3 N., R. 15 E. (O'Brien, 1991), Buttercup Mountain 7.5-minute quadrangle, Camas County, Idaho. Base of section lat 43°33'35" N., long 114°37'25" W. Lower section may be continued above knob 7,856 into the middle and upper members up to knob "Cup" 9,147, although it is structurally thickened.
Section D2	Dry Gulch–Bear Gulch section (intrusive contact in upper part of lower member to axis of syncline in upper member, described by O'Brien, 1991). Composite section. Lower 400 m measured on north slope of Dry Gulch, in northeast-dipping beds from contact with Cretaceous stock at 7,000–8,380 ft elevation; NE¼ sec. 3, NW¼ sec. 4, T. 2 N., R. 16 E. and SW¼ sec. 35, T. 3 N., R. 16 E., Buttercup Mountain 7.5-minute quadrangle, Blaine County, Idaho. Base of section lat 43°32'25" N., long 114°31'35" W. Upper 900 m measured in west-dipping beds in the central headwaters of Bear Gulch, on the east side of knob "Cup" (9,147) from 7,400 ft elevation to core of syncline at 8,900 ft, N½2 sec. 32, T. 3 N., R. 16 E. Base of section lat 43°34'38" N., long 114°32'50" W.
Section D3	Wolf Tone Creek section (covered interval in lower part of lower member to poorly exposed interval near top of middle member, described by Wavra and others, 1986). Measured along northwest side of Wolf Tone Creek in southwest-dipping beds. Section starts at stream fork at 6,058 ft and proceeds upstream to 6,280 ft, opposite prominent stream entering Wolf Tone Creek from the south; SW¼ sec. 8, SE¼ sec. 7, and NE¼ sec. 18, T. 2 N., R. 17 E., Mahoney Butte 7.5-minute quadrangle, Blaine County, Idaho. Base of section lat 43°31'25" N., long 114°27'33" W.

(fig. 6). The member forms slopes and is sparsely exposed east of Bellevue and Hailey, although it is well exposed in alpine ridges of the Boulder Mountains northeast of Ketchum (figs. 1, 9, plate 1). The fine-grained, thin-bedded silty micritic limestone of the Wilson Creek Member weathers to a characteristic reddish-brown regolith. The upper part of the member is dolomitic in part and contains diagenetic chert. The Wilson Creek Member contains a diverse assemblage of bathyal trace fossils that has been interpreted to reflect oxygenation levels controlled by turbidite depositional facies (Burton and Link, 1991). The upper contact of the Wilson Creek Member is eroded below an unconformity with the Eocene Challis Volcanic Group or the present-day topographic surface (fig. 8).

MINERAL DEPOSITS IN THE WOOD RIVER FORMATION

A variety of mineral deposit types are present in the Wood River Formation. The Hailey Member hosts zinc-bearing veins in the Lake Creek mineralized area (Burton and Link, this volume) (fig. 1). Brittle sandstone of the Eagle Creek Member hosts vein deposits in the Boulder Basin (Ratchford, 1989), Galena, and Lake Creek areas (fig. 1). Limestone of the Eagle Creek Member hosts replacement deposits in the Wood River area (fig. 1). Carbonaceous siltstone of the Wilson Creek Member hosts shear-zone vein deposits in the Lake Creek area (Burton and Link, this volume) (fig. 1).

DOLLARHIDE FORMATION

The Dollarhide Formation comprises the southern and southwestern exposures of the Sun Valley Group. Complexly folded Dollarhide Formation is intruded by the Idaho batholith and associated stocks in the Smoky Mountains, west of the Wood River (Whitman, 1990, this volume; Darling and others, this volume) (fig. 1, plate 1). The formation was named by Hall (1985) for a sequence of dark-colored ("sooty") carbonaceous limestone, siltite, fine-grained quartzite, and granule conglomerate exposed on Dollarhide Summit, west of Ketchum (plate 1), but a type section was not designated. Mahoney and others (1991) and O'Brien (1991), following Wavra (1985), Geslin (1986), and Wavra and others (1986), recognized three informal members and designated the principal reference section for the Dollarhide Formation to be east of Willow Creek (section D1, plate 1). Other reference sections for the Dollarhide Formation are south of Wolf Tone Creek (lower and middle members, section D3, plate 1) and at the head of Deer Creek north of Buttercup Mountain (middle and upper members, section D2, plate 1).

Skipp and Hall (1980), Geslin (1986), Darling (1988), and Link and others (1988) described the metasedimentary "Carrietown sequence" as being in thrust contact below the lower member of the Dollarhide Formation in the Smoky Mountains near the edge of the Idaho batholith (fig. 1, plate 1). Whitman (1990, this volume) recognized that this contact is not structural but is a lithologic change accentuated by metamorphism. These metasedimentary rocks are now interpreted as foliated parts of the lower member of the Dollarhide Formation, marginal to the Idaho batholith (Whitman, 1990, this volume; Mahoney and others, 1991; Darling and others, this volume). The name "Carrietown sequence" is abandoned.

The lower member of the Dollarhide Formation (Desmoinesian to lower Wolfcampian) is at least 800 m thick and consists of rhythmically interbedded dark-gray carbonaceous silty micritic limestone and very fine grained light-gray to light-brown micritic sandstone and subordinate medium- to thick-bedded light-brown micritic sandstone and light- to dark-gray lenticular conglomerate containing mainly intrabasinal clasts of siltstone and bioclastic limestone. Extrabasinal clasts (vitreous quartzite and fine-grained sandstone) are rare. The lower member of the Dollarhide Formation includes at the base a minimum of 175 m of quartzite, phyllite, and calc-silicate hornfels formerly assigned to the "Carrietown sequence." The lower member is exposed in the southern Smoky Mountains, where it forms dark-colored slopes punctuated by thin ridges of resistant micritic sandstone. The base of the lower member either is a fault or is masked by the Idaho batholith, and the upper contact is placed above a mappable dark-gray carbonaceous limestone and below thick light-brown micritic sandstone of the middle member (O'Brien, 1991) (fig. 6).

The middle member of the Dollarhide Formation (lower Wolfcampian) consists of about 300 m of fine-grained light-brown micritic sandstone and light-gray sandy micritic limestone and subordinate dark-gray to black carbonaceous siltstone and lenticular conglomerate (fig. 6). The middle member forms prominent light-colored cliffs throughout the central and southern Smoky Mountains. North of Deer Creek (plate 1) it is interpreted to grade into thick sandstone beds of the Eagle Creek Member of the Wood River Formation.

The upper member of the Dollarhide Formation (Wolfcampian and Leonardian) is approximately 900 m thick and is composed of thin-bedded to laminated dark-gray carbonaceous siltstone and light-gray silty micritic limestone and minor light-brown micritic sandstone and conglomerate (fig. 6). The upper member gradationally overlies the middle member, and the upper boundary is not recognized due to erosion. The upper member weathers to a dark-gray regolith, is poorly exposed, and forms slopes throughout the Smoky Mountains, which, as a result, contain large areas of sparsely vegetated "black-shale" regolith (Wavra and Hall, 1989) (fig. 1, plate 1).

MINERAL DEPOSITS IN THE DOLLARHIDE FORMATION

The lower member of the Dollarhide Formation hosts vein and replacement mineral deposits in the Bullion, Carrietown, and Deer Creek stock mineralized areas (fig. 1) and a laminated barite deposit in the northern part of the Bullion area. Sheared vein and replacement mineral deposits are present in the upper member of the Dollarhide Formation in the Bunker Hill and Carrietown mineralized areas (fig. 1). The middle member of the Dollarhide Formation, dominated by sandstone or quartzite, does not host mineral deposits.

Wavra and Hall (1989) designated most dark-colored outcrops of Paleozoic rocks in the southern Smoky Mountains as upper member, Dollarhide Formation (including rocks we map as Milligen Formation in the Minnie Moore mine area) and presented a model whereby the upper member is host to syngenetic mineral deposits including the Deer Creek barite deposit (plate 1). Recent work based on measured sections of the Dollarhide Formation and recognition of distinctive members of the Milligen Formation at the Minnie Moore mine (O'Brien, 1991; Worl and others, 1991; O'Brien and others, 1993; Skipp and others, 1994; Link and Worl, in press) suggests that restriction of mineral deposits to the upper member of the Dollarhide Formation is incorrect. Mineral deposits in the southern Smoky Mountains are hosted by the Milligen Formation, as well as by the lower and upper members of the Dollarhide Formation.

GRAND PRIZE FORMATION

The Grand Prize Formation comprises the northern exposures of the Sun Valley Group and is exposed primarily north of Pole Creek in the White Cloud Peaks area (fig. 1, plate 1). The formation was named by Hall (1985) for a 1,450-meter-thick sequence of light-brown fine-grained quartzite, brown limy siltstone, banded light- and dark-gray siltite, and dark-gray to black carbonaceous silty limestone exposed north of Grand Prize Gulch, a tributary to Pole Creek (section G5, plate 1). Rocks of the Grand Prize Formation had previously been included in the Pole Creek sequence (Skipp and Hall, 1980; Fisher and others, 1983; Sengebush, 1984) or the Wood River Formation (Tschanz and others, 1986). Mahoney and Sengebush (1988) expanded and revised Hall's definition of the Grand Prize Formation, subdivided the formation into four informal members, and designated a composite reference section for the formation north of Fourth of July Creek (section G2, fig. 6, plate 1). Mahoney and others (1991) slightly modified member boundaries within the Grand Prize Formation and described additional reference sections (plate 1) for the formation near the Hoodoo mine (section G1), west of Washington Basin (section G3), above Champion Lakes north of Pole Creek (section G4), and at the headwaters of the Salmon River (section G6).

Member 1 (Desmoinesian) of the Grand Prize Formation is 0–400 m thick and consists of 0–350 m of light-brown to light-gray polymict conglomerate and sandstone overlain by 30–50 m of medium-gray bioclastic limestone (fig. 6). Member 1 is resistant and forms conspicuous cliffs and ledges. The lower contact of member 1 is an erosional unconformity with the underlying Paleozoic Salmon River assemblage. This contact was strongly sheared during Mesozoic compression, and the original stratigraphic thickness of the basal conglomerate is generally not preserved.

Member 2 (Desmoinesian to Wolfcampian) of the Grand Prize Formation consists of 500–1,100 m of thick-bedded to massive light-brown micritic sandstone and subordinate light-gray sandy micritic limestone and gray carbonaceous siltstone. Member 2 gradationally overlies member 1; the contact is placed above the bioclastic limestone of member 1. Member 2 forms discontinuous cliffs and ledges in the northern Smoky Mountains and in the White Cloud Peaks. Member 3 (Wolfcampian to Leonardian) of the Grand Prize Formation consists of 650–1,700 m of fine-grained sandstone and siltstone arranged in rhythmically interbedded couplets 30 cm–3 m thick. The distinctive banded appearance of the member is produced by these couplets of light-gray fine-grained micritic sandstone gradationally overlain by dark-gray carbonaceous siltstone. Interbedded with the couplets is thick-bedded light-brown micritic sandstone and light-gray sandy micritic limestone (fig. 6). Member 3 gradationally overlies member 2. Member 3 has an irregular weathering pattern; the sandier intervals are exposed in bold relief against the more easily weathered finer grained intervals. The member forms distinctly banded light-gray cliffs throughout the White Cloud Peaks area.

Member 4 (Wolfcampian to Leonardian) of the Grand Prize Formation consists of more than 450 m of thin- to medium-bedded carbonaceous siltstone, sandy to silty micrite, and minor micritic sandstone (fig. 6). Carbonaceous siltstone of member 4 gradationally overlies banded sandstone and siltstone of member 3. Member 4 weathers to a dark-gray regolith and forms slopes in the southern part of the White Cloud Peaks area.

MINERAL DEPOSITS IN THE GRAND PRIZE FORMATION

Members 1, 2, and 3 of the Grand Prize Formation host polymetallic vein deposits in Washington Basin (fig. 1). Member 3 hosts a tungsten skarn deposit in the Smiley Creek drainage of the Smoky Mountains mineralized area (fig. 1). Micritic sandstone and sandy micrite of members 2 and 3 host replacement deposits in the Pole Creek and Washington Basin areas (fig. 1). Limestones of member 4 also have potential for replacement deposits in the Galena area.

RELATIONS BETWEEN FORMATIONS OF THE SUN VALLEY GROUP

Our work suggests that the formations of the Sun Valley Group in most places grade laterally and vertically into each other by facies change (fig. 8, plate 1) rather than occupying thrust-fault bounded allochthons. We recognize that, in places, contacts shown as facies changes on figure 2 may appear on the ground to arbitrarily punctuate gradual transitions in color, grain size, or type of cement. Specific relationships between formations are outlined following (after Mahoney and others, 1991).

The Dollarhide Formation-Wood River Formation contact may be observed at five locations:

1. In the Picabo Hills, dark-gray limestone and siltstone of the upper part of the lower member of the Dollarhide Formation grade eastward over a distance of about 2 km into light-gray silty limestone of the Eagle Creek Member of the Wood River Formation, across a set of southwest-vergent folds.
2. Along Deer Creek west of Hailey (Skipp and others, 1994), dark siltstone and limestone of the lower member of the Dollarhide Formation are thrust over light-brown sandstone and limestone of the Eagle Creek Member along the Deer Creek thrust fault.
3. On Kelly Mountain, at the head of Wolf Tone Creek, a section of silicified quartzarenite and light-gray silty limestone of the Eagle Creek Member of the Wood River Formation, several hundreds of meters thick, lies above dark- and light-gray siltstone and sandstone of the middle and lower members of the Dollarhide Formation on a low-angle normal fault.
4. North of Deer Creek, the middle member of the Dollarhide Formation is mapped south of the Challis Volcanic Group near Mahoney Butte, but the Eagle Creek Member of the Wood River Formation is mapped in probably continuous strata to the north. In this area, the middle member of the Dollarhide Formation is a tongue of partially silicified micritic sandstone, petrographically identical to micritic sandstone of the Eagle Creek Member (fig. 8).
5. North of Baker Creek, light-brown calcareous sandstone of the Eagle Creek Member overlies dark-gray siltstone and limestone of the lower member of the Dollarhide Formation along a low-angle normal fault (Stewart, 1987; Stewart and others, 1992).

The Wood River Formation grades northward into the Grand Prize Formation. This contact may be observed in two places and is shown as a line of facies change on plate 1.

1. North of Pole Creek, massive micritic sandstone of the Eagle Creek Member intertongues with thick-bedded micritic sandstone assigned to member 2 of the Grand Prize Formation. This relation was previously shown as a thrust fault by Hall (1985).
2. In Galena Gulch, medium-bedded micritic sandstone, thin-bedded sandy micritic limestone, and silty micritic limestone of the Wilson Creek Member of the Wood River Formation are gradationally overlain by black carbonaceous siltstone and sandy to silty micrite assigned to member 4 of the Grand Prize Formation (fig. 8).

The contact between the upper member of the Dollarhide Formation and member 4 of the Grand Prize Formation is obscured by an Eocene dacite porphyry intrusive body west of the headwaters of the Salmon River (Mahoney, 1992) (plate 1). The upper member of the Dollarhide Formation and member 4 of the Grand Prize Formation are lithologically identical carbonaceous siltstone and are interpreted as lateral equivalents (fig. 8).

MINERAL DEPOSIT MODELS APPLICABLE TO PALEOZOIC ROCKS OF THE EASTERN PART OF THE HAILEY 1°×2° QUADRANGLE

GEOLOGIC COMPLEXITY

The deformational and magmatic history of the eastern part of the Hailey 1°×2° quadrangle is complex (Rodgers and others, this volume). Mineral deposits hosted by Paleozoic strata are of three distinct ages: (1) Devonian syngenetic stratabound deposits, (2) Cretaceous polymetallic veins and skarns associated with intrusion of the Idaho batholith, and (3) Tertiary polymetallic veins and skarn deposits associated with the Challis magmatic episode.

Probable syngenetic metal deposits are in the Devonian Milligen Formation, the Salmon River assemblage, and unnamed Silurian and Devonian strata east of the Boulder Mountains (Hall, 1985, 1987b; Turner and Otto, 1988, this volume; Bruner, 1991; Sanford and Wooden, this volume). These deposits may have formed by hydrothermal circulation systems and normal faults that developed during early phases of the Devonian and Mississippian Antler orogeny (Turner and Otto, this volume).

Magmatically driven hydrothermal systems were active in Late Cretaceous and Eocene time. These hydrothermal systems caused epigenetic recrystallization and remobilization of metals first concentrated in Devonian time. They also tapped new sources of metals in Precambrian granitic basement (Sanford and Wooden, this volume). Several generations of Mesozoic and Tertiary faults cut Paleozoic strata, and ore deposits in almost all districts are cut by postmineralization faults.

MINERAL DEPOSIT MODELS

Several of the mineral deposit models compiled in Cox and Singer (1986) and modified by Worl and Johnson (this volume) are applicable to the central Idaho black-shale mineral belt. The following discussion, although neither a rigorous coverage of the models nor a mine-by-mine review of each mineral district, is meant to summarize present understanding. Complete descriptions of mines and production are contained in Lindgren (1900), Umpleby (1915), Umpleby and others (1930), Anderson and others (1950), Tuchek and Ridenour (1981), Van Noy and others (1986), McIntyre (1985), and Fisher and Johnson (1987).

SKARN DEPOSITS

Skarns have not been major ore producers in the Wood River area, but gold, lead-zinc, and tungsten skarn deposits are present in several places where Paleozoic limestone is intruded by Tertiary or Cretaceous stocks.

At the June Day mine, in Parker Gulch of the Triumph mineralized area (fig. 1), southeast of Ketchum, a gold-bearing skarn is developed in the limestone member of Lucky Coin in the Milligen Formation. The silicified limestone "tactite" contains diopside, garnet, wollastonite, scapolite, clinozoisite, zoisite, and vesuvianite and hosts disseminated gold (Anderson and others, 1950). The intrusive rock is an Eocene dacite porphyry stock.

The Phi Kappa mine of the Summit mineralized area (fig. 1) contains zinc-silver skarn deposits in the Drummond Mine Limestone Member of the Copper Basin Formation (Winkler and others, this volume). Skarn minerals include diopside and grossularite. Sulfide minerals, present in tactite zones, include, in order of decreasing abundance, galena, sphalerite, pyrite, and chalcopyrite (Tuchek and Ridenour, 1981). These skarns are associated with granodiorite of the Eocene Summit Creek stock, which crops out to the west across Phi Kappa Creek.

Tungsten skarn deposits hosted by members 2 and 3 of the Grand Prize Formation are in the Ura group prospects on Smiley Creek (Mahoney and Horn, in press) (fig. 1). Tactite is composed of red garnet, epidote, wollastonite, quartz, and fine-grained scheelite (Van Noy and others, 1986).

Gold-bearing skarns may be present, associated with gold-bearing veins and gossan, in the Black Rock claims of the Washington Basin area (Van Noy and others, 1986) (fig. 1). Granodiorite, of probable Cretaceous age, intrudes limestone, micritic sandstone, and sandy micrite of Grand Prize Formation members 1, 2, and 3 near these claims (Mahoney, this volume).

Skarns are present west of Hailey and Bellevue in the Bunker Hill, Deer Creek stock, and Rooks Creek stock mineralized areas (fig. 1). Silver-lead-zinc deposits in these areas formed during intrusion of Late Cretaceous (about 90 Ma) potassium-rich plutons (Rooks Creek stock, Deer Creek stock, Croesus stock (fig. 1, plate 1) (Link and Worl, in press; Park, in press). These plutons were emplaced east of, and earlier than, the main Atlanta lobe of the Idaho batholith to the west (Lewis, 1989). Tactite in contact metamorphic deposits in the Wood River and Dollarhide Formations contains argentiferous galena in a gangue of grossularite and epidote (Lindgren, 1900; Umpleby and others, 1930; Anderson and others, 1950). Skarns in the Rooks Creek stock area contain disseminated gold and sphalerite in a gangue of garnet and pyroxene (Anderson and others, 1950; Park, in press).

POLYMETALLIC REPLACEMENT DEPOSITS

Calcareous strata in several formations are hosts for replacement deposits. The limestone member of Lucky Coin (Milligen Formation) hosts polymetallic replacement ore in the Triumph and Minnie Moore mineralized areas (fig. 1) (Turner and Otto, this volume; Link and Worl, in press). These lode deposits contain sphalerite, galena, tetrahedrite, and chalcopyrite. They are localized in calcareous beds suitable for replacement and proximal to a fault that acted as a pathway for mineralizing fluids (Kiilsgaard, 1950).

The Dollarhide Formation hosts replacement deposits in several areas west of the Wood River. In the Carrietown mineralized area (fig. 1), the host rock is limestone of the lower member of the Dollarhide Formation (Darling and others, this volume). In the Bunker Hill mineralized area (fig. 1) the host is limestone of the upper member of the Dollarhide Formation. Ore minerals in these areas, both in replacement and vein deposits, are galena, sphalerite, tetrahedrite, pyrite, arsenopyrite, pyrrhotite, and chalcopyrite.

In the Boulder Basin mineralized area on the west side of the Boulder Mountains (fig. 1), the replaced host is calcareous sandstone or quartzite of the Eagle Creek Member of the Wood River Formation. Intrusion of Eocene dikes probably generated ore-bearing circulation systems that deposited pyrite, galena, tetrahedrite, chalcopyrite, and sphalerite (Van Noy and others, 1986; Ratchford, 1989, in press).

In the northern part of the Galena mineralized area, along Pole Creek (fig. 1), calcareous sandstone of member 3 of the Grand Prize Formation is replaced. Gossan provides an exploration guide. Ore minerals include sphalerite, galena, and chalcopyrite (Van Noy and others, 1986). Micritic sandstone and sandy micrite of members 2 and 3 of the Grand Prize Formation host replacement deposits containing detrital quartz grains within a fine-grained to aphanitic galena matrix in the Washington Basin area (Mahoney, this volume) (fig. 1).

Replacement deposits may be present in any calcareous stratum. Thus, such deposits potentially could be present in the limestone member of Lucky Coin of the Milligen Formation, the lower member of the Dollarhide Formation, Grand Prize Formation members 1 to 4, the Hailey, Eagle Creek and Wilson Creek Members of the Wood River Formation, and the Drummond Mine Limestone Member of the Copper Basin Formation.

POLYMETALLIC VEINS

Polymetallic vein mineral deposits in the eastern part of the Hailey quadrangle are commonly associated with, and locally difficult to distinguish from, polymetallic replacement deposits. Vein deposits have been the largest producers and are present throughout the black-shale mineral belt. Generally, veins hosted by carbonaceous siltstone and argillite ("black shale") have siderite gangue, whereas veins in sandstone and quartzite host rock have quartz gangue. Generally, wallrock is not much altered, although some bleaching occurred with formation of sericite and addition of carbonate and pyrite.

CRETACEOUS VEIN DEPOSITS

In general, shear-zone-hosted veins of the Ketchum-Hailey-Bellevue area follow fault systems or shear zones of Cretaceous age and are associated with Cretaceous plutonic activity. The polymetallic veins west of the Wood River were formed by hydrothermal systems driven by heat from the Late Cretaceous plutons that are border phases of the Idaho batholith (Lewis, 1989). These hydrothermal systems derived metals from remobilization of Devonian stratabound syngenetic mineral deposits and from underlying Precambrian crust (Sanford and Wooden, this volume).

In the East Fork Wood River, Bellevue, Minnie Moore, and Triumph mineralized areas (fig. 1), silver-lead-zinc vein deposits of probable Cretaceous age are present in black argillite of the argillite member of Triumph and limestone member of Lucky Coin of the Devonian Milligen Formation (Turner and Otto, 1988, this volume). In the Minnie Moore mineralized area, including the Minnie Moore and Silver Star Queens mines, shear-zone-hosted veins contain galena, pyrite, sphalerite, chalcopyrite, and arsenopyrite in a gangue of siderite, quartz, calcite, and crushed country rock (Anderson and others, 1950). These veins occupy northwest-striking reverse faults or shear zones that apparently formed shortly after intrusion of the adjacent Croesus quartz diorite stock about 90 Ma (Link and Worl, in press).

In the Bullion mineralized area (fig. 1), including the Mayflower and Red Elephant mines, northwest-striking Late Cretaceous veins are hosted by the lower member of the Dollarhide Formation (Link and Worl, in press). The most common ore minerals, in order of increasing abundance, are galena, sphalerite, and tetrahedrite. The most common gangue minerals are siderite, calcite, and quartz. Siderite gangue is associated with ore deposition (Fryklund, 1950).

Many of the areas listed above as containing replacement deposits also contain vein deposits. In general, ore minerals in veins are argentiferous galena, more or less abundant sphalerite and tetrahedrite, and subordinate and variable amounts of pyrite, arsenopyrite, and chalcopyrite (Umpleby and others, 1930; Anderson and others, 1950). The gangue consists of altered and crushed country rock, siderite, calcite, and quartz.

Veins are hosted by both the upper and lower members of the Dollarhide Formation in the Carrietown mineralized area and by the upper member of the Dollarhide Formation in the Bunker Hill mineralized area (fig. 1.

In Washington Basin (fig. 1), polymetallic veins of probable Cretaceous age are present in northeast-trending shear zones or fractures localized along the axial plane of an east-vergent anticline (Mahoney, this volume).

EOCENE VEIN DEPOSITS

North and east of Ketchum, vein deposits of the Smoky Mountains, Galena, Lake Creek, and Summit mineralized areas (fig. 1) probably formed during Eocene plutonism. In the Galena district, mineralized silver-lead-zinc veins follow Tertiary shear zones that strike northeast and dip northwest. In the Lake Creek area, veins are within northwest-striking normal faults in the upper plate of a Paleogene detachment fault (Burton and Link, this volume). In both cases associated igneous rocks are dacite and rhyolite porphyry of Eocene age.

The ore-bearing vein and associated replacement deposits of the Livingston mine on Big Boulder Creek in the Challis 1°×2° quadrangle (plate 1) contain metals likely remobilized from syngenetic stratabound minerals first deposited in the Salmon River assemblage during Devonian time (Hall, 1985). Eocene rhyolite porphyry dikes are associated with the veins. Ore minerals are jamesonite, galena, sphalerite, and tetrahedrite (Van Noy and others, 1986). The gangue is quartz that is locally iron stained.

SYNGENETIC STRATABOUND SILVER-LEAD-ZINC DEPOSITS

An ultimate syngenetic origin for many of the ore metals in the Wood River valley is suggested by location of mines in and near probable source rock of the Devonian Milligen Formation and correlative units, as well as by ore textures and isotopic data (Hall and others, 1972; Hall, 1985; Howe and Hall, 1985). Sanford and Wooden (this volume) use lead-isotope data to suggest that the lead in mineral deposits of the Hailey 1°×2° quadrangle was supplied both from syngenetic stratabound deposits within the Milligen Formation and Salmon River assemblage and from Precambrian crust, but not from detrital lead in Paleozoic black shale.

DEVONIAN MILLIGEN FORMATION

Turner and Otto (this volume) describe mineral deposits from the Milligen Formation in the Triumph mineralized area (fig. 1), which is the best-documented example of syngenetic stratabound mineralization in the black-shale mineral belt. The Triumph stratiform sulfide bodies are within or at the top of the limestone member of Lucky Coin and are associated with limestone turbidite and diamictite. Hall (1985, p. 130) described the stratabound syngenetic ore from the Triumph mine as "fine-grained, thinly-bedded, dark-gray, carbonaceous, micritic, silty limestone containing minute grains of sphalerite, galena, arsenopyrite, and pyrite disseminated along the bedding." The complex intrusive and deformational history of the region has produced a variety of epigenetic mineral deposits by remobilization of these stratabound Devonian sulfide deposits.

SALMON RIVER ASSEMBLAGE

Carbonaceous units of the Salmon River assemblage contain as much as 10,000 ppm V. At the Hoodoo mine on Slate Creek stratabound syngenetic deposits of zinc and barite are present in these rocks (Hall, 1985; Van Noy and others, 1986). The extensive workings follow a mineralized zone along the contact between Salmon River assemblage and member 1 of the Grand Prize Formation (designated as Hailey Conglomerate Member of the Wood River Formation by Van Noy and others, 1986, p. 262). Sphalerite and minor galena form concordant massive lenses from 0.3 to 12 m in thickness. Disseminations and veinlets in brecciated argillite, limestone, and quartzite are cemented by quartz and calcite. Smithsonite is the most abundant secondary mineral. Hall (1987b) speculated that during the time of sedimentation the belt of Salmon River assemblage now exposed along Slate Creek was downdropped and zinc-rich hydrothermal brines were concentrated there. Sphalerite and some galena and pyrite were deposited from these brines.

At several locations, original stratabound syngenetic sulfide deposits within the Salmon River assemblage are thought to have been recrystallized and remobilized by Cretaceous granodiorite plutons (Hall and Hobbs, 1987). Ore at the Livingston mine contains highly deformed black siliceous argillite containing bedded sphalerite, pyrite, and minor galena and jamesonite (Hall, 1985). These minerals are interpreted, based on light-stable-isotope characteristics, to have been syngenetic sulfides deposited with the black argillite (Howe and Hall, 1985). The rock was subsequently shattered and minerals remobilized during emplacement of Eocene rhyolite porphyry dikes. The minerals now occupy small veinlets and matrix between argillite fragments.

ORE TEXTURES IN SILURIAN AND DEVONIAN STRATA, MILLER CANYON AREA

Syngenetic textures suggesting sedimentary exhalative (SEDEX) mineralization are present in diamictite within the lower argillite unit (unit DSal–2 of figs. 3, 4), the unnamed Silurian and Devonian rocks in the Miller Canyon area (Bruner, 1991).

Two types of pyrite were identified in polished sections: (1) dark, brassy, spongy looking pyrite having colloform, orbicular, and framboidal textures and (2) bright, subhedral to euhedral pyrite. The first type of pyrite is thought to be syngenetic because it is present (1) as concentrations in the matrix of breccias and (2) within broken argillite clasts. Its dark, spongy character is attributed to poor crystal development due to rapid growth in open spaces. Such textures are produced by syngenetic precipitates on the ocean floor or in epigenetic veinlets. The presence of these textures in sulfide-bearing argillite clasts in diamictite is strong indication of their syngenetic origin.

The second type of pyrite is thought to be epigenetic because it is present (1) as overgrowths on colloform and framboidal pyrite, (2) in massive occurrences lacking open-space filling textures, and (3) within calcite and quartz veinlets that crosscut all other fabrics in polished section.

Sphalerite, the second most abundant sulfide mineral, forms (1) colloform bands within colloform pyrite and galena and (2) reddish-brown blebs as much as 1.5 cm across within calcite veinlets. Micron-size chalcopyrite blebs are common within the sphalerite and are probably hydrothermal in origin, perhaps remobilized from syngenetic mineral deposits.

Galena, the third most abundant sulfide mineral, forms (1) colloform bands with spongy pyrite and (or) sphalerite in orbicular masses in sedimentary breccia clasts, (2) concordant and discordant veinlets, and (3) isolated anhedral blebs. Large blebs and veinlets of galena have curved cleavage planes that provide evidence of deformation. Deformed galena may have a magmatically driven hydrothermal origin, likely remobilized from syngenetic mineralization.

Pyrrhotite, the fourth most abundant sulfide mineral, forms (1) microscopic veinlets with quartz and (2) small blebs within quartz veinlets. Veinlets containing pyrrhotite crosscut all other sulfide mineral occurrences, indicating that it has a probable magmatically driven hydrothermal origin.

POSSIBLE SYNGENETIC MINERAL DEPOSITS IN THE DOLLARHIDE FORMATION

Hall (1985) and Wavra and Hall (1989) suggested that the presence of laminated barite in the northern part of the Bullion mineralized area (north of Deer Creek) indicates that syngenetic shale-hosted precious-metal deposits are present in the upper member of the Dollarhide Formation. Geologic mapping (Skipp and others, 1994) suggests that the Deer Creek barite deposit contains a lensoidal body of syngenetic or remobilized-syngenetic barite hosted by the lower member of the Dollarhide Formation. Sulfate analysis of the barite (Howe and Hall, 1985) yields a value close to Permian and Pennsylvanian seawater sulfate.

EXPLORATION PROSPECTS

The complex tectonic, magmatic, and hydrothermal systems operative since Paleozoic time in the eastern part of the Hailey 1°×2° quadrangle make exploration for Devonian syngenetic ore deposits difficult but afford the possibility that large mineral deposits remain to be discovered. Some recent models for exploration in the Milligen Formation, the Salmon River assemblage, and correlative unnamed Silurian and Devonian argillite invoke the presence of synsedimentary Devonian fault-bounded "third-order" basins, which during Devonian time were subject to restricted circulation of oceanic water and coeval hydrothermal activity (Turner and Otto, 1988, this volume; Bruner, 1991). Detailed stratigraphic analyses, including biostratigraphic studies, are needed to document Devonian basin evolution.

REFERENCES CITED

Anderson, A.L., Kiilsgaard, T.H., and Fryklund, V.C., Jr., 1950, Detailed geology of certain areas in the Mineral Hill and Warm Springs mining districts, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 90, 73 p.

Batatian, L.D., 1991, Paleozoic stratigraphy and Cenozoic structure, central Boulder Mountains, Blaine and Custer Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 85 p.

Bostwick, D.A., 1955, Stratigraphy of the Wood River Formation, south-central Idaho: Journal of Paleontology, v. 29, p. 941–951.

Brennan, B.D., 1987, Silurian and Devonian rocks along the upper reaches of the North Fork of the Big Lost River, Custer County, Idaho: Milwaukee, University of Wisconsin, M.S. thesis, 72 p.

Breuninger, R.H., Canter, K.L., and Isaacson, P.E., 1988, Pennsylvanian-Permian *Palaeoaplysina* and algal buildups, Snaky Canyon Formation, south-central Idaho, U.S.A., *in* Geldsetzer, H.H.J., James, N.P., and Tebbut, G.E., eds., Reefs—Canada and adjacent areas: Canadian Society of Petroleum Geologists Memoir 13, p. 631–637.

Bruner, D.J., 1991, Origin of mineralization within Devonian-Silurian sedimentary breccias, northeastern Boulder Mountains, Custer County, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 118 p.

Burton, B.R., 1988, Stratigraphy of the Wood River Formation in the eastern Boulder Mountains, Blaine and Custer Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 165 p.

Burton, B.R., and Link, P.K., 1991, Ichnology of fine-grained mixed carbonate-siliciclastic turbidites, Wood River Formation, Pennsylvanian-Permian, south-central Idaho: Palaios, v. 6, p. 291–301.

Burton, B.R., Link, P.K., and Rodgers, D.W., 1989, Death of the Wood River thrust—Structural relations in the Pioneer and Boulder Mountains, south-central Idaho: Geological Society of America Abstracts with Programs, v. 21, no. 5, p. 62.

Carter, Claire, and Churkin, Michael, Jr., 1977, Ordovician and Silurian graptolite succession in the Trail Creek area, central Idaho—A graptolite zone reference section: U.S. Geological Survey Professional Paper 1020, 37 p.

Churkin, Michael, Jr., 1963, Graptolite beds in thrust plates of central Idaho and their correlation with sequences in Nevada: American Association of Petroleum Geologists Bulletin, v. 47, p. 1611–1623.

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

Darling, R.S., 1988, Ore deposits of the Carrietown silver-lead-zinc district, Blaine and Camas Counties, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 193–198.

Dover, J.H., 1969, Bedrock geology of the Pioneer Mountains, Blaine and Custer Counties, central Idaho: Idaho Bureau of Mines and Geology Pamphlet 142, 61 p.

———1981, Geology of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–A, p. 1–75.

———1983, Geologic map and sections of the central Pioneer Mountains, Blaine and Custer Counties, central Idaho: U.S. Geological Survey Miscellaneous Investigations Series Map I–1319, scale 1:48,000.

Dover, J.H., Berry, W.B.N., and Ross, R.J., Jr., 1980, Ordovician and Silurian Phi Kappa and Trail Creek Formations, Pioneer Mountains, central Idaho—Stratigraphic and structural revisions, and new data on graptolite faunas: U.S. Geological Survey Professional Paper 1090, 54 p.

Fisher, F.S., and Johnson, K.M., eds., 1987, Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, 251 p.

Fisher, F.S., McIntyre, D.H., and Johnson, K.M., 1983, Geologic map of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 83–523, scale 1:250,000, 41 p.

Fryklund, V.C., Jr., 1950, The ore deposits of the Mayflower area, *in* Anderson, A.L., Kiilsgaard, T.H., and Fryklund, V.C., Jr., Detailed geology of certain areas in the Mineral Hill and Warm Springs mining districts, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 90, p. 63–73.

Geslin, J.K., 1986, The Permian Dollarhide Formation and Paleozoic Carrietown sequence in the SW¼ of the Buttercup Mountain quadrangle, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 116 p.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

———1987a, Lead-silver-zinc-antimony-tin vein deposits, *in* Fisher, F.S., and Johnson, K.M., eds., 1987, Preliminary manuscript for "Mineral-resource assessment and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 133–140.

———1987b, Stratiform syngenetic deposits of precious and base metals in argillic rocks and micritic limestone, *in* Fisher, F.S., and Johnson, K.M., eds., 1987, Preliminary manuscript for "Mineral-resource assessment and geology of the Challis

1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 205–208.

Hall, W.E., Batchelder, J.H., and Douglass, R.D., 1974, Stratigraphic section of the Wood River Formation, Blaine County, Idaho: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 579–592.

Hall, W.E., and Czamanske, G.K., 1972, Mineralogy and trace element content of the Wood River lead-silver deposits, Blaine County, Idaho: Economic Geology, v. 67, p. 350–361. C

Hall, W.E., and Hobbs, S.W., 1987, Black shale terrane, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource assessment and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 29–42.

Hall, W.E., Rye, R.O, and Doe, B.R., 1978, Wood River mining district—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 579–592.

Hobbs, S.W., 1985, Precambrian and Paleozoic sedimentary terranes in the Bayhorse area of the Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–D, p. 59–68.

Howe, S.S., and Hall, W.E., 1985, Light-stable isotope characteristics of ore systems in central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–P, p. 183–192.

Huerta, A.D., Lovejoy, F., and Link, P.K., 1991, Anomalous Mississippian(?) conglomerates near Summit Creek, northern Boulder Mountains, south-central Idaho: Geological Society of America Abstracts with Programs, v. 23, no. 4, p. 34.

Kiilsgaard, T.H., 1950, Geology and ore deposits of the Triumph-Parker mine mineral belt, *in* Anderson, A.L., Kiilsgaard, T.H., and Fryklund, V.C., Jr., Detailed geology of certain areas in the Mineral Hill and Warm Springs mining districts, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 90, p. 39–62.

Kim, A.Y.H., 1986, The kinematics of brittle polyphase deformation within the Pioneer metamorphic core complex, Pioneer Mountains, Idaho: Bethlehem, Pa., Lehigh University, M.S. thesis, 130 p.

Klapper, B., and Murphy, M.A., 1974, Silurian-Lower Devonian conodont sequence in the Roberts Mountains Formation of central Nevada: University of California Publications in Geological Sciences, v. 3, 62 p.

Kunkel, K.W., 1989, Geology and ore deposits of Garfield Canyon, Blaine County, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 92–94.

Lewis, R.S., 1989, Plutonic rocks in the southeastern part of the Idaho batholith and their relation to mineralization, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles: U.S. Geological Survey Open-File Report 89–639, p. 33–34.

Lindgren, W., 1900, The gold and silver veins of Silver City, DeLamar, and other mining districts in Idaho: U.S. Geological Survey 20th Annual Report, pt. 3, p. 65–256.

Link, P.K., Skipp, Betty, Hait, M.H., Jr., Janecke, Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 5–42.

Link, P.K., and Worl, R.G., in press, Geology and mineral deposits of the Minnie Moore and Bullion mineralized areas, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–Y.

Loeblich, A.R., Jr., and Tappan, Helen, 1988, Foraminiferal genera and their classification: New York, Van Nostrand Reinhold, 970 p.

Mahoney, J.B., 1992, Geologic map of the Frenchman Creek quadrangle, Blaine and Camas Counties, Idaho: Idaho Geological Survey Technical Report 92–2, scale 1:24,000.

Mahoney, J.B., and Horn, M.C., in press, Geology of the Vienna mineralized area, Blaine and Camas Counties, Idaho: U.S. Geological Survey Bulletin 2064–DD.

Mahoney, J.B., Link, P.K., Burton, B.R., Geslin, J.K., and O'Brien, J.P., 1991, Pennsylvanian and Permian Sun Valley Group, Wood River Basin, south-central Idaho, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States–II: Los Angeles, Pacific Section Society of Economic Paleontologists and Mineralogists, Publication 67, v. 2, p. 551–579.

Mahoney J.B., and Sengebush, R.M., 1988, Stratigraphy of the lower Permian Grand Prize Formation, south-central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 169–182.

McIntyre, D.H., ed., 1985, Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658, 227 p.

Nilsen, T.H., 1977, Paleogeography of Mississippian turbidites in south-central Idaho, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 1, p. 275–299.

O'Brien, J.P., 1991, Stratigraphy of the Pennsylvanian-Permian Dollarhide Formation in the Buttercup Mountain quadrangle, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 163 p.

O'Brien, J.P., Lewis, R.S., Link, P.K., Geolin, J.K., Whitman, S.K., and Schmidt, K.L., 1993, Geologic map of the Buttercup Mountain quadrangle, Blaine and Camas Counties, Idaho: Idaho Geological Survey Technical Report 93–1, scale 1:24,000.

Park, W.A., in press, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–CC.

Paull, R.A., and Gruber, D.P., 1977, Little Copper Formation—New name for lowest formation of Mississippian Copper Basin Group, Pioneer Mountains, south-central Idaho: American Association of Petroleum Geologists Bulletin, v. 61, no. 2, p. 256–262.

Paull, R.A., Wolbrink, M.A., Volkman, R.G., and Grover, R.L., 1972, Stratigraphy of the Copper Basin Group, Pioneer Mountains, south-central Idaho: American Association of Petroleum Geologists Bulletin, v. 56, no. 8, p. 1370–1401.

Poole, F.G., 1974, Flysch deposits of the Antler foreland basin, western United States, *in* Dickinson, W.R., ed., Tectonics and sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication 22, p. 58–82.

Poole, F.G., and Sandberg, C.A., 1991, Mississippian paleogeography and tectonics of the Western United States, *in* Stewart J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coaast Paleogeography Symposium 1, p. 67–85.

———1991, Mississippian paleogeography and conodont biostratigraphy of the Western United States, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 1, p. 107–136.

Ratchford, M.E., 1989, Geology of the Boulder Basin, Blaine and Custer Counties, Idaho: Moscow, University of Idaho, M.S. thesis, 154 p.

———in press, Epigenetic lead-tin-silver, antimony-gold veins in Boulder Basin, Blaine and Custer Counties, Idaho—Potential for economic tin mineralization: U.S. Geological Survey Bulletin 2064–JJ.

Ross, C.P., 1960, Diverse interfingering Carboniferous strata in the Mackay quadrangle, Idaho: U.S. Geological Survey Professional Paper 400–B, p. B232–B233.

———1962, Upper Paleozoic rocks in central Idaho: American Association of Petroleum Geologists Bulletin, v. 46, no. 3, p. 384–387.

Sandberg, C.A., 1975, McGowan Creek Formation, new name for Lower Mississippian flysch sequence in east-central Idaho: U.S. Geological Survey Bulletin 1405–E, 11 p.

———1979, Devonian and Lower Mississippian conodont zonation of the Great Basin and Rocky Mountains: Brigham Young University Geology Studies, v. 26, pt. 3, p. 87–106.

Sandberg, C.A., Hall, W.E., Batchelder, J.N., and Axelson, Claus, 1975, Stratigraphy, conodont dating, and paleotectonic interpretation of the type Milligen Formation (Devonian), Wood River area, Idaho: U.S. Geological Survey Journal of Research, v. 3, no. 6, p. 707–720.

Sengebush, R.M., 1984, The geology and tectonic history of the Fourth of July Creek area, White Cloud Peaks, Custer County, Idaho: Missoula, University of Montana, M.S. thesis, 79 p.

Skipp, Betty, 1989, Geologic map of the Mackay 4 (Grouse) NW quadrangle, Butte and Custer Counties, Idaho: U.S. Geological Survey Open-File Report 89–142, scale 1:24,000.

Skipp, Betty, and Hall, W.E., 1975, Structure and Paleozoic stratigraphy of a complex of thrust plates in the Fish Creek Reservoir area, south-central Idaho: U.S. Geological Survey Journal of Research, v. 3, no. 6, p. 671–689.

———1980, Upper Paleozoic paleotectonics and paleogeography of Idaho, *in* Fouch, T.D., and Magathan, E.R., eds., Paleozoic paleogeography of the West-Central United States: Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, West-Central United States Paleogeography Symposium 1, p. 387–422.

Skipp, Betty, Hall, W.E., and Link, P.K., 1986, Pre-middle Eocene extension in south-central Idaho—The Wood River allochthon: Geological Society of America Abstracts with Programs, v. 18, p. 753.

———1994, Geologic map of the Mahoney Butte quadrangle, Blaine County, Idaho: U.S. Geological Survey Geologic Quadrangle Map GQ–1742, scale 1:24,000.

Skipp, Betty, Hoggan, R.D., Schleicher, D.L., and Douglass, R.C., 1979, Upper Paleozoic carbonate bank in east-central Idaho—Snaky Canyon, Bluebird Mountain, and Arco Hills Formations, and their paleotectonic significance: U.S. Geological Survey Bulletin 1496, 78 p.

Skipp, Betty, Kuntz, M.A., and Morgan, L.A., 1990, Geologic map of the Mackay 4 (Grouse) SE quadrangle, Butte County, Idaho: U.S. Geological Survey Open-File Report 89–431, scale 1:24,000.

Skipp, Betty, Sando, W.J., and Hall, W.E., 1979, The Mississippian and Pennsylvanian (Carboniferous) systems in the United States–Idaho: U.S. Geological Survey Professional Paper 1110–AA, p. AA1–AA43.

Spoelhof, R.A., 1972, Structure and stratigraphy of portions of the Meridian Peak and Herd Peak quadrangles, Custer County, south-central Idaho: Golden, Colorado School of Mines, M.S. thesis, 86 p.

Stewart, D.E., 1987, Geology of the northern half of the Baker Peak quadrangle, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 113 p.

Stewart, D.E., Mahoney, J.B., Geslin, J.K., Darling, R.S., and Link, P.K., 1992, Geologic map of the Baker Peak quadrangle, Blaine and Camas Counties, Idaho: Idaho Geological Survey Technical Report 92–1, scale 1:24,000.

Thomasson, M.R., 1959, Late Paleozoic stratigraphy and paleotectonics of central and eastern Idaho: Madison, University of Wisconsin, Ph.D. dissertation, 244 p.

Thompson, M.L., Dodge, H.W., and Youngquist, W., 1958, Fusulinids from the Sublett Range, Idaho: Journal of Paleontology, v. 32, p. 113–125.

Tschanz, C.M., Kiilsgaard, T.H., and Seeland, D.A., 1986, Geology of the eastern part of the Sawtooth National Recreation Area, Idaho, *in* Mineral resources of the eastern part of the Sawtooth National Recreation Area, Custer and Blaine Counties, Idaho: U.S. Geological Survey Bulletin 1545–A, p. 17–43.

Tuchek, E.T., and Ridenour, James, 1981, Economic appraisal of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497, p. 181–292.

Turner, R.J.W., and Otto, B.R., 1988, Stratigraphy and structure of the Milligen Formation, Sun Valley area, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 153–167.

Umpleby, J.B., 1915, Ore deposits in the Sawtooth quadrangle, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 580–K, p. 221–29.

Umpleby, J.B., Westgate, L.G., and Ross, C.P., 1930, Geology and ore deposits of the Wood River region, Idaho: U.S. Geological Survey Bulletin 814, 250 p.

Van Noy, R.M., Ridenour, James, Zilka, N.T., Federspiel, F.E., Evans, R.K., Tuchek, E.T., and McMahan, A.B., 1986, Economic appraisal of the eastern part of the Sawtooth National Recreation Area, Idaho: U.S. Geological Survey Bulletin 1545–E, p. 232–472.

Verville, G.J., Sanderson, G.A., Baesemann, J.F., and Hampton, G.L., III, 1990, Pennsylvanian fusulinids from the Beaverhead Mountains, Morrison Lake area, Beaverhead County, Montana: The Mountain Geologist, v. 27, p. 47–55.

Walliser, O.H., 1964, Conodonten des Silurs: Abhandlungen des Hessischen Landesamtes für Bodenforschung zu Wiesbaden, v. 41, 106 p.

Wavra, C.S., 1985, Structural geology and petrology of the SW¼ of the Mahoney Butte quadrangle, Blaine County, Idaho: Moscow, University of Idaho, M.S. thesis, 103 p.

Wavra, C.S., and Hall, W.E., 1989, Potential for shale-hosted massive sulfides in the Permian Dollarhide Formation, south-central Idaho: U.S. Geological Survey Open-File Report 88–376, 9 p.

Wavra, C.S., Isaacson, P.E., and Hall, W.E., 1986, Studies of the Idaho black shale belt—Stratigraphy, depositional environment, and economic geology of the Permian Dollarhide Formation: Geological Society of America Bulletin, v. 97, p. 1504–1511.

Whitman, S.K., 1990, Metamorphic petrology and structural geology of the Pennsylvanian-Permian Dollarhide Formation, Blaine and Camas Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 108 p.

Wilson, Eric, 1994, Demise of the Glide Mountain thrust, Pioneer Mountains, Blaine and Custer Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 188 p.

Wilson, Eric, Preacher, J.M., and Link, P.K., 1994, New constraints on the nature of the Early Mississippian Antler sedimentary basin in Idaho, *in* Beauchamp, B., Embry, A.F., and Glass, G., eds., Pangea—Global environments and resourses: Canadian Society of Petroleum Geologists Memoir 17.

Winsor, H.C., 1981, The paleogeography and paleoenvironments of the Middle Pennsylvanian part of the Wood River Formation in south-central Idaho: San Jose, Calif., San Jose State University, M.S. thesis, 81 p.

Worl, R.G., Kiilsgaard, T.H., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Mineral Deposits of the Muldoon–Star Hope Area, Southern Pioneer Mountains, South-Central Idaho

By Gary R. Winkler, Kevin W. Kunkel, Richard F. Sanford,
Ronald G. Worl, *and* Falma J. Moye

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND
IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–D

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Mineral Deposits of the Muldoon–Star Hope Area, Southern Pioneer Mountains, South-Central Idaho

By Gary R. Winkler,[1] Kevin W. Kunkel,[2] Richard F. Sanford,[1] Ronald G. Worl,[3] *and* Falma J. Moye[4]

ABSTRACT

Devonian and Mississippian carbonate and clastic rocks of the Milligen(?) Formation, Carey Dolostone, and Copper Basin Formation in the southern Pioneer Mountains host many vein, replacement, and distal skarn deposits of silver-lead-zinc (+copper). In several places, the silver-base metal deposits are associated with stratabound veins and replacements of barite. Base- and precious-metal veins also are present locally in Eocene stocks that intrude the Paleozoic sequence and in rocks of the comagmatic Challis Volcanic Group that overlie the Paleozoic sequence.

The deposits are included in the Little Wood River and Copper Basin mining districts, which were formed in the early 1880's when the first significant discoveries and initial production occurred. The richest metal deposits were mined principally in the 1880's and 1890's and again in the 1940's and 1950's. The richest barite bodies were mined only in the 1950's and 1960's. No mining has taken place in either district for the past 25 years; however, substantial resources have been identified or inferred, and exploration has been renewed in the middle 1970's, the early 1980's, and the late 1980's.

The locations of the silver-base metal deposits are governed by the presence of carbonaceous pelitic source rocks, chemically reactive calcareous trap rocks, concordant and discordant structures that provided channelways for solutions, and intrusive rocks that provided heat and fluids for metasomatism. Argillaceous rocks of the Little Copper Member of the Copper Basin Formation, which are significantly enriched in barium and slightly enriched in several metals, host numerous small and discontinuous veins that show spotty values for silver, lead, zinc, copper, and gold. The best examples are in the vicinity of the Mutual mine; however, such veins seldom were mined. The largest and richest metal deposits are polymetallic replacements or skarns in impure calcareous rocks of the Drummond Mine Limestone Member of the Copper Basin Formation, which overlies the Little Copper Member. The deposits are at several stratigraphic horizons and generally are within 1–3 km of quartz monzonite stocks or swarms of dikes. The best example is the Eagle Bird mine, where high-grade material remains underground in tabular lenses extending 100 m or more along bedding. In addition to the silver, lead, zinc, copper, and gold, some replacements and skarns also contain significant antimony or tin, but none is known to have been produced. Where polymetallic replacements are cut at high angles by faults or fractures, they generally have associated veins. Veins at the Solid Muldoon and Star Hope properties were unusually rich in silver and thus were mined during the early years in the districts.

Significant resources of base- and precious-metals and barite remain in the Muldoon–Star Hope area. Although most individual occurrences are likely to be small, they may be of high grade and closely spaced. Thus, the metal values of a group of occurrences might constitute collectively a minable resource in the future. Furthermore, lower grade resources, particularly of barite, but also of metals, may be relatively widespread in the Copper Basin Formation. Favorable areas for additional exploration include several (or all) of the following features: (1) calcareous, or mixed calcareous and siliceous host rocks; (2) Eocene leucocratic intrusive rocks; (3) carbonaceous pelitic rocks having elevated metal values; (4) ground structurally prepared by steep faults, particularly where associated with bedding-plane flexures or faults; (5) widespread geochemical anomalies of precious- and base-metals and associated elements; and, if available, (6) elevated electrical conductivity measurements, which might indicate the presence of metal-bearing sulfide minerals at shallow depth.

[1]U.S. Geological Survey, Denver Federal Center, Denver, Colorado 80225.

[2]Department of Geology, Idaho State University, Pocatello, Idaho 83209.

[3]U.S. Geological Survey, W. 904 Riverside Avenue, Spokane, Washington 99201.

[4]Box 1145, Challis, Idaho 83226.

INTRODUCTION

In the Muldoon Creek, Garfield Canyon, and Star Hope Creek areas of Blaine and Custer Counties in south-central Idaho (fig. 1), Devonian and Mississippian carbonate and clastic rocks of the Milligen(?) Formation, Carey Dolostone, and Copper Basin Formation host numerous polymetallic deposits containing silver, lead, zinc, copper, and gold. Base- and precious-metal veins also are present in porphyritic quartz monzonite stocks that intrude Paleozoic sedimentary rocks in Muldoon Creek and Garfield Canyon and in rocks of the Eocene Challis Volcanic Group that overlie the Paleozoic sedimentary rocks at the confluence of Copper Creek and Blackspar Canyon. The deposits are present on both flanks of the southeastern Pioneer Mountains, which divide tributaries of the Little Wood River from those of the East Fork of the Big Lost River in Copper Basin. The precious- and base-metals are in vein, replacement, skarn, and stratabound deposits that in several places are associated with stratabound veins and replacements of barite.

HISTORICAL PERSPECTIVE

PAST PRODUCTION

Although prospecting apparently began in the area in the 1860's, the first significant deposits were located in 1881, with the discoveries of silver-base metal deposits at the Muldoon (or Solid Muldoon) and Star Hope claims (fig. 2) on opposite sides of the drainage divide (Umpleby, 1917). Within a few seasons, numerous additional discoveries were made nearby in both drainages. The claims were included in the separate Little Wood River and Copper Basin mining districts, inasmuch as normal access to them was by seasonal roads from Bellevue (or Carey) and Mackay, respectively. Most production in the districts was in the 1880's and again in the 1940's and 1950's, although U.S. Bureau of Mines records indicate some production in every decade from the 1900's through the 1960's (Tuchek and Ridenour, 1981). No mining has taken place in either district for the past 25 years.

LITTLE WOOD RIVER DISTRICT

Most deposits in the district are in a northwest-trending belt about 2 km wide that extends from near the mouth of Muldoon Creek to a short distance north of Little Copper Creek (fig. 2). Clusters of silver-base metal deposits are along upper Muldoon Creek, in Garfield Canyon, in Deep, Boundary, and Mutual Gulches, in the Little Copper Creek drainages, and in Argosy Creek. Base- and precious-metal veins in rocks of the Challis Volcanic Group along Copper Creek and Blackspar Canyon are west of this belt.

The Solid Muldoon claims were located in 1881 and patented in 1886. A concentrating mill and two 40-ton smelters were erected near the claims in late 1881 and early 1882 (E.H. Finch, *in* Umpleby, 1917). About $200,000 in silver-lead ore was produced from the claims between 1881 and 1884 (Wells, 1983). Additional ore-grade material was stockpiled intermittently until 1908, when completion of a new 100-ton mill allowed minor additional production from 1908 to 1911. By late 1911, Muldoon's mining machinery was consigned to another mill near Ketchum (Wells, 1983); however, U.S. Bureau of Mines records show minor additional production until 1948, which probably was chiefly from reprocessed tailings (Tuchek and Ridenour, 1981).

The Idaho Muldoon property on the east side of Muldoon Creek was located in 1908 and was rehabilitated in the 1940's (Anderson and Wagner, 1946). No production records exist, although small shipments of lead-silver ore were sent to mills in Utah for testing. High-grade material remains in underground workings and in surface cuts that extend more than 100 ft south from the main portal (Anderson and Wagner, 1946).

The Eagle Bird claim in Garfield Canyon was located in 1883 and patented in 1889. Numerous nearby claims were located (and some were patented) during the same decade. From 1929 to 1965, the Eagle Bird and nearby claims produced lead, zinc, silver, copper, and gold worth approximately $130,000 (Tuchek and Ridenour, 1981); value of earlier production, if any, is unknown. High-grade material remains in underground workings, and known shoots have not been explored extensively downdip. There are no records of production of base- or precious-metals from the numerous claims between the Solid Muldoon and Eagle Bird mines in Deep, Boundary, and Mutual Gulches. Minor production is likely inasmuch as several dumps are quite extensive, and material apparently has been removed from stopes underground (Tuchek and Ridenour, 1981).

Little published information exists on the Black Spar properties, at the confluence of Blackspar Canyon with Copper Creek. The deposit was discovered in 1887, and more than 300 tons of ore had been mined through 1888 (Wells, 1983). No records of subsequent production exist; however, the survey for patent of 19 claims in 1923 indicates that there are more than 4,500 ft of underground workings, including two shafts, but apparently no stopes (J.J. Jones, U.S. Forest Service, written commun., 1988). There are no records of post-1900 production, although at least one of the tunnels was rehabilitated in the 1960's.

During a renewed pulse of exploration in the Muldoon–Star Hope area in the late 1950's. and 1960's, barite deposits in Deep Gulch were discovered, and a large shipment was made to the Atomic Energy Commission's research facilities near Arco for use as shielding material in nuclear reactors (Tuchek and Ridenour, 1981). Nearby base-metal-bearing veins were explored at the same time. Although the Deep Gulch barite claims were being developed actively in the mid-1970's and additional barite deposits between Muldoon and Argosy Creeks were explored during this time, no production has occurred since the 1970's.

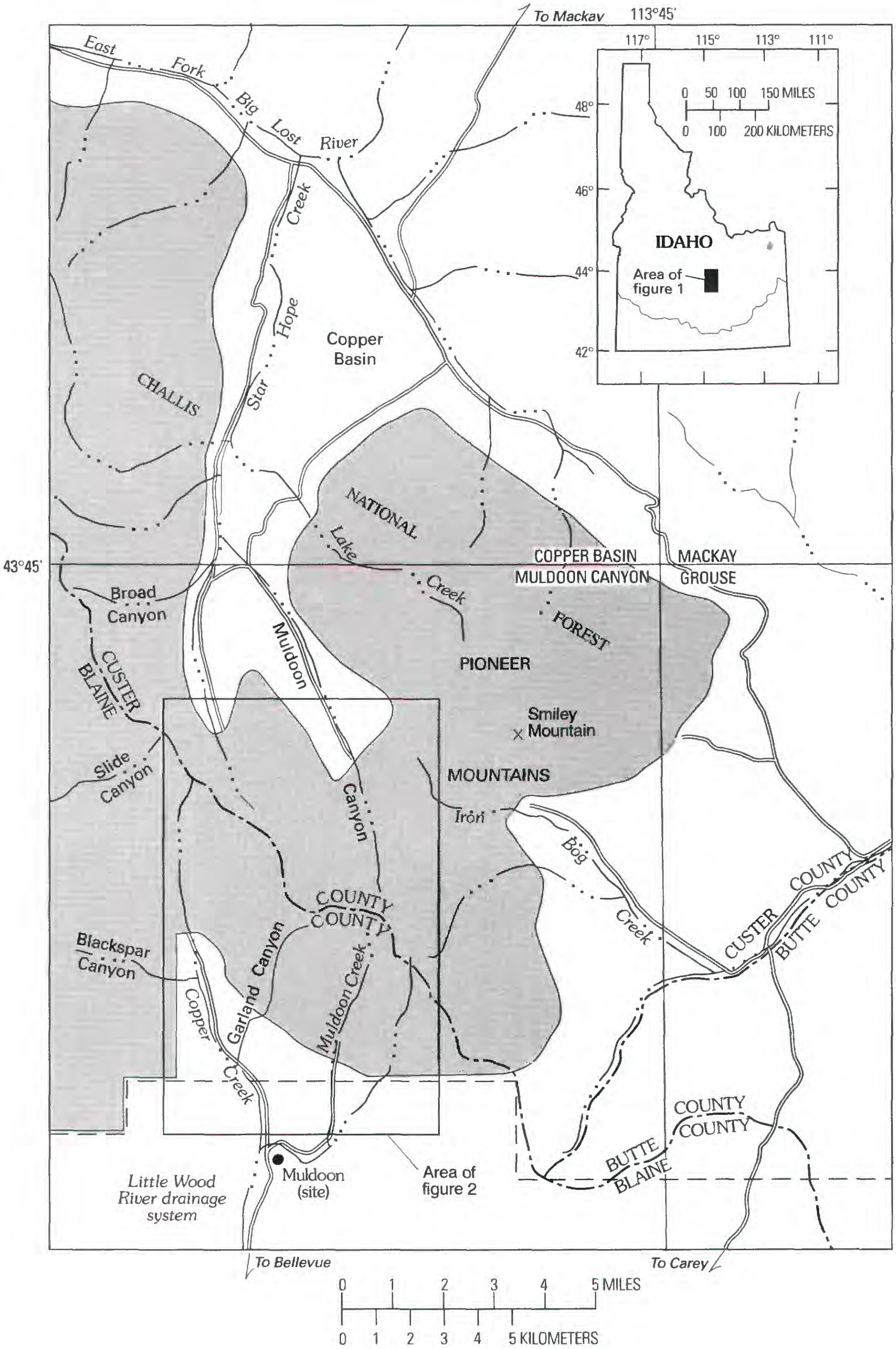
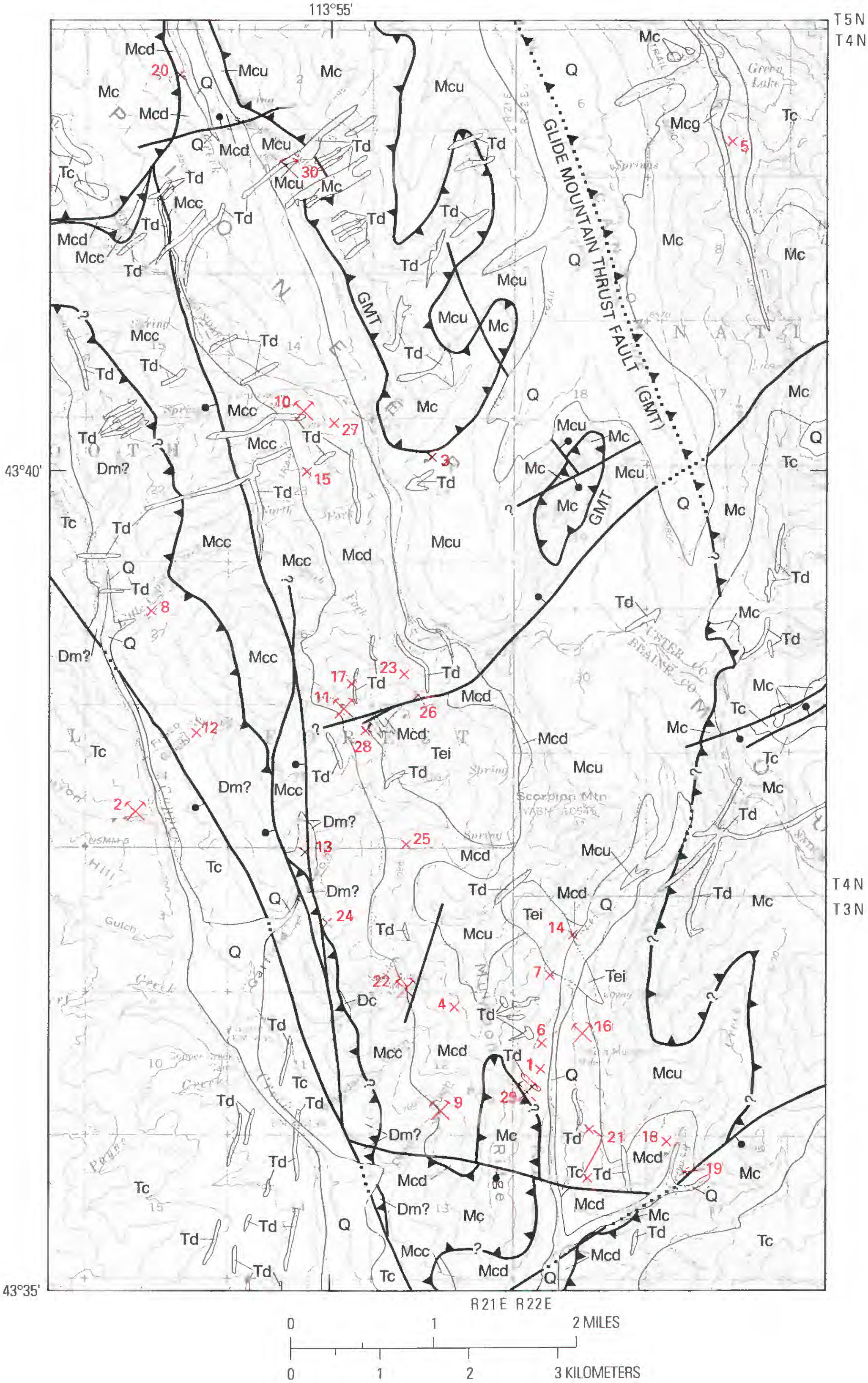


Figure 1. Index map of the Muldoon–Star Hope area in the eastern part of the Boulder-Pioneer Wilderness Study Area, south-central Idaho.


113°55'
T5N
T4N
43°40'
43°35'
T4N
T3N
R21E R22E
GLIDE MOUNTAIN THRUST FAULT (GMT)
GMT
Scorpion Mtn
Green Lake
Mc
Mcd
Mcu
Mcc
Mcg
Td
Tc
Tei
Q
Dm?
Dc
0 1 2 MILES
0 1 2 3 KILOMETERS

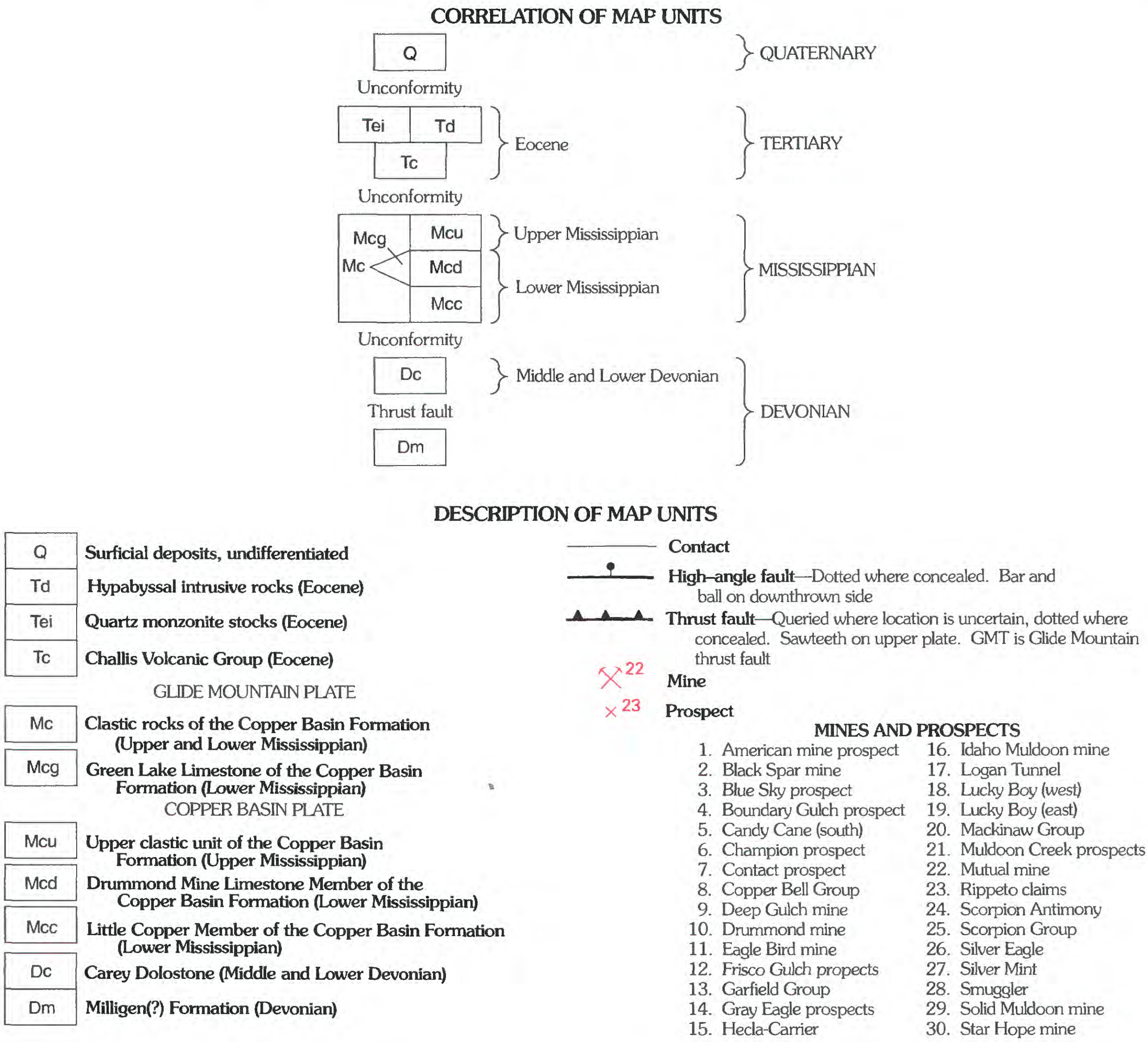


Figure 2 (above and facing page). Geology of the Muldoon–Star Hope area, south-central Idaho. Locations of selected mines and prospects are shown in red. Location of map area is shown in figure 1. Modified from Dover (1981) and Kunkel (1989). Base from U.S. Geological Survey Muldoon (1974) and Star Hope Mine (1974) 7.5-minute quadrangles.

COPPER BASIN DISTRICT

Sulfide-bearing vein deposits of silver, lead, zinc, and copper were discovered beginning in 1881 in the Star Hope Creek drainage in the southwestern corner of Copper Basin (fig. 2). The most promising claims near the head of Star Hope Gulch were patented in 1889. Most development was during the 1890's, when the properties apparently produced approximately $50,000 of lead-silver ore (Umpleby, 1917; Tuchek and Ridenour, 1981). Additional claims were made (and some were patented) in the main drainage between 1899 and 1913. U.S. Bureau of Mines records show latest production, worth about $7,500, in the Star Hope area in 1954, and additional earlier production is inferred from the size of stopes (Tuchek and Ridenour, 1981). There was renewed exploration in the vicinity of the Star Hope properties in 1984, and some claims were staked (A.B. Wilson and S.J. Soulliere, U.S. Geological Survey, written commun., 1990). No additional detailed site evaluation or development has been performed subsequently.

EXPLORATION ACTIVITY IN THE LITTLE WOOD RIVER DISTRICT

Following fieldwork in 1973 and 1974, geologists of the U.S. Bureau of Mines estimated minimum resources of 80,000 tons of high-grade material in the Eagle Bird, Rippeto, Silver Eagle, and nearby claims in Garfield Canyon and recommended that the claims and possible extensions be explored (Tuchek and Ridenour, 1981). U.S. Bureau of Mines assays of samples from these properties contain 3–8 percent Pb, 1–2.5 percent Zn, 3.3–9.4 ounces of silver per ton, and nil to 0.02 ounces of gold per ton (Tuchek and Ridenour, 1981). In the early 1980's, according to documents in the International Archive of Economic Geology at the University of Wyoming, Anaconda Mining Company conducted a regional exploration program for gold that included some sampling in the Muldoon–Star Hope area. In 1987–1989, Hecla Mining Company conducted exploration in Garfield Canyon for precious- and base-metal lodes and skarns. The exploration included detailed geologic mapping and surface geophysical surveys to test the extent and grade of deposits that crop out widely in the canyon through a vertical range of more than 600 m.

GEOLOGIC SETTING

Rocks in the Muldoon–Star Hope area include three fault-bounded Paleozoic sedimentary sequences that are overlain unconformably by flows of the Eocene Challis Volcanic Group (fig. 2). A western sequence of Paleozoic rocks consists of silicified argillaceous rocks that Dover (1981) mapped as the Milligen(?) Formation. These rocks, which comprise the structurally lowest allochthonous sequence above the Pioneer fault system, are undated in the Muldoon–Star Hope area but are correlated with the Milligen Formation of the type area east of the Triumph mine near Ketchum, Idaho, because of their similar strong deformation and suite of rocks. A medial Paleozoic sequence consists of siliceous and calcareous marine basinal facies assemblages of Mississippian age that were deposited unconformably on thinly laminated limy dolomite of Devonian age. An eastern Paleozoic sequence consists of siliceous, shallow-marine to partly terrigenous, basin-margin facies assemblages of Mississippian age and contains only minor calcareous rocks. The Mississippian rocks of the medial and eastern sequences constitute the Copper Basin Formation, and they are roughly coeval; however, they have been juxtaposed by the Glide Mountain thrust fault, which transposed the shallow-marine and nonmarine assemblages onto the basinal assemblages during the Cretaceous Sevier orogeny (Dover, 1981). The tight folding and well-developed axial planar cleavage in Mississippian rocks of the upper Glide Mountain plate are distinct from the generally open folds and homoclinal dips in rocks of the lower Copper Basin plate. The contrast in intensity of deformation may reflect a tectonic transport of upper plate rocks from an original position farther to the west than rocks in the lower plate.

Late Cretaceous and (or) early Tertiary uplift, erosion, and deposition of conglomerate preceded onset of Challis volcanism, which began about 49 Ma. The Eocene magmatic event included coeval volcanism and emplacement of hypabyssal stocks and dikes. The Paleozoic sequences are intruded by porphyritic stocks and by dike swarms and single dikes inferred to be feeders to flows within rocks of the Challis Volcanic Group. Stocks exposed in Garfield Canyon and Muldoon Creek (fig. 2) are strongly discordant porphyritic hornblende-biotite (with or without pyroxene) quartz monzonite having no internal fabric. Biotite from the main phase of the Garfield stock has been dated by $^{40}Ar/^{39}Ar$ at 48.72±0.15 Ma (Moye and others, this volume) and biotite from the Muldoon stock has been dated by K-Ar at 47.0 Ma (R.F. Zartman, *in* Dover, 1981). The two stocks are elongated northwest-southeast. A zone of high magnetic intensity enclosing the stocks indicates that they are connected at depth (Mabey, 1981). A concentrated swarm of dikes and associated locally intense alteration between Muldoon Canyon and Deadman Creek in the Star Hope area is believed to indicate a subjacent pluton at shallow depth (Dover, 1981). Dikes are texturally and compositionally more diverse than the stocks and range from porphyritic high-potassium rhyolite to high-potassium basalt (table 1). Dikes and wallrocks locally are strongly altered. The dikes commonly are oriented parallel with north-northwest and east-northeast fracture systems. In a few places, dikes intrude the stocks; intrusion of the stocks probably was coeval with earlier Challis volcanism.

Throughout the Muldoon–Star Hope area, the largest silver-base metal deposits are in calcareous rocks within 1 km of quartz monzonite stocks (fig. 2). Apparently, weak hydrothermal cells were established during intrusion of the stocks. Analyses of fluid inclusions from quartz and calcite veins in Garfield Canyon indicate a temperature gradient centered on the stock: filling temperatures of fluid inclusions from mineralized veins at the margin of the stock are in excess of 425°C, whereas fluid inclusions in deposits farther from the stock filled at temperatures between 245°C and 280°C (K.W. Kunkel, Idaho State University, unpub. data, 1991). Low salinity values of 2–8 weight percent NaCl indicate dominantly meteoric fluids. Rudimentary zoning of sulfide minerals and alteration assemblages around the Garfield stock may indicate that mineralization was concurrent with intrusion. Chalcopyrite and pyrite are more abundant near the stock and are uncommon at greater distances (Kunkel, 1989). Argillic alteration assemblages (kaolinite-sericite) are present near the stock, whereas propylitic assemblages (chlorite-zeolite) are present outward; however, the assemblages generally are developed only along faults, and unfaulted host rocks generally show little or no alteration.

In many places in the Muldoon–Star Hope area, silver-base metal deposits are in calcareous rocks several kilometers or more from exposed plutons. Although these deposits generally are smaller than those close to plutons, they contain the same equally coarse grained sulfide-mineral assemblages.

All host rocks are intruded by scattered to abundant dikes that probably are coeval with the stocks. In many places, dikes make up more than 10 percent of the outcrop. The dikes apparently occupy structures having the same orientation as those that host mineral deposits. North- to northwest-trending dikes are roughly parallel in strike with enclosing strata but generally dip more steeply. East-northeast-trending dikes generally are almost vertical. Umpleby (1917) doubted whether the dikes alone could have supplied enough heat and introduced enough material to produce the orebodies and metasomatism of the host rocks; however, the dikes and the exposed stocks may have emanated from the same large magma chamber at depth that would have imposed a regional hydrothermal system throughout the Muldoon–Star Hope area.

Steep faults cut the Paleozoic sequences in many places; locally, they cut rocks of the Challis Volcanic Group. The steep faults tend to have three orientations, about N. 20° W., N. 50° E., or N. 80° E., which also are the dominant trends of Eocene dikes. The east-northeast-trending faults and dikes are roughly parallel with the Trans-Challis fault system and likely formed during Eocene regional extension and magmatism. North-northwest-trending faults generally have been attributed to Neogene Basin and Range extension, but their approximate parallelism with trends of some Eocene dikes may indicate an earlier ancestry as a conjugate set to faults that trend about N. 50° E.

HOST ROCKS

In the Muldoon–Star Hope area, a few silver-base metal deposits are in the Devonian Milligen(?) Formation and Carey Dolostone and in the Eocene Challis Volcanic Group, but most deposits having significant resources are hosted by the Mississippian Copper Basin Formation. In the lower reaches of Little Copper Creek, Deadman Creek, and Garfield Canyon, weakly mineralized pyrite-arsenopyrite-chalcopyrite-bearing quartz veins are in argillaceous rocks of the Milligen(?) Formation near porphyritic rhyolite dikes. Near the bottom of Garfield Canyon just east of the Copper Creek fault, silicified limy dolomite of the Carey Dolostone hosts tabular quartz veins and replacements containing galena, pyrite, chalcopyrite, and sphalerite. The sulfide minerals are most abundant at intersections of bedding-parallel and steep north-northwest-trending faults. The latter are subparallel with the Copper Creek fault. The steep faults are marked by mineralized gouge zones as much as 0.5 m wide.

Numerous silver-base metal deposits are present in the Copper Basin Formation (fig. 3), particularly within the marine basinal assemblages, which are designated the Little Copper and Drummond Mine Limestone Members. The members consist chiefly of argillaceous and micritic turbidites, respectively. The Little Copper Member is dominantly a thin- to medium-bedded, blocky weathering, dark-gray argillite and contains some interbeds of medium-gray sandstone and granule conglomerate (Paull and Gruber, 1977). It is 1,100 m or more thick. Analyses of samples from a regional geochemical survey of the Pioneer Mountains indicate that all argillaceous rocks in the Little Copper Member are significantly enriched in barium, many are enriched in boron, and many are slightly enriched in cobalt and manganese (Simons, 1981). Analyses of additional samples from the Muldoon–Star Hope area (table 2) indicate that, in many places, the Little Copper Member also is slightly enriched in zinc. The ubiquity and abundance of barium in the Little Copper imply syngenetic deposition of barite; however, primary detrital or chemical-precipitation textures of barite have not been recognized in the field or in thin sections. Stratabound barite is present as coarsely crystalline, light-colored, bedding-parallel and discordant stringers and lenses at several zones in the Little Copper Member, as well as in the overlying Drummond Mine Limestone Member (fig. 3). In the Hailey and Ketchum areas to the west, data for lead and sulfur isotopes from deposits in argillaceous rocks of the Devonian Milligen Formation have been interpreted as indicating derivation of metals from enclosing host rocks (Hall and others, 1978). By analogy, the inference has been made that the Little Copper Member may have provided a source for metals in mineral deposits higher in the stratigraphic sequence; however, lead-isotope ratios of galena samples from the Mackinaw, Muldoon Barium, and Star Hope mines hosted in the Copper Basin Formation apparently indicate that the primary source of lead is extraformational (Sanford, Wooden, and others, 1989; Sanford and Wooden, this volume). The actual source is speculative at present. Most of the lead is nonradiogenic and may have been derived from a depleted source in the lower crust or upper mantle or from igneous rocks, such as Eocene intrusive rocks, that were derived from them. The lead-isotope ratios for samples from the Muldoon–Star Hope area exhibit characteristics similar to those of samples from the Alta and Lava Creek mining districts 35 km to the northwest and 30 km to the southeast, respectively. The alignment of the three areas coincides with magnetic and gravity anomalies (Kleinkopf and others, 1989) that extend northwestward from the Great Rift shear zone in the Snake River Plain. This coincidence suggests that intrusive rocks along structures subparallel with, but older than, Neogene basin and range faults may have been sources for the lead.

Impure calcareous rocks in the Drummond Mine Limestone Member host the highest grade silver-base metal deposits in the Copper Basin Formation. In the Muldoon–Star Hope area, the Drummond Mine Member is approximately 800 m thick. The lower third of the member consists of medium-to thick-bedded impure micritic limestone and a few thin beds of argillite, chert, and granule conglomerate. The upper two-thirds consists of thin- to medium-bedded micritic limestone and interbedded argillite, calcareous siltstone, and fine-grained calcareous sandstone (Paull and others, 1972). The proportion of limestone in the unit decreases southward, and thick beds of medium- to coarse-grained sandstone and finer grained clastic rocks

Table 1. Major-element-oxide and minor-element composition, lithology, and approximate location of igneous rock samples from the Muldoon-Star Hope area, southern Pioneer Mountains, Idaho.

[Analyses by U.S. Geological Survey, Lakewood, Colo. Major element oxides (weight percent, volatile free) by X-ray spectroscopy; analysts, J.E. Taggart, A.J. Bartel, and D.F. Siems. Fe_2O_3* is total iron as Fe_2O_3; LOI (in weight percent) is loss on ignition, 900°C. Minor elements (parts per million) by optical spectroscopy; analyst, D.L. Fey. < indicates element was detected, but amount is less than lower limit of detection. Elements analyzed for but below limit of detection (given in parentheses, in parts per million) in all samples: Ag (2), Au (8), Bi (10), Cd (2), Ho (4), Mo (2), Sn (10), Ta (40), U (100)]

Sample no.	Wk3A	Wk4A	Wk7A	Wk7B	Wk14A	Wk18F	Wk18G
Major-element oxides							
SiO_2	63.9	78.8	61.6	63.2	62.8	64.1	53.2
Al_2O_3	15.9	12.9	14.7	15.0	15.0	15.8	12.1
Fe_2O_3*	5.90	0.51	6.34	6.01	6.05	5.92	9.30
MgO	3.22	0.27	4.58	3.39	3.43	2.42	9.96
CaO	3.07	1.82	4.44	3.46	4.85	3.33	9.85
Na_2O	3.43	0.17	3.10	3.17	3.78	3.54	1.77
K_2O	3.51	5.38	4.11	4.75	3.05	3.90	2.66
TiO_2	0.76	0.10	0.75	0.69	0.67	0.72	0.72
P_2O_5	0.30	<0.05	0.34	0.33	0.28	0.24	0.27
MnO	0.07	0.02	0.09	0.09	0.09	0.08	0.15
Volatiles							
LOI	4.14	3.59	2.14	2.20	3.84	2.75	4.47
Minor elements							
As	<10	<10	<10	<10	<10	<10	<10
Ba	3,700	1,260	639	1,880	2,330	1,590	1,960
Be	2	6	2	3	2	2	2
Ce	142	43	97	102	92	95	41
Co	25	3	27	22	23	18	51
Cr	81	2	205	110	139	36	527
Cu	14	3	40	38	25	18	90
Eu	2	<2	<2	<2	<2	<2	<2
Ga	21	18	19	20	20	21	17
La	87	23	59	61	56	58	26
Li	16	32	18	19	22	22	53
Nb	18	22	22	21	13	21	<4
Nd	59	16	42	43	41	40	25
Ni	36	6	85	36	42	15	155
Pb	24	37	16	22	20	19	14
Sc	11	<2	14	13	12	11	34
Sr	641	55	604	635	662	681	331
Th	13	34	19	21	19	23	4
V	97	5	116	108	98	102	179
Y	15	22	21	21	20	22	15
Yb	2	3	2	2	2	2	2
Zn	691	20	72	70	72	66	56

Wk3A	Plagioclase feldspar-hornblende porphyritic dacite dike, ridge above Mutual mine.
Wk4A	Coarse-grained quartz-potassium feldspar porphyritic rhyolite dike, ridge south of Garfield Canyon.
Wk7A	Equigranular hornblende-biotite granodiorite, ridge between Garfield and Muldoon Canyons.
Wk7B	Plagioclase feldspar-biotite dacite porphyry pluton, ridge between Garfield and Muldoon Canyons.
Wk14A	Plagioclase feldspar-biotite-hornblende porphyritic dacite dike, Muldoon Ridge.
Wk18F	Hornblende-biotite-plagioclase feldspar porphyritic dacite dike, near Muldoon Creek barite prospect.
Wk18G	Holocrystalline mafic dike (cuts 18F), near Muldoon Creek barite prospect.

Table 1. Major-element-oxide and minor-element composition, lithology, and approximate location of igneous rock samples from the Muldoon-Star Hope area, southern Pioneer Mountains, Idaho—Continued.

[Analyses by U.S. Geological Survey, Lakewood, Colo. Major element oxides (weight percent, volatile free) by X-ray spectroscopy; analysts, J.E. Taggart, A.J. Bartel, and D.F. Siems. Fe_2O_3* is total iron as Fe_2O_3; LOI (in weight percent) is loss on ignition, 900°C. Minor elements (parts per million) by optical spectroscopy; analyst, D.L. Fey. < indicates element was detected, but amount is less than lower limit of detection. Elements analyzed for but below limit of detection (given in parentheses, in parts per million) in all samples: Ag (2), Au (8), Bi (10), Cd (2), Ho (4), Mo (2), Sn (10), Ta (40), U (100)]

Sample no.	Wk27A	Wk33C	Wk40A	Wk40C	Wk44A	Wk47C	Wk48C	Wk49B
Major-element oxides								
SiO_2	77.4	40.0	58.3	66.3	75.0	77.4	64.4	62.4
Al_2O_3	12.7	10.4	12.6	15.0	14.1	12.8	15.0	15.7
Fe_2O_3*	0.98	10.7	8.23	4.61	1.93	0.89	3.72	6.21
MgO	0.12	13.1	6.06	2.21	0.54	<0.10	4.20	3.34
CaO	0.33	15.0	7.02	3.23	0.20	0.22	5.00	4.30
Na_2	3.22	1.13	2.34	3.73	2.95	3.40	3.41	3.72
K_2	5.13	4.67	4.05	4.08	4.91	5.12	3.57	2.93
TiO_2	0.03	2.56	0.89	0.57	0.25	0.03	0.53	0.94
P_2O_5	<0.05	2.32	0.40	0.57	0.08	<0.05	0.19	0.36
MnO	<0.02	0.17	0.10	0.08	<0.02	0.02	0.04	0.07
Volatiles								
LOI	0.99	9.83	4.72	0.54	1.66	0.60	0.66	2.25
Minor elements								
As	<10	<10	<10	<10	<10	<10	10	<10
Ba	138	3,890	1,140	1,330	1,660	252	2,560	1,430
Be	2	2	3	3	3	2	2	2
Ce	25	337	81	107	60	35	81	105
Co	<1	55	32	14	5	<1	16	22
Cr	8	460	374	65	8	4	235	122
Cu	2	85	70	32	7	4	18	6
Eu	<2	6	2	<2	<2	<2	<2	2
Ga	16	17	18	20	20	15	20	21
La	12	187	46	67	37	19	51	64
Li	24	27	55	25	13	11	14	72
Nb	22	120	7	15	12	16	14	10
Nd	12	152	44	42	23	17	34	44
Ni	<2	242	63	25	8	<2	66	44
Pb	22	10	21	31	25	22	24	8
Sc	<2	26	22	9	2	<2	12	11
Sr	50	2,660	597	514	122	49	781	667
Th	21	14	11	30	25	21	11	10
V	<2	209	160	79	14	<2	83	104
Y	19	22	24	20	11	16	14	15
Yb	2	2	3	2	1	2	2	2
Zn	23	75	82	81	207	10	63	66

Wk27A	Quartz porphyritic rhyolite dike, ridge above Rippeto claim.
Wk33C	Biotite-bearing lamprophyric dike, ridge north of Eagle Bird mine.
Wk40A	Fine-grained, holocrystalline intermediate dike, upper Muldoon Creek.
Wk40C	Coarse-grained, equigranular, fresh quartz monzonite (Muldoon stock), upper Muldoon Creek.
Wk44A	Weakly porphyritic rhyolite dike with biotite-bearing groundmass, ridge north of Little Copper Creek.
Wk47C	Quartz porphyritic rhyolite dike, Bent Pine Tree claim (Star Hope Creek).
Wk48C	Plagioclase feldspar porphyritic dacite dike, head of Star Hope Creek.
Wk49B	Plagioclase feldspar, coarsely porphyritic dacite dike (cuts porphyritic rhyolite dikes of sample 47C), north of Drummond mine.

Table 2. Minor-element composition of unmineralized samples from the Little Copper, Drummond Mine Limestone, and Green Lake Limestone Members of the Mississippian Copper Basin Formation, Muldoon-Star Hope area, southern Pioneer Mountains, Idaho.

[Most samples are composite surface samples from a short interval of stratigraphic section. Results in parts per million. Analyses by U.S. Geological Survey, Lakewood, Colo. Au by flame atomic absorption; analyst, B. Roushey. Ag, As, Bi, Cd, Cu, Mo, Pb, Sb, and Zn by partial inductively coupled plasma-atomic emission spectrometry; other elements by total inductively coupled plasma-atomic emission spectrometry; analysts: D.L. Fey and J.M. Motooka. N indicates analysis was performed, but element was not detected (lower limit of detection given in parentheses). < indicates element was detected, but amount is less than lower limit of detection. Elements present but below detection limits in all samples (lower limit of detection given in parentheses): Au (0.05), Bi (0.60), Eu (4), Ho (8), Sn (20), Ta (80), U (200), Yb (2)]

Sample no.	Wk2A	Wk3B	Wk8A	Wk8B	Wk8C	Wk8D	Wk9A	Wk9B	Wk9C	Wk9E	Wk10A	Wk14B	Wk14C
Mn	217	139	76	27	23	38	35	241	164	149	66	176	121
Ag	0.06	0.35	0.19	0.23	0.24	1.1	0.17	0.16	0.14	0.26	0.27	0.11	0.17
As	17	22	17	15	12	17	19	18	9.4	22	28	6.4	8.5
Ba	2,390	3,390	12,000	5,380	4,870	3,410	3,020	12,200	10,600	2,220	3,060	1,160	2,110
Be	<2	<2	<2	3	2	<2	<2	<2	<2	<2	<2	<2	<2
Cd	12	1.3	0.052	0.099	0.032	0.084	N(0.030)	0.090	0.050	0.091	0.14	0.22	0.14
Ce	10	16	<8	49	44	<8	45	17	9	35	31	27	11
Co	8	8	14	8	7	5	6	17	13	9	4	6	7
Cr	58	55	3	104	98	31	75	44	25	52	64	27	42
Cu	19	40	160	28	24	14	15	31	19	31	22	11	16
Ga	<8	<8	<8	14	12	<8	10	<8	<8	<8	9	<8	<8
La	13	12	<4	29	28	6	28	10	7	20	17	12	11
Li	<4	5	<4	<4	<4	6	<4	7	<4	7	7	6	7
Mo	1.2	0.55	0.67	1.1	1.3	0.60	1.3	0.97	0.89	1.5	1.5	0.42	0.62
Nb	<8	<8	<8	9	16	<8	<8	<8	<8	<8	<8	<8	<8
Nd	12	14	<8	25	23	<8	19	13	<8	17	16	8	10
Ni	27	17	10	21	24	15	18	32	13	36	21	14	25
Pb	150	77	34	6.1	9.1	30	4.9	18	11	6.8	11	13	14
Sb	4.0	1.6	2.7	3.6	2.9	2.2	2.0	1.7	0.72	3.5	6.2	N(0.60)	1.0
Sc	6	5	<4	8	9	<4	6	<4	<4	4	6	<4	<4
Sr	233	215	585	59	77	126	44	259	64	31	43	406	184
Th	<8	<8	<8	8	<8	<8	<8	<8	<8	<8	<8	<8	<8
V	101	118	14	260	247	80	165	108	51	130	164	60	106
Y	11	11	<4	14	10	<4	7	8	5	9	9	8	8
Zn	600	110	8.7	12	2.1	15	0.085	14	8.4	37	51	49	55

Wk2A	Sandstone sample from near top of Little Copper Member, roadcut on north side of Mutual Gulch.
Wk3A	Composite sample, lower third of Drummond Mine Limestone Member, upsection from sample 2A.
Wk8A	Barite veins in upper third, Little Copper Member, roadcut on ridge between Garfield Canyon and Mutual Gulch.
Wk8B	Upper third, Little Copper Member, downsection from sample 8A.
Wk8C	Upper third, Little Copper Member, downsection from sample 8B.
Wk8D	Upper third, Little Copper Member, downsection from sample 8C.
Wk9A	Middle third, Little Copper Member, downsection from sample 8C.
Wk9B	Middle third, Little Copper Member, upsection from sample 9A.
Wk9C	Middle third, Little Copper Member, upsection from sample 9B.
Wk9E	Upper third, Little Copper Member, upsection from sample 9C.
Wk10A	Upper third, Little Copper Member, hillside outcrops 0.5 km east of Scorpion Antimony prospect.
Wk14B	Middle third, Drummond Mine Limestone Member, Muldoon Ridge at head of Deep Gulch.
Wk14C	Middle third, Drummond Mine Limestone Member, downsection from sample 14B.

Sample no.	Wk15A	Wk15B	Wk16A	Wk16B	Wk17A	Wk18A	Wk18B	Wk18C	Wk18E	Wk19A	Wk22B	Wk23A	Wk23B
Mn	71	20	201	186	141	463	393	217	581	183	35	58	81
Ag	0.36	0.61	0.12	0.080	0.13	N(0.045)	0.059	0.12	0.046	0.23	0.36	0.15	0.13
As	3.8	3.4	11	2.8	4.6	8.2	11	9.2	5.0	7.7	11	17	12
Ba	728	2,150	1,060	904	958	4,990	4,200	17,800	563	1,730	2,250	2,030	1,700
Be	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2
Cd	0.30	0.086	0.96	0.17	0.17	4.8	0.45	0.39	0.55	0.93	0.39	0.20	0.29
Ce	42	27	20	13	11	11	13	27	<8	10	14	35	61
Co	7	<2	7	5	4	9	8	23	3	6	5	4	7
Cr	42	71	35	26	42	30	33	75	11	28	42	55	63
Cu	20	16	21	6.9	16	9.7	24	17	4.8	58	6.3	21	33
Ga	<8	9	<8	<8	<8	<8	<8	9	<8	<8	<8	<8	9
La	18	19	17	12	9	11	13	17	10	13	12	22	34
Li	8	4	6	5	6	8	14	11	<4	5	<4	6	8
Mo	1.4	2.2	3.1	0.66	0.66	0.73	0.37	0.62	0.17	0.33	5.8	1.1	1.2
Nb	<8	<8	<8	27	<8	<8	<8	<8	<8	<8	<8	<8	<8
Nd	14	15	18	12	10	12	11	18	15	17	12	16	27
Ni	43	11	76	14	18	16	13	76	9	12	28	13	31
Pb	12	64	12	14	13	37	21	12	35	26	27	34	16
Sb	0.83	0.82	1.5	0.82	0.91	N(0.60)	0.66	0.61	N(0.60)	1.5	1.4	2.4	2.1
Sc	<4	6	4	<4	4	<4	<4	4	<4	4	<4	5	6
Sr	24	21	235	338	58	189	159	235	345	770	94	25	29
Th	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8
V	108	213	115	48	106	70	86	115	14	70	234	134	153
Y	11	7	12	8	8	7	11	12	7	9	6	7	10
Zn	120	14	84	44	63	810	200	84	120	130	12	120	43

Wk15A	Middle third, Drummond Mine Limestone Member, downsection from sample 14C.
Wk15B	Lower third, Drummond Mine Limestone Member, downsection from sample 15A.
Wk16A	Lower third, Drummond Mine Limestone Member, spur ridge between Boundary Gulch and Deep Gulch.
Wk16B	Lower third, Drummond Mine Limestone Member, downsection from sample 16A.
Wk17A	Upper third, Little Copper Member, downsection from sample 16B.
Wk18A	Laminated limestone in upper third, Drummond Mine Limestone Member, upsection from upper barite prospect on east side of Muldoon Creek.
Wk18B	Dark siliceous limestone in Drummond Mine Limestone Member immediately above barite lenses at upper prospect, east side of Muldoon Creek.
Wk18C	Barite-bearing turbidite in upper third of Drummond Mine Limestone Member, upper prospect, east of Muldoon Creek.
Wk18E	Pyritized limestone downsection from barite lenses at upper prospect, east side of Muldoon Creek.
Wk19A	Calcareous siltstone in upper third of Drummond Mine Limestone Member, lower barite prospect, east side of Muldoon Creek.
Wk22B	"Sooty" pyritic argillite in lower third of Drummond Mine Limestone Member, near upper adit, Boundary Gulch.
Wk23A	Upper third, Little Copper Member, ridge between Mutual Gulch and Boundary Gulch.
Wk23B	Upper third, Little Copper Member, downsection from sample 23A.

Table 2. Minor-element composition of unmineralized samples from the Little Copper, Drummond Mine Limestone, and Green Lake Limestone Members of the Mississippian Copper Basin Formation, Muldoon-Star Hope area, southern Pioneer Mountains, Idaho—Continued.

[Most samples are composite surface samples from a short interval of stratigraphic section. Results in parts per million. Analyses by U.S. Geological Survey, Lakewood, Colo. Au by flame atomic absorption; analyst, B. Roushey. Ag, As, Bi, Cd, Cu, Mo, Pb, Sb, and Zn by partial inductively coupled plasma-atomic emission spectrometry; other elements by total inductively coupled plasma-atomic emission spectrometry; analysts: D.L. Fey and J.M. Motooka. N indicates analysis was performed, but element was not detected (lower limit of detection given in parentheses). < indicates element was detected, but amount is less than lower limit of detection. Elements present but below detection limits in all samples (lower limit of detection given in parentheses): Au (0.05), Bi (0.60), Eu (4), Ho (8), Sn (20), Ta (80), U (200), Yb (2)]

Sample no.	Wk24A	Wk24B	Wk25A	Wk25B	Wk25C	Wk30A	Wk32B	Wk32C	Wk33A	Wk33B	Wk33D	Wk34A	Wk35A
Mn	151	61	41	127	114	135	145	173	119	268	213	169	292
Ag	0.18	0.11	0.070	0.20	0.11	0.15	0.14	0.14	0.21	0.18	0.065	0.13	0.083
As	16	30	32	28	22	16	7.9	7.5	15	8.1	11	9.8	8.1
Ba	2,360	1,620	1,440	1,800	1,370	1,570	891	1,390	1,320	824	962	993	521
Be	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2
Cd	0.52	0.25	0.31	0.19	0.20	0.062	0.29	0.20	0.65	0.18	0.20	0.18	0.12
Ce	52	39	28	39	37	41	10	18	16	21	20	17	18
Co	4	3	3	5	4	3	5	7	5	8	6	5	6
Cr	80	68	57	74	58	61	34	50	45	34	31	33	32
Cu	38	24	18	30	25	22	11	20	23	17	7.4	16	13
Ga	12	8	<8	10	<8	9	<8	<8	<8	<8	<8	<8	<8
La	28	25	17	23	23	26	12	16	13	18	18	16	16
Li	7	8	7	10	8	10	11	14	9	10	7	14	10
Mo	1.2	0.88	1.1	1.6	1.4	0.94	0.66	0.61	1.0	0.93	0.30	0.45	0.43
Nb	<8	<8	<8	9	<8	<8	<8	11	<8	<8	<8	<8	<8
Nd	25	22	17	18	18	24	9	16	11	16	17	16	15
Ni	28	24	24	28	38	31	20	17	19	24	15	18	15
Pb	15	15	48	9.1	8.6	20	9.7	3.5	34	13	14	7.9	13
Sb	2.3	2.2	4.0	3.6	3.0	7.0	0.74	0.75	1.4	1.5	1.1	N(0.60)	2.2
Sc	8	6	5	7	5	5	<4	5	<4	<4	4	<4	<4
Sr	39	34	23	28	27	40	283	520	106	474	572	478	476
Th	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8
V	199	186	156	188	156	157	81	103	125	72	52	79	61
Y	11	10	8	10	9	11	7	11	9	11	11	8	9
Zn	130	67	67	68	76	28	52	62	170	66	41	48	49

Wk24A	Middle third, Little Copper Member, downsection from sample 23B.
Wk24B	Middle third, Little Copper Member, downsection from sample 24A.
Wk25A	Lower third, Little Copper Member, downsection from sample 24B.
Wk25B	Lower third, Little Copper Member, downsection from sample 25A.
Wk25C	Lower third, Little Copper Member, downsection from sample 25B
Wk30A	Upper third, Little Copper Member, west end of ridge between South Fork of Little Copper Creek and Garfield Canyon.
Wk32B	Upper third, Drummond Mine Limestone Member, ridge between South Fork of Little Copper Creek and Garfield Canyon.
Wk32C	Upper third, Drummond Mine Limestone Member, downsection from sample 32B.
Wk33A	Upper third, Drummond Mine Limestone Member, downsection from sample 32C.
Wk33B	Upper third, Drummond Mine Limestone Member, downsection from sample 33A.
Wk33D	Middle third, Drummond Mine Limestone Member, downsection from sample 33B.
Wk34A	Lower third, Drummond Mine Limestone Member, downsection from sample 33D.
Wk35A	Upper third, Drummond Mine Limestone Member, ridge between Muldoon Canyon and Little Copper Creek.

Sample no.	Wk35B	Wk36A	Wk36B	Wk37A	Wk37B	Wk37C	Wk46A	Wk46B	Wk48A	Wk48B	Wk49A	Wk51A
Mn	220	218	248	339	174	253	211	339	159	166	247	46
Ag	0.13	0.066	0.17	0.091	0.22	0.13	0.090	0.17	0.12	0.30	0.099	0.39
As	6.8	6.7	2.6	25	11	82	4.7	1.2	6.7	21	5.6	23
Ba	1,800	1,190	1,380	2,110	1,960	2,000	2,220	9,210	1,650	3,410	1,960	1,570
Be	<2	<2	<2	<2	<2	2	<2	<2	<2	<2	<2	<2
Cd	0.14	0.18	0.23	0.073	0.069	N(0.030)	0.13	0.16	0.13	0.90	0.16	4.3
Ce	37	20	29	40	28	26	29	37	20	36	26	23
Co	7	7	7	7	8	10	8	13	7	11	8	3
Cr	53	42	59	79	81	91	58	83	51	81	60	63
Cu	13	7.2	12	18	40	26	13	26	21	50	16	37
Ga	<8	<8	<8	12	13	13	8	12	<8	10	9	10
La	25	19	20	24	18	15	22	23	18	25	22	16
Li	14	15	18	21	32	53	14	20	12	12	16	14
Mo	0.53	0.23	0.49	1.6	1.1	1.0	0.49	0.79	0.58	0.94	0.29	3.5
Nb	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8	<8
Nd	27	18	24	20	19	15	23	21	18	20	21	14
Ni	21	14	19	30	36	54	21	33	25	38	20	33
Pb	5.1	8.6	7.9	4.6	14	6.3	5.3	9.9	10	37	11	3.8
Sb	0.88	0.79	N(0.60)	4.0	3.2	3.5	0.70	N(0.60)	N(0.60)	N(0.60)	N(0.60)	0.81
Sc	5	5	7	8	9	9	7	8	5	7	7	5
Sr	701	964	691	67	100	117	577	118	569	454	942	43
Th	<8	<8	<8	9	<8	9	<8	<8	<8	<8	<8	<8
V	104	69	107	204	216	227	118	155	120	218	105	193
Y	11	12	14	10	10	10	13	16	12	13	11	10
Zn	43	36	32	46	36	19	64	88	17	120	31	44

Wk35B Upper third, Drummond Mine Limestone Member, downsection from sample 35A.
Wk36A Middle third, Drummond Mine Limestone Member, downsection from sample 35B.
Wk36B Lower third, Drummond Mine Limestone Member, downsection from sample 36A.
Wk37A Upper third, Little Copper Member, downsection from sample 36B.
Wk37B Upper third, Little Copper Member, downsection from sample 37A.
Wk37C Middle third, Little Copper Member, downsection from sample 37B.
Wk46A Upper half, Green Lake Limestone Member, Copper Basin Formation, west of Green Lake.
Wk46B Lower half, Green Lake Limestone Member, downsection from sample 46A.
Wk48A Upper third, Drummond Mine Limestone Member, head of Star Hope Creek.
Wk48B Upper third, Drummond Mine Limestone Member, downsection from sample 48A.
Wk49A Middle third, Drummond Mine Limestone Member, downsection from sample 48B.
Wk51A Iron-stained argillite of Glide Mountain sequence, Copper Basin Formation near Blue Sky prospect, western tributary to upper Muldoon Canyon.

are present in the middle part of the Drummond Mine Member (Dover, 1981). Geochemical analyses of unmineralized samples (table 2) indicate that, in many places, the member is significantly enriched in barium and zinc, particularly in its lower and upper parts. Analyses of mineralized samples (K.W. Kunkel, Idaho State University, written commun., 1990) (table 3) indicate that, in general, gold and silver and their companion elements bismuth, antimony, and selenium are more abundant near the base of the Drummond Mine Limestone Member. Silver-base metal deposits in the unit are, however, stratabound at several horizons (fig. 3). Although micritic limestone beds are favored hosts, no particular micrite horizon is especially favorable. In detail, deposits are localized or enriched where bedding-plane and crosscutting steep faults intersect, and deposits may extend along faults of either orientation. Many deposits contain coarse-grained sulfide and calc-silicate minerals in contact with virtually unrecrystallized strata, characteristic of zinc-lead skarns that are distant from known or suspected intrusions (Einaudi and others, 1981).

The Lower Mississippian Drummond Mine Limestone Member of the Copper Basin Formation is overlain gradationally by an upper clastic unit more than 1,400 m thick consisting of gritty sandstone, granule conglomerate, and argillite (fig. 3). According to Nilsen (1977), the upper clastic unit includes the Scorpion Mountain Formation and the lower part of the Muldoon Canyon Formation of Paull and others (1972). In the vicinity of the Star Hope mine, the basal part of this unit, which hosts the deposits, consists of carbonaceous argillite and quartzite and thin calcareous interbeds. The strata are strongly folded, brecciated, and silicified and are intruded by swarms of aplite and quartz porphyry dikes that trend northeast.

At the Black Spar properties, brecciated and sheared light-gray andesite flows of the Challis Volcanic Group host weakly mineralized silver-base metal veins. Because the underground workings have been mostly inaccessible for decades, it was not possible during this study to reexamine the character of the host rocks at the main mineralized zones.

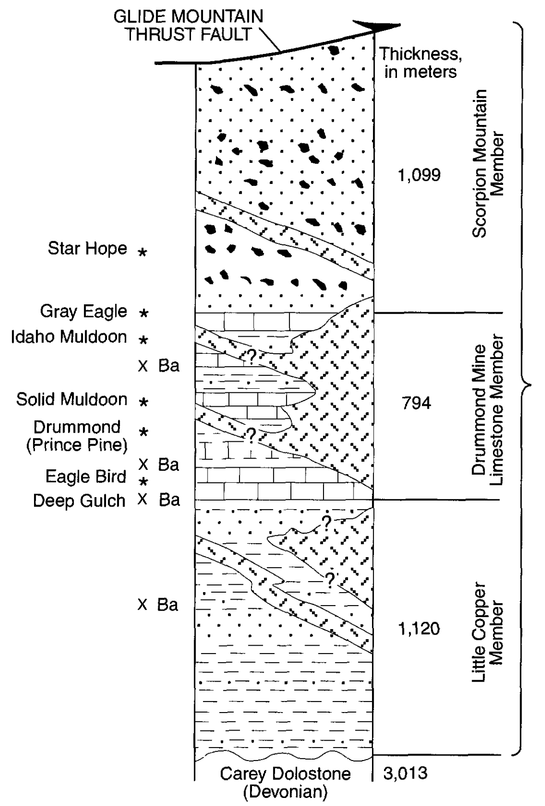


Figure 3. Schematic stratigraphy of the Copper Basin Formation in the Muldoon–Star Hope area, south-central Idaho. Asterisk (*) indicates approximate stratigraphic level of named mine or prospect; xBa indicates approximate level of barite mine or prospect. Locations of mines and prospects are shown in figure 2.

HOST STRUCTURES

Mineral deposits in the Muldoon–Star Hope area are localized by superimposed structures of Cretaceous and Tertiary age. The older structures include asymmetric folds and thrust faults that presumably formed during episodes of east-directed compression of the Cretaceous Sevier orogeny (Dover, 1981). Strata in the hanging wall of the Pioneer fault (the Milligen(?) Formation) and the underlying Copper Basin plate form a broad anticline trends north-northwest. Regionally, broad and gentle folds are typical of the Copper Basin thrust plate. Widespread small-scale flexures that have the same vergence and bedding-parallel faults, which are themselves folded, are probably also Cretaceous in age. The overlying Glide Mountain plate of the Copper Basin Formation is characterized by pervasive folding on small and intermediate scales (Dover, 1981). These folds also generally trend north-northwest, are asymmetrical or overturned to the east, and are probably related to the thrusting. In many places the base of the Glide Mountain plate is a broad zone of intense shearing, tight folding, and silicification, and it locally contains slices and rolled blocks of structurally underlying rocks. At the Blue Sky prospect southwest of Muldoon Canyon (fig. 2), the Glide Mountain fault zone incorporates strongly deformed blocks of Drummond Mine Limestone Member. Fractured fold hinges, bedding-parallel faults, and thrust zones of Cretaceous age provided extensive "prepared ground" for subsequent mineral deposition.

Younger, steep faults cut the folded and thrusted Paleozoic sequences in many places; locally, they also displace rocks of the Challis Volcanic Group. Most are of small to intermediate oblique-slip displacement. The steep faults tend to follow one of three orientations, about N. 20° W., N. 50° E., or N. 80° E., which also are the dominant trends of Eocene dikes that cut the Paleozoic rocks. The faults and dikes likely formed during Eocene regional extension and magmatism. North-northwest-trending faults such as the Copper Creek fault generally have been attributed to Neogene Basin and Range extension, but their approximate parallelism with trends of some Eocene dikes may indicate an earlier ancestry. In addition, the trace of the Copper Creek fault is cut at almost right angles by dikes of probable Eocene age (Nelson and Ross, 1969). Anderson and Wagner (1946) first noted that many silver-base metal deposits in the Little Wood River district are near intersections of steep faults and bedding-plane reverse faults; recent work in Garfield Canyon (Kunkel, 1989) indicates that highest grade replacement ore at the Eagle Bird mine and nearby deposits (fig. 2) is invariably at intersections of bedding-plane and east-northeast-trending steep faults. Lower grade deposits may extend considerable distances along steep fault surfaces away from the richer, but discontinuous replacement deposits.

MINERAL DEPOSITS

DEPOSIT TYPES

Mineral deposits in the Muldoon–Star Hope area are diverse but can be characterized generally as (1) polymetallic replacements or distal skarns (chiefly in carbonate rocks), (2) base- and precious-metal veins and replacements (which can be further subdivided, based on whether the host rock lithology is volcanic, intrusive, or calcareous, arenaceous, or pelitic sedimentary), and (3) bedded replacements and veins of barite. All gradations between these idealized types are present locally. For example, the largest replacement bodies are along structural or lithologic pathways tens to hundreds of meters from intrusive rocks, but in their calc-silicate mineralogy and coarse-grained texture many of the replacement bodies are similar to distal skarns. In addition, in many places where polymetallic replacements have been sheared, base- and precious-metal veins extend into the crosscutting shears.

POLYMETALLIC REPLACEMENT DEPOSITS OR DISTAL SKARNS

Description.—The largest and highest grade deposits in the Muldoon–Star Hope area are polymetallic replacements or skarns. These deposits form tabular elongate lenses or ovoid, pipelike bodies that follow bedding-parallel faults in host rocks. Hosts generally are chemically reactive calcareous rocks, although the largest orebodies at the Solid Muldoon mine replaced sandstone in the footwall and porphyritic rhyolite in the hanging wall. In detail, the deposits are very irregular and discontinuous. At the Eagle Bird mine, replacement by sulfide minerals apparently is more complete where dips of controlling beds and structures are flexed and thus less steep (Anderson and Wagner, 1946). At other deposits, bedding-parallel replacement deposits dip as steeply as 70°. The primary ore is present as massive or partial replacement of country rock, closely spaced veinlets, breccia fillings, or disseminations of sulfide minerals; all primary ore has been oxidized. The deposits have produced (in order of value) silver, lead, zinc, copper, and gold. As much as 4 percent Sb is present in some ore, but no production of antimony has been recorded.

Type example.—Eagle Bird mine, Little Wood River district

Other examples.—Little Wood River district: Drummond mine, Gray Eagle prospect, Hecla-Carrier, Idaho Muldoon mine, and Solid Muldoon (in part). Copper Basin district: Blue Sky prospect and Mackinaw Group

Mineralogy and texture.—Sulfide minerals include galena, sphalerite, pyrite, chalcopyrite, arsenopyrite, marcasite, and complex antimonides. Pyrrhotite may be present in a few deposits. Preliminary electron microprobe data indicate that silver apparently resides principally in galena, although minor tetrahedrite and other discrete silver-bearing minerals may be present locally. Oxidized minerals include cerussite, anglesite, scorodite, malachite, and iron oxides. Gangue minerals include calcite, quartz, diopside, grossularite, epidote, idocrase, augite, sericitized feldspar, and fluorite. In many of the deposits, the sulfide, silicate, and carbonate minerals form very coarse grained euhedral or subhedral mosaics in favorable calcareous beds, whereas adjacent, more argillaceous beds are only slightly silicified and virtually unrecrystallized. In other deposits, the texture is granular, few ore minerals are euhedral, and the wallrocks are bleached and sericitized. Such is the case at the Solid Muldoon, where porphyritic rhyolite wallrocks are silicified and sericitized. In deposits near the margin of the Garfield stock, early formed sulfide minerals such as galena are thoroughly brecciated, whereas later sulfide phases are not.

Characteristics.—Only very limited modern analytical information is available for mineral deposits of the Muldoon–Star Hope area. In general, pyrite and arsenopyrite are paragenetically older and are followed, in sequence, by sphalerite, galena, and chalcopyrite. Preliminary isotopic analyses of galena from replacement and vein deposits indicate a nonsedimentary source for the lead (Sanford, Wooden, and others, 1989; Sanford and Wooden, this volume). As discussed previously in the section on host rocks, the most likely source of lead is the Eocene plutonic rocks that were intruded into the sedimentary rocks. At most replacement deposits, alteration is not extensive in calcareous host rocks, although the rocks may be bleached and silicified. Where host rocks include siliceous igneous rocks, such as at the Solid Muldoon and Star Hope mines, quartz and sericite alteration is widespread. Although no geothermometric or geochronologic measurements have been made on minerals from the replacement deposits, the relative age of mineralization can be constrained by observations at two deposits. At the Blue Sky prospect, blocks of Lower Mississippian

Table 3. Minor-element composition of mineralized samples from selected mines and prospects in the Muldoon-Star Hope area, southern Pioneer Mountains, Idaho.

[All samples are composite surface samples from "best" material exposed in workings or on dumps. In parts per million unless otherwise indicated. Analyses by U.S. Geological Survey, Lakewood, Colo. (P) indicates element by partial inductively coupled plasma-atomic emission spectrometry; all other elements except Au by optical spectroscopy (analysts: D.L. Fey, J.M. Motooka, and P.H. Briggs); Au by flame atomic absorption (analysts: B. Roushey, R.M. O'Leary, and R.H. Hill). N indicates analysis wasperformed, but element wasnot detected (lower limit of detection in parentheses). < indicates element was detected, but amount is less than lower limit of detection. Elements analyzed for but below limit of detection (given in parentheses, in parts per million) in all samples: Eu (4), Ho (8), Sn (20), Ta (80), U (200), Yb (2)]

Sample no.	Wk11	Wk20	Wk21A	Wk22A	Wk34B	Wk40B	Wk40C
Mn	125	78	176	28	14	2,000	582
Ag	62	291	<4	759	61	329	816
As	25,000	12,100	230	35,600	340,000	1,000	300
Ba	18	471	56	924	27	277	787
Be	<2	<2	<2	<2	<2	<2	<2
Bi	<20	30	<20	150	<20	20	20
Cd	85	9	<4	4	16	381	926
Ce	<8	<8	<8	22	<8	17	23
Co	48	9	<2	25	<2	13	17
Cr	18	27	19	56	9	61	50
Cu	994	2,290	19	2,170	638	354	798
Ga	<8	<8	<8	<8	<8	<8	9
La	<4	<4	<4	14	<4	12	13
Li	5	<4	7	6		14	4
Mo	<4	5	<4	47	<4	<4	<4
Nb	<8	<8	<8	<8	<8	<8	9
Nd	<8	<8	<8	11	<8	12	12
Ni	12	4	4	49	6	13	18
Pb	20,000	83,500	360	135,000	5,950	92,400	137,000
Sc	<4	<4	<4	<4	<4	4	<4
Sn	<20	<20	<20	40	20	<20	<20
Sr	10	23	20	59	4	225	131
V	49	55	42	82	18	74	82
Y	<4	<4	<4	<4	<4	9	11
Zn	10,800	988	42	1,010	771	55,800	125,000
Ag (P)	N(1.5)	200	2.9	540	N(1.5)	240	540
As (P)	>24,000	68	190	30,000	>24,000	290	N(6.6)
Bi (P)	N(20)	N(6.6)	N(0.60)	81	N(20)	N(6.6)	N(6.6)
Cd (P)	56	6.4	0.24	1.2	N(1.0)	370	926
Cu (P)	300	1,800	19	1,600	N(1.0)	230	200
Mo (P)	N(3.0)	N(0.99)	1.0	52	N(3.0)	N(0.99)	N(0.99)
Pb (P)	30,000	13,000	820	18,000	8,600	14,000	18,000
Sb (P)	N(20)	160	3.6	320	290	160	410
Zn (P)	9700	N(0.33)	34	790	530	1700	70
Au	0.40	0.50	N(0.05)	0.10	0.15	<0.05	0.05
Pt	<10	<2	1.7	<5	<50	<5	<5
Pd	<20	7.0	<0.8	16	<80	<8	<8
Rh	<10	3.0	<0.5	5.0	10.0	<5	<5
Ru	<10	<6	<2	<20	170	<20	<3
Ir	<10	<2	<0.5	10	<50	<5	<5

Wk11A Idaho Muldoon mine, open cut that extends south from adits. Coarse-grained galena- and arsenopyrite-bearing skarn that has thick scorodite coatings; host rock is calcareous turbidite.

Wk20 Lucky Boy mine, west side of Argosy Creek: vuggy, sulfide-bearing quartz-carbonate veins; host rock is fractured calcareous and arenaceous turbidite cut by pyritized rhyolite porphyry dikes.

Wk21A Surface pits, east side of Argosy Creek, opposite Lucky Boy mine. Weakly iron stained quartz veins in calcareous sandstone that contains widely disseminated galena.

Wk22A Boundary Gulch, highest adit.Granular and vuggy iron- and copper-stained quartz gangue containing abudant galena and arsenopyrite; host rocks are highly cleaved carbonaceous argillite and calcareous siltstone cut by dacite dikes.

Wk34B Eagle Bird mine. Coarse-grained to granular, vuggy to massive, sulfide-rich material including galena and arsenopyrite, heavily coated with scorodite; host rock is calcareous turbidite containing patchy silicification.

Wk40B Grey Eagle claims, upper workings. Multiple horizons of coarse-grained to granular galena- and sphalerite-rich skarn; host rock is thin-bedded, locally silicified calcareous turbidite that is cut by porphyritic rhyolite dikes.

Wk40C Grey Eagle claims, lower adit. Coarse-grained, massive galena- and sphalerite-rich material; host rock is calcareous turbidite intruded by microdiorite dike.

Table 3. Minor-element composition of mineralized samples from selected mines and prospects in the Muldoon-Star Hope area, southern Pioneer Mountains, Idaho—Continued.

Sample no.	Wk45A	Wk47A	Wk50A	Wk53A	RW88-49	RW88-53	RW88-59
Mn	169	21,100	6,270	249	170	170	200
Ag	<4	71	<4	9	<4	170	<4
As	110	20	70	30	60	2300	100
Ba	1,010	222	571	520	2,000	400	2,100
Be	<2	12	<2	<2	<2	<2	<2
Cd	<4	162	<4	<4	<4	12	<4
Ce	53	16	20	<8	21	21	26
Co	19	11	10	26	8	2	9
Cr	76	30	65	22	120	70	78
Cu	100	35	1,180	1,570	41	460	35
Ga	9	37	23	<8	<8	12	12
La	29	14	18	6	15	13	17
Li	17	45	6	9	14	15	12
Mo	<4	<4	5	<4	53	33	<4
Nb	8	<8	<8	<8	<8	<8	<8
Nd	26	12	12	<8	15	13	17
Ni	78	22	17	42	130	12	34
Pb	230	17,000	64	1,860	100	32,000	110
Sc	7	<4	7	<4	7	5	8
Sn	<20	300	250	<20	<20	70	<20
Sr	95	127	243	63	71	41	140
Th	<8	<8	<8	<8	<8	9	8
V	166	65	117	83	580	140	200
Y	18	10	16	<4	13	<4	14
Zn	140	20,100	238	156	550	1,000	130
Ag (P)	0.87	45	1.5	3.3			
As (P)	93	380	54	1,000			
Bi (P)	N(0.60)	120	1.1	N(20)			
Cd (P)	1.0	150	0.95	N(1.0)			
Cu (P)	83	N(1.0)	980	1,200			
Mo (P)	2.1	9.1	0.31	3.9			
Pb (P)	400	25,000	110	2,900			
Sb (P)	1.6	N(20)	1.2	N(20)			
Zn (P)	140	17,000	160	43			
Au	N(0.05)	N(0.05)	N(0.05)	N(0.05)	N(0.05)	0.15	N(0.05)
Pt	<2	<1	0.8	<5			
Pd	<3	<2	1.7	<8			
Rh	<2	<1	<0.5	<5			
Ru	<6	<3	<2	<20			

Wk45A — Main (north) adit, Candy Cane claim group. Weakly iron stained dense argillite , no visible sulfide minerals; adit follows microdiorite dike that cuts contact between argillite and overlying calcareous strata of Green Lake Limestone Member of Copper Basin Formation.

Wk47A — Adit at end of jeep road on east side of Star Hope Creek, opposite Bent Pine Tree claim. "Sooty" granular galena- and sphalerite-bearing material; adit follows contact between porphyritic rhyodacite dike and dark, silicified calcareous turbidite; disseminated chalcopyrite and weak copper staining at contact.

Wk50A — Blue Sky prospect. Coarse-grained, sulfide-bearing calc-silicate skarn showing conspicuous copper staining; host rock is strongly folded and partly silicified tectonic lenses of calcareous turbidite cut by porphyritic rhyolite dikes.

Wk53A — High (south) adit, Candy Cane claim group. Granular pyrite, chalcopyrite, arsenopyrite, and galena in quartz gangue; ore minerals are present as small lenses and fracture coatings along steep shears in silicified lithic sandstone immediately above contact with carbonaceous siltite and argillite.

RW88-49 — Solid Muldoon mine. Chip sample from second dump from bottom, which is composed mainly of dark carbonaceous argillite and very minor amounts of vein quartz and sulfide minerals.

RW88-53 — Solid Muldoon mine. Chip sample from uppermost large dump, which is light colored and composed of silicified and bleached rock and abundant vein quartz. Altered rock contains disseminated pyrite, and vein quartz contains minor amounts of galena.

RW88-58 — Mutual mine. Composite chip sample from lower large dump, which contains a mixture of carbonaceous argillite and altered felsic dike rocks.

Drummond Mine Limestone Member as large as 50 m have been incorporated in the Sevier-age Glide Mountain thrust zone. The blocks are strongly deformed but have been partly silicified and partly replaced by coarse-grained mosaics of sulfide and silicate minerals, indicating that mineralization postdates Cretaceous emplacement of the thrust. At the Solid Muldoon mine, leucocratic dike rocks are strongly altered and weakly mineralized, whereas intermediate dike rocks are virtually unaltered. Although neither set of dikes has been dated isotopically, each is almost certainly Eocene, roughly the same age as the Muldoon stock (47 Ma, K-Ar on biotite; R.E. Zartman, *in* Dover, 1981). Hence, the orebodies likely are also of Eocene age.

Orebodies.—The sulfide replacement bodies generally are short, small, and irregular and form thin tabular seams and stringers of ore-grade material. Typical dimensions are on the order of 10–20 cm thick and a few meters long, but several bodies might be located in close proximity either vertically or laterally where structural and lithologic conditions are favorable. The main orebody at the Eagle Bird mine, which has not been explored completely, is exceptionally large. It consists of several bands and stringers that are, in aggregate, more than 3 m thick, and it extends more than 30 m downdip (Anderson and Wagner, 1946). The outcropping orebody at the Solid Muldoon mine is reported to have been 3.5–6 m thick (E.H. Finch, *in* Umpleby, 1917). It was developed through seven levels totaling more than 610 m of drifts and 275 m of crosscuts that have been inaccessible since early this century (Tuchek and Ridenour, 1981). One stope on the No. 4 level was more than 3 m wide, 15 m long, and 60 m high (Anderson and Wagner, 1946), a shape that may indicate that mineralization was controlled by a steep fault. The orebody (or a series of overlapping bodies) may have extended through the entire vertical distance of workings, 300 m; however, the tenor decreased in lower levels. The Idaho Muldoon mine has two separate replacement deposits, one 0.3-1 m thick and at least 30 m long (Anderson and Wagner, 1946), and a second averaging 0.3 m thick and about 46 m long (Tuchek and Ridenour, 1981). The total tonnage of individual replacement bodies is not large; each of the two orebodies on the Idaho Muldoon property totals approximately 1,000 tons. The Eagle Bird mine deposit is estimated to contain about 33,000 tons (4,436 tons produced to date and 28,500 tons in inferred reserves; Tuchek and Ridenour, 1981), exceptionally large by district standards but small by world standards (Mosier and others, 1986).

The proportion of contained metals varies widely from deposit to deposit. Typical grades are 6–12 ounces of silver per ton, 5–10 percent Pb, 3–7 percent Zn, 0.1–1 percent Cu, and a trace to 0.3 ounces of gold per ton. Several orebodies contain as much as 5 percent As, and the Eagle Bird mine contains as much as 4 percent Sb. Production grades from the Eagle Bird mine were 8.8 ounces of silver per ton, 0.2 percent Cu, 7.6 percent Pb, 3.5 percent Zn, and 0.005 ounces of gold per ton (Tuchek and Ridenour, 1981). Production of ore containing more than 40 ounces of silver per ton and 40 percent Pb from the Solid Muldoon and Mutual mines, and more than 28 percent Cu from the Drummond mine, was exceptional and could not be matched today without hand sorting.

Geochemical signature.—The Muldoon–Star Hope area is within the boundary of a geochemical survey conducted for the Boulder-Pioneer Wilderness Study Area (fig. 1) (Simons, 1981). Many samples of sediments from streams draining the area contained anomalous amounts of Ag, As, Ba, Cu, Pb, Sb, and Zn. Most unmineralized rock samples contained anomalous amounts of barium, many contained anomalous amounts of boron and zinc, and a few contained anomalous amounts of gold, cobalt, copper, and manganese. Each of these elements may be present in a geochemical signature that commonly is associated with polymetallic replacements or skarns.

Chemical analyses of selected sulfide-bearing samples collected during this study from deposits in the Muldoon–Star Hope area (particularly those that are hosted in the Drummond Mine Limestone Member) indicate the following range in metal enrichments: almost all samples contain anomalous amounts of lead (360–137,000 ppm) and zinc (238–125,000 ppm). The majority of samples contain silver (61–816 ppm), copper (354–2,290 ppm), arsenic (300–250,000 ppm), and detectable gold (trace–0.50 ppm). Several samples are enriched in antimony (160–410 ppm) and tin (20–300 ppm); these samples are from deposits that are hosted by, or adjacent to, argillaceous rocks. Barium (727–1,010 ppm) and manganese (2,000–21,100 ppm) each are slightly enriched in three separate samples, and cadmium is enriched in three samples from deposits that are cut by dikes. Bismuth (150–170 ppm) is enriched in two samples, only one of which contains detectable gold. Molybdenum is enriched (47 ppm) in a selected sample from the highest dump in Boundary Gulch; this sample has the broadest spectrum of elevated metal values in the analyzed suite and consists of sulfide minerals in argillaceous rock. Two samples from the upper and lower adits of the Star Hope mine also are enriched in molybdenum (100 and 20 ppm, respectively; Bullock and others, 1990) and were collected from veins cutting arenite.

The chemistry of three composite dump samples reflects the main component of dump material in each of the sampled dumps. Two samples, RGW88–49 and RGW88–58 (table 3), from dumps containing mostly argillite, have metal contents similar to those of unmineralized samples except for higher Zn and slightly higher As, Li, Mo, Ni, Pb, and V. A composite sample from an upper dump at the Solid Muldoon mine that consisted mainly of mineralized and altered rock was significantly enriched in Ag, As, Cd, Cu, Mo, Pb, Th, and Zn and slightly depleted in Ba and Y, as compared to unmineralized samples.

Geophysical signature.—No specific geophysical signatures of polymetallic replacement deposits are known from the Muldoon–Star Hope area, but it is probable that local detailed electrical surveys would detect individual orebodies, which are known to contain conductive sulfide minerals.

Genesis and ore controls.—The locations of polymetallic replacement deposits and skarns are governed by chemically reactive host rocks (principally the Drummond Mine Limestone Member) that are cut by concordant and discordant faults and fractures. The intersections of bedding-parallel northwest-trending faults and steep northeast-trending faults and fractures created open spaces that became sites of later mineral deposition. It is possible that carbonaceous pelitic rocks stratigraphically beneath mineralized horizons provided a source for some of the metals, although available information is equivocal. Stocks and dikes, which are exposed near the deposits, probably provided heat and fluids for metasomatism, although specific isotopic information that would fingerprint their influence is lacking. Locally, silification and sericitization of intrusive and host rocks indicates hydrothermal activity. In addition, many of the polymetallic replacements are coarse-grained mosaics of sulfide, silicate, and carbonate minerals, a characteristic feature of distal skarns. This skarnlike texture and mineralogy is one line of evidence that plutons may be present at shallow depth beneath the workings of most polymetallic deposits.

Exploration guides.—Numerous polymetallic deposits are known in the Drummond Mine Limestone Member, particularly in its southern outcrops in the Muldoon–Star Hope district and in its northern outcrops in the Alta district approximately 30 km to the northwest. Because fault intersections principally localized these deposits, most orebodies are likely to be small, although they may be of high grade. Lower grade polymetallic veins and replacement deposits may extend along high-angle faults away from higher grade deposits for considerable distances, but the likelihood for large-tonnage deposits is low. In specific places where calcareous strata are particularly susceptible to replacement, high-grade orebodies may extend as tabular lenses for 100 m or more along bedding. At the Phi Kappa mine in the Alta district, the main replacement orebody in the Drummond Mine Limestone Member extends almost continuously for more than 1,200 m along strike and is believed to extend at least 300 m downdip (Tuchek and Ridenour, 1981). A concealed orebody of these dimensions in the Muldoon–Star Hope area should be detectable at considerable depth by careful surface electrical surveys. Smaller bodies will be difficult to detect. The proximity of the Phi Kappa mine to the Summit Creek quartz monzonite stock suggests an important magmatic influence on deposit formation. Similarly, the proximity of the Eagle Bird and Solid Muldoon mines to the Garfield and Muldoon stocks is regarded as more than fortuitous, as are the locations of the Star Hope, Mackinaw, and Drummond mines near concentrations of dikes. Thus, additional exploration in the central area of Drummond Mine Limestone Member outcrops between the Star Hope and Phi Kappa mines may not be fruitful, even in areas of northeast-trending high-angle faults, inasmuch as there are no known dike swarms or extensive areas of silicification or bleaching that might indicate leucocratic plutons at shallow depths.

POLYMETALLIC VEINS IN SEDIMENTARY ROCKS

Description.—Polymetallic veins are present in dark, argillaceous rocks of the Milligen(?) Formation and the Little Copper Member of the Copper Basin Formation, in calcareous rocks of the Carey Dolostone and the Drummond Mine Limestone and Green Lake Limestone Members of the Copper Basin Formation, and in arenaceous rocks of the upper clastic units of the Copper Basin Formation (fig. 2). The deposits include small discontinuous lenses, stringers, and pods along steep east-northeast- and north-northwest-trending faults, fractures, and brecciated zones. In places, the veins follow narrow contact metamorphic aureoles close to felsic intrusive bodies. Weathering of the veins forms conspicuous gossans in many places. Where host rocks are calcareous, the veins generally are associated with polymetallic replacements and in places extend along high-angle faults and fractures that cut bedding-parallel replacements. This type of vein is a hybrid and generally incorporates less gangue and therefore is chemically more enriched than veins in noncalcareous rocks. Where host rocks are noncalcareous, the veins generally are quartz rich but also may contain calc-silicate minerals or barite. High metal values in veins of this type are generally very spotty. The silver-rich veins at the Star Hope mine were a notable exception. Polymetallic veins have produced (in order of value) silver, lead, zinc, copper, and gold.

Type example.—Polymetallic veins in carbonate rocks: Rippeto claim, Little Wood River district

Other examples.—Little Wood River district: American mine prospect, Garfield group, Lucky Boy (east), Silver Eagle, Silver Mint, and Smuggler

Type example.—Polymetallic veins in sandstone: Star Hope, Copper Basin district

Other examples.—Little Wood River district: Champion prospect, Logan tunnel, Lucky Boy (west), and Solid Muldoon (part?). Copper Basin district: Candy Cane (south)

Type example.—Polymetallic veins in argillite: Mutual mine, Little Wood River district

Other examples.—Little Wood River district: Copper Bell group, Frisco Gulch prospects, and Scorpion Antimony

Mineralogy and texture.—Sulfide minerals include galena, chalcopyrite, sphalerite, pyrite, and arsenopyrite. Silver values are higher in ore that is rich in galena, although discrete silver minerals including tetrahedrite were identified at the Solid Muldoon and Star Hope mines. Oxidized minerals include anglesite, cerussite, malachite, smithsonite, and iron oxides. Quartz is the predominant gangue mineral in most veins, but clay minerals, calcite, and calc-silicate minerals are present in a few veins.

Characteristics.—No modern laboratory data are available for polymetallic veins of the Muldoon–Star Hope area. The veins share many characteristics with the polymetallic replacements or skarns, however, and probably are coeval and cogenetic.

Orebodies.—Most veins are thin and discontinuous. Typical thicknesses are 0.3–1.0 m, although thinner veins are not uncommon. A mineralized vein 3 m thick in Mutual Gulch (Tuchek and Ridenour, 1981) is exceptionally thick. Typical lengths of veins are 10–30 m; however, the main vein at the Rippeto claim can be followed for 150 m, the main zone at the Silver Eagle claim for 120 m, and a mineralized shear at the Champion property for 140 m (Tuchek and Ridenour, 1981). The inferred tonnages of individual vein deposits are small. Resources on the Rippeto claim, which are exceptionally large for the Muldoon–Star Hope area, were inferred by Tuchek and Ridenour (1981) to be approximately 35,000 tons grading 4.4 ounces of silver per ton, 4.91 percent Pb, 0.9 percent Zn, and a trace of gold. Resources on the Silver Eagle claim are inferred to be 15,000 tons grading 3.3 ounces of silver per ton, 2.83 percent Pb, 2.53 percent Zn, and traces of gold. Other known polymetallic veins have comparable grades but significantly less tonnage; however, the aggregate tonnage in the Scorpion group of claims on the south side of Garfield Canyon (about 15,000 tons) and in the host of claims in Mutual Gulch (about 18,000 tons) each may constitute a minable resource collectively.

Geochemical and geophysical signatures.—In the surrounding region, a geochemical signature commonly associated with polymetallic veins, regardless of host rock, is Zn, Cu, Pb, As, Au, Ag, Mn, Ba, Sb, and Bi (Worl and others, 1989). Many rock and stream-sediment samples from the Muldoon–Star Hope area contain anomalous amounts of one or more of these elements (Simons, 1981). The vein deposits have no direct geophysical expression in the regional magnetic and gravity data available for the Muldoon–Star Hope area. Major steep faults and shallow stocks that may influence localization of some veins are delineated, however, by prominent anomalies and steep gradient zones.

Genesis and ore controls.—East-northeast- and north-northwest-trending steep faults apparently exert the strongest control on the locations of polymetallic veins by providing permeable zones through reactive rocks. Fault offsets of 4 m or less were sufficient to create channelways (Kunkel, 1989). At the Star Hope property, some shallowly dipping veins may follow fractures parallel with the Glide Mountain thrust fault. Silicification and sericitization generally are more intense within the fault zones but in a few places extend laterally a few meters into susceptible country rocks. The most laterally continuous veins in the Muldoon–Star Hope area, those at the Rippeto, Silver Eagle, Champion, and Star Hope properties, are near contacts with quartz monzonite stocks or leucocratic dike swarms believed to emanate from subjacent plutons. The igneous rocks probably provided heat and fluids for metasomatism, although their role has not been substantiated by isotopic or fluid-inclusion analyses. A subtle zoning of sulfide minerals, present in both replacement and vein deposits, probably indicates a hydrothermal influence on ore deposition. Arsenopyrite, pyrite, pyrrhotite, and chalcopyrite are more abundant closer to the Garfield and Muldoon stocks and to the inferred buried stock beneath the Star Hope mine, whereas those sulfide minerals that form at cooler temperatures (galena and sphalerite) are more abundant farther from the igneous rocks.

Exploration guides.—In the Muldoon–Star Hope area, the coincidence of the triad (1) calcareous, or mixed calcareous and siliceous sedimentary host rocks, (2) east-northeast- or north-northwest-trending steep faults, and (3) Tertiary leucocratic igneous rocks indicates prospective ground. Fault control is fundamental for the vein deposits. The veins typically are small and may show only inconspicuous gossans. They are unlikely to have pronounced geophysical expression, although very detailed electrical surveys might detect sulfide-bearing veins hosted in steep fault zones.

POLYMETALLIC VEINS IN IGNEOUS ROCKS

Description.—Polymetallic veins are present near the margin of the Muldoon stock at the Contact prospect along the west side of upper Muldoon Creek and also may be present along the west margin of the Garfield stock in Garfield Canyon. Precious- and base-metal veins also are present in rocks of the Challis Volcanic Group at the Black Spar mine west of Copper Creek (fig. 2). Almost no information has been published on these occurrences. At the Contact prospect, a vein occupies a shear zone at the contact of the quartz monzonite stock and interbedded calcareous and argillaceous rocks of the Drummond Mine Limestone Member, and sulfide minerals also are in irregular pods along joints in the stock. At the Black Spar mine near the confluence of Blackspar Canyon and Copper Creek, extensive workings explore shear zones in rocks of the Challis Volcanic Group. To the west in the Little Wood River area (Sanford, 1988; Sanford, Whitney, and others, 1989), to the south in the Lake Hills (Moye and others, 1988), and to the east in the Grouse quadrangle (Skipp, 1989), andesite, dacite, and rhyodacite flows, breccias, and tuffs constitute the southern field of the Challis Volcanic Group. In the Copper Creek area, rocks of the Challis Volcanic Group consist almost entirely of thick interlayered andesitic tuff-breccias and flows that form the basal Challis unit. About 1.6 km west of the Black Spar mine, these andesitic rocks are overlain by dacitic flows and minor welded ash-flow tuff. The basal andesitic rocks are cut by leucocratic dikes of two orientations: one set trends roughly north-south, and the second trends roughly east-west. The Black Spar underground workings, which are inaccessible, apparently follow silicified shear zones parallel with these trends. West of the Muldoon townsite in the southwestern part of the area, a conspicuous swarm of north-trending sanidine-bearing rhyolite dikes cuts the andesitic rocks (fig. 2); no workings are known to be associated with these dikes.

Mineralogy and texture.—In veins hosted by intrusive rocks, pyrite is the predominant sulfide mineral, but galena, sphalerite, and chalcopyrite also are present. Gossans are weakly developed, and sparse malachite is present. In volcanic-hosted veins, pyrite, galena, sphalerite, and chalcopyrite are present, together with quartz and siderite gangue. Bleaching, silicification, and pyritization extend 2–3 m from veins, and textures in volcanic host rocks are completely obliterated.

Characteristics.—Neither laboratory nor analytical information is available for either type of vein occurrence. At the Contact prospect, the age of mineralization must be Eocene or younger inasmuch as mineralized rock occupies joints within the Muldoon stock, which has been dated at 47 Ma. Although mineralized veins at the Black Spar property are in rocks of the Challis Volcanic Group, they trend north-northwest and east-northeast, similar to faults, dikes, and veins in the underlying Devonian and Mississippian sedimentary rocks. Furthermore, leucocratic dike rocks are slightly to strongly altered, whereas intermediate dike rocks are virtually unaltered. Hence, the age of mineralization is about the same as the age of the dikes, probably Eocene. This age is consistent with ages inferred for volcanic-hosted deposits elsewhere in the Challis volcanic field. According to Moye and others (1989), the deposits are related spatially and temporally to waning stages of volcanism at local centers late in the Eocene.

Orebodies.—At the Contact prospect, the pyritized shear zone is 0.6–1.5 m thick and extends along strike at least 46 m (Tuchek and Ridenour, 1981). Pyrite pods along joints in the Muldoon stock are not extensive but are as thick as 0.9 m. The extent of veins at the Black Spar property is unknown, but the veins are likely to be very irregular and thin, occupying fissures and brecciated zones within the complexly fractured volcanic rocks.

Geochemical and geophysical signatures.—The Contact and Black Spar properties are at the edges of the regional geochemical and aeromagnetic surveys of the Boulder-Pioneer Wilderness Study Area (Mabey, 1981; Simons, 1981) and were not sampled adequately; however, the geochemical sampling included one rock sample near each property and one stream-sediment sample, which was collected near the mouth of Blackspar Canyon. Only lead and uranium showed slightly elevated values near the Contact prospect, and only chromium was slightly enriched in the stream-sediment sample from Blackspar Canyon. Additional rock samples collected in 1987 near the Black Spar property show slightly elevated values of barium, chromium, and manganese but only background values of other elements commonly associated with polymetallic veins (Bullock and others, 1990). A geochemical signature of polymetallic veins normally includes several of the elements Zn, Cu, Pb, As, Au, Ag, Mn, Ba, Sb, and Bi.

Regional magnetic values in the Blackspar Canyon area show a smooth eastward increase toward a positive anomaly in the Garfield Canyon area that is inferred to indicate the subsurface connection and extent of the Garfield and Muldoon stocks. The concealed part of the pluton apparently does not extend westward beneath Blackspar Canyon. A similar, smaller positive aeromagnetic anomaly, west of the former townsite of Muldoon, approximately 6 km south of Blackspar Canyon, also is inferred to indicate a concealed stock (Mabey, 1981) but does not appear to extend northward as far as Blackspar Canyon.

Genesis and ore controls.—Polymetallic veins in igneous rocks almost certainly formed during accumulation of the southern Challis volcanic field, probably during late rhyolitic activity. The genesis of the veins apparently is analogous to that of veins in adjacent Paleozoic sedimentary rocks, and structural controls—particularly intersections of east-northeast- and north-northwest-trending faults and fractures—provided pathways for metal-bearing fluids. The extent of hydrothermal alteration in the Blackspar Canyon area is unknown. It is possible that shallow intermediate to siliceous intrusions are present in rocks of the Challis Volcanic Group of the Blackspar Canyon area, as they are in the Lava Creek district 30 km to the east-southeast. In the latter area, polymetallic vein deposits in rocks of the Challis Volcanic Group are hosted both by the intrusive rocks and the volcanic country rocks and have elevated gold and silver values (Moye and others, 1989). The deposits in the Lava Creek district formed during epithermal brecciation and silicification and have been exposed by younger high-angle faulting.

Exploration guides.—Inasmuch as formation of polymetallic veins accompanied polymetallic replacement in chemically receptive rocks, the search for undiscovered veins in igneous rocks can best follow some of the same guidelines. Most orebodies in igneous rocks are likely to be at or near intersections of steep east-northeast- and north-northwest-trending faults. The relative favorability of the two trends is unknown. If detailed mapping of the Challis Volcanic Group shows shallow, complex intrusive centers, exploration should be expanded to include intrusion margins as well as throughgoing steep faults. The coincidence of magnetic anomalies, leucocratic dikes, and propylitic or argillic alteration suggestive of hydrothermal circulation is highly favorable.

STRATABOUND REPLACEMENT DEPOSITS AND VEINS OF BARITE

Description.—Stratabound barite is present as coarsely crystalline, light-colored, bedding-parallel lenses and stringers in both the Little Copper and Drummond

Mine Limestone Members of the Copper Basin Formation. In many places, veins of barite cut across bedding to connect separate bedding-parallel lenses. Barite is present in several stratigraphic zones in a sequence approximately 1,000 m thick (fig. 3). Argillaceous rocks in the upper 200 m of the Little Copper Member host thin lenses and stringers of barite at many places between Garfield Canyon and Deep Gulch. In the vicinity of the Muldoon Barium Company Deep Gulch mine, several thick replacement deposits of barite are present in thin-bedded calcareous rocks at the contact of the Little Copper and Drummond Mine Limestone Members and in the lower 100 m of the Drummond Mine Limestone Member to the east. Silver-base metal veins are present in the same stratigraphic sequence as the barite replacement deposits but have not been observed in contact with them. Along Muldoon Ridge and to the east of the Eagle Bird mine, thin bedding-parallel stringers and lenses of barite are present in several stratigraphic horizons in thin-bedded calcareous rocks of the Drummond Mine Limestone Member. East of Muldoon Creek, thin and thick barite replacement deposits are within mixed calcareous and noncalcareous rocks in the uppermost 100 m of the Drummond Mine Limestone Member. These deposits are at approximately the same stratigraphic level as polymetallic replacement deposits at the Idaho Muldoon mine and crop out within 0.8 km, but no contact relations have been observed. Thin stringers and lenses of barite are in thin-bedded calcareous and noncalcareous rocks at the Lucky Boy (east) prospects on the east side of Argosy Creek at approximately the same stratigraphic level. Most barite outcrops are weakly iron or copper stained, and analyses show the presence of a few percent zinc, lead, or copper, and traces of silver.

Type example.—Deep Gulch mine, Little Wood River district

Other examples.—Little Wood River district: Lucky Boy group (east) and Muldoon Creek prospects

Mineralogy, texture, and characteristics.—Most replacement deposits and veins contain almost pure barite and only minor carbonate or clastic impurities; however, the deposits develop weak gossans that contain traces of sulfide minerals including pyrite, galena, sphalerite, chalcopyrite, and arsenopyrite. Oxidized minerals include limonite and malachite. The barite is coarse grained, milky white, and almost pure. Although beds adjacent to massive or weakly laminated barite are not obviously recrystallized, chemical analyses of them include significant barium. Measurements of stable isotopes that might pinpoint crystallization conditions of barite are not available.

Orebodies.—Barite replacement deposits and veins are irregular. They form tabular seams and stringers that generally are parallel with bedding in enclosing calcareous or argillaceous host rocks but may abruptly cut across bedding or terminate. The barite bodies are from a few centimeters to 4.3 m in thickness, but most are less than 1 m thick. Intervals in which closely spaced thin barite bodies are present may be as thick as 50 m. One replacement at the Deep Gulch mine may be 1.6 km long (Tuchek and Ridenour, 1981). The barite bodies generally are very pure, from 65 to 95 percent $BaSO_4$, but most assays are greater than 90 percent (Tuchek and Ridenour, 1981). No estimates have been published of the total resources of barite on the Deep Gulch claims, but hundreds of thousands of tons of ore are present, and significant resources also are present in the claims east of Muldoon Creek.

Geochemical signature.—The geochemical survey of the Boulder-Pioneer Wilderness Study Area (Simons, 1981) included the Deep Gulch, Muldoon Creek, and Argosy Creek areas in which barite replacements and veins crop out. Samples of sediments from streams draining these areas contained anomalous amounts of barium, as well as silver, lead, and zinc, elements that are associated with polymetallic replacement deposits and veins. Unmineralized rock samples also contained anomalous amounts of barium, and a few contained anomalous amounts of arsenic, beryllium, and manganese. More extensive sampling of unmineralized rocks from throughout the Little Copper and Drummond Mine Limestone Members indicates anomalous values of barium at all stratigraphic levels and locally anomalous values of arsenic and zinc. These latter analyses indicate that the average background level of barium for the Little Copper Member is 2,750 ppm and for the Drummond Mine Limestone Member 1,460 ppm (table 2).

Genesis and ore controls.—The origin of the stratabound barite deposits is uncertain. Although the texture of the deposits and their local discordances with bedding indicate replacement of country rock sometime after lithification, it is possible that the deposits formed by recrystallization of original syngenetic barite and local remobilization into fractures. The intrinsic richness in barium of the Little Copper and the Drummond Mine Limestone Members of the Copper Basin Formation implies a formational control on barite distribution.

A recent review of stratiform barite in Paleozoic rocks of the Western United States (Poole, 1988) demonstrates that bedding-parallel barite deposits are common in rocks of Cambrian through Pennsylvanian age in the Antler orogenic belt in central Nevada. Barite deposits in Mississippian rocks are among the most easterly barite occurrences in Nevada, a tectonic setting similar to the Idaho occurrences. For the Nevada occurrences, barite deposition is thought to have occurred from the discharge of barium-bearing hydrothermal solutions at the seafloor in marine-basinal or continental-rise settings (Poole, 1988). The barite-rich chemical sediments were subsequently resedimented by turbidity currents. By analogy, deposits in the Muldoon–Star Hope area also may have formed syngenetically; however, their white, coarse crystallinity indicates that they were remobilized during postdepositional deformation and heating, probably during the same Cretaceous and Tertiary events that affected the nearby silver-base metal deposits.

Exploration guides.—Although barite replacement deposits and veins are present in a stratigraphic interval of approximately 1,000 m, the thickest and most continuous known bodies are confined to the uppermost Little Copper and lowermost Drummond Mine Limestone Member in the vicinity of Deep Gulch and to the uppermost Drummond Mine Limestone Member between Muldoon and Argosy Creeks. No stratigraphically confined debris flows, slump features, or hydrothermal haloes that might indicate syndepositional fault systems or the location of exhalative vents have been recognized in either area; however, available information is only reconnaissance in scale. The recognition of such features would be a powerful aid in exploring for stratabound polymetallic deposits as well as barite deposits.

SUMMARY

Numerous mines in the Muldoon–Star Hope area of the southern Pioneer Mountains of south-central Idaho produced silver, lead, copper, zinc, gold, and barium from replacement and vein deposits in Mississippian and Devonian carbonate and clastic rocks; minor production also may have come from vein deposits in igneous rocks that intrude the sedimentary host rocks. The richest metal deposits were mined principally in the 1880's and 1890's and again in the 1940's and 1950's. The richest barite bodies were mined only in the 1950's, and 1960's. Some orebodies were not mined completely, and some known deposits were not explored thoroughly during the episodic activity.

Most mineral deposits are in micritic and argillaceous turbidites of the Mississippian Copper Basin Formation, which are designated the Drummond Mine Limestone and Little Copper Members, respectively. The dark, carbonaceous pelites of the lower member, the Little Copper, are significantly enriched in barium and boron and slightly enriched in cobalt, chromium, manganese, lead, and zinc. These argillaceous rocks contain bedding-parallel stringers and lenses of barite at several stratigraphic horizons and may have served as sources for some metals in mineral deposits in the overlying impure calcareous rocks of the Drummond Mine Limestone Member. The Drummond Mine Limestone Member hosts numerous high-grade sulfide replacement bodies at several stratigraphic levels. These bodies generally are lenticular and of small tonnage; however, they are enriched where bedding-plane and crosscutting faults intersect. This apparent stratigraphic control may indicate some syngenetic metal deposition—an implication that lower grade resources may be relatively widespread in the Copper Basin Formation.

Quartz monzonite stocks and related dike swarms of Eocene age intrude the Copper Basin Formation in many places. Mineral deposits are numerous around the peripheries of several stocks; a crude zoning of sulfide minerals and metal values in these deposits indicates that the intrusions probably provided heat and fluids for metasomatism in bordering rocks. Where chemically reactive calcareous rocks are intruded, the presence of skarn deposits, as well as replacements, merits evaluation.

Significant resources of base- and precious-metals may remain in the Muldoon–Star Hope area. Although most individual occurrences are likely to be small, they may be of high grade and closely spaced. Favorable areas for additional exploration are indicated by the association of some or all of the following features: (1) calcareous or mixed calcareous and siliceous sedimentary host rocks; (2) Tertiary leucocratic igneous rocks; (3) carbonaceous pelitic source rocks having elevated metal values; (4) ground structurally prepared by steep east-northeast- or north-northwest-trending faults, particularly where associated with bedding-plane flexures or faults; and (5) widespread geochemical anomalies of precious- and base-metals and associated elements.

REFERENCES CITED

Anderson, A.L., and Wagner, W.R., 1946, A geological reconnaissance in the Little Wood River (Muldoon) district, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 75, 22 p.

Bullock, J.H., Jr., Wilson, A.B., and Soulliere, S.J., 1990, Analytical results and sample locality maps of rock samples from the northwestern Idaho Falls and the northeastern Hailey 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 90–645, 20 p.

Dover, J.H., 1981, Geology of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–A, p. 15–75.

Einaudi, M.T., Newberry, R.J., and Meinert, L.D., 1981, Skarn deposits, *in* Sims, P.K., and Skinner, B.J., eds., Economic Geology—Seventy-fifth anniversary volume—1905–1980: El Paso, Texas, The Economic Geology Publishing Company, p. 317–391.

Hall, W.E., Rye, R.O., and Doe, B.R., 1978, Wood River mining district, Idaho—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 579–592.

Kleinkopf, M.D., Bankey, Vicki, and McCafferty, A.E., 1989, Gravity and magnetic anomaly patterns applied to mineral resource exploration, Hailey 1°×2° quadrangle, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 26–32.

Kunkel, K.W., 1989, Geology and ore deposits of Garfield Canyon, Blaine County, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2°

quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 92–94.

Mabey, D.R., 1981, Aeromagnetic studies of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–B, p. 77–83.

Mosier, D.L., Morris, H.T., and Singer, D.A., 1986, Grade and tonnage model of polymetallic replacement deposits, *in* Cox, D.P., and Singer, D.A., eds., Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 101–104.

Moye, F.J., Hackett, W.R., Blakley, J.D., and Snider, L.G., 1988, Regional geologic setting and volcanic stratigraphy of the Challis volcanic field, central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 87–98.

Moye, F.J., Hackett, W.R., Snider, L.G., and Mertzman, S.A., in press, Stratigraphy, chronology, and petrology of the southeastern Challis volcanic field, central Idaho: U.S. Geological Survey Bulletin 2064–FF.

Moye, F.J., Sanford, R.F., Erdman, J.A., and Johnson, K.M., 1989, Volcanic-hosted and intrusion-related mineral deposits in the Hailey and western Idaho Falls quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 70–71.

Nelson, W.H., and Ross, C.P., 1969, Geologic map of the Mackay quadrangle, south-central Idaho: U.S. Geological Survey Miscellaneous Geologic Investigations Series Map I–580, scale 1:125,000.

Nilsen, T.H., 1977, Paleogeography of Mississippian turbidites in south-central Idaho, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 1, p. 275–299.

Paull, R.A., and Gruber, D.P., 1977, Little Copper Formation—New name for lowest formation of Mississippian Copper Basin Group, Pioneer Mountains, south-central Idaho: American Association of Petroleum Geologists Bulletin, v. 61, no. 2, p. 256–262.

Paull, R.A., Wolbrink, M.A., Volkmann, R.G., and Grover, R.L., 1972, Stratigraphy of the Copper Basin Group, Pioneer Mountains, south-central Idaho: American Association of Petroleum Geologists Bulletin, v. 56, no. 8, p. 1370–1401.

Poole, F.G., 1988, Stratiform barite in Paleozoic rocks of the western United States: International Association on the Geology of Ore Deposits, Proceedings of the Seventh Quadrennial Symposium, p. 309–319.

Sanford, R.F., 1988, Challis Volcanics, Quaternary basalts, and a Quaternary lake, Little Wood River area, south-central Idaho: Geological Society of America Abstracts with Programs, v. 20, p. 467.

Sanford, R.F., Whitney, H.A., McCafferty, A.E., and Gese, D.D., 1989, Mineral resources of the Little Wood River Wilderness Study Area, Blaine County, Idaho: U.S. Geological Survey Bulletin 1721–D, 15 p.

Sanford, R.F., Wooden, J.L., Doe, B.R., and Worl, R.G., 1989, Origin of ore deposits in the central Idaho black shale belt—Implications from lead isotopes, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 39–52.

Simons, F.S., 1981, A geological and geochemical evaluation of the mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–C, p. 85–180.

Skipp, Betty, 1989, Geologic map of Mackay 4 (Grouse) NW quadrangle, Butte and Custer Counties, Idaho: U.S. Geological Survey Open-File Report 89–142, scale 1:24,000.

Tuchek, E.T., and Ridenour, James, 1981, Economic appraisal of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497, p. 181–292.

Umpleby, J.B., 1917, Geology and ore deposits of the Mackay region, Idaho: U.S. Geological Survey Professional Paper 97, scale 1:375,000, 129 p.

Wells, M.W., 1983, Gold camps and silver cities (2nd ed.): Idaho Bureau of Mines and Geology Bulletin 22, 165 p.

Worl, R.G., Wilson, A.B., Smith, C.L., Kleinkopf, M.D., and Sykes, R.C., 1989, Mineral resource potential and geology of the Challis National Forest, Idaho: U.S. Geological Survey Bulletin 1873, 101 p.

Structural and Stratigraphic Setting of the Triumph Stratiform Zinc-Lead-Silver Deposit, Devonian Milligen Formation, Central Idaho

By Robert J.W. Turner *and* Bruce R. Otto

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–E

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

CONTENTS

FIGURES

Structural and Stratigraphic Setting of the Triumph Stratiform Zinc-Lead-Silver Deposit, Devonian Milligen Formation, Central Idaho

By Robert J.W. Turner[1] *and* Bruce R. Otto[2]

ABSTRACT

The Triumph mine in central Idaho was an important producer of lead, zinc, silver, and gold from 1927 to 1957. The bulk of base-metal production was from stratiform sediment-hosted zinc-lead sulfide bodies interbedded with carbonaceous argillite, limestone, chert, and diamictite of the Devonian Milligen Formation. Near the Triumph deposit, the Milligen Formation is divided into three informal units: lower argillite (black argillite, sandstone, chert), Triumph argillite (black argillite), and Independence sandstone (sandstone and limestone turbidite). Two intertonguing facies are in the transitional contact between the lower argillite and Triumph argillite units; these are the Lucky Coin limestone facies (limestone turbidite, black argillite, diamictite) and the Cait quartzite facies (quartzite turbidite, black argillite).

The Triumph stratiform sulfide bodies are within or at the top of the Lucky Coin limestone unit; structural complication makes their exact stratigraphic position unclear. Near the stratigraphic level of the stratiform sulfide bodies are limestone turbidite, black argillite, diamictite, thin tuff beds, and mafic sills. Diamictite is present as discontinuous lenses that are thickest near the stratiform sulfide body. The poorly sorted nature and similar composition of clasts to underlying Milligen strata suggest local resedimentation of Milligen strata. A contact metamorphic aureole extending northward from the Triumph mine area reflects a buried intrusion. Thin-bedded calcareous units are altered to a quartz, tremolite, clinopyroxene, feldspar, and chloritoid assemblage, whereas mudrock units are variably hornfelsed to a quartz and biotite assemblage. Calc-silicate and hornfels alteration postdates cleavage-forming deformation and obscures sedimentary textures and cleavage.

In the Triumph mine area, Milligen strata generally strike N. 55°–65° W. and dip 30° SW. Milligen strata are deformed into tight to isoclinal folds associated with a variably developed early cleavage (D1). Locally, the early cleavage is folded by east-vergent folds (D2). Southwest-dipping thrust faults repeat Milligen strata; some of these thrust faults host sheared lead-zinc-silver veins and are intruded by granitic and andesitic dikes. The Triumph stratiform orebody is immediately above the Fissure thrust fault in the overturned limb of an east-vergent anticline (D1). Low-angle normal faults (D3) displace Milligen strata and thrust faults to the southwest and juxtapose Milligen rocks against structurally higher Pennsylvanian-Permian Wood River Formation and Eocene Challis Volcanic Group. Northeast-trending high-angle faults offset strata as young as the Challis Volcanic Group. D1 deformation may correlate with either the Early Mississippian Antler orogeny or the Sevier orogeny of Late Jurassic to Cretaceous age, D2 deformation with the Sevier orogeny, and D3 deformation with Eocene extension.

The Triumph stratiform sulfide body is composed of two types of ore known locally as complex ore and siliceous ore. Complex ore is laminated or banded fine-grained pyrite, sphalerite, and galena interbedded with carbonaceous pyritic chert, shale, and siltstone. Siliceous ore is mostly medium grained siderite, sphalerite, and galena interbedded with argillite or carbonate rocks. Along the Fissure thrust zone underlying the stratiform sulfide body, ore known as fissure ore is present as lenses and pods of sheared siderite-quartz-sphalerite-galena within gouge and crushed rock.

Milligen strata record an anoxic depositional basin dominated by hemipelagic and pelagic sedimentation (lower argillite, Triumph argillite) punctuated by two periods of turbidite influx: a Middle to Late Devonian event composed of a carbonate channel complex (Lucky Coin limestone) to the west of a quartz-sand channel system (Cait quartzite), and a more widespread Late Devonian event of mixed siliciclastic and carbonate detritus (Independence sandstone). The

[1]Geological Survey of Canada, 100 W. Pender Street, Vancouver, British Columbia V6B 1R8.

[2]333 Provident Drive, Boise, Idaho 83706.

Triumph stratiform sulfide body was deposited within the Lucky Coin carbonate fan and is associated with diamictite derived from the submarine scarp of an inferred syndepositional fault. This fault is invoked as the conduit for metalliferous fluids that exhaled on the seafloor to form the Triumph stratiform sulfide body.

INTRODUCTION

The Triumph zinc-lead-silver deposit, the major producer in the Warm Springs mining district of central Idaho, has long been recognized as mineralogically and texturally distinct from silver-lead vein deposits in the district (Umpleby and others, 1930; C.W. Merriam and C.N. Bozian, written commun., 1942; Anderson and others, 1950). Bedded ores of the Triumph deposit were interpreted by these authors as replacements of limestone beds adjacent to mineralizing fissures. More recently, the deposit has been reinterpreted as a syngenetic or syndiagenetic deposit coeval with deposition of host Devonian black shale and limestone (Smith, 1977; Hall, 1985). Herein, we describe the structure and stratigraphy of the Devonian Milligen Formation that hosts the Triumph deposit. Our studies are based on exploration field studies from 1983 to 1987 investigating Milligen strata east of Ketchum for Westley Mines Ltd., Mintek Resources Ltd., and Noranda Exploration Inc. (Otto, 1984; Turner and Michelson, 1984; Turner, 1985; Turner and others, 1986). Much of this work was in the Triumph mine area.

The Triumph deposit is in south-central Idaho, approximately 6 mi north of Hailey and 3 mi southeast of Ketchum on the north side of the East Fork of the Wood River (fig. 1). Our studies of the Triumph deposit included the area within the drainages of Triumph, North Star and Courier Gulches, and Independence Canyon, and we refer to this area as the Triumph mine area (fig. 2). Access to the area is by paved all-weather road up the East Fork valley and then by gravel road to the North Star mine in North Star Gulch and to the old Triumph and Independence mines via Triumph Gulch. There is also access to the Independence mine area by gravel road up Independence Canyon from the Elkhorn resort area. Elevations within the mine area vary from 6,000 ft along the East Fork valley to ridges more than 8,000 ft high on the north side of the area. The steep valley sides are mostly sage and grass covered; small groves of aspen are in some valley bottoms.

The surface outcrop of the Triumph orebody in North Star Gulch was discovered in 1881. Mining in the area began at the Independence mine in 1885. The Triumph mine produced from 1927 until 1957. Between 1936 and 1945 the Triumph mine produced about 100,000 tons per year and was the second largest silver producer in Idaho. Production records are incomplete, but from 1936 to 1948 the mine produced 1.06 million tons, and between 1941 and 1948 production averaged 6.57 percent zinc, 4.02 percent lead, 7.8 ounces of silver per ton, and 0.071 ounces of gold per ton (C.W. Merriam and C.N. Bozian, written commun., 1942).

Early access to the Triumph deposit was through the North Star mine at the head of North Star Gulch and a 700-foot-deep shaft of the "Old Triumph mine" in eastern part of upper Triumph Gulch (fig. 3). The mine was worked to the 900-foot level (level elevations are the distance below the collar of the shaft). Ore was taken by aerial tram to the Gimlet railway siding in the Wood River valley, 5 mi to the west. In 1943, a 6,500-foot-long haulage tunnel (Plummer tunnel) was completed connecting the 700-foot level of the mine with the North Star mill in the East Fork valley near the village of Triumph (fig. 3). The Plummer tunnel also provided access to the lower part of the Independence mine. By the 1950's, the high-grade bedded ores were exhausted, and work concentrated on developing reserves within the Fissure shear zone in the lower levels of the northwestern part of the mine.

The possibility of large additional base-metal reserves at the Triumph mine followed the recognition by Clyde Smith that the Triumph deposit was a member of the stratiform sediment-hosted lead-zinc class of ore deposits. Subsequent surface drill programs were conducted by Bear Creek Mining Company in 1978–79, Venture West Minerals in 1981, and Getty Resources in 1984.

Acknowledgments.—We would like to thank Victor Jones of Westley Mines and Rupert House of the Triumph Mining Company for permission to publish this study of the Triumph mine area. This work was conducted under the direction of Dr. Clyde Smith of Westley Mines Ltd., who first recognized the syngenetic nature of the Triumph deposit and who developed a three-dimensional model of the Triumph orebody from mine data. We thank Clyde for his insight and continued encouragement. We were ably assisted in the field by Carl Michelson, Dave Smith, Mary Fitch, Ken Loos, Pat Okita, and Greg Kuzma. James W. Whipple, Don Murphy, Charlie Jefferson, and Clyde Smith provided valuable critiques of the paper that considerably improved its content and style. This paper is GSC Contribution 49790.

GEOLOGIC SETTING

REGIONAL SETTING OF THE MILLIGEN FORMATION, CENTRAL IDAHO

Black argillite and interbedded limestone, sandstone, and chert of the Lower to Upper Devonian Milligen Formation is exposed east of the Wood River valley and U.S. Highway 75 in a belt 35 mi long and as wide as 6 mi (fig. 1). The exposures lie between Cretaceous intrusive rocks of the Idaho batholith to the west and high-grade metamorphic and intrusive rocks of the Eocene Pioneer Mountains core complex to the east. Milligen strata are in thrust contact with the Lower and Upper Mississippian Copper Basin Formation to the east and are structurally overlain by Middle Pennsylvanian to Lower Permian Wood River Formation and Eocene Challis Volcanic Group to the west. This fault-bound package of Milligen strata was named the Pioneer allochthon by Dover (1980).

The Milligen Formation, along with black argillite, chert, and siltstone of the Lower Ordovician to Middle Silurian Phi Kappa and Middle Silurian Trail Creek Formations and an unnamed Silurian and Devonian unit, is interpreted as part of the Roberts Mountains allochthon, a sequence of outer continental margin strata thrust eastward over the continental margin during the Early Mississippian Antler orogeny (Roberts and Thomasson, 1964; Dover, 1980). The sandstone turbidite facies of the Mississippian Copper Basin Formation that is exposed east of the Pioneer Mountains core complex is inferred to have been derived from the exposed Roberts Mountains allochthon ("Antler highlands") to the west (Nilsen, 1977; Poole and Sandberg, 1977). A phyllitic fabric in some clasts within the Copper Basin Formation suggests pre-Mississippian deformation within the strata of the Roberts Mountains allochthon (Davis, 1984). Pennsylvanian and Permian Wood River Formation limestone, siltstone, shale, sandstone, and basal chert conglomerate are inferred to represent the post-Antler overlap assemblage (Poole and Sandberg, 1977). To the north of the Milligen belt rocks of the Challis volcanic field of Eocene age are widespread.

Paleozoic strata are weakly metamorphosed, folded by north- to northwest-trending, east-vergent folds, and cut by east-vergent thrust faults of Mesozoic age (Dover, 1980, 1981). These Mesozoic structures are offset by low-angle extensional faults related to development of the Eocene Pioneer Mountains core complex. Within the Pioneer Mountains, Precambrian and Ordovician metasedimentary rocks and Cretaceous and Eocene plutonic rocks are separated by a low-angle fault from the overlying Paleozoic sedimentary rocks and Eocene volcanic rocks (Wust, 1986; O'Neill and Pavlis, 1988). The direction of displacement of upper plate rocks is N. 65° W. (Wust, 1986).

Zinc-lead-silver deposits and prospects in the Wood River area of the Warm Springs mining district are hosted by the Devonian Milligen Formation and, subordinately, by Cretaceous and Tertiary intrusive rocks and the Pennsylvanian and Permian Wood River Formation. Past production history and geologic setting of these deposits has been described by various authors (Umpleby and others, 1930; Anderson and others, 1950; Hall, Rye, and Doe, 1978).

GEOLOGIC SETTING OF THE MILLIGEN FORMATION, EAST FORK WOOD RIVER VALLEY

The drainage area of the East Fork Wood River is underlain primarily by faulted and folded Milligen strata. On the western margin of the Milligen belt, near the Triumph mine, an overturned sequence of Milligen strata is repeated by northwest-trending thrust faults. Both lower and upper parts of the Milligen Formation are exposed. To the east, a central belt of open, east-vergent folds exposes only the upper part of the Milligen Formation, and thrust faults have not been recognized. On the eastern margin of the Milligen belt, Milligen strata are in fault contact with Wood River Formation.

Within the Milligen belt, a regionally extensive low-angle fault or series of faults separates the Milligen strata into lower and upper structural plates (fig. 2). This fault or fault array offsets earlier northwest-trending thrust and folds and can be mapped throughout an area of 30–40 mi^2. Rocks of the lower plate crop out throughout much of the East Fork valley; upper plate strata cap high northeast-trending ridges of the East Fork valley, creating a map pattern of elongate northeast-trending klippe of upper plate Milligen strata. Klippe of Wood River Formation and Challis Volcanic Group overlie Milligen strata of the upper plate above low-angle normal faults.

GEOLOGIC SETTING OF THE TRIUMPH DEPOSIT

The Triumph stratiform lead-zinc-silver-gold deposit is within the Milligen Formation along the western margin of the outcrop belt of Milligen strata. Near the Triumph mine, the Milligen Formation is a structurally complex sequence of argillite, limestone, and sandstone (figs. 2, 3). Stratiform lead-zinc-silver ore of the Triumph mine is interbedded with carbonaceous shale, limestone, and diamictite. The Milligen strata strike northwest, dip southwest, and are folded, cleaved, and repeated by a series of southwest-dipping thrust faults (figs. 3, 4). These thrust faults locally host sheared lead-zinc-silver-bearing veins. Local calc-silicate alteration and silicification of calcareous beds and induration of fissile shales to nonfissile argillite within Milligen strata is interpreted to represent the metamorphic aureole of an intrusive body that underlies the Triumph mine area (Hall, Batchelder, and Tschanz, 1978). Low-angle normal faults offset Milligen strata and thrust faults and juxtapose Milligen rocks against younger limestone, quartzite, and conglomerate of the Wood River Formation and andesite of the Challis Volcanics. Klippe of Wood River Formation cap some of the higher ridges in the area. West of the Triumph deposit, the Milligen Formation is structurally overlain by rocks of the Challis Volcanic Group.

MILLIGEN FORMATION

Umpleby and others (1930) gave the name Milligen Formation to weakly metamorphosed dark-colored argillaceous rocks of supposed Devonian age in the Wood River

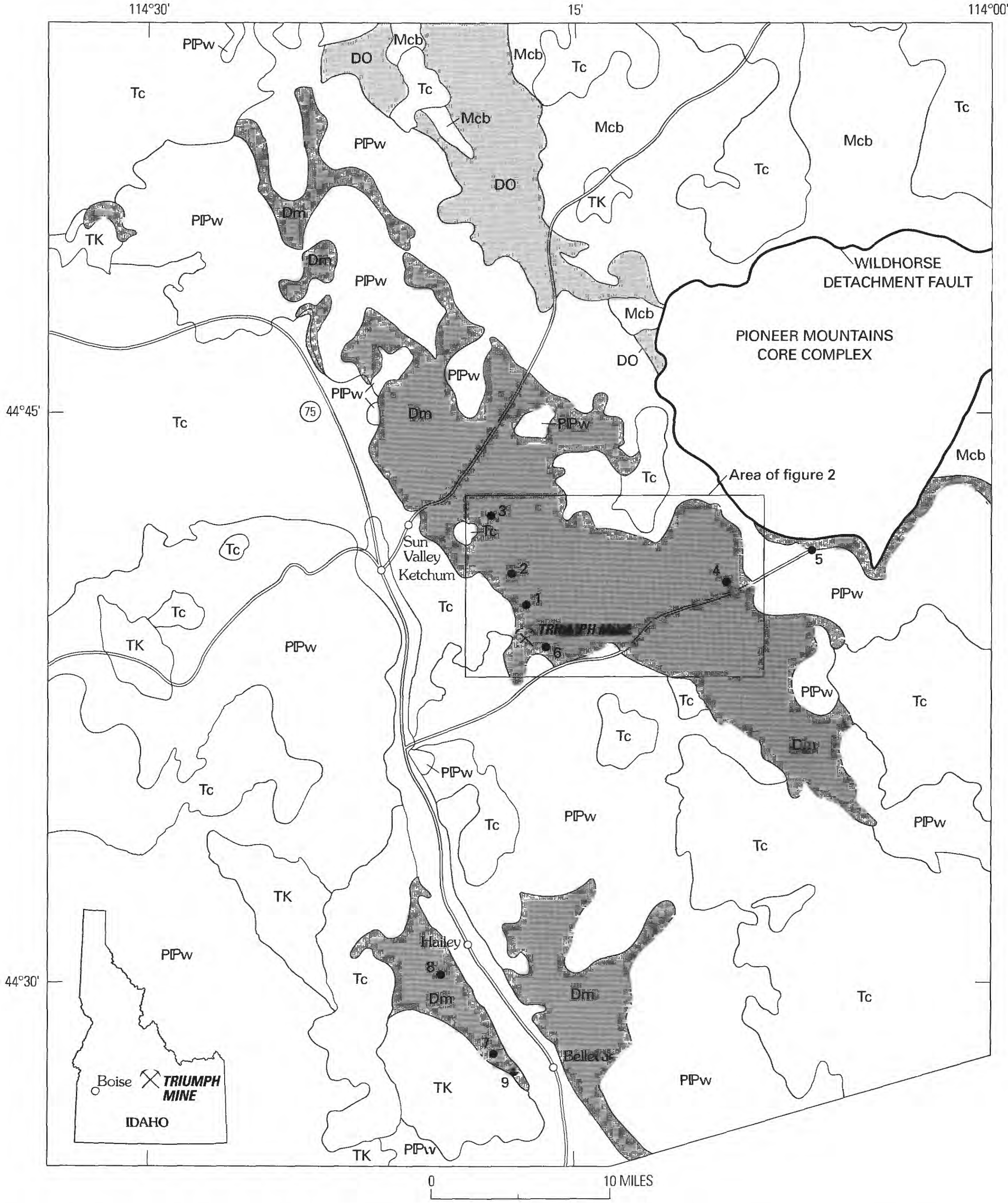


Figure 1 (above and facing column). Geologic map of the Wood River valley area, central Idaho, showing location of the Triumph mine and other lead-zinc-silver mines and prospects within the Milligen Formation. Area of figure 2 is also shown. Modified from Link and Mahoney (1989).

EXPLANATION

Tc	Challis Volcanic Group (Tertiary)
TK	Granodiorite and granite (Tertiary to Cretaceous)
PIPw	Wood River Formation and Dollarhide Formation (Permian-Pennsylvanian)
Mcb	Copper Basin Formation (Mississippian)
DM	Milligen Formation (Devonian)
DO	Phi Kappa and Trail Creek Formations and unamed Silurian-Devonian rocks (Devonian to Ordovican)

Contact—Approximately located

●7 Lead-zinc mine or prospect, labeled or referred to by number
1. Independence mine
2. Baltimore mine
3. Parker mine
4. Cait prospect
5. Mascot mine
6. North Star mine
7. Minnie Moore mine
8. Snoose mine
9. Silver Star Queen mine

area. The formation is present as a fault-bound allochthon (Dover, 1980), and stratigraphic contacts between the Milligen Formation and older or younger strata have not been recognized. The internal stratigraphy of the Milligen Formation is poorly understood due to structural complexity, poor exposure of rock units, and paucity of fossils. Umpleby and others (1930) named the formation after strata exposed in Milligan Gulch, about half a mile east of the Triumph mine. Sandberg and others (1975) divided the Milligen Formation into a thick lower argillite member containing thin quartzite beds (Lower and Middle Devonian) and a thin upper calcareous turbidite member (Upper Devonian). The age determination was based on conodont collections from several measured sections including one in Milligan Gulch.

Otto and Turner (1987) and Turner and Otto (1988) recognized five lithostratigraphic units within Milligen strata exposed in the East Fork Wood River drainage. These were informally named, in ascending order, (1) lower argillite, (2) Lucky Coin limestone, (3) Cait quartzite, (4) Triumph argillite, and (5) Independence sandstone. As noted in this report, however, the Lucky Coin limestone and Cait quartzite are locally developed lithofacies and not always present (fig. 5). Where the Lucky Coin limestone and Cait quartzite are present, they can have an intertonguing relationship with the Triumph argillite and lower argillite. Therefore, we divide these lithostratigraphic units into three informal units (lower argillite, Triumph argillite, Independence sandstone) and two intertonguing facies (Lucky Coin limestone, Cait quartzite).

Stratiform sulfide ore of the Triumph mine is interbedded with limestone and argillite interpreted as an intertonguing zone of Lucky Coin limestone and Triumph argillite; however, structural complications obscure the exact stratigraphic position of the stratiform sulfide orebody.

LOWER ARGILLITE UNIT

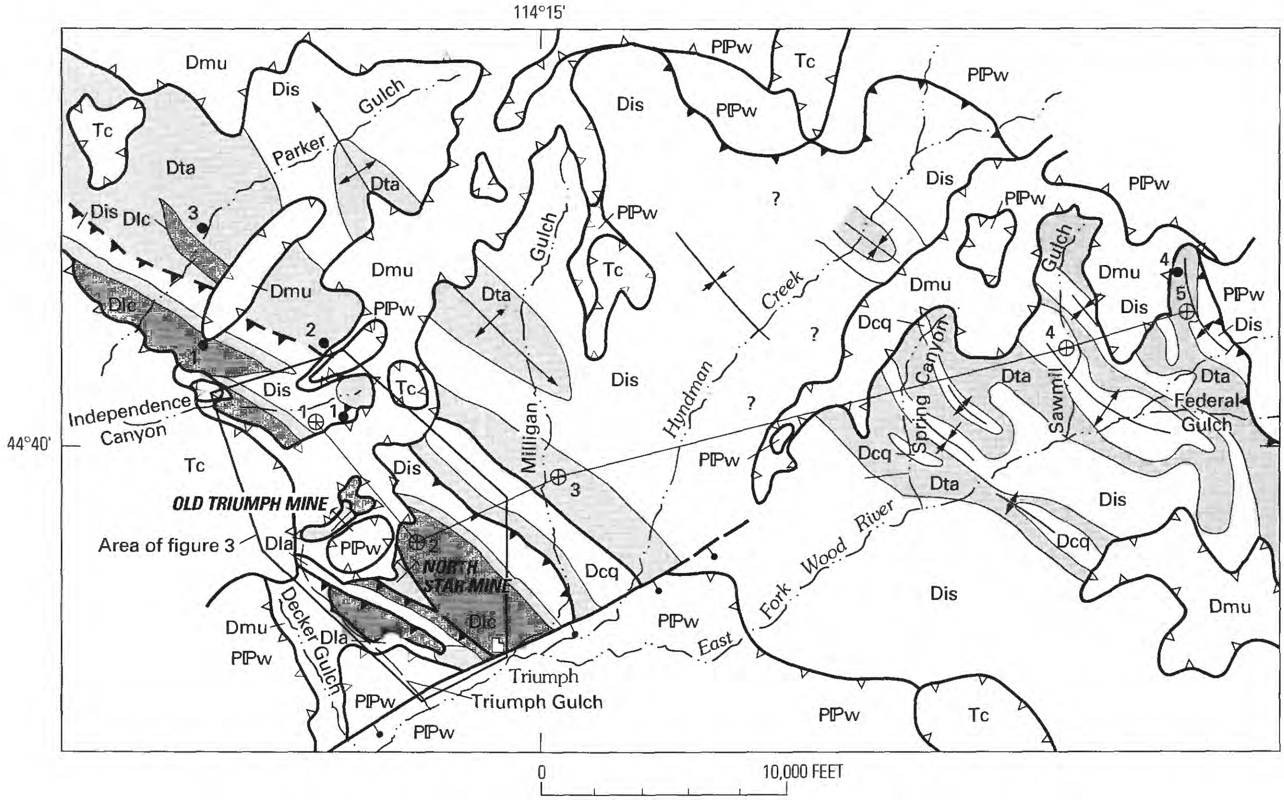
The lower argillite unit consists of siliceous black argillite and minor interbedded sandstone and chert. The lower argillite is best exposed near the Triumph mine where it crops out as resistant ribs between Triumph and North Star gulches (figs. 3, 4*C*). The argillite is massive and characterized by a well-developed southwest-dipping cleavage. Bedding is rarely preserved and typically transposed by cleavage. White chert beds or tremolitic sandstone beds are present locally. The base of the lower argillite is everywhere faulted, and a stratigraphic contact with older strata has not been observed. Our understanding of the lower argillite is hampered by strong cleavage development and limited area of exposure. The lower argillite is distinguished from the Triumph argillite by stratigraphic position with respect to the Lucky Coin limestone or Cait quartzite and by a lack of sooty organic-rich argillite.

Drilling at the Cait property in eastern East Fork valley (fig. 1) intersected a thick sequence of thin-bedded calcareous to noncalcareous, fine-grained sandstone, black argillite, and thin tuff beds below the Lucky Coin limestone (Otto and Turner, 1987). This unit is interpreted as a sand-rich facies of the lower argillite.

LUCKY COIN LIMESTONE FACIES

The Lucky Coin limestone facies is named after well-exposed interbedded limestone and black argillite near the Lucky Coin adit in the lower part of Independence Canyon (fig. 3). In Independence Canyon, the structural thickness of the Lucky Coin limestone exceeds 1,000 ft. Interbedded limestone and argillite below the Fissure fault and exposed on the ridge east of the North Star mine is at least 700 ft thick. Above the Fissure fault, limestone beds are more deformed and some are foliated marbles, precluding an estimate of original stratigraphic thickness.

Early workers in the Triumph mine recognized a lower and upper tremolitic limestone and noted their close association with the lead-zinc-silver ores (C.W. Merriam and C.N. Bozian, written commun., 1942). Kiilsgaard (*in* Anderson and others, 1950) mapped three discontinuous northwest-trending, southwest-dipping limestone units over a distance of 4 mi between the Triumph and Parker mines. Kiilsgaard's threefold subdivision was defined in Independence Canyon, a section we recognize as overturned and cut by faults that repeat the limestone units. Hence



EXPLANATION

- Tc — **Challis Volcanic Group (Eocene)**—Andesite to rhyodacite flows, breccia, and tuff
- PIPw — **Wood River Formation (Lower Permian to Middle Pennsylvanian)**—Sandstone, siltstone, bioclastic limestone, chert, and conglomerate

Milligen Formation (Upper to Lower Devonian)

- Dmu — **Milligen Formation, undifferentiated**
- Dis — **Independence sandstone unit**—Sandstone, argillite, and limestone
- Dta — **Triumph argillite unit**—Siliceous argillite and (or) carbonaceous argillite, sandstone, limestone, and diamictite
- Dcq — **Cait quartzite facies**—Interbedded quartzite and argillite
- Dlc — **Lucky Coin limestone facies**—Interbedded limestone and argillite
- Dla — **Lower argillite unit**—Siliceous argillite and chert and locally thin bedded sandstone

- **Contact**—Approximately located
- **Normal fault, low angle**—Sawteeth on downthrown side
- **Normal fault, high angle**—Ball and bar on downthrown side; dashed where inferred
- **Thrust fault**—Sawteeth on upper plate; dashed where inferred
- **Fold trace of synform**
- **Fold trace of antiform**
- •1 **Lead-zinc mine or prospect, labeled or referred to by number**
  1. Independence mine
  2. Baltimore mine
  3. Parker mine
  4. Cait prospect
- **Line of lithostratigraphic correlation shown in figure 5**—Locations of stratigraphic sections are also shown

Figure 2. Geologic map of the central part of the Milligen belt within the drainage area of the East Fork Wood River, Idaho. Location of map area is shown in figure 1.

Kiilsgaard's "Lower Limestone" lies stratigraphically above his "Middle Limestone" unit, and his "Upper Limestone" is a fault repetition of his "Lower Limestone" (fig. 3). The disappearance and reappearance of individual limestone units along strike reflects offset by low-angle normal faults, as in the gulch immediately north of the Triumph shaft (compare Kiilsgaard, fig. 16, with fig. 3, this paper).

We apply the term Lucky Coin limestone throughout the East Fork Wood River valley to a sequence of interbedded limestone and argillite underlying and interfingering with the Triumph argillite and interfingering with the Cait quartzite (fig. 5). On the Cait property in eastern East Fork valley, the Lucky Coin limestone is a 5–20-foot-thick bed overlying a thin bed of coarse-grained quartzite (Cait quartzite) and within a sequence of argillite.

LIMESTONE LITHOTYPES

The Lucky Coin limestone in the Triumph area is composed of three limestone lithotypes: (1) massive carbonaceous micrite, (2) thin- to medium-bedded limestone and argillite, and (3) foliated limestone or marble.

MASSIVE CARBONACEOUS MICRITE

Massive dark-gray-weathering carbonaceous micrite beds as thick as 40 ft are exposed in the lower part of Independence Canyon, east of the Triumph shaft, and in upper North Star Gulch (fig. 3). In Independence Canyon, angular fragments of mudstone as much as 2 ft in diameter within the micrite suggest a sediment gravity flow origin (Bouma, 1962). Thin interbeds of mud-chip bioclastic conglomerate suggest that the thick micrite units represent deposition of several gravity flow events that deposited a basal lag overlain by carbonate mud. Micrite is composed of fine-grained calcite and an abundance of disseminated micron-size carbonaceous grains. Angular detrital grains of quartz, plagioclase, potassium feldspar, and calcite as much as 0.2 mm in diameter are abundant in some beds. Scattered skeletal poikioblastic diamond-shaped tremolite and lathlike porphyroblastic chloritoid as long as 2 mm are in micritic limestone beds exposed in Triumph and North Star Gulches (fig. 6*A*).

THIN- TO MEDIUM-BEDDED LIMESTONE AND ARGILLITE

An overturned sequence of rhythmically interbedded medium-bedded limestone and calcareous and noncalcareous shale and argillite underlies the ridge east of North Star Gulch and extends west to the Fissure fault in the North Star mine area (figs. 3, 4*A*). This limestone sequence is well exposed on the access road in upper North Star Gulch (fig. 7*A*). Massive to plane-laminated limestone arenite and siltite, commonly having a coarse basal arenite lamination (T_{bde} turbidites), are as thick as 2 ft. These limestone beds are composed of carbonate and quartz grains in a lime silt and mud matrix.

Units as thick as 50 ft of thin-bedded micritic limestone crop out below the uppermost portal (Boiler level) at the North Star mine. These limestone beds are in the footwall of the Fissure fault and are composed of a basal lamina of medium-grained quartz as thick as 5 mm overlain by massive micrite.

A 150-foot-thick overturned and faulted section of Lucky Coin limestone was intersected below the Fissure fault (between 238 and 502 ft, fig. 4) in drill hole 81–2, 400 ft northeast of the Triumph mine shaft (Old Triumph mine in fig. 3). Three sequences, each as much as 50 ft thick, are composed of thinning-upward sandstone beds interbedded with carbonaceous chert and argillite. Sandstone beds, commonly 2–10 in. thick but as thick as 7 ft, have load cast bases and minor cross lamination and are interpreted to be T_{abd}, T_{ad}, and T_{bd} turbidites. These sandstone beds, metamorphosed to a calc-silicate assemblage and referred to as "footwall quartzites" by the miners, are described by Kiilsgaard (*in* Anderson, 1950) as pyritic, fine-grained calcareous quartzite 750 ft thick or more.

FOLIATED LIMESTONE AND MARBLE

Foliated, banded marble and limestone units are in the lower part of Independence Canyon, west and north of the Triumph shaft along the road to Independence Canyon, and on the ridge east of the lower part of Triumph Gulch. Banded marble units are composed of medium- to coarse-grained calcite and concordant and discordant lenses, pods, and veins of coarser calcite and quartz. Some marble units are composed of cigar-shaped calcite grains elongate to the northeast that suggest the principal stretching direction during deformation was in this direction. Banded marble units are transitional; a striped limestone unit is composed of an anastomosing network of pale-colored coarse-grained calcite bands that cut dark-colored fine-grained limestone. Where foliated limestone exhibits calc-silicate alteration, delicate porphyroblasts of tremolite and chloritoid extend undeformed across foliation, indicating calc-silicate mineral growth after deformation (fig. 6*A*).

DIAMICTITE

Diamictite in the Triumph area is present as discontinuous lenses within the Lucky Coin limestone and lower part of the Triumph argillite (figs. 3, 5). Diamictite units are as thick as 50 ft and composed of polymict fragments within a silty mud matrix. Clasts are subangular to subrounded, typically less than an inch in diameter (fig. 7*C*), and composed

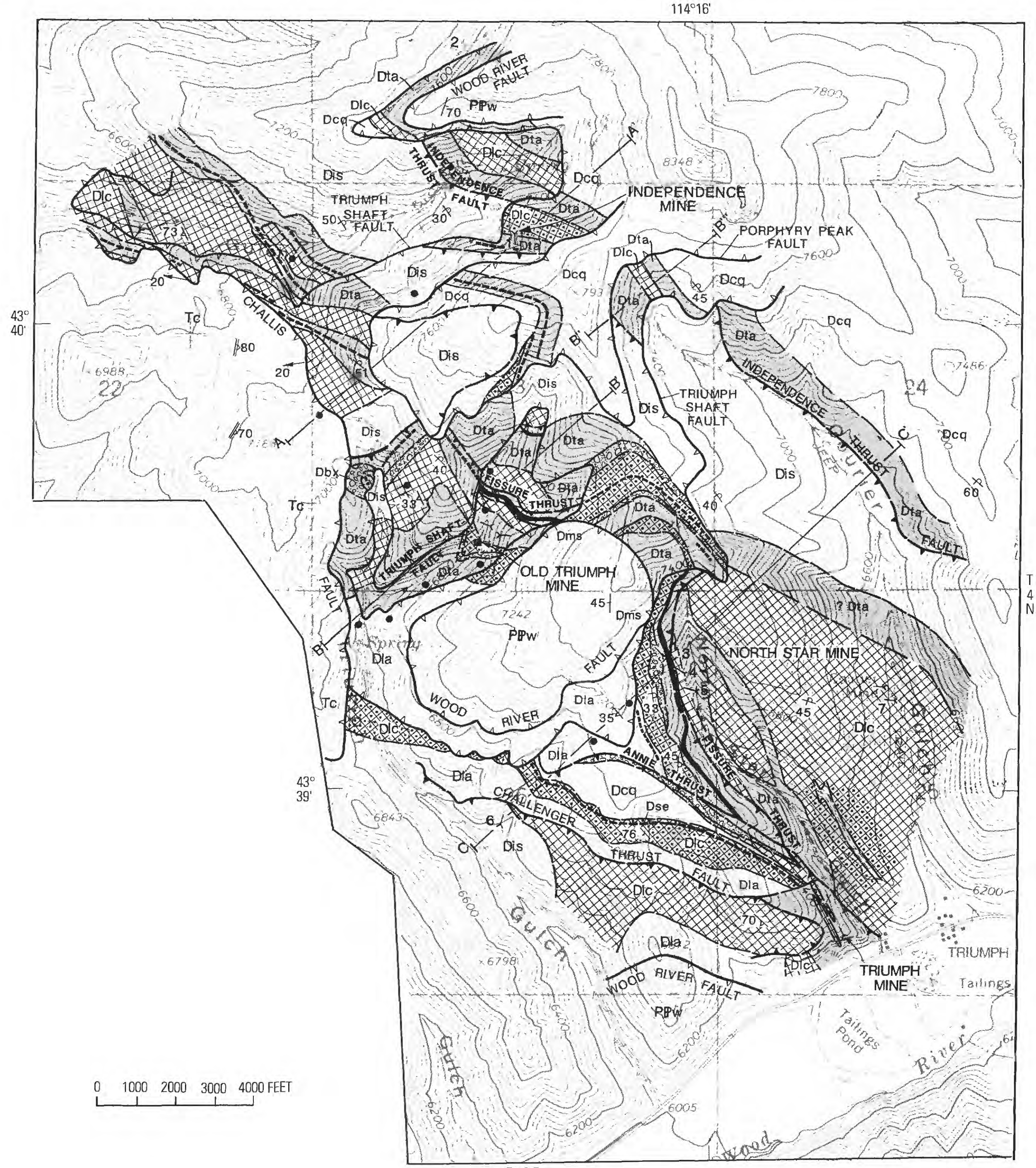


Figure 3 (above and facing column). Generalized geologic map of Triumph mine area, Idaho. Location of map area is shown in figure 2. Numbers refer to mines or prospects listed in figure 1. Lines of sections shown in figure 5 are also indicated. Base is U.S. Geological Survey Sun Valley 7.5-minute topographic quadangle (1967).

EXPLANATION

Tc **Challis Volcanic Group (Eocene)**—Andesite porphyry lava flows

PPw **Wood River Formation (Lower Permian to Middle Pennsylvanian)**—Quartzite, bioclastic limestone, and conglomerate

Milligen Formation (Upper to Lower Devonian)

Dis **Independence sandstone unit**—Sandstone and argillite

Dta **Triumph argillite unit**—Siliceous and (or) carbonaceous argillite, sandstone, and limestone

Dcq **Cait quartzite facies**—Argillite and quartzite

Dbx **Diamictite**—Mixed fragments in argillaceous matrix

Dms **Bedded gossan**—Bedded sulfide, ferroan carbonate

Dse **Bedded siliceous exhalite**

Dlc **Lucky Coin limestone facies**—Limestone and argillite, locally strongly foliated

Dla **Lower argillite unit**—Siliceous argillite and chert

Contact—Approximately located

Low-angle normal fault—Showing dip; sawteeth on downdropped side

Thrust fault—Sawteeth on upper plate

Strike and dip of beds

35 **Upright**

45 **Overturned**

33 **Strike and dip of foliation**

73 **Strike and dip of flow banding**

Diamond drill-hole collar

Mine shaft

Adit

of clasts of calcite-cemented quartz sandstone, tremolitic sandstone, argillite, quartz siltstone, feldspar-sericite rock, black chert, and white chert. The lithology of clasts is similar to strata in the Lucky Coin limestone and lower argillite. Quartz-tremolite-clinopyroxene fragments are interpreted as altered calcareous quartz sandstone fragments of the Lucky Coin limestone. Plagioclase-sericite-quartz clasts may represent altered mafic sills or tuffs.

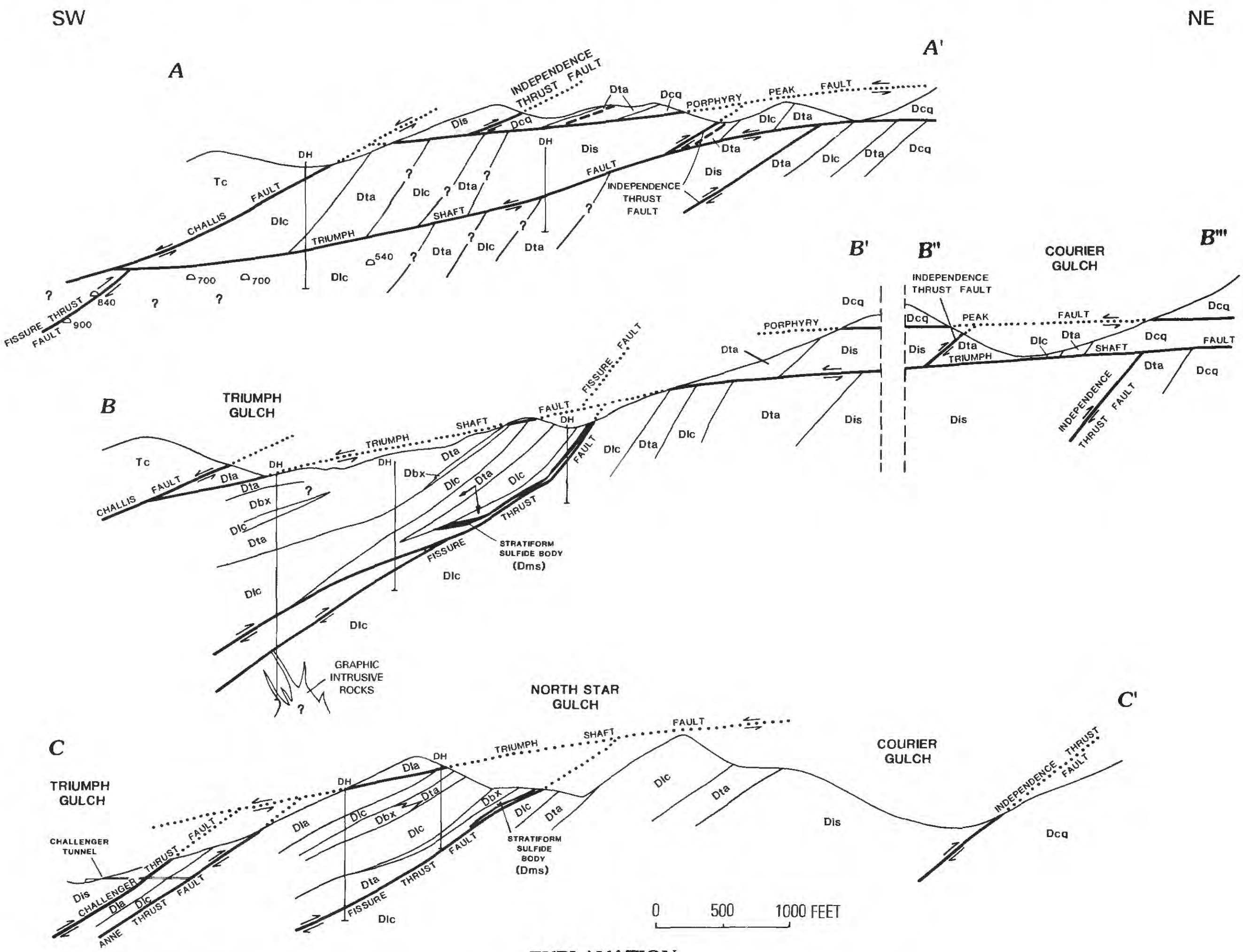
Diamictite lenses are immediately below the stratigraphic level of stratiform sulfides of the Triumph mine (fig. 5). A diamictite more than 20 ft thick is exposed in uppermost North Star Gulch. Based on drill core, this diamictite bed thins rapidly to the south, and a stratigraphically lower diamictite is interbedded with chert and limestone (fig. 5). The lower diamictite is composed of fragments of tremolitic sandstone, argillite, and minor pale chert. Twelve hundred feet north northwest of the Triumph shaft, drill hole 81–1 intersected seventeen feet of diamictite at the stratigraphic base of the overturned stratiform sulfide body (fig. 4). Fragments in this diamictite are predominantly calcite cemented quartz sandstone fragments and minor rounded black chert fragments floating within a calcareous mud matrix. About 1,500 ft southeast of the North Star mine area, and just above the Annie thrust fault (fig. 3), diamictite composed of quartzite and sandstone fragments in an argillite matrix overlies with angular discordance bedded argillite and quartzite. Diamictite interbedded with limestone, carbonaceous argillite, hydrothermal chert, and tremolitic limestone is exposed in a trench in the uppermost part of Triumph Gulch near the faulted contact between the Milligen Formation and Challis Volcanic Group.

Special significance is attached to diamictite because it is present at or near the stratigraphic level of the stratiform sulfide body, is thickest near the orebody, and may be related to syndepositional faults that localized ore formation. The poorly sorted nature and restricted composition of clasts suggest local resedimentation of Milligen sediments as slumps or debris flows, possibly related to the submarine scarps of syndepositional faults. Similar diamictite units are associated with other stratiform sediment-hosted zinc-lead deposits (see, for example, Abbott and Turner, 1990; Turner, 1990). These diamictites are interpreted to represent aprons of seafloor debris shed from syndepositional faults that were also conduits for metalliferous hydrothermal fluids that formed the sulfide deposits.

VOLCANIC TUFF BEDS

Thin, pale limonitic beds as much as an inch or so thick are exposed as interbeds within the Lucky Coin limestone on the east slope of North Star Gulch. These beds are distinct from interbedded quartzose siltstone and sandstone and are interpreted to be volcanic tuffs because of the abundance of plagioclase and potassium feldspar, sericite, and pyrite, the presence of minor amounts of chlorite, rutile, and tourmaline, and the absence of carbonaceous matter.

Tuff beds as thick as 1 ft within the lower argillite have been intersected in drill core at the Cait prospect in the eastern East Fork valley (Turner and Otto, 1988). These pale-green tuff beds are banded to laminated, contain argillite fragments and disseminated pyrite, chalcopyrite, and sphalerite, and are folded by F1 folds. They are composed of an intergrowth of sericite and felsitic plagioclase and lesser amounts of quartz, tourmaline, and altered glass. No primary volcanic textures were recognized.



EXPLANATION

Tc	**Challis Volcanic Group (Eocene)**—Andesite porphyry lava flows
	Milligen Formation (Upper to Lower Devonian)
Dis	**Independence sandstone unit**—Sandstone and argillite
Dta	**Triumph argillite unit**—Siliceous and (or) carbonaceous argillite, sandstone, and limestone
Dcq	**Cait quartzite facies**—Argillite and quartzite
Dbx	**Diamictite**—Mixed fragments in argillaceous matrix
Dms	**Bedded gossan**—Bedded sulfide, ferroan carbonate
Dse	**Bedded siliceous exhalite**
Dlc	**Lucky Coin limestone facies**—Limestone and argillite, locally strongly foliated
Dla	**Lower argillite unit**—Siliceous argillite and chert

Contact—Queried (?) where uncertain

Normal fault—Dotted where projected; arrows indicate direction of movement

Thrust fault—Dotted where projected; arrows indicate direction of movement

900 **Mine working**—Number indicates depth (in feet) below collar of shaft

Figure 4. Structural cross sections, Triumph mine area, Idaho. Lines of section are shown in figure 3. *A–A′*, North of Triumph mine (Independence mine area). *B–B′*, *B″–B‴*, Across the northern part of Triumph mine (Triumph shaft area). *C–C′*, Through the southern Triumph mine (North Star mine area).

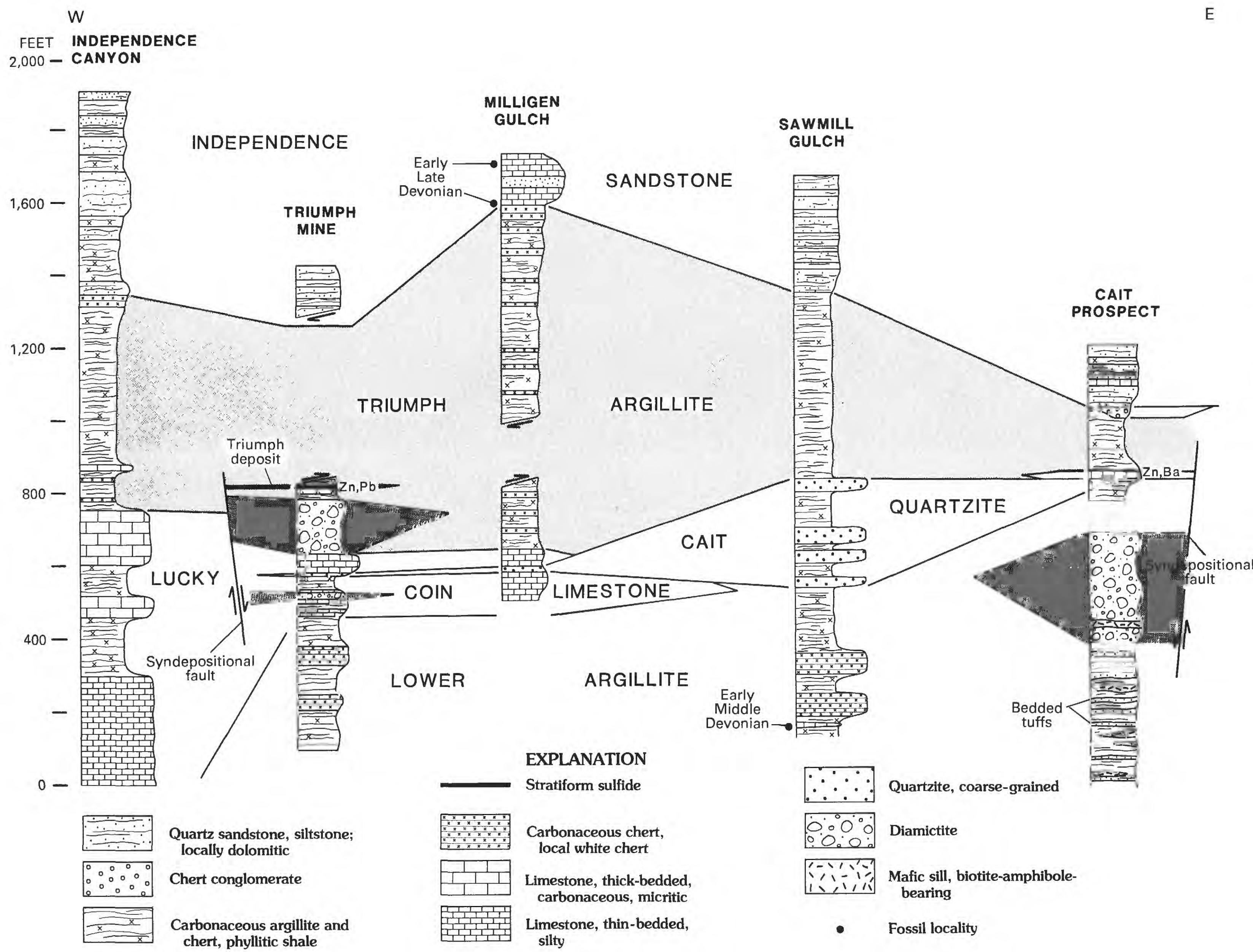


Figure 5. Lithostratigraphic correlation of informal units in the Milligen Formation, East Fork Wood River valley, Idaho. No horizontal scale implied. Dark shaded areas indicate extent of diamictite. Locations of stratigraphic sections are shown in figure 2.

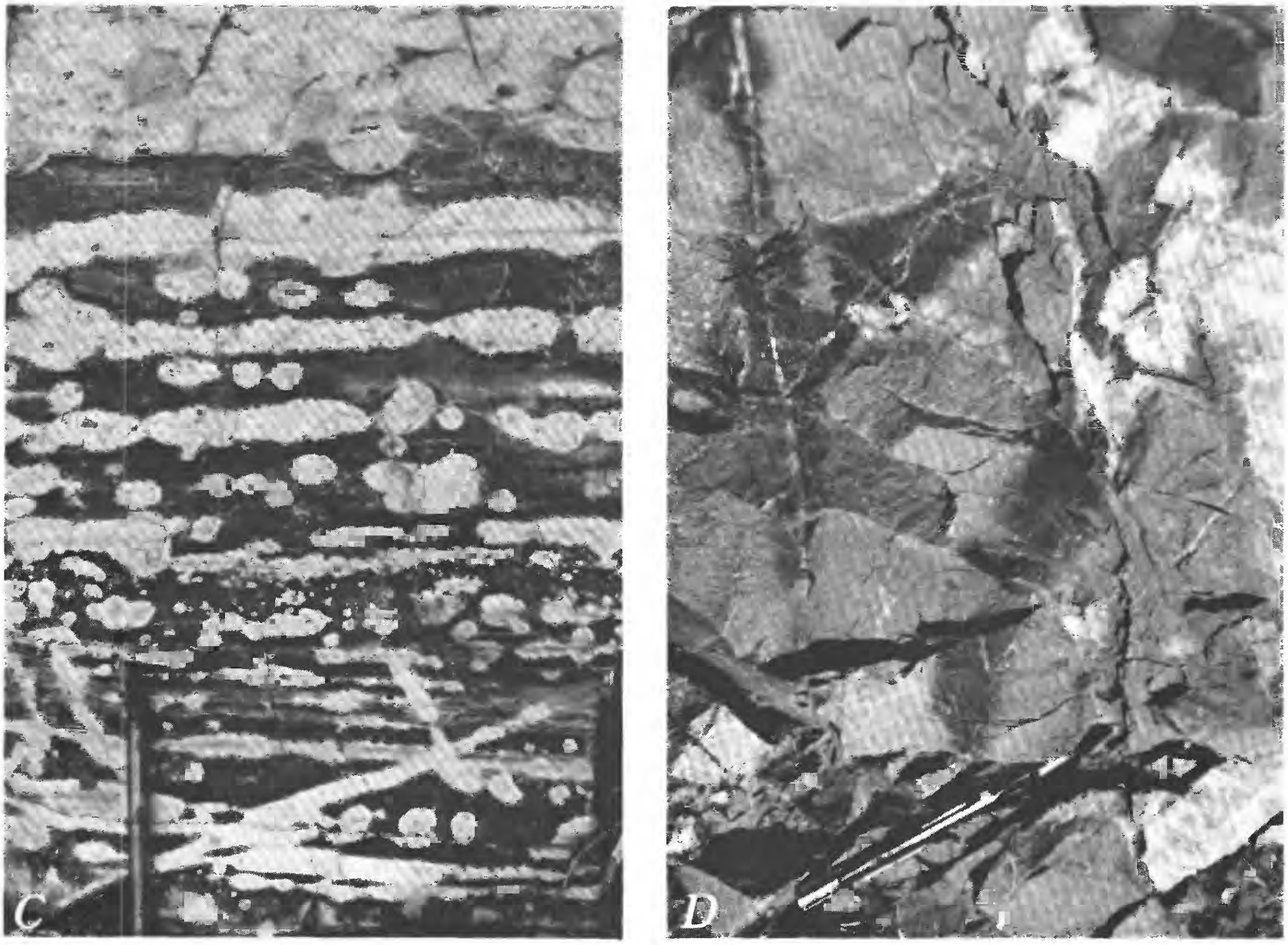

Figure 6 (above and facing page). Rock units and alteration types in the Milligen Formation, Triumph mine area, Idaho. *A*, Lucky Coin limestone facies. Randomly oriented tremolite grains (dark gray) in micritic limestone, Triumph mill area. Pen is shown for scale. *B*, Independence sandstone unit. Thin-bedded, fine-grained sandstone and argillite with wavy parallel laminations, upper part of Courier Gulch area. Hammer is shown for scale. *C*, Independence sandstone unit. Tremolitic calc-silicate alteration (white) as pods, bands, and discordant zones along fractures within calc-turbidite bed (gray), upper Independence Canyon. Pencil is shown for scale. *D*, Independence sandstone unit. Zones of silicification (dark gray) associated with thin quartz veins (white) cut calc-silicate-altered, thick-bedded, calcareous sandstone, Independence Canyon. Pencil is shown for scale.

MAFIC SILLS, FLOWS, AND DIKES

Concordant and discordant mafic igneous units as thick as 4 ft are present within Lucky Coin limestone strata on the east slope of North Star Gulch (fig. 7*D*). Mafic igneous dikes contain hornblende phenocrysts and plagioclase and hornblende microlites in a carbonate-altered matrix and minor amygdules filled with sparry carbonate, quartz, albite, and chlorite. An assemblage of sericite, carbonate, quartz, pyrite, leucoxene, and rutile is present in sheared dike margins. Concordant mafic units commonly display perlitic devitrification textures along one or both of their contacts, lack evidence of contact hornfels in adjacent strata, and may represent volcanic flows or sills. Some mafic units are folded and cut by cleavage and predate at least Mesozoic deformation. If these units are sills, the presence of devitrified glassy margins may indicate emplacement into wet unlithified sediments and hence a Devonian age. The occurrence of thin tuff beds in this same section lends support for an interpreted Devonian age of these mafic rocks.

CAIT QUARTZITE FACIES

The Cait quartzite facies was named for excellent exposures of coarse-grained, dark-gray quartzite and interbedded black argillite on the Cait lode mining claims in and near Sawmill Gulch east of the Triumph mine (fig. 2). The quartzite is composed of well-rounded quartz grains as



much as 1 mm in diameter cemented by quartz. The thickness of the Cait quartzite varies markedly in the East Fork valley area. At the Cait property the Cait quartzite is a single discontinuous bed less than 3 ft thick, whereas in the Sawmill Gulch area individual quartzite beds are more than 30 ft thick and the sequence is more than 300 ft thick.

In the Triumph area, the Cait quartzite is composed of black argillite and subordinate thin quartzite beds; it is thickest to the south and east in the area of Courier Gulch (fig. 3). On the ridge east of Courier Gulch, an overturned Cait quartzite sequence is composed of interbedded black carbonaceous argillite, quartzite beds as thick as 3 ft, minor chert conglomerate, and limestone. Flute marks on the base of some quartzite beds suggest transport to the southwest as turbidity flows. The Cait quartzite thins rapidly to the north and west toward the Triumph mine area where it intertongues with tremolitic sandstone and limestone of the Lucky Coin limestone, massive sulfide, and diamictite (fig. 5). The Cait quartzite probably pinches out northwest of the Triumph shaft, possibly reflecting a channel margin controlled by syndepositional faults (fig. 5).

TRIUMPH ARGILLITE UNIT

The Triumph argillite unit is a black argillite unit between the Lucky Coin limestone and younger Independence sandstone units. The Triumph argillite includes black argillite, carbonaceous siliceous shale, chert, limestone and tremolitic calc-silicate beds, and diamictite lenses (fig. 7*B*). The Triumph argillite is exposed in a northwest-trending belt in the area of the Triumph mine (fig. 5). Poor exposure and structural complexity limit our understanding of the internal stratigraphy and true thickness of the Triumph argillite. The best stratigraphic section of the Triumph argillite is exposed along an old road cut at the base of the south slope of Independence Canyon. Here the Triumph argillite is about 500 ft thick and conformably underlies the Lucky Coin limestone and overlies the Independence sandstone.

The dominant lithology of the member is a black, massive, weakly cleaved, siliceous argillite. In thin section, argillite laminae commonly have a microlenticular texture of aligned 0.05–0.1-millimeter-thick lenticles of turbid, cryptocrystalline chert rimmed by abundant carbonaceous material. Less common is microlaminated argillite composed of

Figure 7 (above and facing page). Rock units of the Milligen Formation, Triumph mine area, Idaho. *A*, Lucky Coin limestone facies. Overturned interbedded limestone and calcareous argillite, North Star mine portal access road. Hammer is shown or scale. *B*, Triumph argillite unit. Thin-bedded black chert, North Star portal area. Hammer is shown for scale. *C*, Diamictite composed of mixed fragments of calcareous sandstone, sandstone and chert. Cait property, East Fork valley. Coin is shown for scale. *D*, Mafic sill within overturned sequence of calcareous turbidites. Lucky Coin limestone facies, North Star mine portal access road. Hammer is shown for scale.

interlaminated carbonaceous material and chert or quartz-feldspar-silt.

The lower part of the Triumph argillite contains very carbonaceous argillite and chert, lenses of diamictite, minor white, gray, and buff chert beds, and the stratiform sulfide body of the Triumph mine. The lower part is exposed at the upper portal of the North Star mine where it is interbedded with bedded gossan of the Triumph orebody, near the pass between Independence Canyon and Triumph Gulch, northwest of the Triumph mine shaft, on the ridge between North Star mine portals and Triumph mine shaft, and on the ridge south of the North Star portal area above the Annie fault (fig. 3).

Near the Triumph mine, the Triumph argillite is more siliceous and indurated and displays less cleavage development than Triumph strata elsewhere in the Milligen belt. This textural difference may reflect a more siliceous nature of the mudrock in the Triumph area but more likely reflects increased induration by contact metamorphism.

INDEPENDENCE SANDSTONE UNIT

The Independence sandstone unit is the uppermost stratigraphic member recognized in the Milligen Formation (Turner and Otto, 1988). The Independence sandstone is named after an overturned sequence of sandstone 1,000 ft or more thick exposed in upper Independence Canyon (fig. 3). The top of the unit is everywhere in fault contact with other Milligen or younger strata. The Independence sandstone is composed of buff- to reddish-brown-weathering, fine-grained sandstone, black to gray-green argillite, silty and locally bioclastic limestone, and minor chert-pebble conglomerate.

Sandberg and others (1975) identified early Late Devonian (Frasnian) conodonts from bioclastic limestone conglomerate within argillaceous limestone exposed on the ridge east of Milligan Gulch; we correlate these strata with similar pale-gray limestone beds within the Independence sandstone and therefore suggest that the age of the Independence sandstone is Late Devonian.

TRIUMPH MINE AREA

Drill hole 81–1 (1,200 ft north-northwest of Triumph shaft) intersected a faulted upright section of Independence sandstone (between 70 and 400 ft thick) above the Fissure fault just north of the Triumph mine shaft (Old Triumph mine of fig. 3). A sandstone-rich section composed of two upward-thinning turbidite sequences, each about 80 ft thick, is overlain by carbonaceous chert and minor thin-bedded sandstone beds. The thinning-upward sequences are composed of sandstone beds as thick as 2 ft interbedded with lesser carbonaceous argillite and chert. Ball-and-pillow structures are common at the base of sandstone beds. Sandstone mineralogy is a calc-silicate assemblage of quartz-clinopyroxene-tremolite/actinolite-pyrite-plagioclase-calcite, relict detrital quartz (0.1–0.5 mm), and minor plagioclase.

The Independence sandstone is well exposed on the east side of the ridge northeast of the Triumph mine shaft in the upper part of Courier Gulch (fig. 4). In this area, sandstone beds as thick as 4 ft display planar lamination, climbing ripples, ball-and-pillow structures, and cross lamination and are interpreted as T_{abcd}, T_{bc}, and T_c turbidites.

EAST FORK VALLEY

The lower part of the Independence sandstone exposed on the east side of Milligan Gulch includes crinoid-rich bioclastic limestone, micritic sandy to silty limestone, quartzite, calcareous sandstone, siltite, argillite, and pebble conglomerate. Sandberg and others (1975) reported an early Frasnian (Late Devonian) fossil assemblage from these beds.

In the eastern part of the East Fork valley, the Independence sandstone is dominated by reddish-brown-weathering, thin-bedded, pyritic sandstone and argillite and lesser interbedded silty limestone and chert conglomerate. Sandstone displays abundant planar lamination, graded bedding, ball-and-pillow structures, and climbing ripples. The color of interbedded argillite is transitional from black near the base upward to pale green. The reddish-brown-weathering color is due to the oxidation of disseminated pyrite, which commonly makes up as much as 5 percent of the rock.

Chert-pebble conglomerate lenses as thick as 75 ft are locally present at the contact between the Independence sandstone and underlying Triumph argillite. Conglomerate is composed of white and gray chert, black argillite, and sandstone clasts as much as 1 in. in diameter and lesser fine- to medium-grained quartz. Thin beds as thick as 1 ft of lithologically similar conglomerate that is interbedded with Cait quartzite have also been noted on the ridge east of Courier gulch.

Limestone units as thick as 20 ft are composed of amalgamated, thin to medium beds of silty limestone.

WOOD RIVER FORMATION

In the area of the Triumph mine, the Wood River Formation is juxtaposed against Milligen strata along shallow, west-dipping normal faults. Wood River strata are exposed as bold outcrops at the mouths of Triumph Gulch and Independence Canyon and as fault klippe capping the peak northwest of North Star mine and the ridge north of Independence Canyon. The Wood River Formation includes blue-gray, sandy bioclastic limestone, gray quartzite, and chert conglomerate (Hall, Batchelder, and Tschanz, 1978). Kiilsgaard (*in* Anderson, 1950) described crinoids, productids, bryozoa, tabulate corals, and fusilinids from bioclastic limestone on North Star Peak. No base-metal-mineralized rocks are present within the Wood River Formation in the area of the Triumph mine.

GRANITE DIKES

Altered granitic dikes are present along the Fissure fault and are well exposed at the portal of the North Star mine. Granite dikes or small stocks within hornfelsed and calc-silicate Milligen strata have been drilled below the Triumph mine (fig. 4). These dikes contain coarse feldspar altered to sericite and clay, about 15 percent anhedral quartz grains (<10 mm), 10 percent biotite (<4 mm), and minor hornblende pseudomorphed by chlorite and pyrite. Elsewhere, similar dikes are highly altered to a granular intergrowth of quartz, muscovite, plagioclase, and as much as 15 percent pyrite. Plagioclase and quartz are intergrown with sutured margins. Muscovite grains (0.05–0.1 mm) are randomly oriented. Veinlike segregations of quartz and lesser pyrite, sphalerite, galena, rutile, and sphene are within the altered dikes.

ANDESITE DIKES

Biotite-hornblende andesite porphyry dikes, similar in composition to rocks of the Challis Volcanic Group, intrude Milligen strata. The dikes are composed of plagioclase (10–20 percent), hornblende (5–20 percent), and biotite (0–2 percent) phenocrysts in an aphanitic matrix. Kiilsgaard (*in* Anderson, 1950) noted that dikes commonly are within northwest-trending, southwest-dipping shear zones and are commonly dismembered, sheared, and variably altered, suggesting that intrusion was prior to last fault movement. In drill hole 81–1 immediately north of Triumph mine shaft, a weakly altered andesite porphyry in contact with highly altered quartz-feldspar porphyry within the Fissure fault zone suggests that intrusion of the andesite dike postdates the emplacement and hydrothermal alteration of the felsic dike. Similar andesite dikes also cut Wood River Formation strata, granitic rocks of the Idaho batholith west of Bellevue (Kiilsgaard *in* Anderson, 1950), and Challis volcanic rocks. Mafic dikes in the nearby Pioneer Mountains dominantly trend N. 50°–70° E. (Dover, 1981).

CHALLIS VOLCANIC GROUP

Andesitic flow rocks of the Challis Volcanic Group structurally overlie Milligen and Wood River strata west of the Triumph mine. These rocks are predominantly dark-greenish-gray hornblende-andesite that weathers rusty brown. Primary layering is rarely observed except immediately west of the saddle between Triumph Gulch and Independence Canyon where vesicle layers and elongate lithic fragments trend north and dip steeply east and west.

CONTACT METAMORPHISM AND ALTERATION IN THE TRIUMPH MINE AREA

A contact metamorphic aureole in Milligen strata first mapped by Batchelder and Hall (1978) extends from the Triumph mine 2 mi north to the Parker mine (fig. 1). In the Triumph mine area, the metamorphic aureole extends to the north from the North Star portal area and to the west to the Challis contact. Contact metamorphism is weak or absent in lower Courier Gulch and lower North Star Gulch. Metamorphosed thin limestone beds contain an assemblage of quartz, tremolite, clinopyroxene, feldspar, and chloritoid. Mudrock is metamorphosed to a massive argillite that locally contains a tremolite-biotite assemblage.

Drilling below the Fissure fault west of the Triumph mine intersected granite dikes as thick as 30 ft that cut bleached pink and green hornfels rock. Similar granitic dikes are within the Fissure fault zone in the area of the

Triumph mine. Within calcareous beds, a commonly noted increase in abundance of clinopyroxene relative to tremolite with increasing depth suggests increasing metamorphic grade. It is likely that the northwest-trending contact metamorphic halo is related to a similarly trending buried granitic intrusion.

Metamorphism is most pervasive and destructive of primary textures in thin-bedded calcareous turbidite. Tremolite, quartz, clinopyroxene, and calcite, variable amounts of muscovite, orthoclase, and minor sphene and pyrite are present as pods, bands, and massive replacement of calcareous turbidite beds (fig. 6*C*). On a microscopic scale, randomly oriented, radial, sheaflike masses of elongate poikilitic tremolite (0.1–1.0 mm) form a semicontinuous matrix for fine-grained carbonate minerals and orthoclase. Prismatic or irregular clinopyroxene and calcite grains (0.1–0.5 mm long) are intergrown with patchlike aggregates of muscovite (as much as 2 mm across), subhedral equant grains of orthoclase (0.1–0.3 mm long), and scattered amoeboid grains of sphene (0.05–0.5 mm). Calc-silicate altered rock is locally cut by zones of silicification and quartz veining (fig. 6*D*). Contacts between limestone and calc-silicate rock are commonly sharp. Metamorphism of thick micritic limestone units is limited to scattered dark-colored porphyroblasts of skeletal poikiloblastic and diamond-shaped tremolite or lath-shaped chloritoid grains (fig. 6*A*). These porphyroblasts are as long as 2 mm and may constitute as much as 5 percent of the rock.

Metamorphism of mudrocks includes silicification, calc-silicate alteration, and biotite hornfelsing. Mudrock altered to biotite hornfels at the head of Courier Gulch is composed of a cryptocrystalline aggregate of quartz, feldspar, and as much as 30 percent biotite and is cut by minor veinlets of potassium feldspar, biotite, and pyrrhotite. Above the Fissure fault northwest of the North Star portal area, mudrock is altered to a fine-grained assemblage of quartz, potassium feldspar, and tremolite and veinlets of tremolite and albite. Interbedded calc-turbidites are converted to calc-silicate assemblages.

Sandstone beds in the Independence sandstone unit are altered to an assemblage of quartz-tremolite-actinolite-clinopyroxene-calcite-pyrite. Tremolite is the dominant calc-silicate mineral. In drill hole 81–1 northwest of the Triumph mine shaft, abundant quartz, clinopyroxene, and actinolite are associated with veinlets of quartz and minor clinopyroxene, actinolite, pyrite, and calcite and trace sphalerite and galena.

Mineral textures indicate that metamorphism occurred after the earliest cleavage-forming deformation. Tremolite is characterized by randomly oriented radial and sheathlike aggregates that cross cleavage. Growth of calc-silicate minerals along cleavage supports a syn- or post-deformation timing of metamorphism (fig. 8*A*). The abrupt disappearance of calc-silicate alteration across the Fissure fault in the upper part of North Star Gulch suggests that some of the fault movement postdates metamorphism.

STRUCTURE OF THE TRIUMPH MINE AREA

GENERAL STRUCTURE

Three major deformational events are recorded in the Milligen strata of the East Fork valley (Turner and Otto, 1988). An early deformation event (D1) that includes tight to isoclinal folds and axial planar cleavage subparallel with bedding may be related to the Early Mississippian Antler orogeny or to younger Mesozoic deformation. Map-scale, northwest-trending, east-vergent folds (D2) that show variably developed cleavage deform D1 structures, as well as Pennsylvanian and Permian strata. Northwest-trending, southwest-dipping thrust faults are likely associated with D2 folds. Gently and steeply dipping extensional faults (D3) offset strata as young as Eocene age and are inferred to be related to formation of the nearby Pioneer Mountains core complex.

In the Triumph mine area, Milligen strata generally strike N. 55°–65° W. and dip 30°–60° SW. Although casual observation of this area suggests a simple homoclinal structure, the rocks are intensely faulted, tightly to isoclinally folded (fig. 8*B*), and generally on the overturned limb of a large east-vergent fold. Locally, D1 cleavage is folded by east-vergent folds (fig. 8*C*). A set of southwest-dipping thrust faults (D2) repeats Milligen strata; some of these thrust faults host sheared lead-zinc-silver vein mineralization. The Triumph stratiform orebody is immediately above a thrust (Fissure fault). Low-angle normal faults displace Milligen strata and thrust faults from east to west and juxtapose Milligen rocks against structurally higher Wood River Formation and Challis Volcanic Group.

DUCTILE DEFORMATION

A penetrative cleavage (D1) associated with tight to isoclinal folds is variably developed in the Triumph mine area. The cleavage generally strikes N. 55°–65° W., dips 30°–60° SW., and cuts bedding at a low angle. D1 cleavage is best developed in unmetamorphosed massive limestone and calcareous shale of the Lucky Coin limestone, shale of the Triumph argillite, and thin-bedded sandstone and shale of the Independence sandstone. Within the contact aureole, this cleavage is commonly absent, presumably due to mineral growth and recrystallization. A lack of similar cleavage within Pennsylvanian and Permian Wood River strata in the Triumph mine area suggests a pre-Pennsylvanian age for the cleavage; however, it may reflect the different rheology of massive limestone and quartzite in the Wood River Formation.

A belt of northwest-trending, southwest-dipping beds of foliated marble and cleaved argillite (D1) is west of the Triumph mine. Most of these highly deformed rocks are

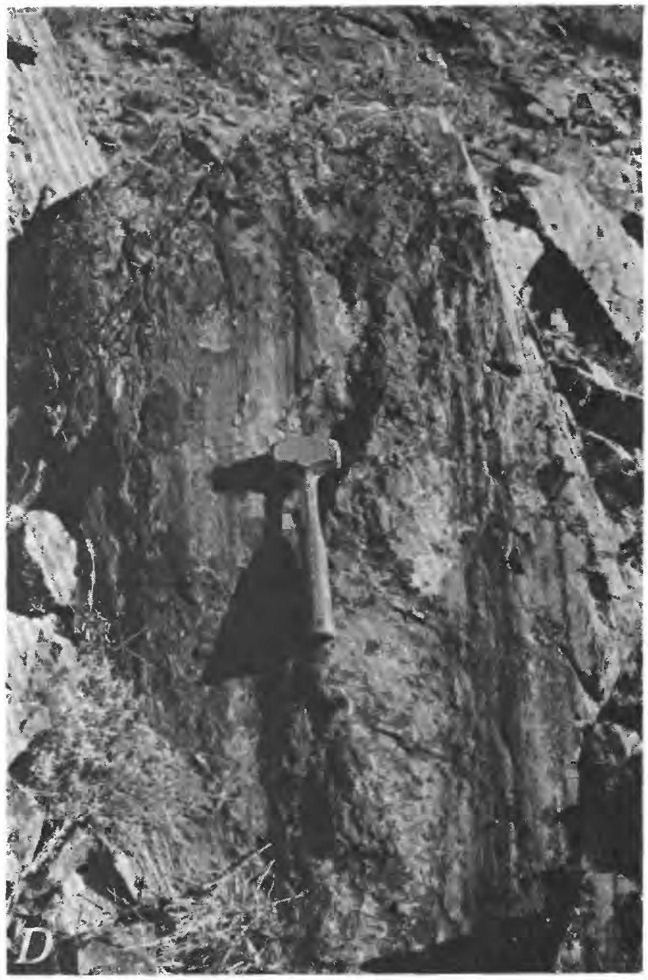
Figure 8 (above and facing page). Alteration types and structures in the Milligen Formation, Triumph mine area, Idaho. *A*, Folded bed of calcareous sandstone interbedded with black argillite. Tremolitic calc-silicate alteration has preferentially developed along cleavage axial planar to fold indicating tremolite formation is younger than deformation. Hammer is shown for scale. *B*, D1 fold. Asymmetric fold in interbedded white and black chert, lower part of Triumph Gulch. Key is shown for scale. *C*, D2 fold. D1 cleavage in calcareous argillite folded into east-vergent fold, North Star Gulch. Coin is shown for scale. *D*, D3 fault. Talus block with slickensides on a polished quartz fault surface, Cait prospect. Similar fault surfaces with northeast-trending slickensides are typical of D3 low-angle faults in the Triumph area and throughout the East Fork Wood River valley. Hammer is shown for scale.

above the Fissure fault. Foliated marble is inferred to have been the locus of early D1 ductile shearing and is truncated by later thrust faults (D2). Just west of the Triumph shaft, banded marble contains augen of coarse quartz and calcite and tightly folded calcite and quartz veins. Calcite grains within the marble strongly elongated in a northeasterly direction suggest that D1 deformation involved northeast transport. Within adjacent hornfelsed Triumph argillite, cleavage is axial planar to isoclinal folds of dismembered sandstone laminae.

Reversals of stratigraphic top direction in drill holes suggest that the structural hanging wall of the Triumph orebody is deformed by isoclinal folds having wavelengths of 10–30 ft. The sense of asymmetry and the cleavage-bedding relationships of outcrop-scale folds suggest that this sequence is on the overturned limb of an east-vergent fold.

Locally, small east-vergent folds deform the early cleavage (fig. 8*C*). Variations in dip of the cleavage, such as northeast-dipping cleavage on the west side of the upper part of Courier Gulch, may reflect folding by map-scale folds.

Interpretation of the structural history of the Triumph mine area is complicated by the contact metamorphic fabric that overprints older rock fabrics. Two interpretations are possible. Either both D1 folds and cleavage and D2 folds are Mesozoic, or the D1 cleavage is pre-Mesozoic (possibly Antler deformation) and younger minor D2 folds are Mesozoic. In the Triumph area there is no reason to believe that both ductile deformation events (D1 and D2) are not



Mesozoic in age; however, in the eastern East Fork valley near the Cait prospect, early D1 folds and cleavage are cut by a younger cleavage, and Turner and Otto (1988) interpreted the early deformation as Devonian Antler deformation based on its ductile soft-sediment style of deformation.

THRUST FAULT ARRAY

Four major northwest-trending, southwest-dipping thrust faults repeat Milligen strata in the area of the Triumph mine. From northeast to southwest and from lower to higher structural position, these faults are the Independence, Fissure, Annie and Challenger thrusts (figs. 3, 5), and the bodies of rock above each thrust fault are referred to as the Independence, Fissure, Annie, and Challenger blocks, respectively. Kiilsgaard (*in* Anderson, 1950) interpreted a reverse sense of displacement on several of these faults based on underground observation of large-scale drag folds adjacent to shear zones, as well as evidence of stratigraphic offset. Although these observations suggest that these faults are thrusts, younger strata are placed on older in the Triumph mine area. This implies that faulting disrupted a previously folded stratigraphic panel. Kiilsgaard also noted that shear zones associated with reverse faults are characterized by a sharp hanging wall fault plane that has slickensides suggesting oblique movement; these structures may reflect reactivation of thrust faults during Eocene extension.

The Challenger, Annie, and Fissure thrust faults are imbricate and converge in the southern part of North Star Gulch. Similar thrust faults have not been recognized outside the Triumph mine area, possibly because of a lack of detailed work in these areas. Regional mapping of Milligen strata to the east indicates that these strata are deformed by map-scale folds (fig. 2).

Much of the lead-zinc-silver ore mined in the Parker-Triumph belt was produced from mineralized thrust faults or "fissures" (Kiilsgaard *in* Anderson, 1950). Ore in the Independence, Baltimore, and Parker mines and the Fissure ore at the Triumph mine are present as pods, streaks, and stringers of vein-type ore within gouge or highly sheared rock. The fault zones were a locus of sulfide deposition, possibly remobilized from stratiform sulfide deposits; latest fault movement postdates mineralization. Thrust faults are offset by low-angle normal faults (figs. 3, 4) and northeast-trending steep normal faults (fig. 2).

INDEPENDENCE THRUST FAULT

The Independence thrust fault is the most northeasterly and structurally lowest fault in the Triumph-Parker thrust array (fig. 2). This important northwest-trending, southwest-dipping structure was mined for lead-silver ore at both the Independence and Baltimore mines and can be traced from north of Independence Canyon to the lower part of Milligan Gulch, a distance of 4 mi (fig. 3). Contractional movement on the Independence fault is indicated by the juxtaposition of overturned strata of the Independence sandstone against underlying overturned argillite and quartzite of the Triumph argillite and Cait quartzite. The Independence fault is offset in several locations by low-angle normal faults. As a result, parts of the Independence fault are in three separate fault-bound panels; the fault is offset progressively further to the west at successively higher structural levels (fig. 5).

The Independence mine was a significant producer of lead and silver and lesser gold from 1883 to about 1923. Umpleby and others (1930) described the Independence fault at the Independence mine as a variably striking zone that has a steplike profile of shallower (20° dip) and steeper (50° dip) segments consisting of crushed carbonaceous argillite. Ore was concentrated within the steeper parts of the structure.

FISSURE THRUST FAULT

The northwest-trending, southwest-dipping Fissure thrust fault subcrops on the west side of North Star Gulch, as well as within a structural window below the Triumph shaft low-angle fault near the Triumph shaft. Sheared and mineralized rock west of the saddle between Triumph Gulch and Independence Canyon may be associated with the Fissure fault, offset to the west by the Triumph shaft fault. Ore in the Triumph mine is within the Fissure fault ("Fissure ore") and as stratiform sulfides overlying the fault. Early miners interpreted the faults at the Triumph and North Star mines as separate structures; however, we interpret them as a single structure offset by a down-to-the-northwest normal fault shown on underground mine maps (Triumph Mining Company, unpublished maps).

C.W. Merriam and C.N. Bozian (written commun., 1942) described the Fissure fault in the Triumph mine as a 5–20-foot-thick zone of crushed rock, clay gouge, graphitic flour, and pods and fragments that have multiple slickensided surfaces and include sulfide-quartz-carbonate, altered quartz porphyry granite dikes, and altered mafic dikes. The hanging-wall surface of the fault zone is well defined and has mullion structure. Drag folds are numerous, and their fold sense suggests that the principal movement was reverse dip-slip; however, slickenside striae suggest several directions of fault movement. The fault surface generally strikes northwest and dips 20°–40° SW., but in the northwestern part of the Triumph mine the strike swings westerly and then northerly, forming a large-scale bend in the fault plane.

The Fissure fault has a complex history that includes early contractional movement, deposition of lead-zinc-silver-gold ore, intrusion by granitic and andesitic dikes, and reactivation as a normal fault coincident with or following intrusion and mineralization (unpublished data). Contractional movement on the Fissure fault is suggested by repetition of the Lucky Coin limestone and Triumph argillite units, repetition of the Triumph stratiform sulfide body above the 200-foot level in the Triumph mine along imbricate splays of the fault, and small-displacement northeast-directed thrust faults noted in outcrop adjacent to the fault. Dikes and sulfide-quartz-carbonate mineralized rocks within the Fissure fault are cut by normal faults and have foliated margins. The Fissure fault is offset by a series of northeast-trending, down-to-the-northwest normal faults.

ANNIE THRUST FAULT

The Annie thrust fault, exposed on the ridge between North Star and Triumph Gulches, juxtaposes overturned Cait quartzite and Lucky Coin limestone against lower argillite and Lucky Coin limestone (fig. 3). The more westerly trending Annie fault merges with the underlying Fissure fault in southern North Star Gulch. The Annie fault is cut by the Triumph shaft fault, a low-angle normal fault (fig. 4).

CHALLENGER THRUST FAULT

The Challenger thrust fault has the highest structural position of the four thrusts in the Triumph mine area. It is named for exposures in the Challenger tunnel in the lower part of Triumph Gulch where it is a 50-foot-wide zone of chloritic and graphitic slickensided gouge (fig. 4). Contractional movement on the Challenger fault is indicated by the juxtaposition of overturned strata of the Lucky Coin limestone and Independence sandstone units against underlying overturned lower argillite. The fault cuts upsection from southeast to northwest in the hanging wall.

LOW-ANGLE NORMAL FAULTS

A major contribution of this study has been recognition of the importance of low-angle normal faults. Three shallowly west dipping normal faults offset Milligen strata and older thrust faults and juxtapose younger rocks of the Wood River Formation and Challis Volcanic Group against Milligen strata. In successively higher structural position, these faults are the Triumph shaft, Porphyry Peak–Wood River, and Challis (figs. 3, 4). Offset of Milligen units and thrust faults indicates a net westerly displacement of upper plate rocks on the Triumph Shaft and Porphyry Peak faults. Slickenside striae in adjacent parallel subsidiary faults indicate movement to the northwest (fig. 8*D*).

These low-angle faults are not mineralized, and they offset lead-zinc-silver-gold-bearing thrust faults. Low-angle faults both offset and are offset by northeast-trending high-angle normal faults of small displacement.

TRIUMPH SHAFT FAULT

The Triumph Shaft fault, the structurally lowest major low-angle fault recognized in the Triumph area (fig. 4*B*), is partially exposed in the upper parts of North Star, Triumph and Independence drainages (fig. 3). The Triumph Shaft fault dips shallowly to the west at 5°–15° and appears to flatten to the east (fig. 4*B*). The Triumph Shaft fault zone is poorly exposed, but subsidiary low-angle faults below the Triumph Shaft fault are well exposed in cliffs immediately northwest of the Triumph Shaft. Slickenside striae on these and other faults of the Triumph Shaft fault zone indicate a primary top-to-the-northwest sense of displacement and a secondary west-southwest trend of movement. The Independence thrust fault is offset more than 1,200 ft to the west by the Triumph Shaft fault; total offset could be considerably more than this westerly component of displacement (fig. 3). An updip portion of the Fissure fault and Triumph stratiform body is above the Triumph Shaft fault west of the saddle between Triumph and Independence drainages (figs. 3, 4*B*).

We correlate the Triumph shaft fault with a regionally extensive flat fault that can be traced to the northeast for 4 mi along the divide between Parker and Milligan Gulches (fig. 2). This structure may be equivalent to structures exposed on similar northeast-trending ridges between Milligan Gulch and Hyndman Creek and between Hyndman Creek and East Fork of Wood River (fig. 2). If this correlation is correct, the areal extent of this structure exceeds 30–40 mi^2.

PORPHYRY PEAK–WOOD RIVER FAULTS

The Porphyry Peak fault, about 300 ft above and subparallel with the Triumph shaft fault, subcrops on the ridgeline between the Independence, Triumph, and Courier drainages (figs. 3, 4*A*). The Porphyry Peak fault displaces strata of the Fissure and Independence blocks and the Independence fault 2,000–2,500 ft to the west. Given evidence for northwesterly displacement on other low-angle faults, we suspect a northwestly displacement on the Porphyry Peak fault; however, total displacement in a northwesterly direction is difficult to estimate because this direction is subparallel with the trend of Milligen strata and thrust faults.

The low-angle Wood River fault underlies a klippe of Wood River Formation on the peak south of the Triumph mine shaft ("Old Triumph mine" in fig. 3). South of Triumph mine shaft this fault zone dips about 40° to the southwest, contains northwest-trending slickensides, and truncates the Fissure thrust fault. The Wood River fault has structural position similar to that of the Porphyry Peak fault. North of Independence Canyon near the Baltimore mine, Milligen Formation is in thrust contact with underlying Wood River Formation and both are truncated by the Wood River fault (M.E. Ratchford, University of Idaho, oral commun., 1989). Here the Wood River fault is a silicified and brecciated fault zone that has east-trending slickenside striae.

The Porphyry Peak and Wood River faults were originally considered different faults because of the differing rock types in the upper plate (Milligen Formation versus Wood River Formation); however, the occurrence of structurally juxtaposed Milligen and Wood River strata above the Wood River fault north of Independence Canyon suggests that what we refer to as the Wood River fault is the same structure as the Porphyry Peak fault.

CHALLIS FAULT

Rocks of the Challis Volcanic Group of Eocene age are juxtaposed against underlying Milligen and Wood River strata on the west side of the Triumph mine area along the shallowly west dipping Challis fault (fig. 2). The Challis fault exposed in surface trenches in the upper part of Triumph Gulch is a broad zone of altered and brecciated volcanic rock above well-foliated Milligen argillite; overlying Challis volcanic rocks dip steeply into the fault zone. In a drill hole several hundred feet north of the trench, the Challis fault is a 3-foot-thick gouge zone that contains mixed Challis and Milligen fragments and is underlain by 30 ft of brecciated Milligen strata. Challis volcanic rocks are also faulted against Milligen strata in the lowest levels in the northwestern part of the Triumph mine (R. House, Hailey, Idaho, oral commun., 1985). This interpretation that the Milligen-Challis contact is a fault differs from that of Kiilsgaard (*in* Anderson, 1950), who interpreted the rubbly and weathered nature of the contact in the Lucky Coin adit and Triumph mine as an unconformity.

Small low-angle faults having west-trending slickenside striae were noted in several places within rocks of the Challis Volcanic Group.

STEEPLY DIPPING OBLIQUE-SLIP FAULTS

A widespread set of steeply dipping (60°–90°), northeast-trending (N. 55°–80° E.), small-displacement faults that offset low-angle normal faults may be the youngest structures in the Triumph mine area. Slickenside striae on nine such faults trend N. 40°–90° E. and plunge shallowly northeast. Offset of marker beds on some of these faults suggests right-lateral oblique-slip displacement. In the Triumph mine, northeast-striking, steeply west dipping faults offset the orebody and the Fissure fault (C.W. Merriam and C.N. Bozian, written commun., 1942) with apparent down-to-the-north or right-lateral displacement. The largest such fault has an apparent normal offset of 200 ft and separates the Triumph orebody into the structurally higher North Star orebody to the southeast and the structurally lower Old Triumph orebody to the northwest. These faults parallel the lower drainage of the East Fork valley. Batchelder and Hall (1978) interpreted right-lateral offset on a fault under the lower East Fork valley based on apparent right-lateral offset of northwest-trending normal faults of the Wood River graben system.

TRIUMPH OREBODY

HISTORY OF MINE DEVELOPMENT

The Triumph orebody includes a southwest-dipping stratiform sulfide body locally referred to as bedded "complex" and "siliceous" ores, as well as underlying faulted vein-mineralized rock or "fissure ore" within the southwest-dipping Fissure fault zone. Early production was mostly from the high-grade "complex ores" of the stratiform sulfide body. With the exhaustion of those high-grade ores, production included lower grade "siliceous ores" and the widespread but discontinuous "fissure ores."

Production at the Triumph mine spanned thirty years from 1927 to 1957. Initial development of the Triumph orebody was from adits at the head of North Star Gulch (North Star mine) and in the eastern part of upper Triumph Gulch (Old Triumph mine). During the early years, production came from a 700-foot-deep shaft at the Old Triumph mine. Later, an interior shaft extended development to the 850-foot level (level elevations are the depth below the surface shaft collar). Ore was taken by aerial tram to the Gimlet railway siding, 5 mi to the west. In 1943 a 6,500-foot-long haulage tunnel (Plummer tunnel) was completed and connected the 700-foot level of the mine with the North Star mill in the East Fork valley near the village of Triumph (fig. 3). A total of 10 mi of tunnels, excluding stopes, was developed. By the 1950's, the high-grade bedded ore was exhausted and work concentrated on developing "fissure ore" reserves in the deeper northwestern part of the mine.

STRUCTURE AND STRATIGRAPHY OF STRATIFORM OREBODY

The bulk of the production from the Triumph mine came from a stratiform deposit containing bedded pyrite-sphalerite-galena-ferroan carbonate-quartz ore. C.W. Merriam and C.N. Bozian (written commun., 1942) inferred this bedded ore to be the result of sulfide replacement of thin limy partings interbedded with argillite beds. The stratiform sulfide body is more than 4,000 ft long along its northwest strike and is within an overturned sequence of interbedded limestone, carbonaceous argillite, and diamictite immediately above the Fissure fault. Strata younger than the bedded ores have been offset by the Fissure fault, and, because only thin, faulted slivers of argillite and limestone remain, the stratigraphic position of the Triumph deposit is somewhat ambiguous. It may lie within the Lucky Coin limestone, or, more probably, it is near the contact of the Lucky Coin limestone and overlying Triumph argillite.

Workers in the Triumph mine described two types of bedded ore: high-grade massive pyrite-sphalerite-galena "complex ores" and lower grade ferroan carbonate-quartz-sphalerite-galena "siliceous ores" (C.W. Merriam and C.N. Bozian, written commun., 1942). The term complex ore, used to describe massive banded fine-grained pyrite, sphalerite, and galena, reflected the metallurgical problems associated with these zinc-rich ores. A typical assay of complex ore was 6–13 percent zinc, 3–6 percent lead, 6–7 ounces of silver per ton, and 0.01–0.02 ounces of gold per ton. C.W. Merriam and C.N. Bozian (written commun., 1942) described complex ore as commonly bedded or laminated, dense, fine-grained pyrite and lesser amounts of sphalerite and galena. Sulfide bands are interbedded with carbonaceous, pyritic chert and siltstone. Contacts between sulfide beds and interbeds are sharp to gradational. Major production from the Triumph mine was from complex ore. Kiilsgaard (*in* Anderson and others, 1950) described an ore shoot of complex ore in the Triumph mine that was 700 ft long, 170 ft wide, and 6–50 ft thick.

Siliceous ore was described by C.W. Merriam and C.N. Bozian (written commun., 1942) as bedded siderite, galena, sphalerite, and quartz in argillite or limestone. Relative to complex ores from the Triumph mine, siliceous ore was commonly marginal in grade and had comparable zinc grades but below average gold, silver, and lead values. C.W. Merriam and C.N. Bozian (written commun., 1942) reported that the average grade of 30,000 tons of siliceous ore produced from the Triumph mine in 1942 was 6.4 percent zinc, 3.4 percent lead, 4.1 ounces of silver per ton, and 0.01 ounces of gold per ton.

FISSURE ORE

The fissure ore along the Fissure fault extended from the surface to the deepest workings below the 850-foot level in the Triumph mine and was much more extensive than the stratiform sulfide ore. Fissure ore is present as lenses and pods of sulfide-carbonate-quartz as much as 30 ft thick within gouge and crushed rock of the Fissure fault zone. Average metal grade of the fissure ore was elevated in gold relative to complex and siliceous ores; 46,500 tons of Triumph fissure ore averaged 3.3 percent zinc, 3.3 percent lead, 6.7 ounces of silver per ton, and 0.13 ounces of gold per ton, whereas 20,000 tons of fissure ore from the North Star mine averaged 9.3 percent zinc, 5.8 percent lead, 9.2 ounces of silver per ton, and 0.18 ounces of gold per ton (C.W. Merriam and C.N. Bozian, written commun., 1942).

C.W. Merriam and C.N. Bozian (written commun., 1942) reported that fissure ore contains galena, sphalerite, arsenopyrite, tetrahedrite, boulangerite, and pyrrhotite in a gangue of siderite, ankerite, and quartz. Locally coarse grained galena and other sulfide minerals are commonly massive and granular due to shearing. The fissure ore is similar to tetrahedrite-galena-siderite veins elsewhere in the Wood River area but has a higher arsenopyrite to tetrahedrite ratio (Hall, Rye, and Doe, 1978).

DEVONIAN SETTING OF STRATIFORM SULFIDE ORES

DEPOSITIONAL SETTING OF REGIONAL MILLIGEN STRATA

The internal stratigraphy of the Milligen Formation in the central part of the Milligen belt reflects deposition in an anoxic basinal environment of Devonian age dominated by hemipelagic and pelagic sedimentation (lower argillite and Triumph argillite units) and punctuated by periods of turbidite influx. A Middle to Late Devonian turbidite event coincident with local syndepositional faulting and stratiform sulfide formation at the Triumph deposit was composed of two separate distributary systems: a carbonate depocenter to the west (Lucky Coin limestone facies) that interfingers with a quartz-sand depocenter to the east (Cait quartzite facies). A younger and more widespread Late Devonian turbidite system deposited mixed siliciclastic and carbonate detritus across the Milligen basin (Independence sandstone unit).

Beds of the Lucky Coin limestone are interpreted to have been deposited by turbidity flows in an anoxic basin at a depth below wave base on the basis of graded nature of beds, lack of sediment reworking or bioturbation, and association with carbonaceous, siliceous mudrock. Thick massive micrite beds were deposited as lime-mud debris flows within a channelized deposystem; evidence for this is based on the presence of coarse mudclasts probably derived from incised channel walls and on a high sedimentation rate as indicated by the amalgamated character of beds. The anomalous thickness and predominance of thick-bedded micritic limestone suggest that the exposures in Independence Canyon were deposited near the center of this channelized system. Rhythmically interbedded limestone and argillite sequences likely represent turbidite fan or apron deposits lateral to, or distal to, the channel system. Thin-bedded amalgamated calcareous turbidite could reflect more active sedimentation in channel levee or interchannel areas.

The lack of significant coarse bioclastic material and paleocurrent indicators limits what can be deduced about the source area of the Lucky Coin limestone. The minor bioclastic component suggests that the source area was, at least in part, shallow water. A coeval Devonian carbonate platform, presently exposed several tens of miles to the east of the Triumph area in the Lost River and Lemhi Ranges, was a likely source area.

Diamictite is interpreted as resedimented Milligen derived from a nearby source. The close spatial association of diamictite with the Triumph stratiform sulfide body suggests association with a synsedimentary fault. Though not recognized, this fault is likely within the immediate area of the deposit because the diamictite does not extend far beyond the extent of the deposit. Similar relations have been recognized at other Devonian stratiform zinc-lead deposits in the Cordillera (Winn and Bailes, 1987; Turner, 1990).

The Cait quartzite reflects a sudden influx of a small volume of well-sorted quartz sand into the Milligen basin. Although the dispersal system was channelized, thin deposits of quartzite blanketed much of the Milligen basin. Sands deposited in the vicinity of the East Fork Wood River were transported in a channelized system subparallel with the axis of the present drainage. A likely source of these sands was Ordovician or Proterozoic shelf sandstone presently exposed in the Pioneer, Lost River, and Lemhi Ranges to the east.

The Independence sandstone represents progradation of a major turbidite fan complex into the Milligen basin, likely from the east. The upward-thickening sequence of upward-thinning cycles of sandstone beds in the lower part of the Independence sandstone suggests prograding channelized deposition of an inner fan environment. The finer grained nature of the Independence sandstone elsewhere in the Milligen Formation suggests deposition dominated by channel-lateral, middle fan, or outer fan environments.

Conglomerate lenses in the lowermost part of the Independence sandstone noted near the Cait property in the eastern part of the East Fork valley are well-sorted, clast-supported chert-pebble conglomerate. The well-sorted, clast-supported nature implies significant transport by traction or turbidity flow mechanism, perhaps from a subaerial source. The source for chert may have been the lower argillite or older Ordovician to Devonian chert presently exposed northeast of the Milligen belt (fig. 1). The chert conglomerate suggests uplift of an intrabasinal source. Because the conglomerate was deposited only slightly after the diamictite and stratiform sulfides, uplift was probably related to intrabasinal faulting within the Milligen basin. The coincidence of chert conglomerate deposition and the onset of siliciclastic turbidite deposition may suggest association with a common tectonic event.

EXHALATIVE ORIGIN AND DEPOSITIONAL SETTING OF STRATIFORM SULFIDE BODY

Stratiform ore was inferred by early workers to be the result of the replacement of limestone beds adjacent to a mineralizing fissure, probably during Mesozoic time (C.W. Merriam and C.N. Bozian, written commun., 1942). We agree, however, with the interpretation of Smith (1977) and Hall (1985) that the stratiform sulfide ore is of syngenetic origin and formed during the Middle to Late Devonian. The finely bedded and laminated nature of the complex ore is suggestive of a sedimentary origin. The association of diamictite with stratiform ore can be explained as detritus derived from the scarp of a syndepositional fault that was also the structural conduit for metalliferous fluids discharging on the seafloor (Winn and Bailes, 1987; Turner, 1990). White and buff chert beds at the ore horizon can be traced for more than 3 mi along strike and are interpreted as an exhalative silica facies distal to the sulfide depocenter. Such an exhalative origin for the Triumph sulfide body places it within the family of Devonian to Mississippian shale-hosted exhalative zinc-lead deposits widespread in British Columbia, Yukon, and Alaska (Turner, 1988). These deposits share a similar mineral assemblage (chert, pyrite, sphalerite, and galena with or without barite and ferroan carbonate), mineral facies (massive or banded ferroan carbonate, laminated chert-pyrite-sphalerite-galena-barite), and host lithology (carbonaceous mudrock, chert, turbidite).

The Triumph stratiform sulfide body is interbedded with carbonaceous mudrock, chert, and limestone that lack evidence for bioturbation or wave reworking. Thin tuff beds and mafic sills interbedded with Lucky Coin limestone strata near the Triumph deposit suggest that sulfide deposition coincided with minor volcanism and local emplacement of sills. Basin sedimentation during sulfide formation was dominated by hemipelagic and pelagic sediments interrupted by an influx of carbonate, mixed carbonate-siliciclastic, and quartzose sediments. Coincidence of the location of the sulfide body and a thick channel complex of limestone turbidite suggests that the sulfide body formed within a carbonate turbidite inner fan setting.

Diamictite deposits are at and near the stratigraphic level of the sulfide body; they are thickest near the sulfide body and thin to the north and south of the deposit. The similar composition of diamictite clasts and underlying strata and the poorly sorted character of the diamictite suggest local derivation.

EAST FORK GRABEN AND ASSOCIATED STRATIFORM ZINC-LEAD SULFIDE DEPOSITS

The trend formed by stratiform base-metal deposits at the Triumph mine, Cait prospect, and Mascot mine is almost east-west (fig. 1). All of these deposits are associated with diamictite or conglomerate units interpreted as locally derived detritus from eroded Milligen strata. Stratiform sulfide deposits at both the Triumph mine and Cait prospect are within interfingering Lucky Coin limestone and Cait quartzite near the contact with the overlying Triumph argillite. The stratigraphic level of the Mascot zinc-lead mineralized rock is unclear due to structural complication but may be within the Cait quartzite as well. We interpret these three stratiform deposits to have formed adjacent to an east-trending Devonian fault or fault system. The presence of the west-trending channel system of Cait quartzite just to the south of this proposed Devonian structure suggests structural control of the Cait distributary system. We refer to this structurally controlled Devonian depocenter along the present drainage of the East Fork Wood River as the "East Fork graben." Whether there is a structural margin on the south side of this East Fork graben is unclear because Milligen strata younger than the stratiform sulfide horizons (Independence sandstone) cover much of this area.

PALEOTECTONIC SETTING OF THE MILLIGEN FORMATION

Strata of the Milligen Formation are part of a north-trending belt of Paleozoic fine-grained clastic rocks and chert in central Idaho that is referred to as the Central Idaho shale belt (Hall, 1985). Near Ketchum, such strata include black shale, chert, and siltstone of Ordovician to Devonian age (fig. 1). Together with the Milligen Formation, these Paleozoic shale-rich strata have been interpreted as an outer continental margin sequence (Sandberg and others, 1975). These strata correlate with similar shale- and chert-dominant strata along the western Cordillera of North America from Yukon to Nevada (Turner and others, 1989). Based on distinctive internal stratigraphy, Turner and others suggested a common tectonic, eustatic, and depositional history from Cambrian to Late Devonian time for these outer continental margin sequences.

The Triumph deposit is part of a family of Middle Devonian to Mississippian stratiform sediment-hosted zinc-lead and stratiform barite deposits that are present in outer continental margin strata from Alaska to Sonora, Mexico. This family of syngenetic deposits is everywhere interpreted as related to local block faulting (MacIntyre, 1983; Abbott and others, 1986; Winn and Bailes, 1987; Abbott and Turner, 1990) that likely reflects an extensional, transextensional, or transcurrent regime along the entire western margin of western North America during Middle Devonian to Mississippian time (Eisbacher, 1983; Turner, 1988; Turner and others, 1989).

DEFORMATION AND METAMORPHISM OF THE TRIUMPH OREBODY

Three major deformational events are recorded in the Milligen strata (Turner and Otto, 1988). D1 structures include tight folds and axial-planar cleavage that may be correlated with the Antler orogeny based on soft-sediment-style of some F1 folds and the absence of similar structures in younger Pennsylvanian and Permian strata. D1 structures may also be of Mesozoic age. Map-scale north-trending, east-vergent folds (fig. 2) and variably developed axial-planar cleavage deform D1 structures and are inferred to be Mesozoic in age based on similarities to structures in younger Pennsylvanian and Permian strata. Northwest-trending, southwest-dipping thrust faults are likely associated with D2 folds of Mesozoic age. D3 low-angle west-dipping extensional faults having displacements to the northwest and high-angle normal faults offset strata as young as Eocene and are interpreted to be related to Eocene extension and formation of the nearby Pioneer Mountains core complex.

EARLY MISSISSIPPIAN ANTLER OROGENY (D1)

Strata of the Milligen Formation and black shale, chert, and siltstone of Ordovician to Devonian age of the Roberts Mountains allochthon have been interpreted as continental-rise deposits thrust eastward onto the continental margin of North America during the Mississippian Antler orogeny (Roberts and Thomasson, 1964; Poole and Sandberg, 1977; Dover, 1980). Strata of the Roberts Mountains allochthon structurally overlie westerly derived flysch deposits of the Mississippian Copper Basin Formation and are interpreted as synorogenic sediments (Nilsen, 1977).

The strongest evidence for such a history is in the stratigraphic record of the Copper Basin Formation. Defining the structural fabric associated with this orogenic event has been more problematic. At issue is the age of deformational fabrics within the Milligen Formation. An Antler age for cleavage in the Milligen Formation was inferred by Sandberg and others (1975) and Skipp and Hall (1975); however, Dover (1980) pointed out that, because these studies lacked detailed fabric analysis or rigorous comparison of fabrics between the Milligen and younger post-Antler stratigraphic units, a Mesozoic age could not be ruled out. Davis (1984) documented cleaved chert and argillite clasts within the Mississippian Copper Basin Formation in the Pioneer Mountains (Davis, 1984), clasts presumably derived from the deformed Antler highlands to the west (Nilsen, 1977). Turner and Otto (1988) described Milligen strata at the Cait property in which early D1 penetrative cleavage axial to isoclinal folds is refolded by upright folds that have weak axial cleavage. Based on the ductile style of these early folds, Turner and Otto inferred that this deformation took place prior to sediment lithification and suggested that the folds are of Antler age. It is uncertain whether the early foliation and folding in the Triumph mine area are related to the Antler deformation or instead are an intense development of D2 strain.

MESOZOIC FOLDING AND THRUSTING (D2)

East-vergent map-scale folds, locally developed cleavage, and a stack of older-over-younger thrust sheets in the Pioneer Mountains and areas to the east have been attributed to the Late Jurassic and Cretaceous Sevier orogeny (Skipp and Hall, 1975; Skipp and Hait, 1977; Dover, 1980). The westernmost such thrust, the Pioneer thrust fault, places Milligen Formation, as well as lower Paleozoic shale- and chert-rich strata, over Mississippian Copper Basin Formation (Skipp and Hall, 1975; Skipp and Hait, 1977; Dover, 1981). Link and Mahoney (1989) noted the lack of major thrust faults and the predominance of east-vergent folds west of the Pioneer Mountains core complex.

We correlate the northwest-trending belt of thrust faults in the area of the Triumph mine with this Mesozoic deformation. East of this thrust belt, compressive deformation is expressed as map-scale east-vergent folds that suggest greater amounts of shortening in rocks on the west side of the Milligen belt. D1 folds and cleavage in the Triumph area may be related to an early phase of Mesozoic deformation.

SILVER-LEAD-ZINC VEIN MINERALIZATION AND CONTACT METAMORPHISM

Hall, Rye, and Doe (1978) interpreted silver-lead-zinc vein mineralization in the Wood River area to have been related to intrusion of the Cretaceous Idaho batholith. The small-displacement D2 thrusts in the Triumph area probably were permeable and possibly active structures during younger Cretaceous plutonism and the site of sulfide deposition (fissure ore) within a large-scale hydrothermal system (Hall, Rye, and Doe, 1978). Contact metamorphism in the Triumph area may have been coincident with this plutonism and mineralization. Reactivation of the mineralized faults during the Tertiary sheared and dismembered veins.

Contact metamorphism in the Triumph area has been interpreted to be related to a buried intrusion below the Triumph mine (Batchelder and Hall, 1978). Granite dikes associated with hornfelsed rock below the Triumph mine are more similar in composition to the Tertiary granitic suite of the Pioneer Mountains than to quartz diorite and granodiorite intrusive rocks of Cretaceous age. For this reason, and because silver-lead-zinc vein mineralization is related to Cretaceous plutonism, granite dike formation and contact metamorphism postdate fissure mineralization at the Triumph deposit.

TERTIARY EXTENSIONAL FAULTING (D3)

Paleogene crustal extension has affected much of east-central Idaho. In the Pioneer Mountains just east of the Triumph area, metasedimentary rocks and Cretaceous and Eocene intrusive rocks are separated by the Eocene Wildhorse detachment fault from an upper plate of Paleozoic and Tertiary sedimentary and volcanic rocks (Wust, 1986; O'Neill and Pavlis, 1988). Displacement on the Wildhorse detachment fault is estimated to be at least 10.5 mi in a top-to-the-northwest direction (Wust, 1986). Widespread younger-over-older low-angle faults in the upper plate of the Wildhorse detachment throughout the Wood River area were originally interpreted as thrust faults (Hall, Batchelder, and Tschanz, 1978; Dover, 1981) but have been reinterpreted as low-angle normal faults (Link and others, 1988; Turner and Otto, 1988). Northeast-trending high-angle faults that are widespread throughout the Wood River area are parallel with the Trans-Challis fault system northwest of the Wood River area and may also have accommodated northwest-southeast extension (Link and others, 1988). Southwest-northeast Neogene extension resulted in the northwest-trending basin and range faults and the present topography.

In the Triumph area, Paleogene extension resulted in movement to the northwest along a series of shallowly dipping normal faults. These faults dismembered the Mesozoic thrust belt and displaced the updip part of the Triumph stratiform sulfide body. Reactivation and normal or oblique-slip movement on the Mesozoic thrust faults resulted in deformation and dismemberment of sulfide-carbonate-quartz veins and granite and andesite dikes within the fault zones.

REFERENCES CITED

Abbott, J.G., Gordey, S.P., and Tempelman-Kluit, D.J., 1986, Setting of sedex, sediment-hosted Pb-Zn deposits in Yukon and northeastern British Columbia, *in* Morin, J., ed., Mineral deposits of the northern Cordillera: Canadian Institute of Mining and Metallurgy Special Volume 37, p. 1–18.

Abbott, J.G., and Turner, R.J.W., 1990, Character and paleotectonic setting of Devonian stratiform sediment-hosted Zn-Pb-barite deposits, Macmillan Fold Belt, Yukon, *in* Abbott, J.G., and Turner, R.J.W., eds., Mineral deposits of the northern Canadian Cordillera, Yukon-British Columbia: International Association on the Genesis of Ore Deposits Symposium, 8th, Field Trip Guidebook, Field Trip 14, Geological Survey of Canada Open File 2169, p. 99–136.

Anderson, A.L., Kiilsgaard, T.H., and Fryklund, V.C., Jr., 1950, Detailed geology of certain areas in the Mineral Hill and Warm Springs mining districts, Blaine County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 90, 73 p.

Batchelder, J.N., and Hall, W.E., 1978, Preliminary geologic map of the Hailey 7.5-minute quadrangle, Idaho: U.S. Geological Survey Open-File Report 78–545, scale 1:24,000.

Bouma, A.H., 1962, Sedimentology of some flysch deposits: Amsterdam, Elsevier, 168 p.

Davis, S., 1984, A microstructural perspective of orogenesis in the Pioneer Mountains, central Idaho: Missoula, University of Montana, M.S. thesis, 55 p.

Dover, J.H., 1980, Status of Antler orogeny in central Idaho, clarification and constraints from the Pioneer Mountains, *in* Fouch, T.D. and Magathan, E.R., eds., Paleozoic paleogeography of the West-Central United States: Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, West-Central United States Paleogeography Symposium 1, p. 371–381.

———1981, Geology of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–A, p. 15–75.

Eisbacher, G.H., 1983, Devonian-Mississippian sinistral transcurrent faulting along the cratonic margin of western North America—A hypothesis: Geology, v. 11, p. 7–10.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

Hall, W.E., Batchelder, J.H., and Tschanz, C.M., 1978, Geologic map of the Sun Valley 7.5-minute quadrangle, Blaine County, Idaho: U.S. Geological Survey Open-File Report 78–1058, scale 1:24,000.

Hall, W.E., Rye, R.O., and Doe, B.R., 1978, Wood River mining district—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 579–592.

Link, P.K., and Mahoney, J.B., 1989, Stratigraphic setting of sediment-hosted mineralization in the eastern Hailey 1°×2° quadrangle, Blaine, Custer, and Camas Counties, south-central Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 53–69.

Link, P.K., Skipp, Betty, Hait, M.H., Jr., Janecke, Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 5–42.

MacIntyre, D.G., 1983, Geology and sedex barite-sulfide deposits of the Gataga District, northeastern British Columbia, *in* Sangster, D.F., ed., Sediment-hosted stratiform lead-zinc deposits: Mineralogical Association of Canada Short Course Handbook, v. 9, p. 85–120.

Nilsen, T.H., 1977, Paleogeography of Mississippian turbidites in south-central Idaho, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic

Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 1, p. 275–299.

O'Neill, R.L., and Pavlis, T.L., 1988, Superposition of Cenozoic extension on Mesozoic compressional structures in the Pioneer Mountains metamorphic core complex, central Idaho: Geological Society of America Bulletin, v. 100, p. 1833–1845.

Otto, B.R., 1984, Internal company report: Missoula, Montana, Noranda Exploration.

Otto, B.R., and Turner, R.J.W., 1987, Stratigraphy and structure of the Milligen Formation, Sun Valley area, Idaho: Northwest Geology, v. 16, p. 95–103.

Poole, F.G., and Sandberg, C.A., 1977, Mississippian paleogeography and tectonics of the Western United States, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 1, p. 67–85.

Roberts, R.J., and Thomasson, M.R., 1964, Comparison of late Paleozoic depositional history of northern Nevada and central Idaho: U.S. Geological Survey Professional Paper 475–D, p. D1–D6.

Sandberg, C.A., Hall, W.E., Batchelder, J.N., and Axelson, Claus, 1975, Stratigraphy, conodont dating, and paleotectonic interpretation of the type Milligen Formation (Devonian), Wood River area, Idaho: U.S. Geological Survey Journal of Research, v. 3, no. 6, p. 707–720.

Skipp, Betty, and Hait, M.H., Jr., 1977, Allochthons along the northeast margin of the Snake River Plain, Idaho: Wyoming Geological Association Guidebook, Twenty-Ninth Annual Field Conference, p. 499–515.

Skipp, Betty, and Hall, W.E., 1975, Structure and Paleozoic stratigraphy of a complex of thrust plates in the Fish Creek Reservoir area, south-central Idaho: U.S. Geological Survey Journal of Research, v. 3, no. 6, p. 671–689.

Smith, C.L., 1977, Internal company report: Westley Mines Ltd.

Turner, R.J.W., 1985, Summary report for 1985 field work, Milligen Project, Blaine County, Idaho: Private company report, Mintek Resources Ltd., Vancouver, B.C.

———1988, Massive sulphide deposits of mid-Paleozoic age: Mining Review, v. 8, no. 6, p. 43–47.

———1990, Jason stratiform Zn-Pb-barite deposit, Selwyn Basin, Yukon—Geological setting, hydrothermal facies and genesis, *in* Abbott, J.G., and Turner, R.J.W., eds., Mineral deposits of the northern Canadian Cordillera, Yukon-British Columbia: International Association on the Genesis of Ore Deposits Symposium, 8th, Field Trip Guidebook, Field Trip 14; Geological Survey of Canada Open File 2169, p. 137–176.

Turner, R.J.W., Madrid, R., and Miller, E.L., 1989, Roberts Mountains allochthon—Stratigraphic comparison with lower Paleozoic outer continental margin strata of the northern Canadian Cordillera: Geology, v. 17, p. 341–344.

Turner, R.J.W., and Michelson, C., 1984, The geological character and setting of the stratiform Ag-Zn-Pb mineralization, Triumph-Annie properties, Blaine County, Idaho: Private company report, Mintek Resources Ltd., Vancouver, B.C.

Turner, R.J.W., and Otto, B.R., 1988, Stratigraphy and structure of the Milligen Formation, Sun Valley area, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 153–167.

Turner, R.J.W., Otto, B.R., and Smith, D., 1986, Summary report, Triumph-Annie properties, Blaine County, Idaho: Private company report, Mintek Resources Ltd., Vancouver, B.C.

Winn, R.D., Jr., and Bailes, R.J., 1987, Stratiform lead-zinc sulfides, mudflows, and turbidites—Devonian sedimentation along a submarine fault scarp of extensional origin, Jason deposit, Yukon Territory, Canada: Geological Society of America Bulletin, v. 98, p. 528–539.

Wust, S.L., 1986, Extensional deformation with northwest vergence, Pioneer core complex, central Idaho: Geology, v. 14, p. 712–714.

Structural Setting of Ore Deposits in the Lake Creek Mineralized Area, Blaine County, South-Central Idaho

By Bradford R. Burton *and* Paul Karl Link

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–F

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

Structural Setting of Ore Deposits in the Lake Creek Mineralized Area, Blaine County, South-Central Idaho

By Bradford R. Burton[1] *and* Paul Karl Link[2]

ABSTRACT

The Lake Creek mineralized area is on the west slope of the Boulder Mountains, northwest of Ketchum, Blaine County, Idaho. In this area silver-lead-zinc mineral deposits are present in fine-grained calcareous rocks of the Middle Pennsylvanian to Lower Permian Wood River Formation. The source of the metals in the Lake Creek mineralized area is thought to be syngenetic ore deposits in the unconformably underlying Milligen Formation.

Large-scale northeast-vergent overturned folds in both the Wood River and Milligen Formations probably formed during the late Mesozoic Sevier orogeny. In Paleogene time Mesozoic folds were cut by gently west dipping dextral-normal faults that show N. 60° W. mean transport direction. The previously named Wood River thrust fault crops out within the Lake Creek mineralized area. We reinterpret the Wood River thrust fault to be one of several faults in a southwest-dipping imbricate stack of Paleogene extensional faults in the area and rename it the Trail Creek fault. These extensional faults are interpreted to result from crustal thinning in the upper plate of the Pioneer Mountains metamorphic core complex.

Mineral deposits in the Lake Creek area are in two geologic settings. Higher grade ore concentrations are in vein-filled fissures and replacement deposits along shear zones in the hanging wall of the Lake Creek fault. Lower grade deposits are not related to the Lake Creek fault but are in bedding-parallel brecciated pods above the locally sheared unconformity between the Milligen and Wood River Formations. The dominant ore minerals are zinc and lead carbonate minerals, sphalerite, galena, and chalcopyrite. Gangue minerals include quartz, calcite, hematite, and pyrite. Hydrothermal activity associated with magmatism of the Eocene Challis Volcanic Group was probably responsible for the remobilization and concentration of metals. A new exploration model, based on the recognition of mineralized rock in the hanging wall of low-angle faults, can be applied to an extensive area of the Boulder and Pioneer Mountains.

[1]Department of Geology and Geophysics, University of Wyoming, Laramie, Wyoming 82071.

[2]U.S. Geological Survey and Department of Geology, Idaho State University, Pocatello, Idaho 83209.

INTRODUCTION

The Lake Creek mineralized area is about 16 km northwest of Ketchum, Blaine County, Idaho, on the west flank of the Boulder Mountains (fig. 1). Five inactive or abandoned mines and numerous prospects are in the Lake Creek, Eagle Creek, and Trail Creek drainage basins. The most extensive workings are in the Lake Creek drainage and include the Homestake, Long Grade, High Grade, Lake Creek, and Price Group properties (figs. 1, 2). The area is accessible from gravel roads along Lake, Eagle, and Trail Creeks. Steep northeast-trending valleys and intervening tree-covered ridges characterize the topography; elevations range from 2,182 to 2,961 m.

Ore deposits of Lake Creek mineralized area were described by Umpleby and others (1930), Service and Kellum (1956), and Tuchek and Ridenour (1981). The geology was described in general by Dover (1981) and Link and others (1988) and in detail by Burton (1988). The stratigraphy of the Wood River Formation in the Lake Creek mineralized area was described by Hall and others (1974, 1978), Burton (1988), and Mahoney and others (1991). The Milligen Formation in neighboring areas was described by Sandberg and others (1975), Otto and Turner (1987), Turner and Otto (1988, this volume), and Ratchford (1989, this volume) but has not been studied in detail in the Lake Creek mineralized area. Detailed geologic mapping of the Lake Creek mineralized area was conducted by W.E. Hall (unpublished U.S. Geological Survey maps), C.M. Tschanz (unpublished U.S. Geological Survey map), and Burton (1988). Recently completed, unpublished geologic mapping (B.R. Burton, P.K. Link, D.W. Rodgers, and A.D. Huerta) and stratigraphic studies (Burton, 1988) enable us to improve on previously published descriptions of the occurrence, structural setting, and host-rock characteristics of these mineral deposits. We

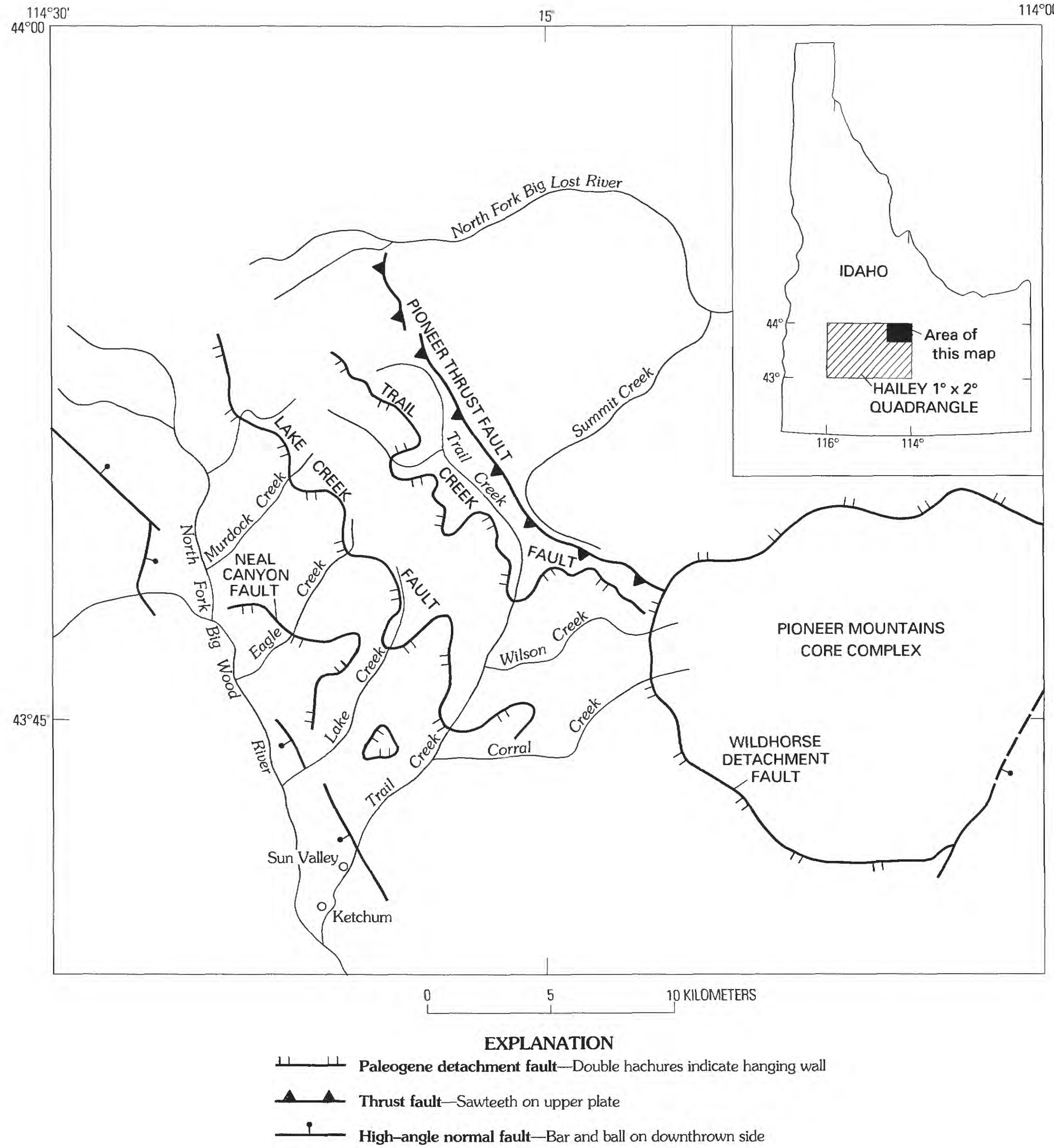


Figure 1. Map showing major structural features of the Lake Creek mineralized area, Blaine County, Idaho.

also present data that affect regional structural models (see Rodgers and others, this volume).

The Lake Creek mineralized area is between the Idaho batholith on the west and the Pioneer Mountains metamorphic core complex on the east (fig. 1). The Lake Creek mineralized area is in the western part of the Sevier orogenic belt; 8 km to the northeast the Pioneer thrust fault is mapped above a major frontal ramp (Rodgers and Janecke, 1992). The Pioneer thrust fault juxtaposes Silurian and Ordovician rocks over Mississippian rocks. Paleozoic rocks in the Lake Creek mineralized area were transported to the east above this major thrust fault during the Late Cretaceous Sevier orogeny.

Paleozoic rocks of the central Idaho black-shale mineral belt (Hall, 1985) are the principal formations exposed in the Lake Creek mineralized area. These include carbonaceous argillite, siltite, and sandstone of the Milligen Formation (Devonian) and fine-grained mixed carbonate and siliciclastic strata of the Wood River Formation (Middle Pennsylvanian to Lower Permian). The Late Cretaceous Idaho batholith, to the west of the Lake Creek mineralized area, consists of numerous biotite and hornblende-biotite granitic plutons that have yielded radiometric ages of 95–70 Ma (Lewis and others, 1987; Kiilsgaard and others, in press; Lewis, in press). Hydrothermal activity that accompanied the Late Cretaceous magmatism is thought to have remobilized metals from Paleozoic syngenetic ores to produce most of the silver-lead-zinc deposits of the Wood River valley and Smoky Mountains (Hall and others, 1978; Hall, 1985; Darling and others, this volume; Sanford and Wooden, this volume; Park, in press). Recent work (Worl and Johnson, this volume) suggests, however, that Eocene hydrothermal activity associated with the Challis volcanic episode is responsible for mineralization in the Lake Creek mineralized area and other areas.

The Pioneer Mountains core complex (Dover, 1983; Kim, 1986; Wust, 1986; O'Neill and Pavlis, 1988; Wust and Link, 1988; Silverberg, 1990), east of the Lake Creek mineralized area, formed in response to Paleogene regional extension. In the core complex, a major fault (the Wildhorse detachment fault) separates high-grade, ductilely deformed rocks in the footwall from low-grade metamorphic and unmetamorphosed rocks in the hanging wall. Radiometric age determinations (Silverberg, 1990) suggest that the infrastructure (footwall) of the core complex was subject to rapid erosion or active tectonic unroofing at a rate of 1.5 km per million years from 48 to 45 Ma. Volcanic rocks of the Eocene Challis Volcanic Group, also related to Paleogene extension, crop out immediately south of and within 5 km to the north of the Lake Creek mineralized area. Ductile deformation along the Wildhorse detachment system persisted to Oligocene time (36–33 Ma) (Silverberg, 1990).

Acknowledgments.—Research and mapping presented in this report was conducted for the U.S. Geological Survey Hailey 1°×2° quadrangle CUSMAP (Conterminous United States Mineral Assessment Project) under the direction of R.G. Worl. Part of this study was conducted as a Master of Science thesis at Idaho State University by Burton under the direction of Link and D.W. Rodgers. This report was partially funded by Idaho State Board of Education Grant 89–56 to P.K. Link. We acknowledge useful reviews by S. Soulliere, R. Tysdal, and D.W. Rodgers.

GEOLOGY OF THE LAKE CREEK MINERALIZED AREA

STRATIGRAPHY

Paleozoic rocks of the Milligen and Wood River Formations in the Lake Creek mineralized area are dominantly fine grained, carbonaceous sedimentary rocks that were deposited in deep-water environments. The Milligen Formation consists of dark-gray to black subphyllitic carbonaceous argillite and argillitic chert and subordinate interbeds of fine-grained quartzite and dolomitic siltstone. Due to structural complexity, the thickness of the Milligen Formation is not known but is estimated to be at least 1,200 m (Sandberg and others, 1975).

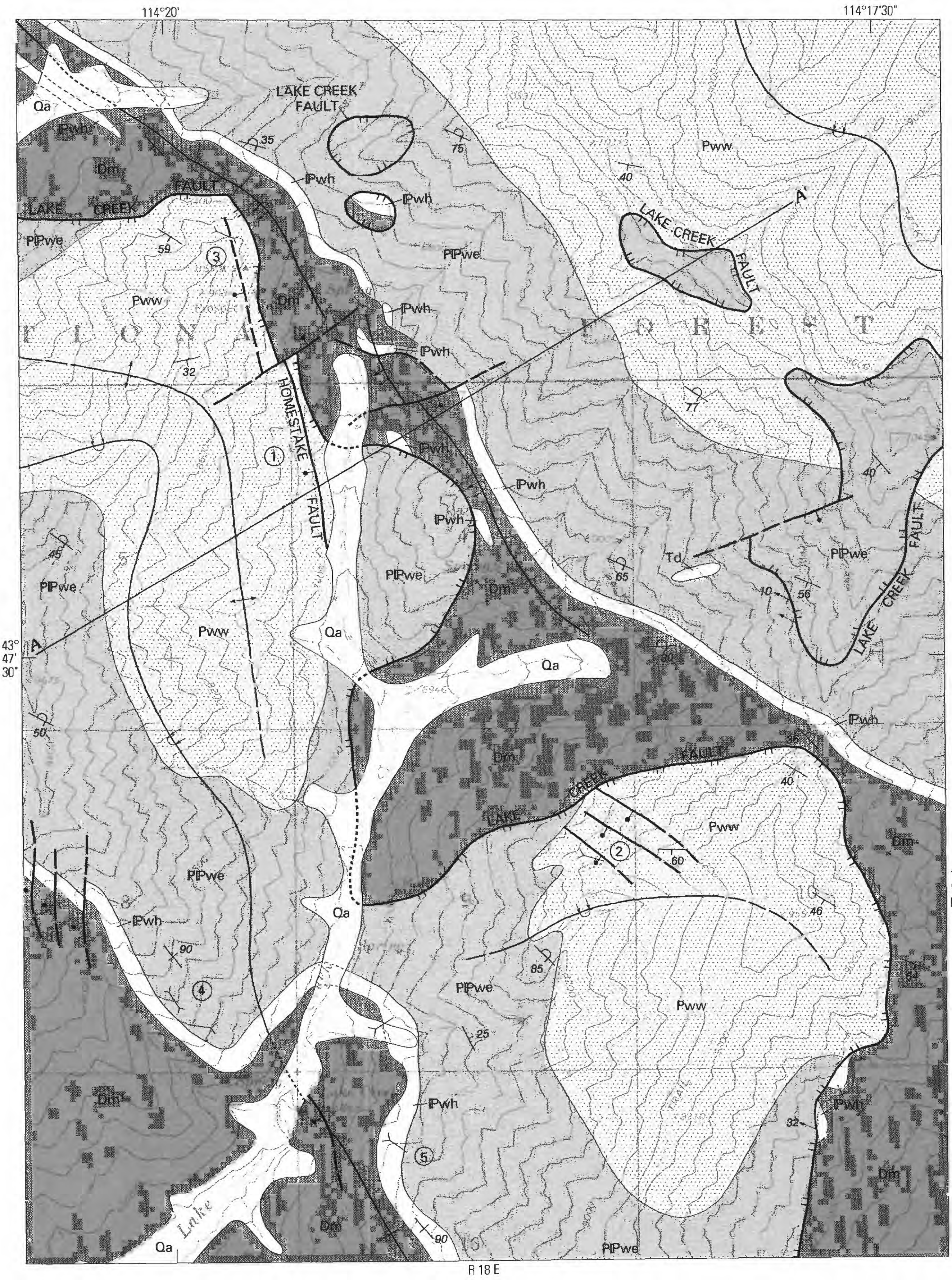
The Milligen Formation is overlain by the Wood River Formation of the Middle Pennsylvanian to Lower Permian Sun Valley Group (Mahoney and others, 1991). An angular unconformity crosscuts both bedding and a pre-existing structural fabric of the Milligen Formation and separates the Milligen from the Sun Valley Group (fig. 3). The unconformity is marked by a paleoregolith composed of angular clasts derived from the Milligen Formation, identified by petrographic comparison of lithology and microstructures of the Milligen Formation and clasts of the paleoregolith (Mahoney and others, 1991). In other areas of the Wood River valley this contact has been sheared and injected with quartz veins and was previously mapped as the Wood River thrust fault (Hall and others, 1974; Dover, 1981, 1983; Hall, 1985).

Above the unconformity the Wood River Formation is composed of conglomerate, fossiliferous limestone, silty micrite, micritic sandstone, siliceous sandstone, and cherty micritic siltstone (fig. 3). The Wood River Formation is at least 2,200 m thick; the uppermost part of the formation is truncated by the present-day erosional surface.

The Milligen and Wood River Formations are cut by dikes and sills related to the Eocene Challis Volcanic Group (Stewart and others, in press). Contact metamorphism associated with this intrusive activity is limited. In scattered, isolated areas within several kilometers of the Lake Creek mineralized area, Paleozoic sedimentary rocks are unconformably overlain by boulder and cobble conglomerate that correlates with the conglomerate of Smiley Creek (Paull, 1974), which contains Eocene palynomorphs (Burton and Blakley, 1988; E. Davies, Branta Biostratigraphic Ltd., Calgary, Alberta, written commun., 1989). The conglomerate of Smiley Creek grades upward into volcaniclastic conglomerate and volcanic rocks of the Challis Volcanic Group (51–40 Ma) (McIntyre and others, 1982; Ekren, 1985; Moye and others, in press). Quaternary alluvial sediments that include Pleistocene glacial deposits of the Boulder Creek and Prairie Creek advances (Pearce and others, 1988) are present in the valley floors and headwaters of the Trail, Lake, and Eagle Creek drainages. Late Quaternary landslide deposits and landslide-dammed lacustrine deposits are also present.

STRUCTURAL GEOLOGY

The Lake Creek mineralized area has undergone four episodes of deformation, including two episodes of compressional deformation (D1 and D2) and two episodes of extensional deformation (D3 and D4). Evidence for these multiple episodes of deformation is discussed following.



114°20'
114°17'30"
43° 47' 30"
LAKE CREEK FAULT
HOMESTAKE FAULT
Qa
Dm
Pww
PIPwe
IPwh
Td
A
A'
T 5 N
R 18 E

EXPLANATION

Qa **Alluvium (Quaternary)**

Td **Dacitic dike (Paleogene)**

Pww **Wilson Creek Member of the Wood River Formation (Lower Permian)**

PIPwe **Eagle Creek Member of the Wood River Formation (Lower Permian to Middle Pennsylvanian)**

IPwh **Hailey Member of the Wood River Formation (Middle Pennsylvanian)**

Dm **Milligen Formation (Devonian)**

Contact—Dashed where approximately located; dotted where concealed

32 **Paleogene detachment fault**—Showing trend and plunge of fault striae. Double hachures show hanging wall. Dashed where approximately located; dotted where concealed

High-angle normal fault—Ball and bar on downthrown side. Dashed where approximately located; dotted where concealed

Anticline—Dashed where approximately located

Overturned anticline—Showing direction of dip of limbs. Dashed where approximately located; dotted where concealed

Syncline—Dashed where approximately located

Overturned syncline—Showing direction of dip of limbs. Dashed where approximately located; dotted where concealed

46 **Strike and dip of bedding**—Showing amount of dip

77 **Strike and dip of overturned bedding**—Showing amount of dip

Adit

Trench

4 **Mine or prospect**—Shown by circled number on map
1. Homestake mine
2. Long Grade prospect
3. High Grade prospect
4. Deposits of the Lake Creek group
5. Deposits of the Price group

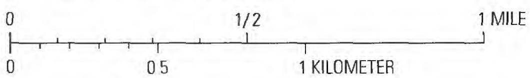


Figure 2 (above and facing page). Map showing geology of the Lake Creek mineralized area, Blaine County, Idaho. Line of section A–A′ (fig. 6) is also shown. Modified from Burton (1988). Base is U.S. Geological Survey Rock Roll Canyon 7.5-minute topographic quadrangle (1967).

D1 DEFORMATION

The first phase of compressional deformation (D1) is recognized only in the Milligen Formation. D1 features include (1) a penetrative slaty cleavage (S1) in pelitic rocks and a spaced cleavage in coarser grained rocks, (2) small-scale (10 cm to several meters amplitude) isoclinal folds (F1), and (3) north-directed folding and tilting about a west-northwest-oriented fold axis. These features are observed throughout the Milligen Formation in the Lake Creek mineralized area. In the surrounding region, similar features are observed in the Milligen Formation (Turner and Otto, 1988, this volume) but are not present in all areas. The regional distribution of D1 features in the Milligen Formation may be related to proximity to a thrust fault formed during the Early Mississippian Antler orogeny but is not well understood.

The tilted and folded rocks of the Devonian Milligen Formation were beveled by pre-Middle Pennsylvanian erosion, marked by an unconformity below the Wood River Formation. In several locations, notably just below the Long Grade mines and at the outcrop of the Hailey Member of the Wood River Formation on the central western edge of the map area (SE¼NW¼ sec. 8, T. 5 N., R. 18 E.) (fig. 2), the unconformity crosscuts S1 cleavage and small-scale F1 folds. No strata in the Wood River Formation exhibit slaty S1 cleavage nor does the Wood River Formation contain small-scale F1 folds. These relationships constrain D1 deformation to pre-Middle Pennsylvanian time. We infer that D1 deformation was produced by the Early Mississippian Antler orogeny (Roberts and others, 1958), which is the only large-scale compressive tectonic event known to have affected this area between Late Devonian and Middle Pennsylvanian time (Dover, 1980; Skipp and Hall, 1980; Turner and Otto, this volume).

Equal-area projections of poles-to-bedding in the Milligen Formation are shown in figure 4. These data were measured in Devonian strata preserved in the upright lower limb of a map-scale F2 fold in the Lake Creek mineralized area. Considerable scatter is evident in the data, the source of which may be that reliable facing-direction indicators are seldom present in the Milligen Formation and therefore some data from near the hinge of the F2 fold may be for overturned beds. The data show a point maximum 52°, N. 56° E. (labeled **D** on fig. 4*D*), which lies along a best-fit great circle oriented N. 24° E., 68° SE. To determine the orientation of bedding in the Milligen Formation subsequent to D1 deformation but prior to D2 deformation, the point maximum and F1 fold axis were rotated to restore bedding in the Wood River Formation to horizontal (fig. 4*E*). The plot shows that D1 deformation resulted in folding of the Milligen Formation about an axis 2°, N. 84° W., and northward tilting of Milligen Formation strata. D1 deformation is not coaxial with D2 deformation.

D2 DEFORMATION

The second deformational event (D2) produced map-scale (hundreds of meters to 1,000 m amplitude) east-vergent asymmetric folds (F2). D2 deformation affects both the Wood River and Milligen Formations and is recognized throughout the eastern part of the Hailey 1°×2° quadrangle.

F2 folds have open, concentric synclines and tight, chevron or kink anticlines (figs. 2, 5, 6). Bedding dips steeply to the southwest in both limbs of these folds and is overturned in the lower limbs of anticlines and the upper limbs of synclines. Equal-area projection of poles-to-bedding in the Wood River Formation define a F2 fold axis of 10°, S. 47° E., implying that the shortening direction during F2 deformation was oriented approximately N. 43° E. (fig. 4).

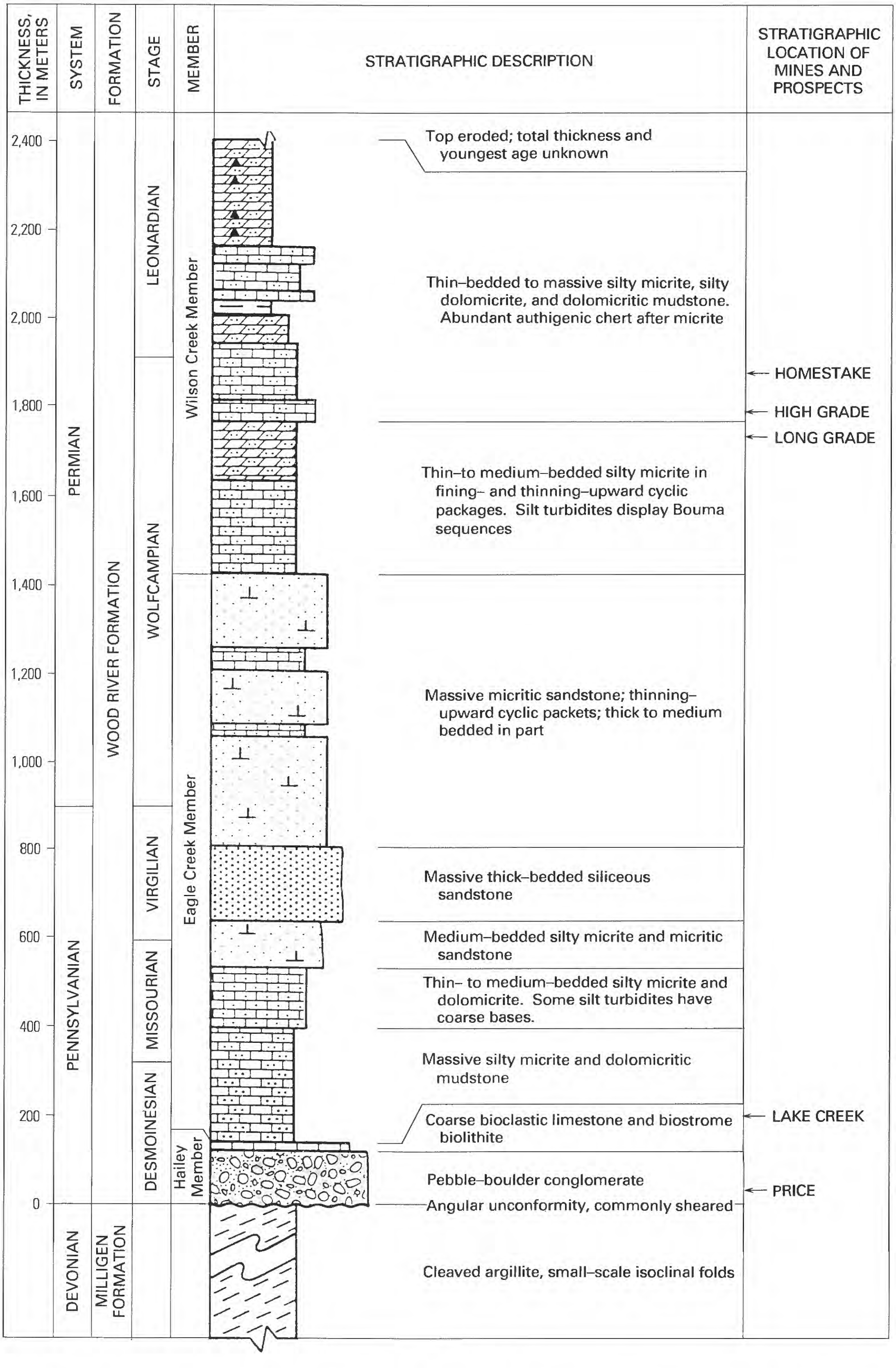


Figure 3. Stratigraphy of the Wood River Formation. The stratigraphic locations of mines and prospects discussed in the text are also shown. Modified from Mahoney and others (1991).

The tight, overturned anticlines are interpreted to be fault propagation folds cored by small-displacement splay thrusts above a basal detachment (fig. 6). These faults are generally not exposed at the surface in the Lake Creek mineralized area, but at least one small-offset thrust fault was mapped along an anticlinal trace in Murdock Creek, 2 km north of the headwaters of Eagle Creek beyond the northwestern corner of the mineralized area.

Numerous faults were previously mapped as thrust faults by other workers in Lake Creek mineralized area (Dover, 1983; W.E. Hall, unpublished mapping), but we reinterpret these faults as oblique-slip normal faults associated with D3 deformation (see later discussion). We identified no major thrust faults in the Lake Creek mineralized area. Instead, shortening was here accommodated dominantly by folding. Depth-to-detachment calculations, using the line-length area-balance method of Woodward and others (1985), indicate that a basal detachment fault in the lower part of the Milligen Formation (about 625 m above sea level) is necessary to explain the amount of shortening observed in the megascopically folded Wood River Formation.

We infer that D2 deformation in the Lake Creek mineralized area was produced by the Sevier orogeny (Armstrong, 1975) during Late Cretaceous time. We found no direct evidence for the age of D2 deformation in the Lake Creek mineralized area but inferred the age from evidence in adjacent areas. In the Smoky Mountains, on the eastern margin of the Idaho batholith, approximately 40 km southwest of the Lake Creek mineralized area, D2 deformation is broadly synchronous with intrusion of the Idaho batholith and metamorphism at 83.9±3.4 Ma (Whitman, 1990, this volume). In the Mahoney Butte quadrangle, 25 km southwest of the Lake Creek mineralized area (Skipp and others, 1994), the Deer Creek thrust fault was folded and then intruded by the Deer Creek stock about 90 Ma. Thus, regionally recognized D2 deformation in the area of the Lake Creek mineralized area must have occurred prior to and partly synchronous with emplacement of the Idaho batholith 90–85 Ma.

D3 DEFORMATION

The third phase of deformation in the Lake Creek mineralized area (D3) consists of extensional deformation that produced gently dipping corrugated dextral-normal faults with top-to-the-northwest displacement. Smaller displacement, synthetic, steeply dipping normal faults in the hanging wall of the larger scale, gently dipping faults are interpreted to flatten into the underlying detachment faults. D3 deformation affects Paleozoic and Mesozoic rocks and crosscuts D2 features (fig. 5).

Northwest-striking oblique-slip normal faults that have shallow (10°–30°) southwest dip crosscut the F2 folds. These faults generally place younger rocks of the Eagle Creek and Wilson Creek Members of the Wood River Formation on older strata of the Hailey Member and Milligen Formation, although, locally, older-on-younger relations are present (figs. 2, 6). Slip indicators on the faults include fault striae or slickenlines and chatter marks, and they show that last movement of the hanging wall was toward the northwest (fig. 2) (Huerta, 1992). The faults display corrugated or curviplanar geometry and thus vary locally from dip-slip normal faults to strike-slip faults. They are related to Paleogene extension in the upper plate of the Pioneer Mountains core complex (Wust, 1986; Wust and Link, 1988; Burton and others, 1989; Batatian, 1991; Huerta, 1992; Rodgers and others, this volume). These faults were mapped by previous workers (Umpleby and others, 1930; Dover, 1981, 1983) as thrust faults.

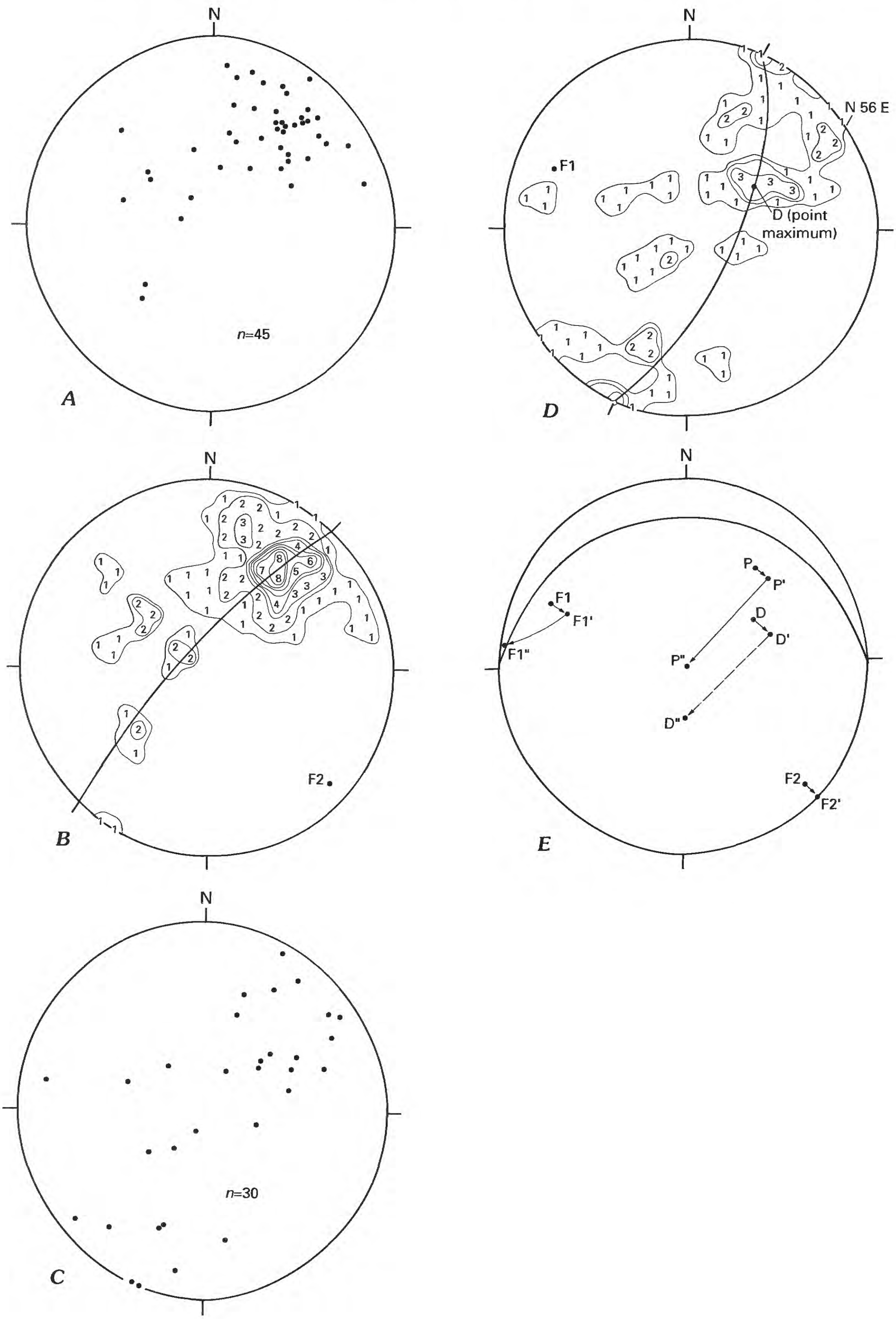
Within the hanging wall of the shallowly dipping D3 detachment faults are northwest-striking, moderately to steeply southwest dipping normal faults. These faults accommodated only tens of meters of displacement and, because their separation can only be mapped in single detachment-fault plates, are interpreted to sole into the underlying detachment faults. We suggest that they formed during movement on the underlying detachment in response to dextral-normal displacement above the curviplanar detachment fault surface. These minor structures are important because they host several of the mineral deposits in the Lake Creek mineralized area. The Lake Creek fault (figs. 1, 2) is a D3 structure that hosts the most significant mineral deposits of the Lake Creek mineralized area.

D4 DEFORMATION

The last deformational event recognized in the Lake Creek mineralized area (D4) is manifested by steeply dipping dip-slip normal faults that strike north to northeast and accommodated down-to-the-west displacement. These faults cut Paleozoic, Mesozoic, and Paleogene rocks and include the Sun Valley fault (fig. 1). Northerly striking, steeply west dipping dip-slip normal faults (D4) that accommodated tens to possibly hundreds of meters of down-to-the-west displacement have been mapped to the south of Lake Creek mineralized area (Hall and others, 1978; Burton, 1988). The northwest-striking D4 faults are a product of Neogene crustal extension of the Basin and Range structural province of the western United States (Link and others, 1988; Rodgers and others, this volume). D4 features are not associated with economic mineral deposits of the Lake Creek mineralized area.

REINTERPRETATION OF THE WOOD RIVER THRUST FAULT

In the area north and east of the Lake Creek mineralized area, in the Boulder and Pioneer Mountains, Dover (1983) mapped two major structures, the Wood River thrust fault


N
n=45
A
N
F1
N 56 E
D (point maximum)
D
N
F2
B
N
P
P'
F1
F1'
D
D'
F1"
P"
D"
F2
F2'
E
N
n=30
C

and the Pioneer thrust system. Subsequent work, summarized here and in Rodgers and others (this volume), reinterprets the Wood River thrust fault as a normal fault associated with Paleogene crustal extension and early stages of formation of the Pioneer Mountains core complex. Part of the Pioneer thrust system that lies south of lat 43°47′30″ N. was similarly reinterpreted and was renamed the Wildhorse detachment (Wust, 1986). North of this latitude, the Pioneer thrust fault juxtaposed Ordovician rocks over Mississippian rocks and is recognized as a major thrust fault (Dover, 1981, 1983; Worl and others, 1991; Rodgers and Janecke, 1992).

The Wood River thrust fault as defined by Dover (1981, 1983) included all locations where Wood River Formation lies directly stratigraphically or structurally above the Milligen Formation. Dover, following Hall and others (1978), applied the term "Wood River thrust" to areas where the unconformable contact of the Milligen and Wood River Formations is sheared due to flexural slip associated with F2 folding. This contact is exposed in many locations in the Boulder and Pioneer Mountains (Worl and others, 1991). The contact demonstrates several geologic relations: (1) paleoregolith of Milligen Formation below conglomerate of the Hailey Member with no evidence of shearing, (2) a shear zone filled with quartz veins below the Hailey Member, as in the High Grade and Lake Creek mines (fig. 2), and (3) sheared and attenuated Hailey and lower part of the Eagle Creek Members above the Milligen Formation. The contact does not contain evidence of structural thickening, repeated strata, or older-on-younger relationships associated with thrust faulting. Our work suggests that the contact is not a regional thrust fault but rather an unconformity that experienced varying amounts of flexural slip between rocks of different competence.

Figure 4 (facing page). Equal-area, lower hemisphere projections of poles-to-bedding in the Milligen and Wood River Formations in the Lake Creek mineralized area, Blaine County, Idaho. A, Poles-to-bedding in the Wood River Formation (data are from the upright limb of an F2 fold). B, Contour plot of data shown in A, showing best-fit great circle N. 43° E., 80° NW., which describes a fold axis, F2, oriented 10°, S. 47° E. C, Poles-to-bedding in the Milligen Formation. D, Contour plot of data shown in C, showing best-fit great circle N. 24° E., 68° SE., which intersects the point maximum at 52°, N. 56° E., and describes a fold axis, F1, oriented 22°, N. 66° W. E, Plot showing rotation of F1 fold axis to remove the effects of D2 deformation. Points shown are Wood River Formation (P) and Milligen Formation (D) poles-to-bedding point maxima and F1 and F2 fold axes. The F2 fold axis was rotated to horizontal (solid arrows) along with the F1 fold axis and point maxima to yield F2′, F1′, P′, D′, then Wood River Formation bedding was restored to horizontal (dashed arrows). The resulting orientation of F1″ is 2°, N. 84° W., and of D″ is 68°, S. 3° E.

TRAIL CREEK FAULT

We interpret the Wood River thrust as mapped by Dover (1981, 1983) on the northeast side of the Lake Creek mineralized area to be the lowest exposed, shallowly dipping dextral-normal fault in the hanging wall of the Wildhorse detachment fault, and we rename it the Trail Creek fault (figs. 1, 6). Above the Trail Creek fault, we identify and name the Lake Creek fault and the Neal Canyon fault as other, structurally higher faults that are also associated with the Wildhorse detachment fault (fig. 1).

The recognition that the Trail Creek fault is a D3 transtensional feature is based on the following criteria: (1) the Trail Creek fault juxtaposes younger rocks in the hanging wall against older rocks in the footwall, (2) displacement on the Trail Creek fault is not coaxial with D2 shortening (F2 folds in both the hanging wall and footwall display a N. 43° E. vergence direction associated with Sevier orogenic compression, but striae on the Trail Creek fault surface trend N. 50°–65° W., indicating fault displacement toward the northwest), and (3) the Trail Creek fault cuts across earlier formed F2 folds and must be a product of younger tectonism.

The cross sections of Umpleby and others (1930) and Dover (1981, 1983) also show that the Trail Creek fault truncates F2 folds, but the interpretation by Umpleby and others and by Dover of this relationship was that both the folds and the thrust fault were produced by Sevier orogenic compression. This interpretation would require that the Wood River thrust fault be a late stage, out-of-sequence fault that cut previously folded rocks. Kinematic data presented in this report and by Burton and others (1989) and Huerta (1992) favor the simpler interpretation that the Trail Creek fault is a younger transtensional feature.

LAKE CREEK FAULT

At the headwaters of Lake Creek (fig. 2), a northwest-striking fault, here named the Lake Creek fault, dips 30° SW. and places F2 folded Eagle Creek and Wilson Creek Members of the Wood River Formation on F2 folded rocks of the subjacent Milligen and Wood River Formations (fig. 6). In most places, this results in a younger-over-older relationship. Klippen of the Lake Creek fault are on the top of the ridge east of the Homestake mine and in several places on the southwest slope of the Boulder Mountains. In these excellent exposures of the fault surface, the Lake Creek fault has consistent N. 50°–65° W. kinematic indicators (fault striae or slickenlines on the highly polished fault surface) indicating last movement to the northwest. The fault has a curviplanar geometry including long-wavelength, low-amplitude

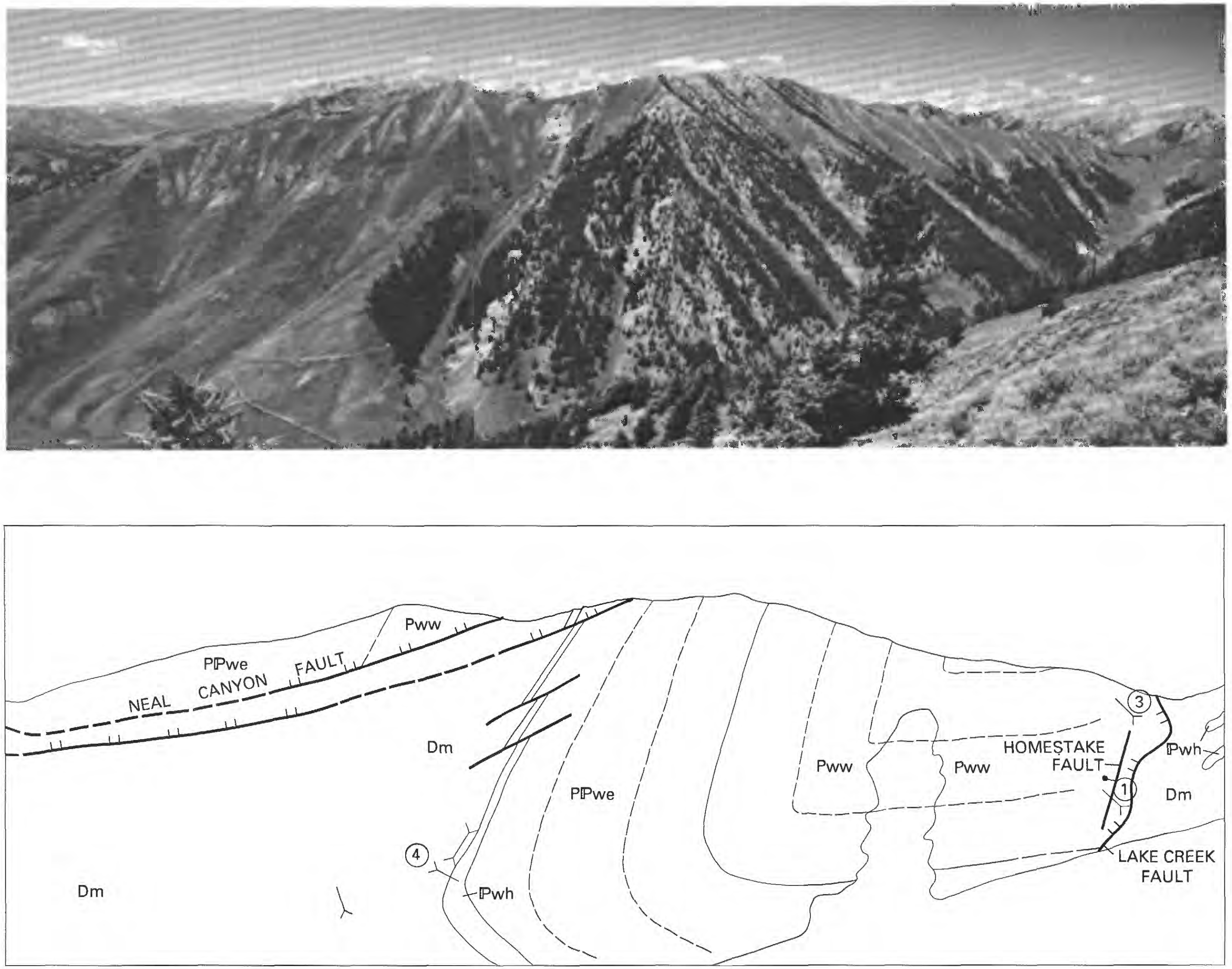


Figure 5. Photomosaic of a view to the northwest of the ridge between Lake and Eagle Creeks, Lake Creek mineralized area, Blaine County, Idaho. The location of Paleogene extensional faults associated with mineralization of the area and the location of the Homestake mine (circled number 1), High Grade prospect (circled number 3), and the Lake Creek group of deposits (circled number 4) are shown on the photo-overlay. Stratigraphic units are the Wilson Creek Member of the Wood River Formation (Pww), the Eagle Creek Member of the Wood River formation (PIPwe), the Hailey Member of the Wood River Formation (IPwh), and the Milligen Formation (Dm).

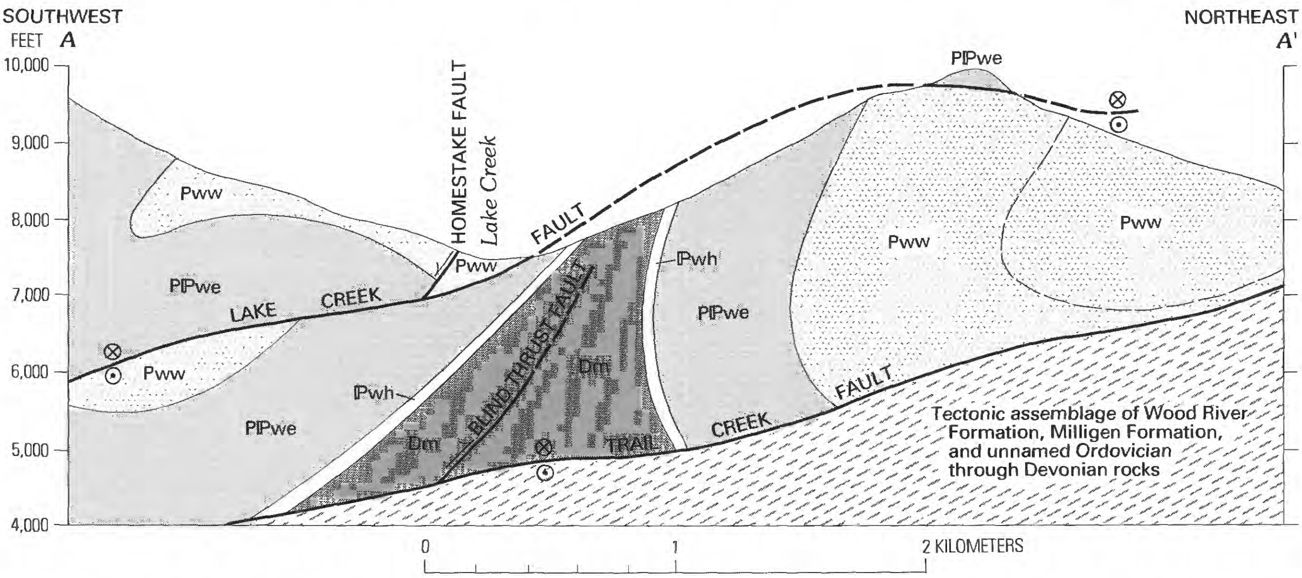


Figure 6. Cross section showing geology of the Lake Creek mineralized area, Blaine County, Idaho. Line of section is shown in figure 2. The Homestake mine is in a shear zone of the Homestake fault, a high-angle normal fault synthetic to the Lake Creek fault. Relative movement on the Lake Creek fault was dextral-normal slip with hanging wall toward N. 60° W. The fault displays a curviplanar geometry with axes of corrugations parallel with the movement direction. Circled x indicates movement direction is into plane of cross section; circled period indicates movement direction is out of plane of cross section. Tight anticlines in the Devonian Milligen Formation and the Pennsylvanian to Lower Permian Wood River Formation are thought to be thrust fault propagation folds. Stratigraphic units are the Wilson Creek Member of the Wood River Formation (Pww), the Eagle Creek Member of the Wood River Formation (PIPwe), the Hailey Member of the Wood River Formation (IPwh), and the Milligen Formation (Dm).

sinusoidal undulations. The axes of these corrugations are approximately parallel with the measured northwest-directed displacement on the fault. The fault cuts both up- and down-section through F2 folds in both the hanging wall and footwall. The Lake Creek fault is shown on the cross sections of Dover (1981, 1983) as a folded thrust fault.

Lead-isotope data (Sanford and Wooden, this volume) show that galena collected from the Homestake mine contains Triumph-type lead, similar to syngenetic galena from the Devonian Milligen Formation at the Triumph mine 5 km south of the Lake Creek mineralized area. Tightly folded and cleaved gray argillite of the Milligen Formation in the footwall of the Lake Creek fault is suspected to be the major source of the ore metals in the Lake Creek mineralized area.

ECONOMIC GEOLOGY

HOST-ROCK CHARACTERISTICS

Host rocks in the Homestake mine are carbonaceous silty micrite and calcareous mudstone of the Eagle Creek and Wilson Creek Members of the Wood River Formation (fig. 3). The largest deposits replace micritic wallrock adjacent to structurally controlled fissure veins. Silty micritic limestone wallrock has been replaced by ferruginous, porous zones that host ore deposits within a quartz matrix. Noncalcareous sandstone and quartzite are generally not mineralized.

The host rocks in the Lake Creek and Price Group prospects are conglomerate and limestone of the Hailey Member of the Wood River Formation (fig. 3). Here, replacement-type mineral deposits are within a few tens of meters of the underlying Devonian Milligen Formation.

ALTERATION

Alteration of the host rock consists of (1) oxidation of iron phases to a brick-red gossan that is present in a thin zone, notably around the Lake Creek and Price Group deposits, (2) oxidation of primary galena to cerussite, and (3) oxidation of primary sphalerite to smithsonite and hemimorphite. Breccia and gouge in the mineralized fissures are bleached and have wispy, discontinuous hematite staining.

GEOLOGIC SETTING OF MINERAL DEPOSITS

Mineral deposits in the area are in at least two geologic settings: (1) steeply southwest dipping normal faults host vein-filled fissures or replacement deposits (Homestake mine and Long Grade and High Grade prospects), and (2) the sheared unconformity between the Milligen Formation and overlying Hailey Member of the Wood River Formation hosts pods of mineralized rock (deposits of the Lake Creek and Price groups).

HOMESTAKE MINE AND LONG GRADE AND HIGH GRADE PROSPECTS

The Homestake mine is near the headwaters of the west fork of Lake Creek (figs. 2, 5). The mine is in sections 3, 4, and 5, T. 5 N., R. 18 E., and in sections 32 and 33, T. 6 N., R. 18 E., Warm Springs Creek mining district, Blaine County, Idaho. Carbonaceous argillite and silty micrite of the Wilson Creek Member of the Wood River Formation host mineral deposits along a major northwest-striking (N. 26° W.) shear zone (the Homestake fault, figs. 2, 6) that dips 45°–75° SW. (Tuchek and Ridenour, 1981, p. 223). The shear zone can be traced on the surface for approximately 1.6 km and can be inferred from discontinuous exposure for more than 4.8 km. The shear zone accommodated S. 70° W. extension and is synthetic to the underlying Lake Creek fault, which forms a dip slope north of the mine in the northeast headwaters of Lake Creek (figs. 2, 6).

The ore minerals formed by oxidation of primary galena and sphalerite. Altered remnants of galena are in the wallrock. The dominant ore minerals in veins and replaced limestone are zinc and lead carbonates and silicates, mainly smithsonite, hemimorphite, and cerussite. Silver is present as trace amounts in primary galena and sphalerite. On the margins of mineralized veins, the sheared, altered, brecciated, and veined country rock contains quartz, calcite, and hematite. Ore shoots are from 7 cm to 2 m in thickness and from 3 to more than 30 m in length. Production records from 1923 to 1961 indicate that the average ore grade from filled-fissure and replacement deposits was 6.5 ounces per ton silver, 4.5 percent lead, and 9.5 percent zinc (Tuchek and Ridenour, 1981). The mine today is mainly caved and inaccessible.

At the Homestake mine, ore is preferentially in carbonaceous silty micrite and calcareous mudstone of the Eagle Creek and Wilson Creek Members of the Wood River Formation. Ore is not present where the shear zone cuts quartzite or siliceous mudstone of the formation. The Homestake workings were described by Tuchek and Ridenour (1981), who estimated that at least 730,000 tons of submarginal reserves of silver, lead, and zinc ore remain in unexplored parts of the shear zone.

The Long Grade prospect is east of Lake Creek, about 1.9 km south of the Homestake workings (fig. 2). Five adits expose northwest-trending, breccia-filled mineralized shear zones that cut gray silty mudstone and tan calcareous sandstone of the Wilson Creek Member of the Wood River Formation (fig. 3). The shear zones occupy the hanging wall of the Lake Creek fault and are directly analogous in geologic setting to the Homestake structure (Tuchek and Ridenour, 1981). The shear zones strike N. 10°–30° W. and dip 30°–50° SW. Ore minerals include lead and zinc carbonate minerals and minor amounts of galena, sphalerite, and pyrite in a gangue of calcite, hematite, and crushed limestone wallrock. Tuchek and Ridenour (1981) estimated that the Long Grade property has 32,500 tons of submarginal reserves of lead and zinc ore.

The High Grade prospect (figs. 2, 5) is near the ridge top between Lake and Eagle Creeks, along the northward extension of the Homestake shear zone (Tuchek and Ridenour, 1981). The controls on mineralization are the same as those for the Homestake mine.

Mineral deposits of the Homestake mine and the Long Grade and High Grade prospects are confined to shear zones that are synthetic to, and in the hanging wall of, the Lake Creek fault (D3). The mineral deposits are therefore probably Eocene or younger in age. The metals are interpreted to have been remobilized from the Milligen Formation by hydrothermal convection cells associated with dacite porphyry dikes that are present throughout the area. These intrusive rocks are part of the regionally extensive Eocene magmatic rock suite associated with the Challis Volcanic Group and subjacent intrusive rocks (Moye and others, in press; Stewart and others, in press). Coeval extensional faulting related to uplift of the Pioneer Mountains core complex provided pathways for migration of metals and for formation of veins. The metals were deposited in veins and as replacements of calcareous rocks of the Wood River Formation.

DEPOSITS OF THE LAKE CREEK AND PRICE GROUPS

Deposits of the Lake Creek group of prospects (figs. 2, 5), described in detail by Tuchek and Ridenour (1981), are about 245 m northwest of lower Lake Creek Lake, 280 m above the valley floor. The workings follow a zinc-bearing and calcite-filled shear zone 0.1–1.1 m thick within silty micrite of the lower part of the Eagle Creek Member of the Wood River Formation (fig. 3). The shear zone strikes N. 40° W. and dips from 80° SW. to vertical. The shear zone is parallel with the strike of beds in the country rock and probably represents bedding-parallel shear produced during Mesozoic folding.

Deposits of the Price group (fig. 2) are about 405 m southeast of lower Lake Creek Lake, 312–499 m above the valley floor. Veins are developed in calcite-filled shear zones as long as 312 m that strike N. 40° W. and dip from 80° SW. to vertical. The host rock is conglomerate of the Hailey Member of the Wood River Formation. Umpleby and others (1930, p. 196) noted that mineral deposits are present as pods along irregular zones of brecciation and in joints and discontinuous fractures. These silver-lead-zinc-rich pods formed along the angular unconformity between the Hailey Member of the Wood River Formation and the underlying Milligen Formation. The unconformity has accommodated interstratal slip in the upper limb of a concentric, northeast-verging, overturned F2 fold. A shear zone produced by this interstratal slip hosts the mineral-rich pods.

Structures that host deposits of the Lake Creek and Price groups are F2 folds, and thus mineralization occurred either during Late Cretaceous plutonism or during Eocene magmatism (coincident with mineralization in the Homestake, Long Grade, and High Grade prospects). The small size and podlike nature of the F2-hosted mineral deposits limit their value for future mineral development.

EXPLORATION MODEL

Source of metals.—A source of metals is available from metamorphosed pelitic rocks of the Devonian Milligen Formation.

Remobilization.—Mesozoic plutonism and associated hydrothermal activity may have remobilized and concentrated some metals in Mesozoic compressional structures.

Structural conduits.—Paleogene detachment faults in the upper plate of the Pioneer Mountains core complex and associated shear zones served as sites of formation of vein and replacement ore.

Remobilization and concentration.—The presence of Eocene hypabyssal intrusive rocks suggests that remobilization and concentration of metals by hydrothermal convection cells occurred during Eocene magmatism. Deposition of ore minerals during the Cretaceous is rejected because there are no Cretaceous intrusive rocks within several tens of kilometers of the Lake Creek mineralized area and because the extensional host structures are demonstrably of Eocene age.

Host rock.—Calcareous fine-grained sedimentary rocks of the Wood River Formation are available as hosts for mineralization.

The Homestake and Long Grade mineral deposits previously were thought to be associated with D2 Mesozoic compressional structures; however, the recognition of Paleogene extensional tectonism (D3) (Link and others, 1988; Burton and others, 1989) in the eastern Boulder Mountains suggests that a new structural model should be used for silver-lead-zinc exploration in the region.

The Lake Creek deposits are similar in tectonic setting to silver-lead-zinc deposits in the southern Trigo Mountains of Yuma County, Arizona (Garner and others, 1982) and to gold deposits at the Picacho mine, Imperial County, California (Liebler, 1988). In these areas, mineralized rock is in the hanging wall of extensional faults, just as it is in the Boulder Mountains of Idaho. An areally extensive network of vertically stacked extensional faults is around the south, west, and northwest sides of the Pioneer Mountains core complex, an area that makes up most of the Boulder and Pioneer Mountains.

REFERENCES CITED

Armstrong, R.L., 1975, Sevier orogenic belt in Nevada and Utah: Geological Society of America Bulletin, v. 79, p. 429–458.

Batatian, L.D., 1991, Paleozoic stratigraphy and Cenozoic structure, central Boulder Mountains, Blaine and Custer Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 85 p.

Burton, B.R., 1988, Stratigraphy of the Wood River Formation in the eastern Boulder Mountains, Blaine and Custer Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 165 p.

Burton, B.R., and Blakley, J.D., 1988, The basal Challis conglomerate, Custer County, south-central Idaho—Implications for the initiation of Challis volcanism: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 408–409.

Burton, B.R., Link, P.K., and Rodgers, D.W., 1989, Death of the Wood River thrust—Structural relations in the Pioneer and Boulder Mountains, south-central Idaho: Geological Society of America Abstracts with Programs, v. 21, no. 5, p. 62.

Dover, J.H., 1980, Status of the Antler orogeny in central Idaho, clarification and constraints from the Pioneer Mountains, *in* Fouch, T.D. and Magathan, E.R., eds., Paleozoic paleogeography of the West-Central United States: Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, West-Central United States Paleogeography Symposium 1, p. 371–381.

———1981, Geology of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–A, p. 1–75.

———1983, Geologic map and sections of the central Pioneer Mountains, Blaine and Custer Counties, central Idaho: U.S. Geological Survey Miscellaneous Investigations Series Map I–1319, scale 1:48,000.

Ekren, E.B., 1985, Eocene cauldron-related volcanic events in the Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–C, p. 43–58.

Garner, W.E., Frost, E.G., Tanges, S.E., and Germinario, M.P., 1982, Mid-Tertiary detachment faulting and mineralization in the Trigo Mountains, Yuma County, Arizona, *in* Frost, E.G., and Martin, D.L., eds., Mesozoic-Cenozoic tectonic evolution of the Colorado River region, California, Arizona, and Nevada: San Diego, Calif., Cordilleran Publishers, p. 158–172.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

Hall, W.E., Batchelder, J.H., and Douglass, R.D., 1974, Stratigraphic section of the Wood River Formation, Blaine County, Idaho: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 579–592.

Hall, W.E., Rye, R.O., and Doe, B.R., 1978, Wood River mining district, Idaho—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, p. 579–592.

Huerta, A.D., 1992, Lake Creek fault—Evidence of pre-Challis shear within south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 57 p.

Kiilsgaard, T.K., Lewis, R.S., and Bennett, E.H., in press, Plutonic and hypabyssal rocks of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 2064–U.

Kim, A.Y.H., 1986, The kinematics of brittle polyphase deformation within the Pioneer metamorphic core complex, Pioneer Mountains, Idaho: Bethlehem, Pa., Lehigh University, M.S. thesis, 130 p.

Lewis, R.S., in press, Alteration and mineralization in the eastern part of the Soldier Mountains, Camas County, Idaho: U.S. Geological Survey Bulletin 2064–V.

Lewis, R.S., Kiilsgaard, T.H., Bennett, E.H., and Hall, W.E., 1987, Lithologic and geochemical characteristics of the central and southeastern part of the southern lobe of the Idaho batholith, *in* Vallier, T.L., ed., Geology of the Blue Mountains region of Oregon, Idaho, and Washington; the Idaho batholith and its border zone: U.S. Geological Survey Professional Paper 1436, p. 171–196.

Liebler, G.S., 1988, Geology and gold mineralization at the Picacho Mine, Imperial County, California, *in* Schafer, R.W., Cooper, J.J., and Vikre, P.G., eds., Bulk mineable precious metal deposits of the western United States: Geological Society of Nevada, p. 453–472.

Link, P.K., Skipp, Betty, Hait, M.H., Jr., Janecke., Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 5–42.

Mahoney, J.B., Link, P.K., Burton, B.R., Geslin, J.K., and O'Brien, J.P., 1991, Pennsylvanian and Permian Sun Valley Group, Wood River Basin, south-central Idaho, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 2, p. 551–580.

McIntyre, D.H., Ekren, E.D., and Hardyman, R.F., 1982, Stratigraphic and structural framework of the Challis volcanics in the eastern half of the Challis 1°×2° quadrangle, Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 3–22.

Moye, F.J., Hackett, W.R., Snider, L.G., and Mertzman, S.A., in press, Stratigraphy, chronology, and petrology of the southeastern Challis volcanic field, central Idaho: U.S. Geological Survey Bulletin 2064–FF.

O'Neill, R.L., and Pavlis, T.L., 1988, Superposition of Cenozoic extension on Mesozoic compressional structures in the Pioneer Mountains metamorphic core complex, central Idaho: Geological Society of America Bulletin, v. 100, p. 1833–1845.

Otto, B.R., and Turner, R.J.W., 1987, Stratigraphy and structure of the Milligen Formation, Sun Valley area, Idaho: Northwest Geology, v. 16, p. 95–103.

Park, W.A., in press, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–DD.

Paull, R.A., 1974, Upper Cretaceous(?) to lower Eocene post-orogenic conglomerate, south-central Idaho: Geological Society of America Abstracts with Programs, v. 6, no. 5, p. 466.

Pearce, S., Schlieder, G., and Evenson, E.B., 1988, Glacial deposits of the Big Wood River Valley, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 203–207.

Ratchford, M.E., 1989, Geology of the Boulder Basin, Blaine and Custer Counties, Idaho: Moscow, University of Idaho, M.S. thesis, 154 p.

———in press, Epigenetic lead-tin-silver-antimony-gold veins in Boulder Basin, Blaine and Custer Counties, Idaho—Potential for economic tin mineralization: U.S. Geological Survey Bulletin 2064–JJ.

Roberts, R.J., Hotz, P.E., Gilluly, J., and Furgusan, H.G., 1958, Paleozoic rocks of north central Nevada: American Association of Petroleum Geologists Bulletin, v. 42, p. 2813–2857.

Rodgers, D.W., and Janecke, S.U., 1992, Tertiary paleogeologic maps of the western Idaho-Wyoming-Montana thrust belt, *in* Link, P.K., Rodgers, D.W., and Platt, L.B., eds., Regional geology of eastern Idaho and western Wyoming: Geological Society of America Memoir 179, p. 83–94.

Sandberg, C.A., Hall, W.E., Batchelder, J.N., and Axelson, Claus, 1975, Stratigraphy, conodont dating, and paleotectonic interpretation of the type Milligen Formation (Devonian), Wood River area, Idaho: U.S. Geological Survey Journal of Research, v. 3, p. 707–720.

Service, A.L., and Kellum, D.F., 1956, Final report, contract Idm–E18, Homestake Mine (unpublished): Spokane, Wash., Defense Minerals Exploration Administration, 12 p.

Silverberg, D.S., 1990, The tectonic evolution of the Pioneer metamorphic core complex, south-central Idaho: Cambridge, Massachusetts Institute of Technology, Ph.D. dissertation, 279 p.

Skipp, B., and Hall, W.E., 1980, Upper paleozoic paleotectonics and paleogeography of Idaho, *in* Fouch, T.D., and Magathan, E.R., eds., Paleozoic paleogeography of the West-Central United States: Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, West-Central United States Paleogeography Symposium 1, p. 387–422.

Skipp, B., Hall, W.E., and Link, P.K., 1994, Geologic map of the Mahoney Butte quadrangle, Blaine County, Idaho: U.S. Geological Survey Geologic Quadrangle Map GQ–1742, scale 1:24,000.

Stewart, D.E., Schmidt, K.L., Moye, F.J., Mahoney, J.B., and Link, P.K., in press, Eocene intrusive complexes in the Smoky and Boulder Mountains, south-central Idaho: U.S. Geological Survey Bulletin 2064–II.

Tuchek, E.T., and Ridenour, James, 1981, Economic appraisal of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497, p. 181–292.

Turner, R.J.W., and Otto, B.R., 1988, Stratigraphy and structure of the Milligen Formation, Sun Valley area, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 153–167.

Umpleby, J.B., Westgate, L.G., and Ross, C.P., 1930, Geology and ore deposits of the Wood River region, Idaho: U.S. Geological Survey Bulletin 814, 250 p.

Whitman, S.K., 1990, Metamorphic petrology and structural geology of the Pennsylvanian-Permian Dollarhide Formation, Blaine and Camas Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 108 p.

Woodward, N.B., Boyer, S.E., and Suppe, John, 1985, An outline of balanced cross sections (2nd ed.): University of Tennessee Studies in Geology 11, 170 p.

Worl, R.G., Kiilsgaard, T.H., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Wust, S.L., 1986, Extensional deformation with northwest vergence, Pioneer core complex, central Idaho: Geology, v. 14, p. 712–714.

Wust, S.L., and Link, P.K., 1988, Field guide to the Pioneer Mountains core complex, south-central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 43–54.

Geologic Setting of Mineral Deposits in the Washington Basin Area, Custer County, South-Central Idaho

By J. Brian Mahoney

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064-G

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

Geologic Setting of Mineral Deposits in the Washington Basin Area, Custer County, South-Central Idaho

By J. Brian Mahoney[1]

ABSTRACT

In the Washington Basin area of south-central Idaho, mineral deposits of lead, silver, and zinc are contained within folded and faulted Paleozoic strata that have been intruded by Cretaceous and Eocene stocks and dikes. Micritic sandstone, siltstone, sandy micrite, and polymict conglomerate of the Middle Pennsylvanian to Lower Permian Sun Valley Group unconformably overlie black argillite of the Paleozoic Salmon River assemblage. These strata are contained in a northeast-trending, south-plunging, east-vergent anticline that is the dominant structure in the area. Deformation associated with Mesozoic folding led to the development of fractures parallel with the direction of the fold hinge and the interstratal slip that created intense shearing along the unconformable contact. The anticline was intruded by Cretaceous granodiorite and biotite quartz diorite in its hinge region, and Eocene dacite porphyry dikes are localized in one area near the Cretaceous granodiorite-Paleozoic contact. Paleozoic strata near the intrusive rocks are metamorphosed to hornblende hornfels facies and commonly bleached white. Northeast-trending high-angle faults offset folded Paleozoic rocks and Eocene andesitic to dactitic flow rocks of the Challis Volcanic Group in the southeastern part of the area.

Mineral deposit types in the Washington Basin area include polymetallic veins, shear-zone-hosted polymetallic veins, and replacement deposits. Sulfide ores contain pyrrhotite, sphalerite, chalcopyrite, pyrite, and galena and minor amounts of jamesonite, arsenopyrite, and telluride minerals. These deposits reflect at least two distinct phases of mineralization. Polymetallic vein deposits near the Cretaceous granodiorite-Paleozoic contact and in fold-hinge parallel fractures are interpreted to be the result of late-stage hydrothermal fluid circulation associated with intrusion of Cretaceous granodiorite. Massive replacement deposits near Red Hill formed when Cretaceous(?) biotite quartz diorite intruded the Cretaceous granodiorite and Middle Pennsylvanian to Lower Permian Grand Prize Formation. Polymetallic vein deposits in high- and low-angle shear zones on Bible Back Mountain are of uncertain affinity but may be associated with Late Cretaceous mineralization. Lead-isotope signatures and the spatial association of the Paleozoic Salmon River assemblage and mineral deposits suggest that the Salmon River assemblage may be the source of metals in the mineral deposits of the Washington Basin area.

INTRODUCTION

Washington Basin is at the southern end of the White Cloud Peaks in the southeastern corner of Custer County, Idaho. The basin is a spectacular glacial cirque that has a southeasterly aspect, and it is drained by Washington Creek, a tributary to Germania Creek (fig. 1). The terrain is rugged; elevations range from 3,187 m (10,519 ft) on Washington Peak to 2,545 m (8,400 ft) at the east end of the cirque. In this report, the term "Washington Basin area" refers to both Washington Basin itself and to Germania Basin, directly to the south (fig. 2). The area is bounded on the north by the headwaters of Fourth of July Creek, on the west by Champion Lakes Basin, on the east by Washington Lake Creek, and on the south by Germania Creek. The area is reached by following the Pole Creek road 22 km (13.2 mi) from State Highway 75, then continuing 3.5 km (2.1 mi) along a rough jeep trail into the basin itself. The Washington Basin area is within the Sawtooth National Recreation Area and is withdrawn from new mineral entries; current exploration is limited to existing claims. The White Cloud Peaks adjacent to Washington Basin are currently under review for inclusion in the proposed White Cloud–Boulder Wilderness Area.

The Washington Basin area is on the boundary between the Hailey and Challis 1°×2° quadrangles. The geology of Washington Basin was briefly described in conjunction with regional geologic studies of the White Cloud Peaks by Ross (1963), Fisher and others (1983), Sengebush (1984), Hall (1985), Tschanz and others (1986), and Hall and Hobbs

[1]U.S. Geological Survey and University of British Columbia, Vancouver, British Columbia V6T 2B4.

(1987). The most recent geologic maps of the area include those by Fisher and others (1983), and Worl and others (1991). I previously (1989) described the structure and stratigraphy of the area, and this report is a modification and expansion of those ideas.

Mineral deposits of the Washington Basin area were described in detail by Van Noy and others (1986) of the U.S. Bureau of Mines, who included the Washington Basin area in the Germania Creek mining district. Teepen (1985) described the content and genesis of sulfide mineral veins in a part of the area. I incorporated the mineral deposit evaluations of Van Noy and others (1986) into a description of mineral deposit types of the area (Mahoney, 1989).

Acknowledgments.—I thank Paul K. Link and David W. Rodgers of Idaho State University for their assistance in the field and for numerous thought-provoking discussions about the geology of the region. Jeff Jones of the Sawtooth National Recreation Area was a valuable source of information about the mining history and geology of the area. This manuscript benefited from reviews by Reed S. Lewis, Paul K. Link, and Ronald G. Worl.

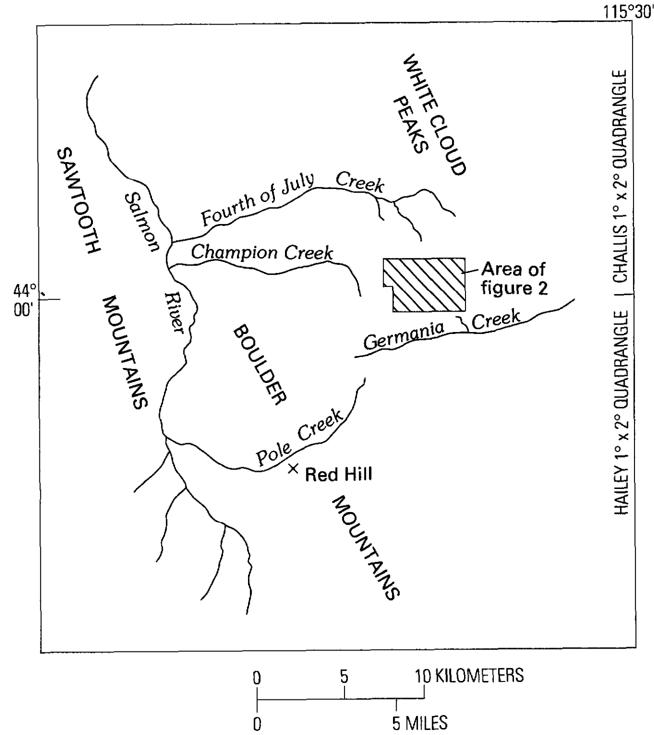


Figure 1. Map showing location of the Washington Basin area, Custer County, Idaho. Location of map area of figure 2 is also shown.

HISTORICAL PERSPECTIVE

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

Although records of production are limited, the major producing mines in the area were apparently on the northwestern flank of Bible Back Mountain (the Idahoan and Old Bible Back mines) and at the southwestern end of Washington Basin (Black Rock mine). Quantitative estimates of ore removed are nonexistent, but monetary values (1915 prices) exceed $250,000 from Bible Back Mountain and $125,000 from the Black Rock mine (Umpleby, 1915). The production in the Washington Basin area has been primarily in lead-silver ores, although some gold values have been reported (Van Noy and others, 1986). Small amounts of copper and zinc may have also been produced. Other reported commodities include Cu, Zn, Sb, W, Bi, Te, and Se (Van Noy and others, 1986). The area has been the subject of recent gold exploration activity.

GEOLOGIC SETTING

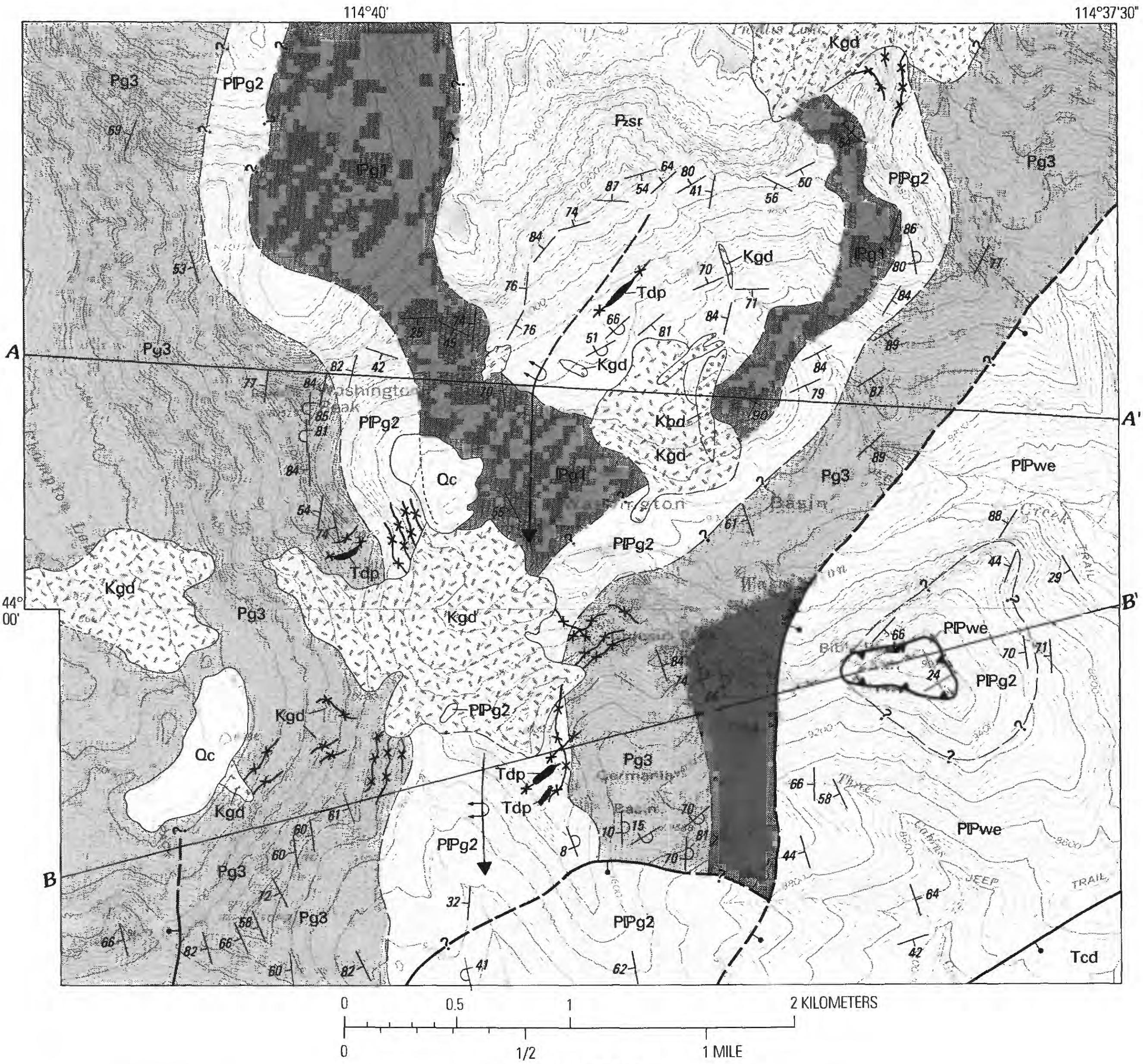
The geology of the Washington Basin area consists of complexly folded and faulted Paleozoic strata that have been intruded by at least two different types of igneous stocks and dikes. The mineral deposits in the area display a strong structural control, and at least two separate mineralization events are believed to have occurred. Recent mapping (Mahoney, 1989, this report, unpublished data) has resulted in significant revision of previously postulated stratigraphic and structural models.

STRATIGRAPHY

Washington Basin is underlain by three Paleozoic sedimentary units: the Paleozoic Salmon River assemblage and the Middle Pennsylvanian to Lower Permian Wood River and Grand Prize Formations of the Sun Valley Group (Mahoney and others, 1991) (fig. 2). Mineral deposits are hosted by each of these units in the Washington Basin area, but the most important host is the Grand Prize Formation, which underlies most of the basin.

Pennsylvanian and Permian sedimentary rocks in the area previously were assigned to variety of stratigraphic units including the rocks at Pole Creek (Fisher and others,

Figure 2 (facing page). Map showing geology of the Washington Basin area, Custer County, Idaho. Location of map area is shown in figure 1. Lines of section of figure 3 are also shown. Base from U.S. Geological Survey 7.5-minute topographic series maps, Washington Peak (1964) and Horton Peak (1970).



EXPLANATION

Qc — **Quaternary colluvium**

CHALLIS VOLCANIC GROUP (EOCENE)

Tcd — **Dacite lava**

Tdp — **Dacite porphyry**

CRETACEOUS INTRUSIVE ROCKS

Kgd — **Quartz diorite (Cretaceous?)**

Kbd — **Biotite granodiorite**

SUN VALLEY GROUP (LOWER PERMIAN TO MIDDLE PENNSYLVANIAN)

Grand Prize Formation

Pg4 — **Member 4 (Lower Permian)**

Pg3 — **Member 3 (Lower Permian)**

PPg2 — **Member 2 (Lower Permian to Middle Pennsylvanian)**

PPg1 — **Member 1 (Middle Pennsylvanian)**

Wood River Formation

PPwe — **Eagle Creek Member (Lower Permian to Middle Pennsylvanian)**

OLDER PALEOZOIC STRATA

Pzsr — **Salmon River assemblage (Paleozoic)**

Contact—Dashed where approximately located; dotted where concealed; queried where uncertain

High-angle fault—Ball and bar on downthrown side; dashed where approximately located; queried where uncertain

Low-angle fault—Sawteeth on upper plate

Dike

Overturned anticline—Dashed where approximately located

40 **Strike and dip of bedding**

15 **Strike and dip of overturned bedding**

1983), the Wood River Formation (Tschanz and others, 1986), and the Grand Prize Formation (Hall, 1985; Mahoney and Sengebush, 1988; Mahoney and others, 1991). Mahoney and others (1991) included the Wood River and Grand Prize Formations in the Middle Pennsylvanian to Lower Permian Sun Valley Group. The unit designations presented in this report clarify stratigraphic nomenclature in the Washington Basin area.

SALMON RIVER ASSEMBLAGE

The Salmon River assemblage in the Washington Basin area consists of dark carbonaceous argillite, limestone, siltstone, and minor sandstone exposed on the ridge north of Washington Basin proper (fig. 2). Rocks of the assemblage extend northward from Washington Basin, underlie all of the Phyllis Lake cirque and continue north across Fourth of July Creek (figs. 1, 2). The Salmon River assemblage forms the core of a south-plunging, east-vergent anticline that dominates the structural setting of the area. Its upper contact forms an arcuate map pattern with the overlying Grand Prize Formation in Washington Basin (figs. 2, 3). The lower contact of the Salmon River assemblage has not been identified (Hall and Hobbs, 1987).

The Salmon River assemblage in the Washington Basin area is primarily thin-bedded, dark-gray to black, carbonaceous argillite that is locally iron stained and commonly has a phyllitic sheen. The dark argillite is intercalated with medium-bedded blue-gray sandy limestone, thin-bedded parallel- to cross-laminated fine-grained sandstone, and thin-bedded dark-gray siltstone. The unit is structurally deformed with mesoscopic tight to isoclinal folds, near-bedding parallel cleavage, quartz veining, and abundant fractures. Primary bedding is difficult to recognize in many exposures. The Salmon River assemblage is locally contact metamorphosed to calc-hornfels and is commonly silicified. Carbonaceous units within the assemblage locally host mineral deposits, and the assemblage contains numerous mines and prospects throughout the White Cloud Peaks (Van Noy and others, 1986).

The age of the Salmon River assemblage is problematic; age determinations range from Late Cambrian to Late Mississippian, and the assemblage is currently simply designated as Paleozoic (Hall and Hobbs, 1987). The Salmon River assemblage in Washington Basin is, however, lithologically similar to the Devonian Milligen Formation to the south and occupies a similar structural position (Link and others, this volume). Both the Paleozoic Salmon River assemblage and the Devonian Milligen Formation contain thick sections of homogeneous black argillite, abundant thin-bedded fine-grained turbidite sequences, and intercalated blue-gray, locally tremolitic, limestone. They also both contain well-developed near-bedding-parallel cleavage and mesoscopic tight to isoclinal folds and display a characteristic phyllitic sheen. They both host inferred syngenetic sulfide deposits having identical lead-isotope signatures (Sanford and Wooden, this volume). The Salmon River assemblage in the White Cloud Peaks has yielded Middle to Late Devonian conodonts (Hobbs, 1985; Hall and Hobbs, 1987); the reported Cambrian and Mississippian dates are obtained from rocks exposed to the north of the main White Cloud Peaks that are interpreted as structural imbrications within the assemblage (Hall, 1985). I interpret the Salmon River assemblage exposed in the White Cloud Peaks as primarily Devonian in age and suggest that the Salmon River assemblage is the lateral equivalent of the Devonian Milligen Formation exposed to the south.

GRAND PRIZE FORMATION

The Pennsylvanian and Permian Grand Prize Formation unconformably overlies the Salmon River assemblage in the Washington Basin area. The Grand Prize Formation is at least 3,000 m (9,840 ft) thick and is subdivided into four informal members, all of which are exposed in the Washington Basin area. A partial stratigraphic section of the Grand Prize Formation, measured on the east flank of Washington Peak, is shown in figure 4. Complete stratigraphic descriptions of the formation are in Mahoney and Sengebush (1988) and Mahoney and others (1991).

The contact between the Grand Prize Formation and the Salmon River assemblage has been the subject of dispute among workers in the region (Hall, 1985; Hall and Hobbs, 1987; Mahoney and Sengebush, 1988; Mahoney and others, 1991). Hall (1985) described the contact as a thrust fault in which the Grand Prize Formation is thrust over the Salmon River assemblage along an intensely brecciated fault zone. Sengebush (1984) suggested that the contact is actually a sheared unconformity and that the brecciated zone is a tectonically deformed clast-supported conglomerate at the base of the Grand Prize Formation. Mahoney and Sengebush (1988) and Mahoney and others (1991) described regional relations that support a sheared erosional unconformity between the Grand Prize Formation and the underlying Salmon River assemblage. Examination of the contact on the ridge north of Washington Basin strongly supports the sheared unconformity hypothesis. At this location, the contact consists of matrix-supported (locally clast supported) conglomerate and intercalated fine- to coarse-grained quartzite overlying black argillite with a low-angle structural discordance. The conglomerate contains intercalated quartzite beds, sedimentary structures including graded bedding and weak clast imbrication, and subrounded black argillite clasts strikingly similar to the underlying Salmon River assemblage. These features support a sedimentary, not tectonic, origin for the lowest member of the Grand Prize Formation in this area and suggest that the contact with the underlying Salmon River assemblage is an erosional unconformity. The

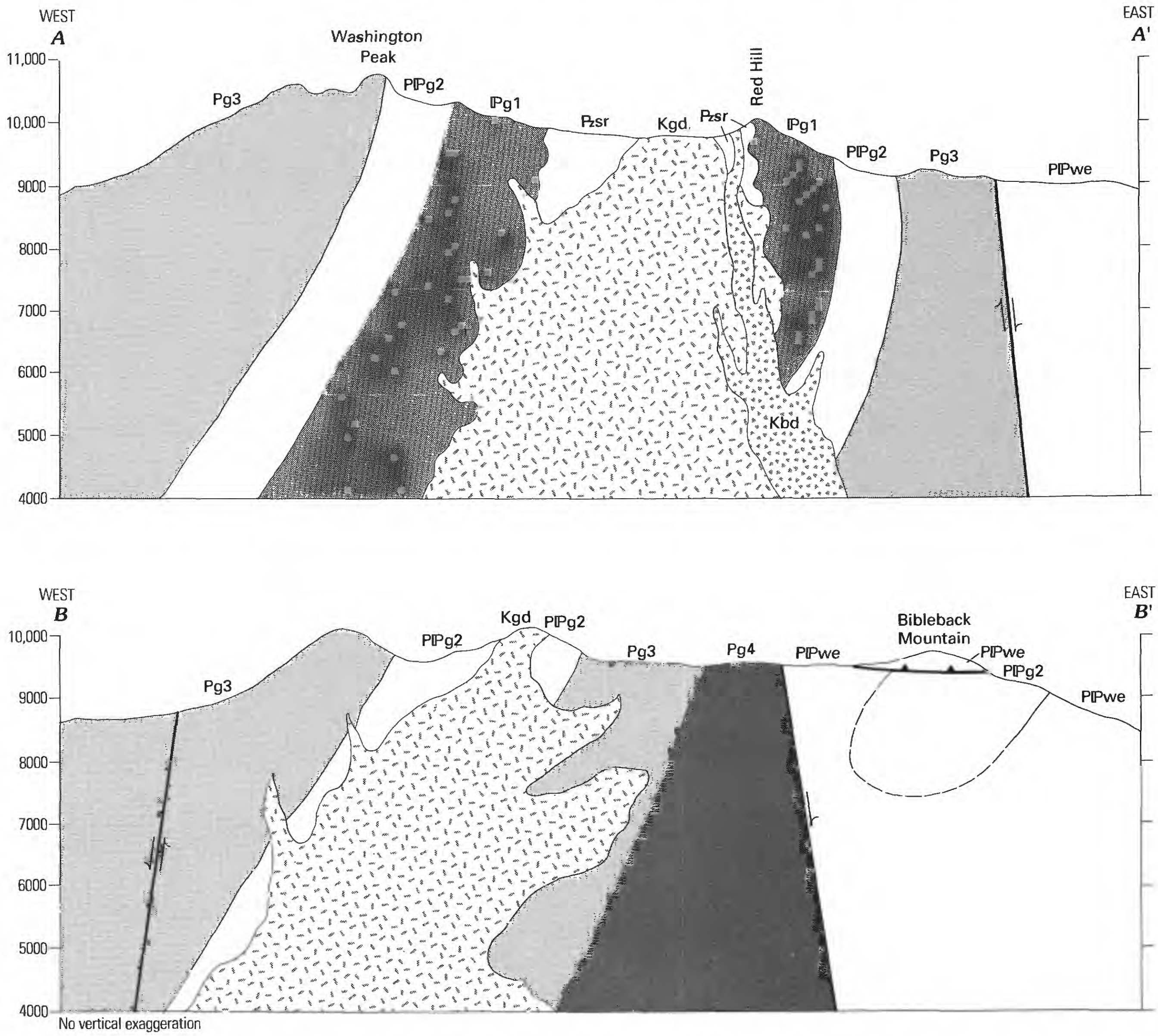


Figure 3. Cross sections showing geology of the Washington Basin area, Custer County, Idaho. Lines of section are shown in figure 2. Stratigraphic units: Kgd, quartz diorite (Cretaceous?); Kbd, biotite granodiorite (Cretaceous?); Grand Prize Formation of the Sun Valley Group including Pg4, member 4 (Lower Permian); Pg3, member 3 (Lower Permian); PIPg2, member 2 (Lower Permian to Middle Pennsylvanian); and IPg1, member 1 (Middle Pennsylvanian); PIPwe, Eagle Creek Member of the Wood River Formation (Lower Permian to Middle Pennsylvanian); Pzsr, Salmon River assemblage (Paleozoic).

conglomerate displays local brecciation, quartz veining, and a preferential alignment of elongate clasts, particularly near the base of the unit. This preferred clast orientation and brecciation is interpreted as the result of shearing due to interstratal slip along the unconformable contact during megascopic folding in the Mesozoic and is not the result of intense brecciation during thrust displacement.

In the Washington Basin area, the Grand Prize Formation is contained in both limbs of a south-plunging, east-vergent anticline (figs. 2, 3). In the western limb of the anticline, member 1 of the Grand Prize Formation consists of approximately 175 m (574 ft) of matrix- to clast-supported conglomerate and minor intercalated coarse-grained quartzite overlain by 175 m (574 ft) of thin- to medium-bedded, locally parallel to wavy laminated, fine-to medium-grained quartzite (fig. 4). Overlying this quartzite is approximately 75 m (246 ft) of clast-supported conglomerate that is overlain by about 25 m (82 ft) of blue-gray bioclastic (fossil hash) sandy limestone.

Member 2 of the Grand Prize Formation is in conformable contact with the underlying member and, in the Washington Basin area, consists of about 250 m (820 ft) of

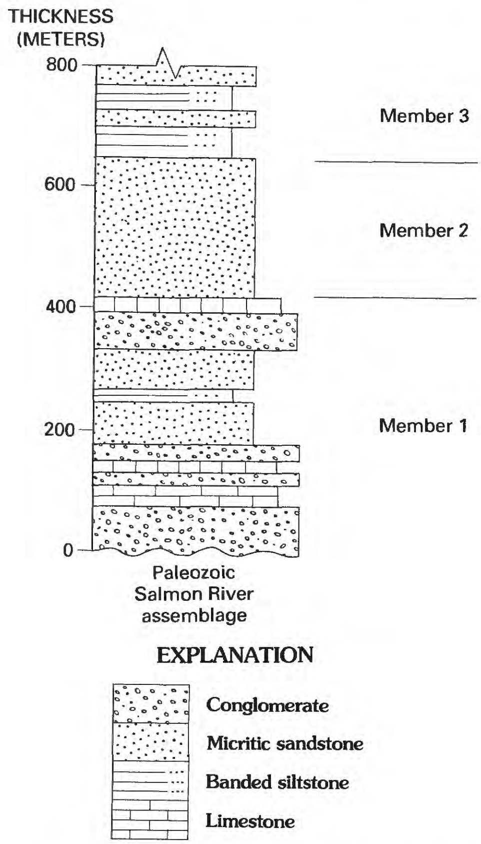


Figure 4. Stratigraphic section of part of the Middle Pennsylvanian to Lower Permian Grand Prize Formation, Washington Basin area, Custer County, Idaho. Measured from east to west on the ridge on the north side of Washington Basin in steeply west dipping beds on the east flank of peak 10,519 ft, Washington Peak U.S. Geological Survey topographic map (scale 1:24,000). Section starts at 10,200 ft and goes along ridge to top of peak 10,519 ft. Base of section is at lat 44°15′00″N., long 114°39′38″ W.

thick-bedded to massive, fine-grained, calcareous sandstone intercalated with minor thin- to medium-bedded, parallel- to cross-laminated, fine-grained sandstone and dark-gray calcareous siltstone (figs. 2, 4). Member 2 has a gradational contact with overlying member 3 (figs. 2, 4).

Member 3 of the Grand Prize Formation consists of thin-bedded, fine-grained sandstone and dark siltstone forming distinctive banded couplets, 15–45 cm (5.9–18 in.) thick, that are well exposed in the cirque headwall south of Washington Peak (fig. 2). Intercalated with these banded couplets are thick-bedded to massive sandstone beds as thick as 25 m (82 ft). Regionally, member 3 has an estimated thickness of 650–1,700 m (2,130–5,580 ft); however, the thickness of member 3 in the Washington Basin area is uncertain due to faulting in the Champion Lakes cirque and intrusion-related deformation in the Washington Basin area (Mahoney and others, 1991) (fig. 2). The member underlies most of the Champion Lakes cirque; its gradational upper contact with member 4 is exposed below the ridge west of Champion Lakes (Mahoney, 1991). In the Champion Lakes cirque, member 3 contains a much higher percentage of sandstone and quartzite than previously reported in the Grand Prize Formation (Mahoney and Sengebush, 1988; Mahoney and others, 1991).

In the Washington Basin area, member 4 is exposed in the eastern limb of the anticline on the east side of Croesus Peak. Member 4 consists of thin- to medium-bedded carbonaceous limestone, siltstone, and mudstone and forms a distinctive dark-gray unit. The siltstone and limestone contain abundant sedimentary structures including graded bedding, cross laminae, parallel laminae, and soft-sediment deformation features.

The Grand Prize Formation is Middle Pennsylvanian to Early Permian (Desmoinesian to Leonardian) in age, based on conodonts collected from the type section approximately 8 km (4.8 mi) southwest of Washington Basin (Hall, 1985) and on lithostratigraphic correlations with the well-dated Wood River and Dollarhide Formations of the Sun Valley Group (Mahoney and others, 1991).

In Washington Basin, the Grand Prize Formation is locally contact metamorphosed to calc-silicate hornfels and is commonly bleached and silicified, resulting in a bright-white appearance in outcrop. It hosts mineral deposits in members 2, 3, and 4 both in Washington Basin and in Germania Basin to the south; mineral deposits are in polymetallic veins in fracture systems and in replacement deposits. Regionally, the Grand Prize Formation contains numerous mines and prospects.

WOOD RIVER FORMATION

The Wood River Formation consists of more than 2,500 m (8,200 ft) of micritic sandstone, siltstone, and sandy micrite and is subdivided into three members (Mahoney and others, 1991). Only the Eagle Creek Member of the Wood River Formation is exposed in the study area, most notably on Bible Back Mountain, where it is host to polymetallic veins in shear zones. The formation is exposed in a series of fault blocks to the southeast of Washington Basin (fig. 2).

In the Washington Basin area, the Eagle Creek Member of the Wood River Formation consists primarily of thick-bedded to massive, fine-grained micritic sandstone intercalated with minor thin- to medium-bedded, blue-gray sandy micrite and micritic siltstone. Regionally, the unit has an average thickness of 880–1,300 m (2,887–4,265 ft), but in the Washington Basin area extensive brecciation prevents an accurate thickness determination.

The Eagle Creek Member of the Wood River Formation is well constrained as Middle Pennsylvanian to Early

Permian (Desmoinesian to Wolfcampian) in age and is interpreted as a lateral equivalent of member 2 of the Grand Prize Formation (Mahoney and others, 1991; Link and others, this volume). South of Bible Back Mountain, the Eagle Creek Member of the Wood River Formation is gradationally overlain by member 2 of the Grand Prize Formation. South of Germania Basin, outside the study area, a thin sequence (~100 m, 328 ft) of the Wilson Creek Member of the Wood River Formation is gradationally overlain by member 4 of the Grand Prize Formation. These intertonguing relations indicate that the Grand Prize Formation passes laterally into the Wood River Formation to the south and that the Grand Prize Formation represents a northern, sandy facies of the Sun Valley Group (Mahoney and others, 1991; Link and others, this volume).

The Wood River Formation is in high-angle fault contact with the Grand Prize Formation in the saddle west of Bible Back Mountain (figs. 2, 3). On top of Bible Back Mountain, the Wood River Formation is metamorphosed to a siliceous quartzite and is extensively brecciated and iron stained; the formation is cut by numerous north-trending shear zones that host polymetallic lead-silver veins. This iron-stained unit is in low-angle fault contact with underlying, essentially unfractured Grand Prize Formation; this relationship is well exposed on the north face of Bible Back Mountain (fig. 2).

STRUCTURE

The dominant structure in the Washington Basin area is a northeast-trending, south-plunging, east-vergent anticline within the Salmon River assemblage and the overlying Grand Prize Formation (figs. 2, 3). This anticline and its associated hinge-parallel fractures controlled the location of mineralization in the area and probably localized Cretaceous intrusions (figs. 2, 5). The Salmon River assemblage forms the core of the anticline, which plunges south-southeast at a moderate angle (~50°–60°) (fig. 2), as indicated by stereonet fold solutions. The west limb of the anticline consists of steeply southwest dipping to near vertical Grand Prize Formation in unconformable contact with the underlying Salmon River assemblage. Competency differences between the two units at the unconformity have resulted in ductile deformation of the upper part of the Salmon River assemblage and brecciation and shearing in the more competent conglomeratic member 1 of the Grand Prize Formation. The eastern limb of the anticline consists of near vertical to overturned, steeply southwest dipping Grand Prize Formation. The eastern limb of the fold is difficult to recognize due to widespread igneous intrusive activity. The structural orientation of the eastern limb is easily recognized on the east side of Croesus Peak, where steeply west dipping, overturned members 3 and 4 of the Grand Prize Formation are exposed (fig. 3).

Several northeast-trending high-angle faults that have minor displacement (tens of meters) are exposed in the Washington Basin area. Two high-angle faults in the area are interpreted to have significant displacement (tens to hundreds of meters) (fig. 2). The fault in the saddle west of Bible Back Mountain is steeply dipping and juxtaposes Grand Prize Formation (members 3 and 4) to the west with Wood River Formation (Eagle Creek Member) to the east. This fault has been reported to be a large northeast-trending, west-dipping reverse fault with numerous parallel shear zones (Van Noy and others, 1986). The map pattern of the fault indicates that it is a near vertical to steeply east dipping normal fault of significant displacement (hundreds of meters). The western side of this fault exposes medium-grained Cretaceous granodiorite and contact-metamorphosed, ductilely deformed Grand Prize Formation in the eastern limb of a south-plunging anticline. The eastern side of the fault exposes brittlely deformed, moderately dipping Eagle Creek Member of the Wood River Formation and overlying member 2 of the Grand Prize Formation, both of which are structurally overlain in low-angle fault contact by fractured, brecciated, and iron-stained quartzite of the Eagle Creek Member (figs. 2, 3). The difference in styles of structural deformation, intrusion, and metamorphism across the fault suggests that this fault is down to the east, an interpretation supported by three-point fault solutions indicating normal offset on a steeply east dipping fault. Several mineralized veins are parallel with this fault on the north side of Washington Basin, and Van Noy and others (1986) suggested that there may be mineralized zones along this fault.

The second fault of significant displacement in the Washington Basin area is in the southeastern corner of the map area, where a northeast-trending normal fault places the Eagle Creek Member of the Wood River Formation against andesitic to dacitic flow rocks of the Challis Volcanic Group (fig. 2). The lack of Paleozoic sedimentary units and the presence of more than 500 m (1,640 ft) of volcanic rocks to the east of this fault, together with the absence of volcanic rocks on the west side of the structure, suggest that the fault has a displacement of at least 500 m (1,640 ft).

A prominent low-angle structure is exposed on the north face of Bible Back Mountain, where massive quartzite and calc-silicate of the Eagle Creek Member of the Wood River Formation structurally overlie thin- to thick-bedded quartzite, siltite, and limestone of member 2 of the Grand Prize Formation and the Eagle Creek Member of the Wood River Formation. The structure strikes to the northeast and dips about 18° SE. (Van Noy and others, 1986). Rocks on both sides of the structure are heavily brecciated and fractured, but the upper plate rocks are much more silicified and iron stained than lower plate rocks. Van Noy and others (1986) described the low-angle structure as a thrust fault, although sense of movement indicators are lacking. This structure contains polymetallic shear-zone-hosted mineral deposits (Van Noy and others, 1986).

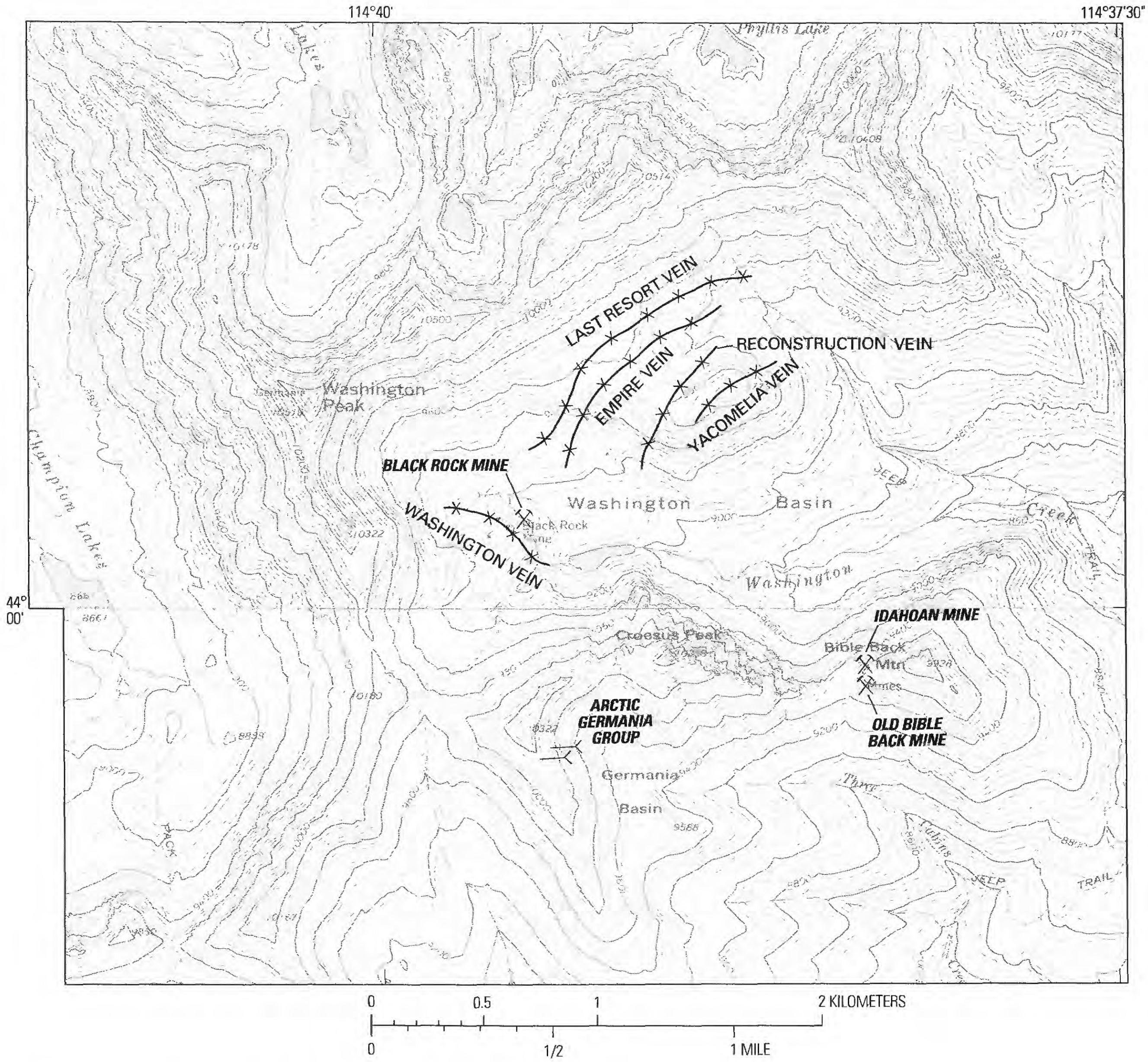


Figure 5. Map showing mines and major mineralized veins in the Washington Basin area, Custer County, Idaho. Modified from E. Oberbilling (unpublished Defense Minerals Exploration Administration report, 1962). Base from U.S. Geological Survey 7.5-minute topographic series maps, Washington Peak (1964) and Horton Peak (1970).

IGNEOUS INTRUSIVE ROCKS

The Washington Basin area contains at least three separate igneous rock units. The most extensive unit is porphyritic biotite granodiorite that makes up several stocks within the Washington Basin area (fig. 2). The granodiorite also forms dikes on the north side of Croesus Peak and in the headwall of the Washington Basin cirque. The biotite granodiorite is medium to coarse grained, has a grussy weathered appearance, and locally contains phenocrysts of alkali feldspar 3–10 cm (1.2–3.9 in.) long. The porphyritic texture of the granodiorite decreases with depth and is believed to be the result of potassium metasomatism near intrusive contacts. The granodiorite intrudes the Grand Prize Formation and the Salmon River assemblage and locally contains metasedimentary roof pendants. Randomly oriented aplite dikes and mafic xenoliths are common in the granodiorite. The granodiorite is locally iron stained near contact zones, particularly in exposures on the west side of peak 9776 ft (Red Hill) (fig. 2). Because of its proximity to the Cretaceous White Cloud stock and because it is cut by Eocene dacite porphyry dikes, the granodiorite is believed to be Cretaceous in age. The location of several of the Cretaceous stocks near the hinge region of the anticline suggests that stock emplacement may have been structurally controlled (figs. 2, 3).

A fine-grained, holocrystalline, biotite-rich quartz diorite dike intrudes the porphyritic granodiorite and the Grand Prize Formation on Red Hill (figs. 2, 3). The dike is 10–15 m (33–49 ft) wide on Red Hill and splits into three separate dikes on the north side of Red Hill. Fine-grained biotite makes up more than 50 percent of the unit. The dike is probably Cretaceous in age and is believed to be a late-stage intrusive feature associated with the porphyritic granodiorite; however, the biotite quartz diorite dike crosscuts the porphyritic granodiorite and may therefore be as young as Eocene in age. Chloritic alteration of biotite precludes a more accurate age determination. Quartzite of the Grand Prize Formation hosts replacement mineralization adjacent to the diorite dike. Similar dikes are adjacent to massive sulfide replacement deposits in quartzite of the Grand Prize Formation at the Pole prospect approximately 8 km (4.8 mi) to the southeast along Pole Creek.

Eocene dacite porphyry dikes intrude both the Grand Prize Formation and the Cretaceous granodiorite in the head of Germania Basin. Emplacement of the dikes was primarily along the contact between the granodiorite and the Paleozoic rocks. A mafic dike of uncertain age cuts the quartzite on the west end of Bible Back Mountain and is spatially related to mineralized rocks; similar lamprophyric dikes elsewhere in the White Clouds are interpreted as Cretaceous in age (R.G. Worl, oral commun., 1989).

MINERAL DEPOSITS

In the Washington Basin area, mineral deposits are hosted by members 2, 3, and 4 of the Grand Prize Formation, by the Eagle Creek Member of the Wood River Formation, and by the Cretaceous porphyritic granodiorite. Mineralization was primarily structurally controlled, as suggested by mineralized north-trending shear zones on Bible Back Mountain and by northeast-trending (fold-hinge parallel) mineralized veins in Washington Basin (fig. 5). Alteration of host rocks is pervasive and includes oxidation in both the igneous and sedimentary hosts and sericitic alteration in the igneous host. Alteration products include limonite, hematite, goethite, and sericite (Van Noy and others, 1986).

Teepen (1985) described the mineral deposits and paragenetic sequence of the Empire and Black Rock claims in Washington Basin (fig. 4). Van Noy and others (1986) provided a complete description of mines and prospects in the Washington Basin area. Information from these reports is included in the following descriptions.

Three types of mineral deposits are present in the Washington Basin area: (1) polymetallic veins and replacement deposits associated with the Cretaceous porphyritic granodiorite, (2) shear-zone-hosted polymetallic veins in sedimentary rocks, and (3) massive replacement deposits in quartzitic rocks.

POLYMETALLIC VEINS

Two types of polymetallic veins are present in Washington Basin: (1) northeast-trending veins localized along fold hinge-parallel fractures and (2) veins along the contact between the Cretaceous porphyritic granodiorite and calc-silicate rocks of the intruded Grand Prize Formation.

The most prominent mineralized structure in the Washington Basin area is a system of northeast-trending quartz-pyrrhotite-pyrite veins on the floor of Washington Basin. These veins are contained in shear zones and fracture systems that are subparallel with the fold axial plane of the asymmetric anticline that dominates the structure of the area, and they are probably the result of tensional stresses in the axis of the fold (figs. 2, 5). The vein system includes the Last Resort, Empire, Reconstruction, and Yacomella veins and is easily recognized by heavily oxidized sulfide minerals at the surface (fig. 5). The veins consist primarily of quartz and pyrrhotite and lesser amounts of pyrite, chalcopyrite, sphalerite, and galena. Small amounts of scheelite, telluride minerals, and jamesonite have also been reported (Van Noy and others, 1986). Samples assayed from the Empire vein, in the center of Washington Basin, average less than 0.01 ounces of gold per ton, 0.59 ounces of silver per ton, 0.28 percent Pb, 0.40 percent As, and less than 0.01 percent WO_3 (Van Noy and others, 1986). Minor amounts of gold have documented from all the veins (Van Noy and others, 1986), and disseminated gold has been reported from wallrock near the vein systems (J. Jones, oral commun., 1988).

Teepen (1985) conducted petrographic, geochemical, and geothermometry studies on sulfide minerals from the vein systems in the Empire and Black Rock claims and documented a paragenetic sequence of sulfide mineralization of pyrrhotite>pyrite>sphalerite>chalcopyrite. Teepen disagreed with the analytical results of Van Noy and others (1986) and was unable to detect bismuth-, lead-, arsenic-, or gold-bearing minerals in petrographic, X-ray, or geochemical studies of the Empire and Black Rock claims (fig. 4). This discrepancy is problematic, but the higher sampling density and broader sampling techniques of Van Noy and others (1986) lend credence to their observations.

The Grand Prize Formation and the porphyritic granodiorite are hydrothermally altered and mineralized with finely disseminated sulfide minerals along the vein system in Washington Basin, suggesting post-intrusion mineralization that is probably the result of late-stage circulation of metal-bearing fluids associated with Cretaceous plutonic activity.

The second type of polymetallic veins in the Washington Basin area is along the contact between the Cretaceous porphyritic granodiorite and calc-silicate rocks of the intruded Grand Prize Formation. Intrusion of the granodiorite resulted in contact metamorphism of members 1, 2, and 3 of the Grand Prize Formation and mineralization of both the Paleozoic sedimentary rocks and the granodiorite itself. Examples of this type of veins are in the Black Rock mine in the southwestern part of Washington Basin and in prospects of the Germania group and Arctic group in the head of Germania Basin.

At the Black Rock mine, mineral deposits consist of a northwest-striking, southwest-dipping quartz-pyrite zone (Washington vein) that cuts both granodiorite and the Grand Prize Formation (Van Noy and others, 1986). The mineralized zone is subparallel with the intrusive contact. Sulfide minerals are contained in pods as much as 0.3 m (0.98 ft) in diameter and 0.6 m (2.0 ft) in length in the granodiorite and in tactite zones in the Paleozoic sedimentary rocks (Van Noy and others, 1986). In the granodiorite, the wallrock is sericitized and disseminated pyrite is common; Paleozoic rocks are contact metamorphosed to calc-silicate hornfels. The entire contact zone is heavily iron stained. Sulfide minerals include pyrrhotite, arsenopyrite, pyrite, and minor sphalerite. Assays of the vein material averaged 0.005 ounces of gold per ton, 1.3 ounces of silver per ton, 1.27 percent Pb, and 0.32 percent Sb (Van Noy and others, 1986).

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

SHEAR-ZONE-HOSTED POLYMETALLIC VEINS

Two types of shear-zone-hosted polymetallic veins are in the Washington Basin area: (1) a thick mineralized breccia zone associated with the low-angle structure on Bible Back Mountain and (2) north-trending veins on Bible Back Mountain containing finely disseminated, oxidized sulfide minerals (Van Noy and others, 1986). In addition, Van Noy and others (1986) reported polymetallic veins contained in minor shear zones associated with northeast-trending brittle structures throughout the area.

The most important shear-zone-hosted polymetallic veins in the Washington Basin area are within the low-angle structure exposed on the north face of Bible Back Mountain. The fault zone is extensively mineralized and has been responsible for much of the production from the Washington Basin area (Umpleby, 1915; Van Noy and others, 1986). The brecciated fault zone is 1–1.5 m (3.3–4.9 ft) thick, strikes to the northeast, and dips about 18° to the southeast (Van Noy and others, 1986). The fault zone is highly oxidized, and iron staining is common; visible sulfides are rare. Van Noy and others (1986) reported small, irregular replacement-type ore shoots within the fault zone that consist of small amounts of pyrite, galena, and jamesonite in a quartz and calcite gangue. Samples from the mineralized zone collected by the U.S. Bureau of Mines (Van Noy and others, 1986) averaged 0.06 ounces of gold per ton, 12.6 ounces of silver per ton, 4.27 percent Pb, 1.02 percent Zn, 0.58 percent Cu, and 0.45 percent Sb. The U.S. Bureau of Mines estimated that approximately 8,600 tons of paramarginal resources are present within the old workings on Bible Back Mountain and that an additional 30,000 tons of paramarginal resources may be present in unexplored parts of the fault zone.

The second type of polymetallic vein system consists of north-trending shear zones that cut the brecciated Eagle Creek Member of the Wood River Formation on Bible Back Mountain. The Eagle Creek Member on Bible Back Mountain is pervasively sheared and brecciated and comprises the upper plate of the low-angle structure previously discussed. The shear zones are almost vertical and are generally 0.5–1.5 m (1.6–4.9 ft) wide. Visible sulfide minerals are rarely observed, but the entire mountain is pervasively iron stained (limonite), and oxidized gouge is common in the shear zones. Both quartz and calcite stringers are present. Shear zone samples analyzed by Van Noy and others (1986) averaged 0.08 ounces of gold per ton, 13.7 ounces of silver per ton, 21.0 percent Pb, 0.4 percent Sb, and minor amounts of copper and zinc.

MASSIVE REPLACEMENT DEPOSITS

A small replacement-type deposit is exposed in a bulldozer cut on the north side of Red Hill (fig. 1). The deposit consists of almost complete replacement of massive quartzite, probably member 2 of the Grand Prize Formation, by massive sulfides. The ore zone is highly oxidized and iron stained, and the deposit has a distinct sulfurous odor to it. Within the ore zone, massive sulfide material, including galena and sphalerite, has completely replaced the calcareous cement between the fine to medium sand-sized quartz grains comprising the quartzite. Mineralized veins are absent, although small stringers of quartz are associated with the deposit. The replacement deposit apparently grades into unaltered sedimentary rock, although exposure in the area is poor. The bulldozer cut is exploratory, and no evidence of production exists. The replacement deposit is at the north end of a biotite quartz diorite dike associated with the main biotite quartz diorite on Red Hill and is probably genetically related to that intrusion.

GENESIS AND ORE CONTROLS

The most prominent mineral deposits in the Washington Basin area are in veins that occupy fractures along the axial trace of the asymmetric anticline. The anticline formed as a result of northwest- to southeast-directed compressive stress related to the Cretaceous Sevier orogenic event; northeast-trending fractures formed parallel with the fold hinge. The anticline apparently controlled emplacement of the Cretaceous porphyritic granodiorite, and hydrothermal systems driven by this intrusion remobilized metals from the Paleozoic sedimentary rocks and deposited them within northeast-trending fractures. The intrusive event also resulted in polymetallic vein mineralization in the calcareous Paleozoic country rock. The most probable source of the metals is the Paleozoic Salmon River assemblage, which forms the core of the anticline and is spatially associated with each of the mineralized areas. Lead-isotope studies also suggest that the Salmon River assemblage was a probable source of metals (Sanford and Wooden, this volume).

Mineral deposits in the Washington Basin area formed during at least two separate mineralization events. The earlier event produced polymetallic vein deposits believed to have formed from late-stage hydrothermal fluid circulation associated with the intrusion of Cretaceous granodiorite. These polymetallic veins were emplaced in fold-hinge-parallel fractures or along the contact between the Grand Prize Formation and the Cretaceous granodiorite. A later mineralization event formed replacement deposits within quartzite of the Grand Prize Formation. This event is associated with intrusion of the biotite quartz diorite dike into Cretaceous granodiorite and the Grand Prize Formation on Red Hill. Crosscutting relations clearly show that the biotite quartz diorite dike is younger than the granodiorite, but the timing of the mineralization events is less clear. It is possible that intrusion of the biotite quartz diorite dike and both replacement and polymetallic vein mineralization are genetically related and therefore coeval; however, the presence of polymetallic veins along the granodiorite-Grand Prize Formation contact and pervasive alteration of both granodiorite and Grand Prize Formation throughout the area suggest that the polymetallic vein mineralization was associated with intrusion of the granodiorite. In addition, the lack of polymetallic veins associated with the biotite quartz diorite and the absence of biotite quartz diorite and replacement deposits elsewhere in the area also suggest that the replacement mineralization and dike intrusion postdate an earlier polymetallic vein mineralization event.

The timing of emplacement of the polymetallic shear-zone-hosted veins on Bible Back Mountain is problematic. As stated previously, replacement mineralization on Red Hill is believed to postdate polymetallic vein mineralization associated with the Cretaceous intrusive activity. The temporal relationship between the shear-zone-hosted polymetallic veins on Bible Back Mountain and the other mineral deposit types is difficult to determine. The shear-zone-hosted veins may represent a higher structural level of mineralization associated with Cretaceous intrusive activity, or they may be the result of a younger event associated with younger (Eocene?) high- and low-angle fault displacement. More detailed structural analysis of Bible Back Mountain and dating of alteration minerals are needed to resolve the timing of mineralization in that area.

MINERAL POTENTIAL

The Washington Basin area has potential for minable lead-silver ores in and adjacent to the well-documented northeast-trending mineralized rocks and moderate potential for gold skarn deposits. The U.S. Bureau of Mines (Van Noy and others, 1986) reported that between 25,000 and 100,000 tons of submarginal ore-grade material (1974 prices) may be present in the Washington Basin area and that there is potential for the discovery of minable ore shoots in the area.

Future exploration activities should concentrate on exploiting the prominent northeast-trending vein system in Washington Basin. The localization of veins along fold-hinge fractures suggests that veins continue at deeper levels beneath Germania Basin to the south. Additional polymetallic vein deposits may be present along the contact between porphyritic granodiorite and Paleozoic sedimentary rocks, particularly in Germania Basin. The deposits on Bible Back Mountain are probably of too low grade to warrant further exploration.

REFERENCES CITED

Fisher, F.S., McIntyre D.H., and Johnson, K.M., 1983, Geologic map of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 83–523, scale 1:250,000, 41 p.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

Hall, W.E., and Hobbs, S.W., 1987, Black shale terrane, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 29–42.

Hobbs, S.W., 1985, Precambrian and Paleozoic sedimentary terranes in the Bayhorse area of the Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–D, p. 59–68.

Mahoney, J.B., 1989, Geology and mineral deposits of the Washington Basin area, Custer County, Idaho, *in* Winkler, G.P., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 95–110.

Mahoney, J.B., Link, P.K., Burton, B.R., Geslin, J.K., and O'Brien, J.P., 1991, Pennsylvanian and Permian Sun Valley Group, Wood River Basin, south-central Idaho, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 2, p. 551–580.

Mahoney, J.B., and Sengebush, R.M., 1988, Stratigraphy of the Lower Permian Grand Prize Formation, south-central Idaho, in Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 169–182.

Ross, C.P., 1963, Geology along U.S. Highway 93 in Idaho: Idaho Bureau of Mines and Geology Pamphlet 130, 98 p.

Sengebush, R.M., 1984, The geology and tectonic history of the Fourth of July Creek area, White Cloud Peaks, Custer Country, Idaho: Missoula, University of Montana, M.S. thesis, 79 p.

Teepen, K.L., 1985, Sulfide mineralization in Washington Basin, Idaho: Bowling Green, Ohio, Kent State University, M.S. thesis, 84 p.

Tschanz, C.M., Kiilsgaard, T.H., and Seeland, D.A., 1986, Geology of the eastern part of the Sawtooth National Recreation Area, Idaho, *in* Mineral resources of the eastern part of the Sawtooth National Recreation Area, Custer and Blaine Counties, Idaho: U.S. Geological Survey Bulletin 1545–A, p. 17–43.

Umpleby, J.B., 1915, Ore deposits in the Sawtooth quadrangle, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 580–K, p. 221–249.

Van Noy, R.M., Ridenour, James, Zilka, N.T., Federspiel, F.E., Evans, R.K., Tuchek, E.T., and McMahan, A.B., 1986, Economic appraisal of the eastern part of the Sawtooth National Recreation Area, Idaho: U.S. Geological Survey Bulletin 1545–E, p. 232–472.

Worl, R.G., Kiilsgaard, T.H., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Mineral Deposits at Leadbelt, Lava Creek Mining District, South-Central Idaho

By Sandra J. Soulliere, Anna B. Wilson, *and* Betty Skipp

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–H

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Mineral Deposits at Leadbelt, Lava Creek Mining District, South-Central Idaho

By Sandra J. Soulliere,[1] Anna B. Wilson,[1] *and* Betty Skipp[1]

ABSTRACT

The Leadbelt mines of the Lava Creek mining district of south-central Idaho produced lead, silver, and zinc from polymetallic vein deposits in Mississippian terrigenous and carbonate rocks. Ore is present as discontinuous lenses and isolated clots within veins cutting carbonate rocks. Veins of this type have little or no surface expression in the area but are known to be associated with north-northwest-trending faults or with fractures related to folding.

INTRODUCTION

The Leadbelt mines of the Lava Creek mining district are about 12 mi west of Arco and 3 mi south of Grouse in Butte County, Idaho (fig. 1). The mines are on the southeastern flank of the Pioneer Mountains, south of Antelope Creek and near the heads of Leadbelt and Dry Fork Creeks (fig. 2). In this area, Mississippian terrigenous rocks of the McGowan Creek Formation and carbonate rocks of the Middle Canyon and Scott Peak Formations host polymetallic vein deposits containing lead, silver, and zinc.

The Leadbelt deposits and study area are within the Challis National Forest or on adjacent land administered by the Bureau of Land Management; the area is adjacent to privately owned land on its eastern, western, and southern sides. Access is by unimproved roads along Leadbelt Creek, south of the road along Antelope Creek. A four-wheel-drive road from the Leadbelt mine provides access to Cave Rock (fig. 2). Permission to use private roads on the eastern and southern boundaries of the study area was denied.

Topography of the area is characterized by broad ridges and steep valleys, and elevations range from 6,800 to 9,178 ft. Tributaries of Dry Fork and Leadbelt Creeks drain the area.

[1]U.S. Geological Survey, Denver Federal Center, Denver, Colorado 80225.

Descriptions of the early mine workings and production estimates are given in Umpleby (1917) and Anderson (1929). Worl and others (1989) assessed the potential for undiscovered mineral resources of the Challis National Forest including the Leadbelt Creek area. Skipp and Hait (1977), Skipp and Hall (1980), and Link and others (1988) described the structure and stratigraphy of south-central Idaho. Recent geologic mapping of the Mackay 4 SE and Blizzard Mountain North 7.5-minute quadrangles by Skipp and others (1990) and Skipp and Bollman (1992), respectively, includes the Leadbelt area and was used to prepare figure 3 of this report.

EXPLORATION PERSPECTIVE

Although prospecting began in the area in 1890, significant lead and silver deposits at Leadbelt were not mined until 1905. Only two deposits, the Leadbelt and Butte-Antelope, were extensively developed (Anderson, 1929). From 1905 to 1913, several shipments of lead-silver ore, having an estimated total value of $100,000 (1917 dollars), were hauled from the mines (Umpleby, 1917). Anderson (1929) reported that the Leadbelt mine produced 1,050 tons of ore in 1913. Production values for the deposits from 1913 to 1941 are given in table 1 (R.G. Worl, written commun., 1991).

The portal of the Leadbelt mine is at the head of Leadbelt Creek at an elevation of about 7,200 ft (fig. 2). The stope is reportedly less than 40 ft along the north-trending strike of the vein, which apparently pinched with depth (Anderson, 1929). Principal production from the mine was in 1913 when 21 cars of 50 tons each were shipped; ore averaged 16 percent lead and 16 ounces of silver per ton (Umpleby, 1917). The Leadbelt mine is currently overgrown, although several open shafts were located, and no ore was found on the dump.

The portal of the Butte-Antelope mine is also at the head of Leadbelt Creek, about 0.5 mi southwest of the Leadbelt mine, at an elevation of about 7,250 ft (fig. 2). The old workings consist of open cuts, short tunnels and drifts scattered along the hillside. In 1928, a tunnel was cut

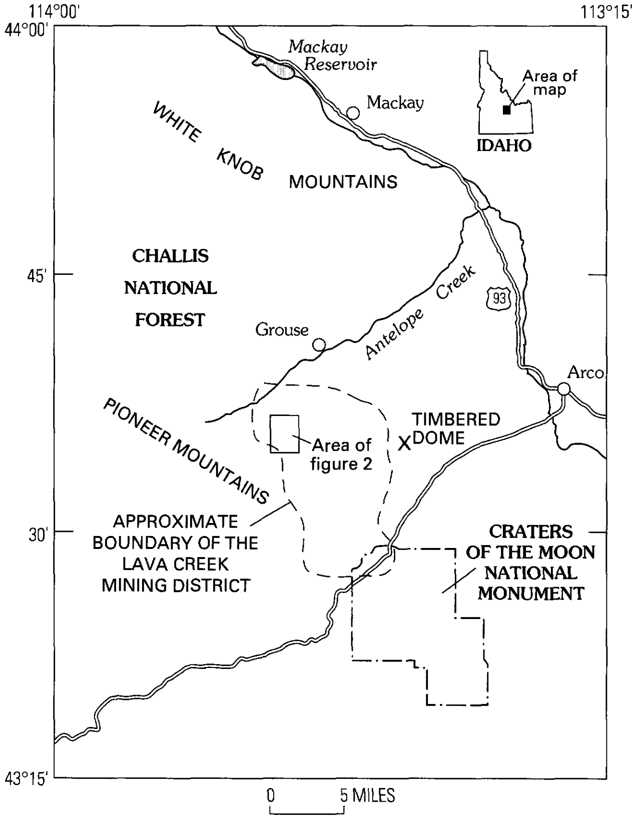


Figure 1. Map showing location of the Leadbelt area, Lava Creek mining district, south-central Idaho.

through the old workings and extended about 300 ft through black carbonaceous shale. Anderson (1929) reported assay values of 31 ounces of silver per ton from selected samples of galena at the mine. A visit to the site in 1987 found the original Butte-Antelope mine workings virtually obliterated by recent excavations. All that remains is a steep cutbank with no sign of dumps or workings.

A number of old prospects and adits are also scattered along the hillside at the head of Dry Fork Creek; no production values are known for these.

At the time of this study in 1987–88, there was active exploration at an adit between the Leadbelt and Butte-Antelope mines. Rehabilitation of the adit and a nearby shaft indicated that a small-scale mining venture was operating in the area.

A mining operation with heap leach pads was underway in 1988 on privately owned land about 2 mi southwest of the Leadbelt mines. This mine is in argillite of the Copper Basin Formation, on the Copper Basin thrust plate (fig. 3), and was not studied by the authors. No other information was available for this site, and permission for access was denied.

Table 1. Production data for deposits in the Leadbelt area, Lava Creek mining district, south-central Idaho.

[Data collected by R.G. Worl (U.S. Geological Survey) from U.S. Bureau of Mines files, Spokane Field Office. Crude ore is in dry tons; Au and Ag in ounces (troy); Cu and Pb in pounds. Zero (0) indicates no production; leaders (--) indicate no recorded production]

Year	Ore	Au	Ag	Cu	Pb
Antelope Creek district					
1935	456	14.00	373	1	68
1939	401	--	201	198	--
1940	814	--	475	339	--
Total	1,671	14.00	1,049	538	68
Unnamed mine, Antelope Creek					
1924	16	--	205	--	774
Mine total	16	--	205	--	774
Butte (Antelope Creek) mine					
1935	40	1.00	519	24	1,763
Mine total	40	1.00	519	24	1,763
Leadbelt mine					
1913	34	--	514	--	8,409
1914	353	6.00	5,619	136	123,810
1915	205	4.00	1,902	--	43,615
1916	83	2.00	1,162	--	23,522
1917	557	22.00	4,788	736	130,074
1918	416	6.00	5,369	2,030	111,038
1919	276	3.00	3,742	995	93,619
1921	0	0	0	0	0
1922	39	1.00	1,242	65	23,767
1923	0	0	0	0	0
1924	24	--	348	--	6,935
1925–28	0	0	0	0	0
1929	57	3.00	511	--	12,031
1941	13	--	121	34	2,217
Mine total	2,057	47.00	25,318	3,996	579,037
District total	3,784	62.00	27,091	4,558	581,642
			Short tons	2.3	290.8

GEOLOGIC SETTING

Geologic units exposed in the Leadbelt area are mainly Mississippian sedimentary rocks unconformably overlain to the northwest by flows of the Eocene Challis Volcanic Group (fig. 3). Several north-northwest- and east-trending faults and folds transect the area.

In south-central Idaho, several major thrust plates separate distinct stratigraphic rock sequences (Skipp and Hait, 1977; Skipp, 1987; Link and others, 1988). Paleozoic sedimentary rocks in the Leadbelt area are part of two such thrust plates, the Copper Basin plate to the west of the Copper Basin thrust fault and the Grouse plate to the east (fig. 3). Both plates were probably emplaced during east-west-directed compression in mid-Cretaceous time (Skipp, 1987). Most of the Paleozoic rocks in the Leadbelt area are on the Grouse thrust plate. The Copper Basin thrust fault, exposed along the western edge of the Leadbelt area (fig. 3), forms

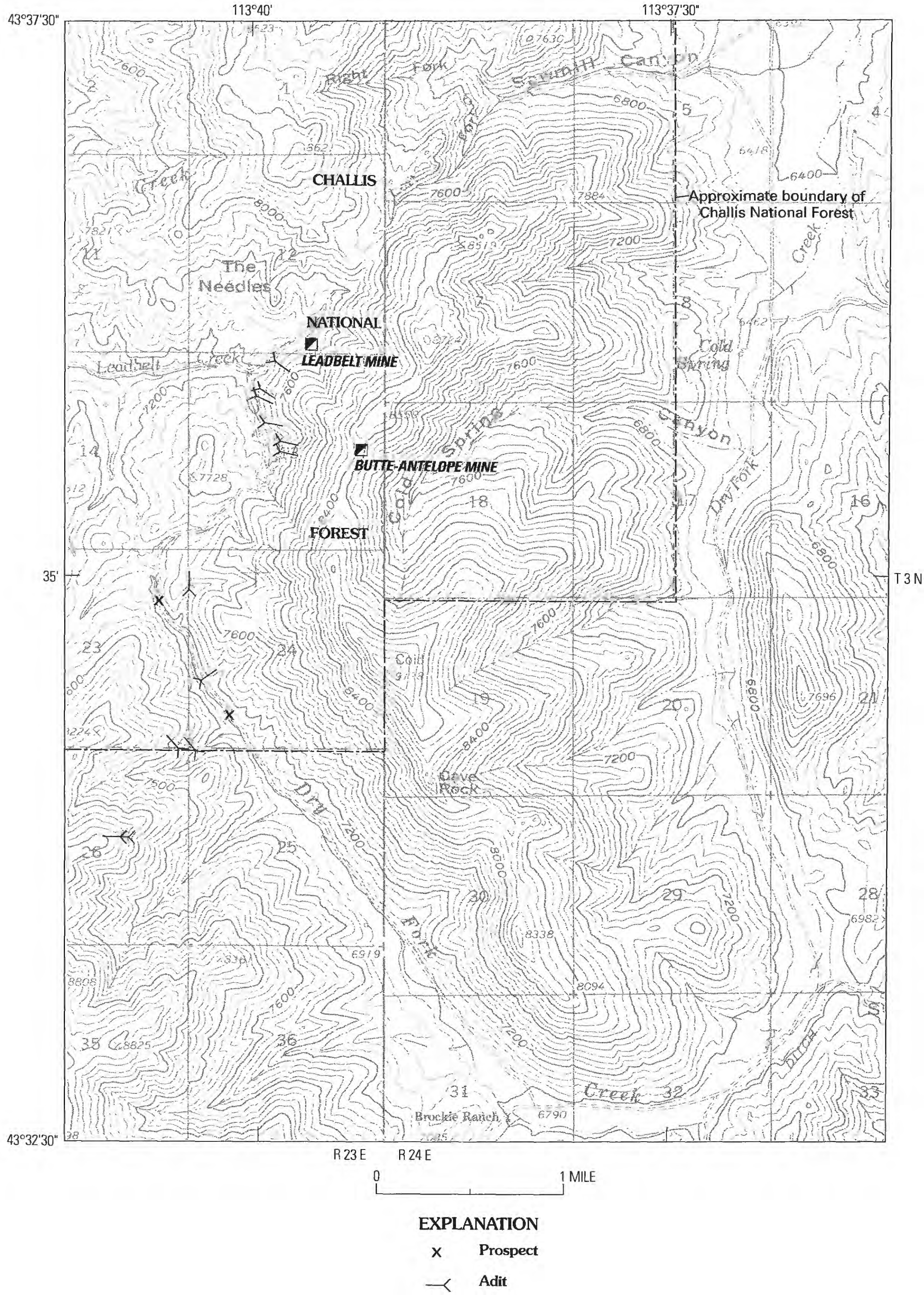


Figure 2. Map showing location of Leadbelt and Butte-Antelope mines, south-central Idaho. Topographic base U.S. Geological Survey Grouse 15-minute quadrangle; contour interval 80 ft.

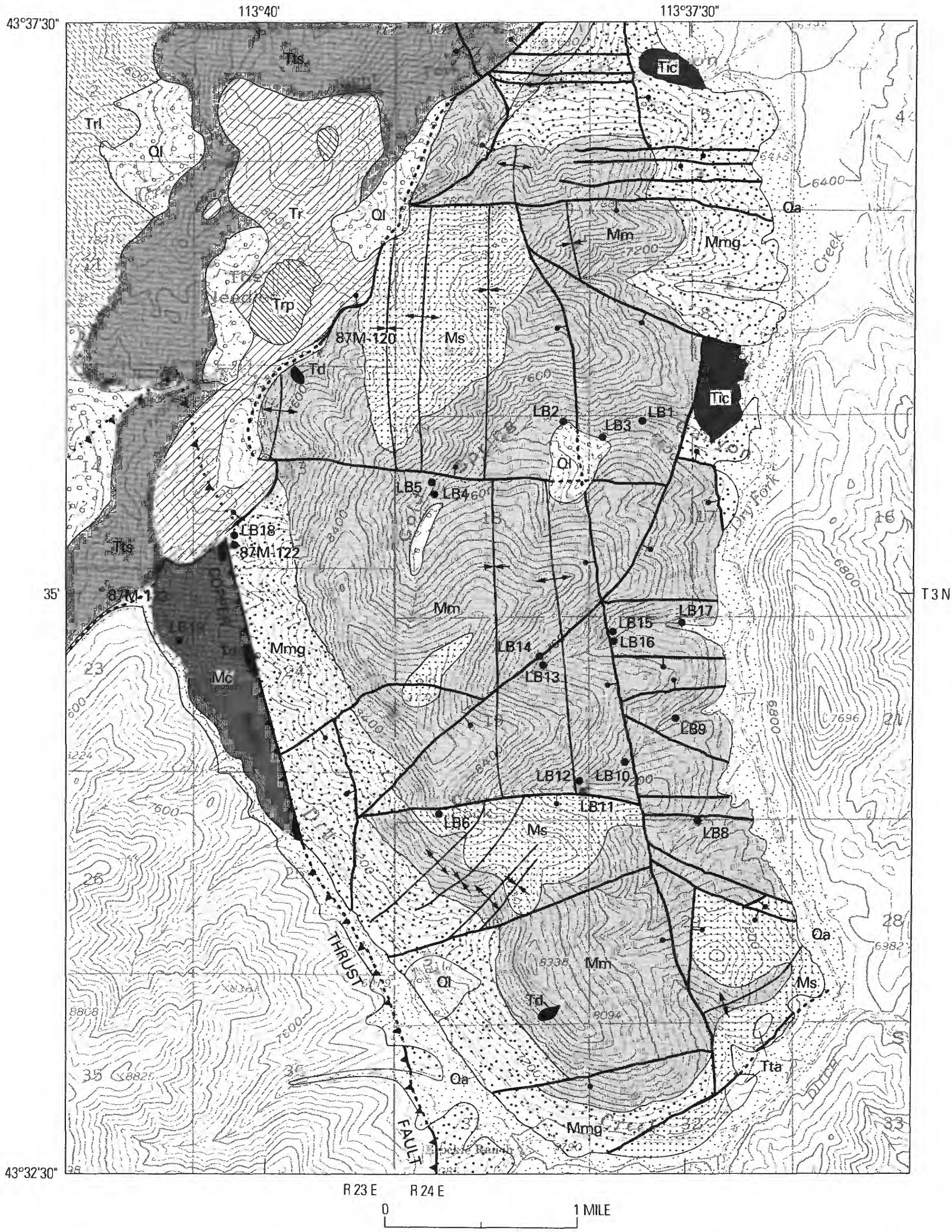
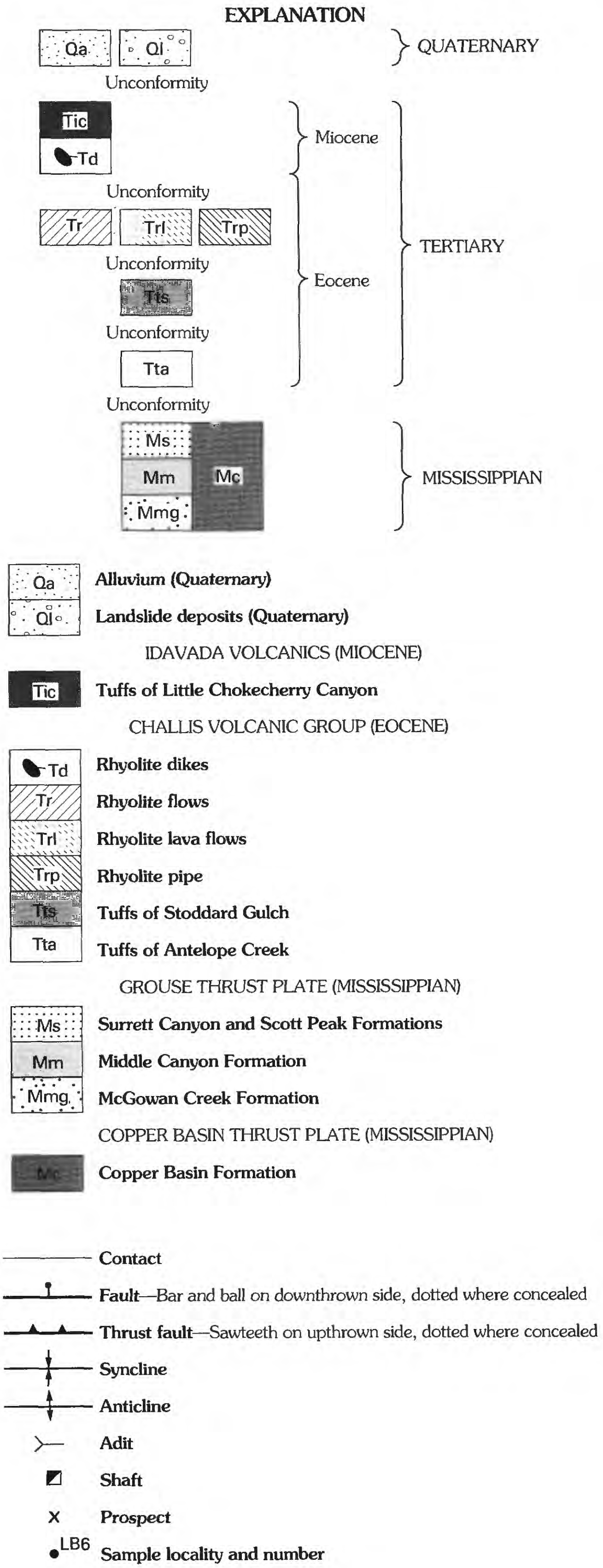


Figure 3 (above and facing column). Map showing geology of the Leadbelt area. Sample localities are also shown. Modified from Skipp and others (1990) and Skipp and Bollmann (1992).



the western boundary of the Grouse plate and juxtaposes Lower Mississippian clastic rocks of the Little Copper Member of the Copper Basin Formation against the Lower Mississippian McGowan Creek Formation (Skipp and others, 1990; Skipp and Bollmann, 1991, unpub. mapping).

In the Grouse plate, carbonate bank and forebank deposits of the Lower and Upper Mississippian Middle Canyon and Upper Mississippian Scott Peak Formations (Skipp and others, 1979) overlie turbidite of the Lower Mississippian McGowan Creek Formation. The McGowan Creek Formation was deposited in the eastern part of a foreland basin (Nilsen, 1977; Skipp and others, 1979). Sedimentary rocks of the McGowan Creek Formation generally dip to the northeast and locally are folded. The Middle Canyon and Scott Peak Formations constitute part of a prograding carbonate-bank and forebank complex.

Late Cretaceous and (or) early Tertiary uplift, erosion, and deposition of conglomerate preceded Eocene Challis volcanism. Rocks of the Challis Volcanic Group are widespread north of the Leadbelt area, where they unconformably overlie Paleozoic sedimentary rocks. In the immediate area of the mines, all contacts between sedimentary rocks and volcanic rocks are faults. The volcanic rocks consist mainly of rhyolitic flows, pipes, and dikes and andesitic tuffs, breccias, and flows. Intrusion of hypabyssal stocks and dikes accompanied Eocene volcanism. Several rhyolite dikes are present in the Leadbelt area, and one dike crops out near the Leadbelt mine. A positive magnetic anomaly northeast of the mines may indicate an intrusive body at depth (Worl and others, 1989).

High-angle faults are widespread in the area and formed during three periods of Tertiary extension (Skipp and others, 1990). Isotope data suggest that Late Cretaceous extension may also have been important in the region (Snee and Kunk, 1989). East-trending faults represent a period of local north-south extension that followed Mesozoic thrusting and preceded Challis volcanism. North-trending rhyolite dikes are considered to be the result of syn-Challis extension (Anderson, 1929; Skipp and others, 1990). Post-Challis extension is represented north of the mines where rocks of the Challis Volcanic Group are faulted against Paleozoic rocks (Skipp and others, 1990).

North- and northwest-trending anticlines and synclines are also widespread. These folds are locally fractured and steep and have limbs that dip 25°–45°. Several north-trending folds are displaced by east-trending faults.

MINERAL DEPOSITS

The deposits in the Leadbelt area are polymetallic veins in carbonate terrane. They are irregularly shaped orebodies of base- and precious-metals in veins along fractures and

Table 2. List of elements and their limits of detection for panned-concentrate samples.

[All values in parts per million except gold, which is in parts per billion. Detection limits are computed for a standard size sample of nominal weight. If the actual weight of sample is greater than the nominal weight of standard, then the limit of detection may be lower; if the acutal weight is lower, then the limit of detection may be higher. Analyses by Geochemical Services, Inc., Sparks, Nevada]

Element	Lower limit	Upper limit
Silver (Ag)	0.015	12,500
Arsenic (As)	1.0	25,000
Gold (Au)	0.5	6,250
Bismuth (Bi)	0.25	12,500
Cadmium (Cd)	0.25	12,500
Copper (Cu)	0.025	50,000
Gallium (Ga)	0.50	12,500
Mercury (Hg)	0.10	6,250
Molybdenum (Mo)	0.10	12,500
Lead (Pb)	0.25	25,000
Antimony (Sb)	0.25	12,500
Selenium (Se)	1.0	100
Thallium (Tl)	0.5	12,500
Zinc (Zn)	1.0	50,000

faults in carbonate rocks. The sulfide minerals—argentiferous galena, pyrite, tetrahedrite, and sphalerite—are present as disseminations, clots, and discontinuous lenses in calcite veins 2–3 ft wide (Anderson, 1929). Gangue minerals in the veins include calcite, rhodochrosite, and, in places, quartz. Individual orebodies vary in size and were mostly worked in the oxide zones where iron and manganese oxide minerals, cerussite, and cerargyrite are common.

Mines and prospects of the Leadbelt area are in or near terrigenous rocks of the McGowan Creek Formation and carbonate rocks of the Middle Canyon and Scott Peak Formations (fig. 3). The McGowan Creek Formation (Lower Mississippian) consists of distal turbidite, calcareous siltstone, and minor silty limestone (Sandberg, 1975). The Middle Canyon Formation (Lower and Upper Mississippian) overlies the McGowan Creek and consists of medium-gray, fine- to medium-grained, spiculitic, thin- to medium-bedded, silty limestone (Skipp and others, 1990). Thin, noncalcareous siltstone beds are present in the basal few feet. Medium- to dark-gray, thick-bedded, variably cherty, fossiliferous limestone of the Scott Peak Formation (Upper Mississippian) overlies the Middle Canyon Formation. Fossils in the Scott Peak are varied and abundant and include corals, brachiopods, mollusks, bryozoans, algae, calcareous foraminifers, trilobites, ostracodes, and shark teeth (Skipp and others, 1979). Locally, all of these units are cut by dikes.

Host rocks generally are not visibly altered; a narrow zone of calcite, rhodochrosite, and calc-silicate minerals is present along some of the veins (Umpleby, 1917; Anderson, 1929). Primary ore is locally oxidized. Because the underground workings are inaccessible and the mineralized zones are not exposed, it was not possible to examine alteration at the mineralized zones. Rock samples were collected and examined at mine dumps and prospect pits.

Previous studies (Anderson, 1929; Worl and others, 1989) suggest that the location of orebodies is controlled by fractures and faults in favorable carbonate host rocks. Ore-bearing veins at the Leadbelt and Butte-Antelope mines fill fractures that trend north and dip steeply west (Anderson, 1929). These fractures may be related to folding of the nearby anticline (fig. 3). The Butte-Antelope mine adit follows an east-trending high-angle fault (fig. 3), and the intersection of northwest- and east-trending faults may also be an important control for deposition of ore minerals. Some deposits are along bedding-plane faults and fractures in folded carbonate rocks. The Leadbelt mine adit is on the western flank of a north-trending anticline (fig. 3).

Analyses of mineralized rocks provide useful geochemical information about the trace-element assemblages associated with a mineralizing solution. In 1987, four rock-chip samples of mineralized rock were collected from mine dumps and prospect pits in the Leadbelt area (fig. 3). At each site a 5-lb bag of rock chips and one or more hand samples were collected. The samples were analyzed for 34 elements using the semiquantitative emission spectrographic methods of Golightly and others (1987).

In 1988, 17 stream-sediment samples were collected in a reconnaissance-scale survey of tributary streams that drain the area (fig. 3). Permission to collect samples on private land south and east of the mines was denied. At each sample site a 10-lb bag of bulk stream sediment was collected. Stream sediment was then panned in the field to a more concentrated form and analyzed in the laboratory for 14 elements using inductively coupled emission spectrographic methods. Table 2 lists the elements and their detection limits. Analyses of stream-sediment samples are useful in identifying those drainage basins that contain concentrations of elements that may be derived from mineral deposits.

Zinc, copper, lead, arsenic, gold, silver, manganese, and barium are elements common to polymetallic vein deposits regardless of host rock type (Cox and Singer, 1986). Many of these elements were detected in rock-chip and panned-concentrate samples from the Leadbelt area.

Table 3 presents the results of the analyses of rock samples collected from mine dumps and prospect pits. Rock-chip sample LB6, collected from outcrop in a small prospect pit near Cave Rock (fig. 3), had the highest concentrations of Ag (1.95 ppm), Au (0.005 ppm), Cd (15.5 ppm), Mo (33.1 ppm), Sb (31.8 ppm), Zn (334 ppm), Tl (7 ppm), and Se (97.4 ppm).

The suite of elements Ag, As, Cu, Hg, Mo, Pb, Sb, Zn, Cd, Ga, and Se was detected in a majority of the panned-concentrate samples (table 4). Sample LB13, collected downstream from a northeast-trending fault in the Middle

Table 3. Results of geochemical analyses of rock samples from the Leadbelt area, Lava Creek mining district, south-central Idaho.
[Ca, Fe, Mg, Na, P, and Ti in percent; all other elements in parts per million. Lower limit of detection is shown in parentheses above element symbol. L indicates less than lower limit of detection; N indicates not detected; B indicates that the sample was not analyzed for this element. Sample LB6 was analyzed by Geochemical Services, Inc.; detection limits are given in table 2]

	(0.05) Ca	(0.05) Fe	(0.02) Mg	(0.2) Na	(0.002) Ti	(0.5) Ag	(200) As	(10) Au	(10) B	(20) Ba	(1) Be	(10) Bi
87M–120	10	2	0.7	0.5	0.03	1.5	1500	N	L	200	N	N
87M–122	L	1.5	0.5	N	0.2	0.7	N	N	70	2000	L	N
87M–123	N	0.5	0.3	N	0.15	1	N	N	30	700	N	N
LB6	B	B	B	B	B	1.95	332	0.005	B	B	B	<2.35

	(20) Cd	(10) Co	(10) Cr	(5) Cu	(5) Ga	(10) Ge	(50) La	(10) Mn	(5) Mo	(20) Nb	(5) Ni	(10) Pb
87M–120	N	N	N	10	15	L	N	300	N	N	5	150
87M–122	N	N	N	70	30	5	N	N	N	N	10	N
87M–123	N	N	N	20	10	5	N	15	N	L	5	20
LB6	15.5	B	B	35.0	<4.71	B	B	B	33.1	B	B	<2.35

	(100) Sb	(5) Sc	(10) Sn	(100) Sr	(100) Th	(10) V	(20) W	(10) Y	(200) Zn	(10) Zr	Tl	Se
87M–120	N	N	N	200	N	20	N	N	200	10	B	B
87M–122	N	L	N	N	N	200	N	N	N	50	B	B
87M–123	N	N	N	N	N	150	N	N	N	70	B	B
LB6	31.8	B	B	B	B	B	B	B	334	B	7.00	97.4

Canyon Formation, had the highest concentrations of As (123 ppm), Cu (32.1 ppm), Mo (28.7 ppm), Zn (494 ppm), and Se (5.98 ppm). Sample LB1 was collected downstream from the intersection of east-trending and northeast-trending faults in Middle Canyon Formation and has the highest Hg (1.73 ppm) and Pb (46.7 ppm) concentrations. The highest concentrations of Sb (18.80 ppm) and Cd (2.5 ppm) are from sample LB5, which was collected downstream from the intersection of east- and north-trending faults in Middle Canyon Formation. No significant concentrations of silver were detected in the panned-concentrate samples. The highest concentration of Ag (0.286 ppm) was from sample LB19, collected downstream from a mineralized prospect pit. Table 5 presents a summary of analytical results for panned-concentrate samples.

Correlation analysis of the panned-concentrate data shows Mo-As-Zn-Cu and Sb-Cd associations (correlation coefficient >0.90), a Cu-As-Mo-Zn-Sb-Cd-Ga association (correlation coefficient >0.85), and a Cu-As-Mo-Zn-Sb-Cd-Ga-Pb-Hg association (correlation coefficient >0.75) (table 6).

No geophysical signatures specific to polymetallic replacement vein deposits have been identified in the Leadbelt area. Regional aeromagnetic surveys of the Hailey and Idaho Falls 1°×2° quadrangles provide criteria for mapping buried intrusive complexes, delineating major fault and shear zones, and identifying areas that might be hydrothermally altered or mineralized (Kleinkopf and others, 1988a, b). A belt of high-intensity magnetic gradients, known as the Great Rift shear zone, trends northwest from about the Craters of the Moon National Monument to Sun Valley (Kleinkopf and others, 1989). The Leadbelt area is in the center of this shear zone, and a magnetic high or positive anomaly is northeast of the mines (Worl and others, 1989). This magnetic high may indicate a buried intrusive body at depth that could represent a thermal source for a regional hydrothermal system. In addition, parallel magnetic contours east of the magnetic high indicate a north-northwest-trending fault and fold zone through the Leadbelt area.

In simplistic terms, ore deposition was from hydrothermal fluids probably associated with Eocene igneous intrusive activity. Three possible sources for the fluid are (1) hydrothermal water from magma at depth, (2) convecting meteoric water, or (3) connate water from the Paleozoic rocks. Fractures and faults formed by extensional tectonism provided conduits for mineralizing solutions. Calcite, silica, magnesium carbonate minerals, and lead-, zinc-, and silver-bearing sulfide minerals were deposited as veins in carbonate rocks.

The influence of local intrusive activity on the formation of the base-metal deposits is uncertain. Intrusive activity at depth may have been the source of a regional hydrothermal system that either supplied the metals or acted as a heat source for transport of metals from sedimentary rocks. Lead-isotope data from vein deposits near Leadbelt (Sanford and others, 1989; Sanford and Wooden, this volume) suggest that either the Tertiary intrusive bodies or Precambrian basement rocks may be the source of the lead; however, until additional analyses are performed, the source of the metals at Leadbelt is uncertain.

Table 4. Results of geochemical analysis of panned-concentrate samples from the Leadbelt area, Lava Creek mining district, south-central Idaho.

[Samples were also analyzed for Tl, Bi, Pd, and Te, but all values were below the limits of detection. All values are in parts per million; L indicates a value less than the detection limit shown]

Sample number	Ag	As	Au	Cu	Hg	Mo	Pb	Sb	Zn	Cd	Ga	Se
LB1	0.109	71.9	0.001	30.6	1.73	13.8	46.7	9.66	436	1.89	4.25	1.34
LB2	0.049	26.1	0.0005L	12.2	0.488	4.64	15.5	3.70	176	0.780	2.91	0.967L
LB3	0.074	21.6	0.0007L	9.49	0.210	2.77	11.8	4.80	144	0.706	0.832	1.36L
LB4	0.160	30.2	0.001	19.2	0.095L	4.89	14.0	4.71	207	0.493	1.62	2.59
LB5	0.098	98.6	0.001	28.9	0.426	12.2	29.6	18.8	447	2.50	4.31	5.40
LB8	0.047	35.0	0.002	24.6	0.319	4.64	16.8	4.46	239	1.03	2.77	2.60
LB9	0.064	46.0	0.0007L	18.4	0.138	8.22	25.4	6.64	236	1.37	3.96	1.98
LB10	0.107	66.3	0.0005L	28.5	0.314	13.5	20.3	7.86	343	1.48	4.14	2.05
LB11	0.084	74.0	0.0006L	22.7	0.430	8.55	22.5	8.68	273	1.31	2.78	1.41
LB12	0.140	86.0	0.0007L	28.1	0.230	13.4	25.6	10.6	395	1.62	2.98	5.58
LB13	0.084	123.0	0.0007L	32.1	0.322	28.7	29.4	12.1	494	1.76	4.36	5.98
LB14	0.071	44.7	0.0005L	20.0	0.325	7.31	19.1	5.48	237	1.07	2.17	2.28
LB15	0.113	53.9	0.0004L	25.8	0.859	7.03	21.6	6.35	304	0.928	2.00	2.45
LB16	0.126	102.0	0.001	30.8	0.537	21.0	32.7	11.2	470	2.01	4.24	2.60
LB17	0.087	47.6	0.001	19.4	0.549	12.1	24.0	6.56	286	1.24	2.99	2.68
LB18	0.146	34.2	0.0006L	22.3	0.119	6.17	16.3	7.06	259	0.722	1.81	4.34
LB19	0.286	16.1	0.001	15.4	0.097L	1.66	10.5	2.37	125	0.379	1.18	2.06

Table 5. Summary of analytical results of panned-concentrate samples from the Leadbelt area, Lava Creek mining district, south-central Idaho

[In parts per million. Samples were analyzed for Tl, Bi, Pd, and Te, but all values were below the limits of detection]

	Minimum value detected	Maximum value detected	Mean of values	Deviation of values	Number of samples having detectable values
Ag	0.047	0.29	0.11	0.056	17
As	16.1	123.0	57.48	31.14	17
Au	<0.0004	0.002	0.0008	0.0004	10
Cu	9.49	32.10	22.85	6.7	17
Hg	<0.095	1.73	0.423	0.39	15
Mo	1.66	28.70	10.03	6.89	17
Pb	10.50	46.70	22.46	8.93	17
Sb	2.37	18.80	7.666	3.989	17
Zn	125.0	494.0	298.9	115.76	17
Cd	0.379	2.50	1.252	0.577	17
Ga	0.832	4.360	2.900	1.168	17
Se	<0.967	5.980	2.805	1.557	15

Table 6. Correlation coefficients for each pair of selected elements for panned-concentrate samples from the Leadbelt area, Lava Creek mining district, south-central Idaho

[Data analyzed by using USGS STATPAC statistical computer package (Van Trump and Miesch, 1976) to determine element associations. These elements were selected because they were detected in more than 50 percent of the samples. Results of geochemical analysis and summary of results are given in tables 4 and 5]

Ag	1.00											
As	–0.17	1.00										
Au	0.03	0.03	1.00									
Cu	–0.02	0.85	0.25	1.00								
Hg	–0.18	0.26	0.04	0.40	1.00							
Mo	–0.16	0.91	0.03	0.77	0.24	1.00						
Pb	–0.21	0.72	0.06	0.73	0.75	0.68	1.00					
Sb	–0.14	0.87	0.05	0.73	0.22	0.68	0.67	1.00				
Zn	–0.14	0.95	0.10	0.92	0.43	0.89	0.83	0.86	1.00			
Cd	–0.30	0.88	0.05	0.76	0.39	0.74	0.83	0.92	0.89	1.00		
Ga	–0.36	0.78	0.07	0.72	0.38	0.77	0.79	0.71	0.81	0.87	1.00	
Se	0.11	0.60	0.13	0.55	–0.26	0.54	0.19	0.66	0.58	0.44	0.28	1.00
	Ag	As	Au	Cu	Hg	Mo	Pb	Sb	Zn	Cd	Ga	Se

SUMMARY

Silver, lead, and zinc production in the Lava Creek mining district of south-central Idaho was from small isolated clots and discontinuous lenses in veins cutting carbonate rocks. Veins of this type have little or no surface expression in the area but are known to be associated with north-northwest-trending faults or with fractures related to folding. Exploration for hidden orebodies should therefore concentrate on the structural fabric of the area and favorable host rocks. Further detailed geochemical and geophysical surveys would also be useful in locating hidden orebodies. Detailed electrical surveys would facilitate detection of individual orebodies containing conductive sulfide minerals. Isotope data for ore in the Leadbelt area would help determine the source of the metals in these deposits.

REFERENCES CITED

Anderson, A.L., 1929, Geology and ore deposits of the Lava Creek district, Idaho: Idaho Bureau of Mines and Geology Pamphlet 32, 70 p.

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

Golightly, D.W., Dorrzapf, A.F., Jr., Mays, R.E., Fries, T.L., and Conklin, N.M., 1987, Analysis of geologic materials by direct-current arc emission spectrography and spectrometry, *in* Baedecker, P.A., ed., Methods for geochemical analysis: U.S. Geological Survey Bulletin 1770, p. A1–A13.

Kleinkopf, M.D., Bankey, Vicki, and McCafferty, A.E., 1988a, Aeromagnetic and gravity anomaly maps of the Hailey and part of the Idaho Falls two-degree quadrangles, Idaho: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 424–425.

———1988b, The application of regional geophysical data in mineral resource assessments in the Hailey and Idaho Falls quadrangles, Idaho: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 425.

———1989, Gravity and magnetic anomaly patterns applied to mineral resource exploration, Hailey 1°×2° quadrangle, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 26–32.

Link, P.K., Skipp, Betty, Hait, M.H., Jr., Janecke, Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 5–42.

Nilsen, T.H., 1977, Paleogeography of Mississippian turbidites in south-central Idaho, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 1, p. 275–299.

Sandberg, C.A., 1975, McGowan Creek Formation, new name for Lower Mississippian flysch sequence in east-central Idaho: U.S. Geological Survey Bulletin 1405–E, 11 p.

Sanford, R.F., Wooden, J.L., Doe, B.R., and Worl, R.G., 1989, Origin of ore deposits in the central Idaho black shale belt—Implications from lead isotopes, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 39–52.

Skipp, Betty, 1987, Basement thrust sheets in the Clearwater orogenic zone, central Idaho and western Montana: Geology, v. 15, p. 220–224.

Skipp, Betty, and Bollmann, D.D., 1992, Geologic map of Blizzard Mountain North quadrangle, Blaine and Butte Counties, Idaho:

U.S. Geological Survey Open-File Report 92–280, scale 1:24,000.

Skipp, Betty, and Hait, M.H., Jr., 1977, Allochthons along the northeast margin of the Snake River Plain, Idaho: Wyoming Geological Association Guidebook, Twenty-Ninth Annual Field Conference, p. 499–515.

Skipp, Betty, and Hall, W.E., 1980, Upper Paleozoic paleotectonics and paleogeography of Idaho, *in* Fouch, T.D., and Magathan, E.R., eds., Paleozoic paleogeography of the West-Central United States: Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, West-Central United States Paleogeography Symposium 1, p. 387–422.

Skipp, Betty, Kuntz, M.A., and Morgan, L.A., 1990, Geologic map of Mackay 4 (Grouse) SE quadrangle, Butte County, Idaho: U.S. Geological Survey Open-File Report 89–431, scale 1:24,000.

Skipp, Betty, Sando, W.J., and Hall, W.E., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States–Idaho: U.S. Geological Survey Professional Paper 1110–AA, p. AA1–AA42.

Snee, L.W., and Kunk, M.J., 1989, $^{40}Ar/^{39}Ar$ thermochronology of mineral deposits in the southern part of the Idaho batholith, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 37–38.

Umpleby, J.B., 1917, Geology and ore deposits of the Mackay region, Idaho: U.S. Geological Survey Professional Paper, scale 1:375,000, 129 p.

Van Trump, George, Jr., and Miesch, A.T., 1976, The U.S. Geological Survey RASS-STATPAC system for management and statistical reduction of geochemical data: Computers and Geosciences, v. 3, p. 475–488.

Worl, R.G., Wilson, A.B., Smith, C.L., Kleinkopf, M.D., and Sykes, R.C., 1989, Mineral resource potential and geology of the Challis National Forest, Idaho: U.S. Geological Survey Bulletin 1873, 101 p.

Mineral Deposits of the Mackay and Copper Basin Mineralized Areas, White Knob Mountains, South-Central Idaho

By Anna B. Wilson, Sandra J. Soulliere, *and* Betty Skipp

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–I

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Mineral Deposits of the Mackay and Copper Basin Mineralized Areas, White Knob Mountains, South-Central Idaho

By Anna B. Wilson,[1] Sandra J. Soulliere,[1] *and* Betty Skipp[1]

ABSTRACT

Between 1879 and 1982 about 1 million tons of ore were produced from skarn deposits and associated replacement veins near Mackay and Copper Basin, Idaho. The deposits contain copper, lead, zinc, silver, and gold and formed during contact metamorphism and metasomatism of the Mississippian White Knob Limestone by intrusion of the Eocene Mackay stock and related rocks. Field observations and statistical analysis of geochemical data indicate that the deposits are typical of zoned copper skarns. Copper-gold-silver is the geochemical signature of the inner zone, grading outward to gold-silver, and then to lead-zinc-silver in the outer zone. Deposits of the inner zone typically are in metamorphosed and metasomatized limestone along the contacts of the Mackay Granite and leucogranite porphyry phases of the Mackay stock, and copper is the chief product. In the outer zone, the limestone is generally unaltered, contact metamorphism is limited, and lead, zinc, and silver are the commodities mined.

In 1988 the area was explored for sediment-hosted, jasperoid-associated, precious-metal deposits and for hot-springs-type, volcanic-hosted, disseminated gold deposits.

INTRODUCTION

Polymetallic skarn deposits are prevalent in the Mackay and Copper Basin mineralized areas, southwest of the town of Mackay, Idaho (fig. 1). These copper, lead, zinc, silver, and gold deposits of probable Paleogene age were formed by replacement of carbonate-bearing rocks during contact metamorphism and metasomatism. The Upper Mississippian White Knob Limestone adjacent to the Eocene Mackay stock (fig. 2) is the host rock. Total production for the Mackay and Copper Basin mineralized areas is estimated to have been about 1 million tons of ore (table 1). Potential for as yet undiscovered, sediment-hosted, jasperoid-associated, precious-metal deposits near the Mackay stock is high.

The area described herein extends from the town of Mackay to Copper Basin (fig. 1) in a wedge-shaped northeast-trending block, approximately 13 mi long by 2½ mi wide at the northeast end and 8 mi wide at the southwest end (fig. 2). It contains the highest peaks in the central White Knob Mountains including Shelly (11,278 ft), Redbird (11,273 ft), Cabin (11,244 ft), and Lime (11,179 ft) Mountains, White Knob (10,529 ft), and Mackay Peak (10,273 ft). The lowest elevations in the area are 6,000 ft near Mackay and 7,700 ft near Copper Basin. Both the Alder Creek mining district and the eastern part of the Copper Basin mining district are in the study area (fig. 1).

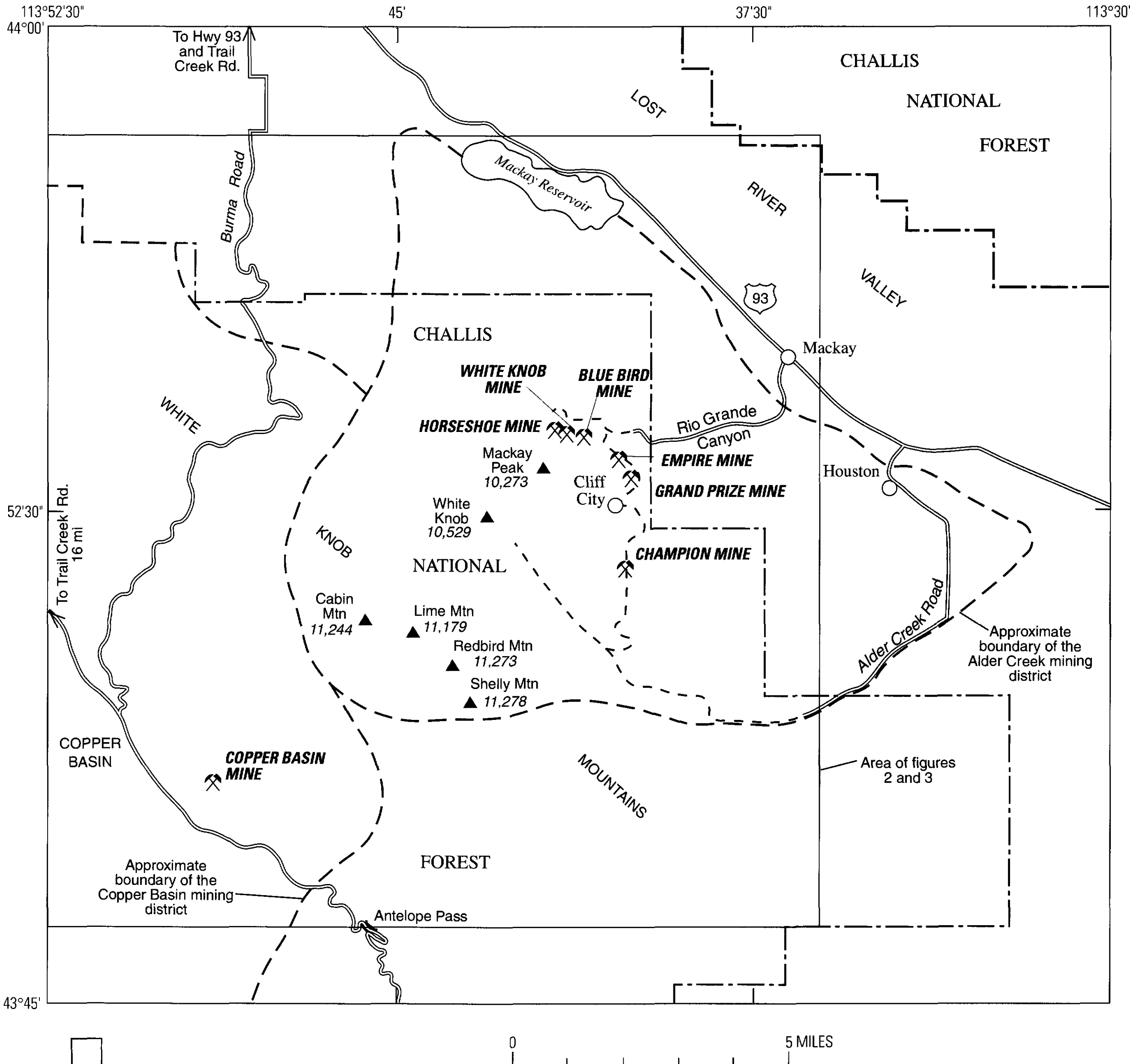
The study area is within or immediately adjacent to the Challis National Forest. Access to the Alder Creek mining district is by graded roads from Mackay via Rio Grande Canyon and Alder Creek and by various four-wheel-drive roads near the mines. Access to the Copper Basin area is by graded Forest Service roads from Antelope Pass, Trail Creek Road, or Burma Road. Other than old four-wheel-drive roads to the Copper Basin mine, no roads access the western part of the study area east of the mine.

PREVIOUS WORK

Nelson and Ross (1968, 1969a) prepared geologic maps of the Mackay and Copper Basin mineralized areas at 1:24,000 and 1:125,000 scales. Skipp and Hait (1977), Skipp and Harding (1985), and Wilson and Skipp (1994) updated and reinterpreted the geology as previously presented but at a much smaller scale. Unpublished detailed maps of several mines by other authors included in the Anaconda Geological Documents Collection (International Archive of Economic Geology, American Heritage Center, University of Wyoming, Laramie) were useful.

Kemp and Gunther (1907) and Umpleby (1917) published the first comprehensive studies of the ore deposits in the Mackay region. Other studies include those of Ross (1930), Nelson and Ross (1968, 1969a, b), and Leland

[1]U.S. Geological Survey, Denver Federal Center, Denver, Colorado 80225.



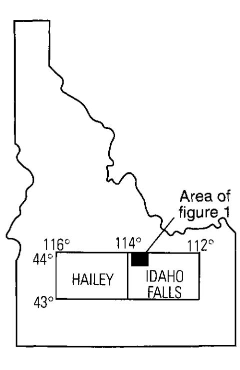
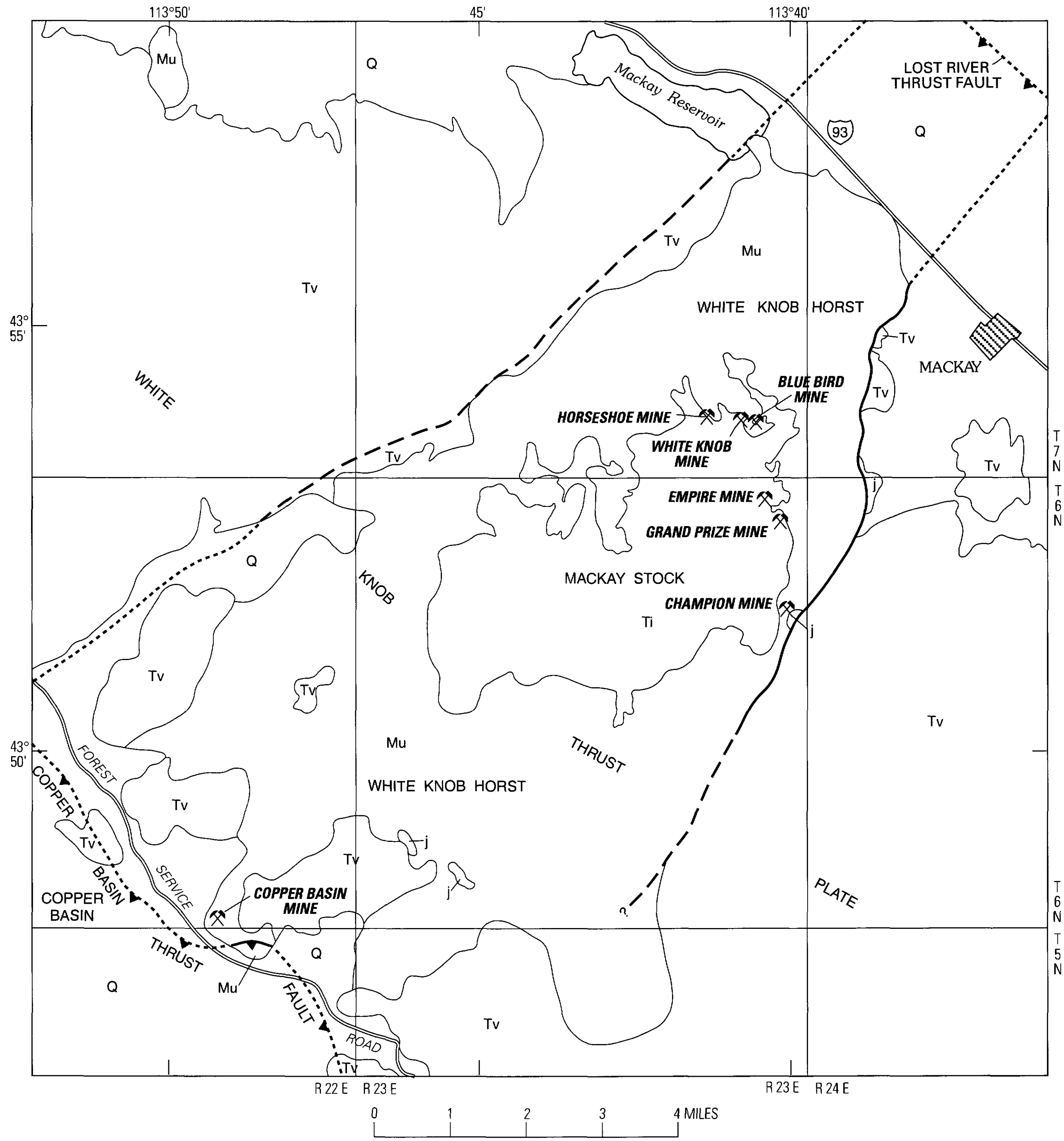


Figure 1. Map showing approximate boundaries of the Copper Basin and Alder Creek mining districts, location and elevation (in feet) of prominent mountains, access roads to the mines, and mines mentioned in the text, Mackay and Copper Basin mineralized areas, White Knob Mountains, Idaho.



EXPLANATION

Q	Quaternary deposits, undifferentiated
j	Jasperoid
Ti	Mackay stock intrusive rocks (Eocene)
Tv	Challis Volcanic Group (Eocene)
Mu	Sedimentary rocks, undifferentiated (Mississippian)

Contact

Fault—Dashed where inferred, dotted where concealed, and queried where unknown

Thrust fault—Dotted where concealed; sawteeth on upper plate

Mine

Figure 2. Map showing simplified geology, structural elements, and location of mines, Mackay and Copper Basin mineralized areas, White Knob Mountains, Idaho. Geology modified from Nelson and Ross (1968, 1969a) and Dover (1981).

Table 1. Production data for mines in the Mackay and Copper Basin mineralized areas, White Knob Mountains, Idaho.
[Compiled from Ross (1930), Farwell and Full (1944), Leland (1957), and Nelson and Ross (1968). Asterisk (*) indicates summary of data collected by R.G. Worl and T.H. Kiilsgaard (U.S. Geological Survey) from U.S. Bureau of Mines files, Spokane Field Office. Values for the Champion mine were computed from assay data and tonnage reported by Leland (1957). Crude ore is in dry tons; Au and Ag in ounces (troy); Cu, Pb, and Zn in pounds. Leaders (--) indicate no value on record. Total for Alder Creek mining district does not include the Copper Basin mine]

Year	Ore	Au	Ag	Cu	Pb	Zn
			Copper Basin mine			
*1912	60	0.60	163	13,654	--	--
*1915–19	5,685	88.96	16,064	682,406	2,662	--
*1953	429	13.00	1,758	27,344	--	--
Total	6,174	102.56	17,985	723,404	2,662	--
			Blue Bird mine			
*1924–28	571	8.89	5,472	574	135,038	3,111
*1935–39	365	7.55	4,525	90	143,931	--
Total	936	16.44	9,997	664	278,969	3,111
			Champion mine			
*1908	26	--	126	--	2,812	--
1938	18	0.23	152	234	4,608	504
1943	339	3.39	993	2,373	86,106	20,340
1944	200	2.00	532	1,400	42,000	13,200
1947	65	0.81	954	1,105	13,650	6,890
1948	161	1.21	576	386	42,504	10,304
1949	106	1.59	212	1,102	18,656	4,452
1950	--	--	--	--	--	--
1951	481	4.81	1,443	3,175	126,984	28,860
1952	107	0.93	251	835	22,042	6,206
1953	--	--	--	--	--	--
1954	617	5.55	1,462	8,761	159,186	34,552
1955	459	2.30	1,193	1,102	115,668	24,786
Total	2,579	22.82	7,668	20,473	634,216	150,094
			Horseshoe mine			
1916	196	0.88	1,468	3,093	55,587	--
1917	1,462	6.99	17,293	2,997	678,074	--
1918	1,087	8.64	11,440	1,522	446,628	--
1919	1,128	4.76	12,011	2,333	361,937	--
1920	2,319	5.96	21,165	2,021	386,352	--
1921	116	2.25	1,811	376	55,902	--
1922	651	7.61	9,297	902	246,659	--
1923	660	6.98	5,014	1,603	169,872	--
1924	79	4.33	1,012	287	36,538	--
1925	75	1.50	1,776	348	59,310	--
1926	865	11.72	10,236	3,198	332,597	18,942
1927	182	1.52	2,275	495	58,256	18,163
1928	49	0.43	123	1,159	791	2,366
*1936–51	1,691	20.32	21,366	6,211	837,355	79,822
*1954	31	--	233	100	11,200	--
*1970–71	3,271	36.00	13,210	12,100	280,700	330,420
Total	13,862	119.89	129,726	38,745	4,017,758	449,713

Table 1. Production data for mines in the Mackay and Copper Basin mineralized areas, White Knob Mountains, Idaho—Continued.

[Compiled from Ross (1930), Farwell and Full (1944), Leland (1957), and Nelson and Ross (1968). Asterisk (*) indicates summary of data collected by R.G. Worl and T.H. Kiilsgaard (U.S. Geological Survey) from U.S. Bureau of Mines files, Spokane Field Office. Values for the Champion mine were computed from assay data and tonnage reported by Leland (1957). Crude ore is in dry tons; Au and Ag in ounces (troy); Cu, Pb, and Zn in pounds. Leaders (--) indicate no value on record. Total for Alder Creek mining district does not include the Copper Basin mine]

Year	Ore	Au	Ag	Cu	Pb	Zn
			Empire mine			
1902	1,721	14.00	607	14,966	--	--
1903	15,681	241.00	12,658	441,286	--	--
1904	67,850	85.00	3,500	2,700,000	--	--
1905	13,000	385.00	22,065	684,134	--	--
1906	40,838	1,842.00	71,854	2,807,926	--	--
1907	37,141	1,823.00	70,222	2,895,881	--	--
1908	382	16.00	673	38,698	--	--
1909	1,436	28.00	2,236	90,347	--	--
1910	7,206	265.00	28,754	919,492	--	--
1911	11,057	663.00	40,900	1,415,314	--	--
1912	26,227	1,766.00	69,942	2,854,281	--	--
1913	35,950	1,892.00	106,463	3,962,125	--	--
1914	17,801	971.00	59,243	2,106,441	--	--
1915	54,295	3,155.00	125,134	4,702,119	--	--
1916	69,907	2,875.00	123,453	5,006,291	--	--
1917	66,808	2,530.00	74,645	4,208,401	--	--
1918	53,211	2,476.00	56,014	3,404,161	--	--
1919	12,904	673.00	31,833	1,300,518	--	--
1920	15,755	1,369.00	29,888	1,480,678	--	--
1921	9,992	1,236.00	23,354	1,088,148	--	--
1922	16,717	2,019.00	33,988	1,843,200	--	--
1923	15,791	1,458.00	25,908	1,449,838	--	--
1924	11,775	1,245.00	18,808	1,137,771	--	--
1925	24,753	2,096.00	35,439	2,352,306	--	--
1926	3,635	234.00	6,453	239,785	--	--
1927	13,627	761.00	9,734	684,154	--	--
1928	11,532	495.00	9,776	514,697	--	--
1929	66,573	2,282.00	60,883	2,824,032	1,000	--
1930	26,214	755.00	22,925	1,121,586	--	--
1935	190	10.00	1,510	26,518	--	--
1936	173	9.00	639	18,897	--	--
1937	22	1.00	306	3,876	--	--
1939	996	207.00	2,465	175,940	--	--
1940	4,484	526.00	11,300	632,217	--	--
1941	3,169	381.00	7,013	380,469	--	--
1942	1,274	141.00	1,874	104,000	--	--
*1943–49	15,404	1,009.00	16,223	822,272	--	--
*1954–65	85,486	2,549.00	62,431	4,245,260	4,200	41,247
*1970–71	26,443	563.00	9,579	710,800	--	--
*1974–75	11,651	113.00	1,615	51,271	18,642	870,768
*1982	446	--	901	--	29,537	--
Total	899,517	41,159.00	1,293,208	61,460,098	53,379	912,015
			Alder Creek district			
Total	916,894	41,318.15	1,440,825	61,519,978	4,984,322	1,514,933
			Short tons	30,760	2,492	757

(1957) on the Alder Creek district and that of Farwell and Full (1944) on the Empire mine. Except for Ross (1930), detailed information on the Copper Basin mine is proprietary and is not available for public use. A study of the mineral resource potential and general geology of the Challis National Forest (Worl and others, 1989) and recent studies on jasperoid deposits (Soulliere and others, 1988; Wilson and others, 1988) provide a summary of current (as of 1988) activity and geologic interpretations. A mining history of the area (Wells, 1983) makes for enjoyable reading.

In 1990 geologists from Idaho State University were conducting detailed mapping and studies of the ore deposits in the volcanic rocks of the Challis Volcanic Group in the Mackay and Copper Basin mineralized areas.

HISTORICAL PERSPECTIVE

MACKAY MINERALIZED AREA

Prospectors first discovered copper in Alder Creek in 1879, but it was not until the 1884 discovery of ore at what is now known as the Empire mine that the Alder Creek mining district was established. Mining camps were settled at Houston and Cliff City (fig. 1) in 1884 (Wells, 1983, p. 124), and a small copper smelter constructed at Cliff City operated intermittently for several years with varying success (Umpleby, 1917, p. 93).

A railroad was built from Blackfoot to service the Empire mine in 1901, and the new mining center at its terminus, Mackay, lured settlers from the mining camps. A smelter built at Mackay operated from 1902 to 1907 but was not successful. After 1907 ore was shipped to Bingham, Utah (T.H. Kiilsgaard, written commun., January 1991) or Salt Lake (Umpleby, 1917, p. 93) for smelting.

Early mining operations at the Empire mine had problems (Umpleby, 1917, p. 13–14, 93; Wells, 1983, p. 125), but Umpleby (1917, p. 14, 93) noted that conditions at the mine improved when, in 1907, the property was acquired by the Empire Copper Company, and a leasing system was established that allowed individual miners to be paid for the ore they produced. About this time, improvements were made to the Empire mine, including replacement of the 12-mile-long "expensive electric railway with a Shay steam locomotive" (Wells, 1983, p. 125) that was 7¾ mi long (Umpleby, 1917, p. 100). The steam locomotive was replaced in 1918 by an overhead tram (Wells, 1983, p. 126), whose towers were still visible in Rio Grande Canyon (fig. 1) in 1990. In 1924, a flotation mill was constructed to handle the ore from the Empire mine (Farwell and Full, 1944, p. 4; Wells, 1983, p. 126). Mining operations continued more or less steadily until 1930 and resumed on a smaller scale in the mid-1930's after one of the largest orebodies in the mine was discovered (T.H. Kiilsgaard, written commun., January 1991).

The entire 1,100-foot level of the Empire mine was driven in the 1960's, and ore was mined from it (T.H. Kiilsgaard, written commun., 1991). A large reworked open-pit mine on the southern edge of the Empire mine property is evidence of a mining revival in the early 1970's. All that remains of the operation are skeletons of heavy machinery, deteriorating platforms, and empty tanks indicative of a heap leach operation. Records for this operation could not be located. Production records (table 1) indicate production from the Empire mine as recently as 1982.

COPPER BASIN MINERALIZED AREA

In 1888, copper ore containing lesser amounts of lead and silver was discovered in Copper Basin. Until a smelter was completed at the mine site in 1901, the ore was hauled to Ketchum for shipment (Wells, 1983, p. 124). The Copper Basin smelter was not a success; it proved cheaper to haul the ore to the railroad at Mackay (Umpleby, 1917, p. 103) for shipment. Mining at the Copper Basin mine has been intermittent since its discovery, and only some of its ore is accounted for in official records (table 1).

DEVELOPMENT

MACKAY MINERALIZED AREA

None of the underground workings at any of the mines was examined by the authors of this paper. Previous reports on the Empire mine (Farwell and Full, 1944; Nelson and Ross, 1968) indicate that there were more than 60,000 ft of underground workings driven on grade and numerous stopes in an area that extends two-thirds of a mile and is as wide as 400 ft. The lowest level of the mine is the Cossack Tunnel (the 1,600-foot level) at 7,010 ft elevation from which no ore has been mined. The lowest level that produced ore was the 1,100-foot level (about 7,500 ft elevation). The highest level is at about 8,500 ft, and the main workings are at about 7,700 ft elevation. Surface mining extends as high as 8,880 ft.

Workings at the Horseshoe mine on six levels covered a vertical distance of 350 ft (Nelson and Ross, 1968). Both the major developments and the ore zones trend about N. 30° W. (Ross, 1930). The Champion mine was developed on four levels and had almost 2,100 ft of workings. Most of the ore was mined from the No. 2 level, and the lower levels were never productive (T.H. Kiilsgaard, written commun., April 23, 1952, U.S. Geological Survey (USGS) files, Spokane Field Office).

COPPER BASIN MINERALIZED AREA

The mine at Copper Basin was considerably smaller than the Empire mine. As of 1917, it had five adits, a 265-foot-long shaft, and more than 3,000 ft of workings

(Umpleby, 1917). A map from 1968 (USGS files, Spokane Field Office) shows a glory hole and more than 1,600 ft of workings in only four adits.

PRODUCTION

MACKAY MINERALIZED AREA

Detailed production records for the mines in the Alder Creek mining district are summarized in table 1. Some of the early figures can be verified in Ross (1930, p. 8, 9), Farwell and Full (1944, p. 5, 6), Leland (1957, p. 17–19), and Nelson and Ross (1968, p. A28–A30). Additional data are summarized from files at the U.S. Bureau of Mines Western Field Office, Spokane. The district produced at least 916,000 tons of crude ore that yielded more than 41,159 ounces of gold, 1.44 million ounces of silver, 30,760 tons of copper, 2,490 tons of lead, and 750 tons of zinc.

Oxidized copper ore in skarn was produced from the Empire mine, the main producer in the Alder Creek mining district. From 1902 through 1982, the Empire mine produced at least 30,730 tons of copper, 41,159 ounces of gold, and 1,293,208 ounces of silver from 899,517 tons of crude ore (table 1). Most of the oxidized copper ore averaged 4–5 percent Cu, whereas the sulfide ore contained half that amount of copper (Ross, 1930, p. 15).

Oxidized lead-zinc bearing veins containing lesser amounts of copper, silver, and gold have also been exploited from deposits near the Empire mine, notably the Blue Bird, Champion, Horseshoe, and White Knob mines (fig. 1).

From 1916 through 1971, the Horseshoe mine produced almost 14,000 tons of ore (table 1). From this, almost 120 ounces of gold, 130,000 ounces of silver, 19 tons of copper, 2,000 tons of lead, and 225 tons of zinc were recovered (table 1). Approximately 4,500 tons of ore is reported to have come from the Champion mine prior to 1960 (Nelson and Ross, 1968, p. A28), although Leland (1957) reports only 2,579 tons (table 1) and records from the U.S. Bureau of Mines show only 2,000 tons. The 936 tons of ore produced from the Blue Bird mine yielded 139 tons of lead (table 1). The only production figure found for the White Knob mine is 2,100 tons of 17-percent zinc ore in 1941–42 (Farwell and Full, 1944, p. 6). T.H. Kiilsgaard (written commun., January 1991) recalled that the White Knob was producing smithsonite zinc ore in 1946. No production records for the other mines are available.

COPPER BASIN MINERALIZED AREA

Estimates for the value of ore produced at Copper Basin vary widely and cover different time periods. Umpleby (1917, p. 103) estimated $40,000 by 1914, whereas Ross (1930, p. 10) estimated $230,000 by the late 1920's. Other estimates of production are proprietary but are within this range (USGS files, Spokane Field Office). Only scanty production data are preserved in U.S. Bureau of Mines files: almost 6,200 tons of ore yielded 102 ounces of gold, 18,000 ounces of silver, 360 tons of copper, and 1.3 tons of zinc from 1912 to 1919 and in 1953 (table 1).

EXPLORATION ACTIVITY

Exploration interest in 1989–90 focused on jasperoid-associated, low-grade, large-tonnage, precious-metal deposits and gold skarns in the sedimentary terrane of the region. Claims were staked on and surrounding silicified outcrops, and assessment work was kept current on many old claims. Numerous small jasperoid outcrops were staked between Mackay and the Empire mine. Newly located claims were staked in the Copper Basin mine area. Representatives from large and small mining companies were active in the area, but no extensive rehabilitation of the old workings was undertaken nor were any of the jasperoid deposits developed.

Concurrently, there was exploration for hot-springs-type, volcanic-hosted, disseminated gold deposits in the volcanic rocks adjacent to the area. New prospects and exploration targets were identified, but none was developed. Interest was focused on newly mapped intersections of northeast- and northwest-trending structures (L.G. Snider, Idaho State University, oral commun., May 1990). Volcanic-hosted ore deposits are discussed by Moye and others (in press).

GEOLOGIC SETTING

The Mackay and Copper Basin mineralized areas are east of the Idaho batholith and north of the Snake River Plain. They are in the Cordilleran thrust belt, at the edge of the Basin and Range structural province. Thrust plates composed predominantly of Paleozoic quartzite, carbonate, and shale were emplaced from west to east during Mesozoic time. The thrust plates are offset by northeast-striking normal faults of probable Eocene age, which in turn truncate less well developed, northwest-striking Eocene faults (Southworth, 1988). The northeast-striking faults are truncated by younger, northwest-striking basin and range faults (Skipp and Harding, 1985).

A pair of the northeast-striking Eocene faults forms the margins of the White Knob horst, which includes the Mackay and Copper Basin mineralized areas (Skipp and Harding, 1985). Exposed on the horst are Mississippian sedimentary rocks, a middle Eocene pluton (the Mackay stock), and, locally, volcanic rocks. Adjacent to the horst, volcanic rocks predominate (fig. 2).

DESCRIPTION OF ROCK UNITS

Rocks exposed in the Mackay and Copper Basin mineralized areas are primarily carbonate rocks and minor clastic units, granitic intrusive rocks, and skarn. In addition, jasperoid is present locally. For more complete descriptions refer to the cited publications and references contained therein.

McGOWAN CREEK FORMATION

The Lower Mississippian McGowan Creek Formation (mapped as the Copper Basin Formation of Nelson and Ross (1968, 1969a) in this area) is the oldest rock unit exposed on the White Knob horst (fig. 3). It is an argillite sequence more than 4,000 ft thick composed of distal thin-bedded turbidite and interlayered mudstone, siltstone, and limestone deposited as flysch in a foreland basin (Nilsen, 1977; Skipp and others, 1979).

WHITE KNOB LIMESTONE

The Upper Mississippian White Knob Limestone conformably overlies the McGowan Creek Formation (fig. 3). It consists of more than 5,500 ft of folded and faulted, variably fossiliferous, blue-gray to black, chiefly thick bedded, locally dolomitic limestone that contains abundant nodules and lenses of chert in some beds. The upper 3,000 ft contains conglomerate, sandstone, and mudstone interbeds. The formation was deposited in a gradually shoaling marine environment ranging from deep slope at the base (exposed immediately west of Mackay) to shallow turbulent shelf margin in the upper part (exposed at the Copper Basin mine) (Skipp and others, 1979; Neely and Isaacson, 1982). The White Knob Limestone, locally altered to marble and (or) skarn (described following), is the primary host rock for ore. The quartzite and quartzite conglomerate described near the Copper Basin mine, and mineralized locally, are probably in the upper part of the formation. The formation has not been mapped in sufficient detail to determine which layers are likely to host ore.

MACKAY STOCK

The Eocene Mackay stock intrudes the Mississippian formations (fig. 2). It includes the Mackay Granite, quartz monzonite, leucogranite porphyry, and dikes of quartz latite, rhyolite, and porphyritic rhyolite (Nelson and Ross, 1968). None of the intrusive phases is mineralized. The descriptions that follow are based on the work of Nelson and Ross (1968) and supplemented with our reconnaissance-scale observations. Nelson and Ross (1968) reported that all of the intrusive rocks, and possibly some of the extrusive rocks, were derived from a single parent magma. Detailed studies of the intrusive rocks and their genetic relationships are described by Doyle (1989).

The Mackay Granite (fig. 3), technically a granite porphyry, is the most prominent of the intrusive phases and is exposed in an area of more than 11½ mi^2. Although most outcrops are weathered to a rusty pink, fresh surfaces and fine-grained chill margins are greenish gray. Phenocrysts of pinkish, subhedral to euhedral orthoclase as long as 1 cm and medium-gray, rounded quartz (as long as 0.5 cm) make up 25–35 percent and 5–15 percent of the granite, respectively. The groundmass is composed of quartz, orthoclase, and plagioclase and lesser biotite, hornblende, chlorite, magnetite, apatite, zircon, and rutile (Nelson and Ross, 1968). Recent argon data bracket the age of the granite between 49 and 47 Ma (L.W. Snee, oral commun., May 1990).

Quartz monzonite as described by Nelson and Ross (1968) is exposed in three places at the northern end of the stock (fig. 3) and is reported in several places in the underground workings. The quartz monzonite is light to medium gray and is generally darker and finer grained than the Mackay Granite. Phenocrysts of oligoclase as long as 0.15 cm make up almost 25 percent of the rock. Locally, tiny, rounded phenocrysts of quartz are present. The groundmass consists of fine- to medium-grained orthoclase, quartz, oligoclase, diopside, hornblende, and biotite. Apatite, sphene, magnetite, and chlorite are common (Nelson and Ross, 1968).

Leucogranite porphyry (equivalent to quartz porphyry of Kemp and Gunther, 1908) crops out at the northeastern end of the main intrusive complex (fig. 3). Euhedral to subhedral phenocrysts of orthoclase as long as 2 cm, oligoclase as long as 0.4 cm, and rounded and embayed quartz as long as 0.8 cm are conspicuous. The very pale gray groundmass is composed of very fine grained quartz and feldspar (primarily orthoclase). Diopside, calcite, and sphene are accessory minerals. Based on crosscutting relations, the porphyry is assumed to be younger than the Mackay Granite (Nelson and Ross, 1968).

Quartz latite dikes (mapped as unit Tmr, fig. 3) are light to dark greenish gray. In hand sample, the quartz latite contains small phenocrysts of altered oligoclase, patches of chlorite (with calcite, magnetite, limonite, and quartz), and, locally, pyroxene, amphibole, or biotite. Quartz grains are not visible. The groundmass is orthoclase, oligoclase, and quartz (Nelson and Ross, 1968).

Abundant rhyolite dikes trend northeast across the length of the stock and extend into the sedimentary country rocks (fig. 3). Approximately 25 percent of the dike rock is composed of euhedral to subhedral phenocrysts as long as 0.5 cm of clear and smoky quartz, oligoclase, and orthoclase. The groundmass is very fine grained and yellowish to greenish, light to medium gray. The rhyolite may be contemporaneous with the leucogranite porphyry because these units tend to grade into each other at the northeastern end of the stock (Nelson and Ross, 1968). Preliminary age data indicate that the rhyolite dikes are approximately 48–47 Ma (L.W. Snee, oral commun., May 1990).

Porphyritic rhyolite dikes contain conspicuous euhedral phenocrysts (1 cm) of white oligoclase and fewer phenocrysts (1.5–2.5 cm) of orthoclase, rounded and embayed quartz (1 cm), and minor secondary chlorite after biotite, epidote, and calcite (Nelson and Ross, 1968). These dikes have also been described as porphyritic granodiorite and trachyte porphyry (Farwell and Full, 1944). Commonly, the margins of the porphyritic rhyolite dikes are bordered by dark fine-grained dikes whose composition and relationship were not studied. Some porphyritic rhyolite dikes were emplaced after mineralization (Farwell and Full, 1944, p. 14).

Dikes of several other compositions such as granite, aplite, and andesite porphyry mentioned in the literature and on unpublished maps are not shown in figure 3.

CHALLIS VOLCANIC GROUP

Rocks of the Challis Volcanic Group are not widely exposed on the White Knob horst. Most exposures (figs. 2, 3) on the horst are thin remnants of flows, indurated and welded tuff, and tuff breccia (Nelson and Ross, 1968). Rhyolite, quartz latite, porphyritic obsidian, andesite, and dacite are distinctive shades of reddish and purplish brown, greenish gray to gray, and light tan. The volcanic rocks were extruded 51–44 Ma (Moye and others, 1988).

Geologic mapping by Nelson and Ross (1968, 1969a) does not differentiate volcanic rock types. Detailed mapping of volcanic rocks on either side of the White Knob horst was in progress by geologists, primarily from Idaho State University, in 1990. The new mapping (Moye and others, in press) correlates volcanic units across the horst, locates eruptive centers, and identifies many more structures, especially northwest-striking fault zones, than were previously known.

MARBLE

At the margins of the intrusive rocks, the White Knob Limestone is metamorphosed locally to a very pure, white, coarsely crystalline marble bordered by a zone of tremolite adjacent to unbleached limestone. The marble is finer grained closer to the unmetamorphosed limestone. The limestone probably was metamorphosed to marble in Eocene time during intrusion of the plutonic rocks.

SKARN

Skarn, considered Eocene in age, developed locally along the margins of the Mackay stock (fig. 3) where the stock is in contact with the White Knob Limestone or carbonate-bearing beds of the McGowan Creek Formation. The skarn may be tens of feet thick adjacent to the leucogranite porphyry, especially where engulfed or partly surrounded by the leucogranite porphyry. Early reports noted that the skarn is in the intrusive rocks (Umpleby, 1917), but later studies (Nelson and Ross, 1968) and our observations show that these skarns are totally altered xenoliths, septa, and roof pendants of limestone within the intrusive rock. Endoskarn is scarce.

As much as 75 percent of the skarn consists of andradite and grossularite garnet; diopside, hedenbergite, magnetite, hematite, actinolite, tremolite, scapolite, wollastonite, epidote, vesuvianite, fluorite, calcite, chlorite, gypsum, and quartz are also present. Locally the skarn is almost entirely composed of magnetite and hematite. The texture of the skarn is crystalline and granular, but in places relict bedding is preserved. Exposures of skarn are sporadic at the surface; without access to underground exposures we could not determine if zoning is present within the skarn bodies.

JASPEROID

Jasperoid crops out along the southern fault boundary of the horst block and also locally is present as scattered lenses throughout the White Knob Limestone peripheral to skarn (fig. 3). The jasperoid is commonly gray to brown, very fine grained, brecciated, and ferruginous. Hematite and limonite stain exterior weathered surfaces. The degree of silicification varies from patchy and diffuse to massive and dense. Textures commonly indicate more than one episode of brecciation and silicification. Veins of quartz and calcite are common, and barite, fluorite, jarosite, and stibiconite are less prevalent. Petrographic studies show no grains of identifiable minerals nor any fluid inclusions large enough for analysis. Detailed descriptions and geochemistry of jasperoid samples from the Mackay area are given in Wilson and others (1988) and Soulliere and others (1988).

STRUCTURE

In south-central Idaho, several Cretaceous thrust plates separate distinct stratigraphic rock sequences (Skipp and Hait, 1977; Skipp, 1987; Link and others, 1988). The Mackay and Copper Basin mineralized areas are on the White Knob thrust plate. The Copper Basin thrust fault, exposed on the hillside immediately below, and southwest of, the Copper Basin mine, forms the western margin of the thrust plate (fig. 2). The eastern margin of the White Knob thrust plate, the Lost River thrust fault, is not exposed in the study area but is postulated to be buried under alluvium in the Lost River Valley east of Mackay (Skipp and Harding, 1985). Thrust faults are probably Cretaceous in age and have been used to explain abrupt facies changes in Mississippian rocks (Skipp and Hait, 1977; Skipp and Harding, 1985).

Paleozoic rock units in the area are folded as a result of the Cretaceous thrusting. Locally the folding varies, but in

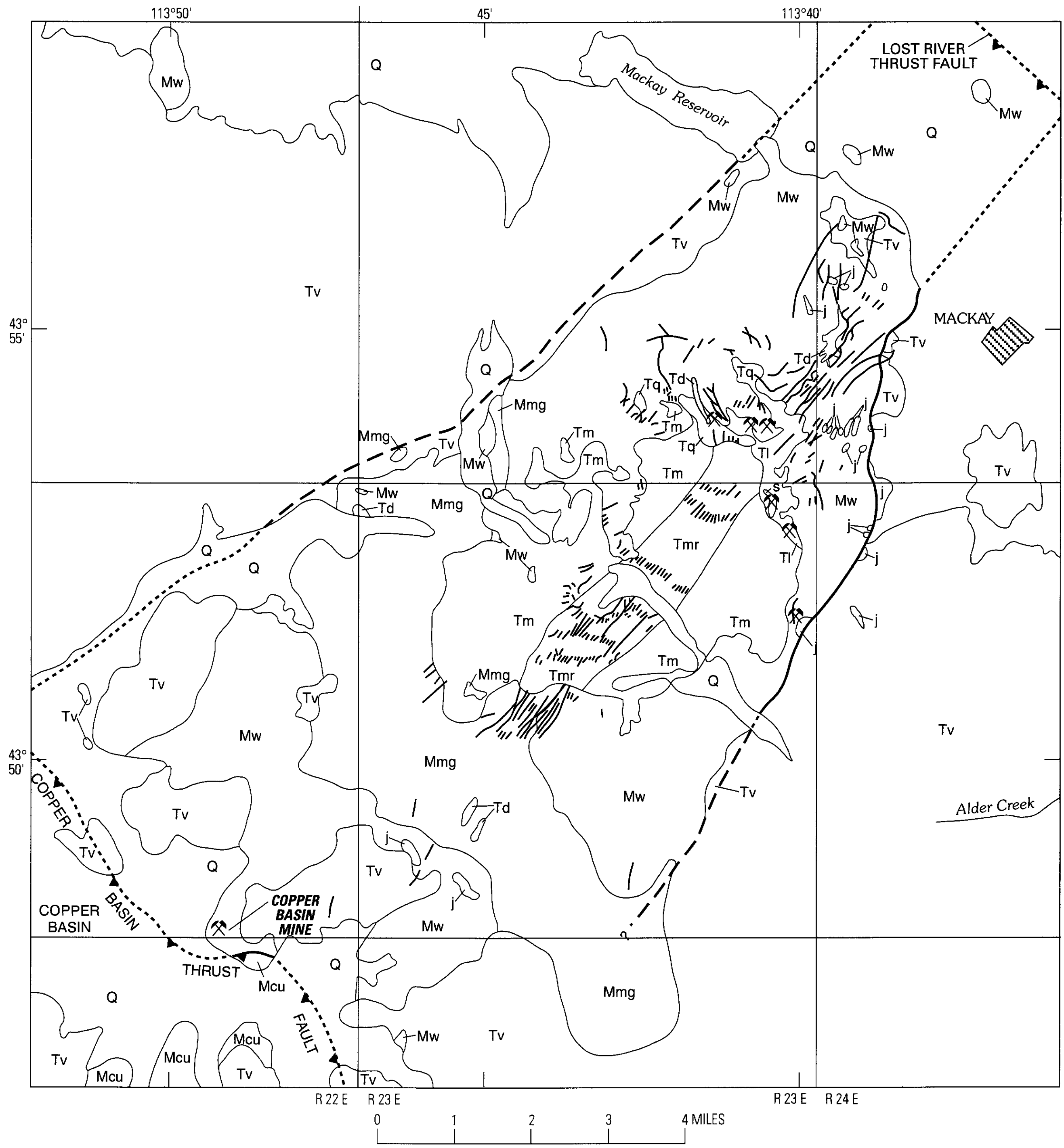


Figure 3 (above and facing column). Map showing geology of the Mackay and Copper Basin mineralized areas, White Knob Mountains, Idaho. Geology modified from Nelson and Ross (1968, 1969a) and Dover (1981).

EXPLANATION

Q	Quaternary deposits, undifferentiated
j	Jasperoid
s	Skarn
Tv	Challis Volcanic Group (Eocene)
Td	Dikes, undifferentiated (Eocene)
Tl	Leucogranite porphyry (Eocene)
Tm	Mackay Granite (Eocene)
Tmr	Zone of rhyolite dikes in the Mackay Granite (Eocene)
Tq	Quartz monzonite (Eocene)
Mw	White Knob Limestone (Mississippian)
Mmg	McGowan Creek Formation (Mississippian)
Mcu	Copper Basin Formation (Mississippian)

Contact

Fault—Dashed where inferred; dotted where concealed; queried where unknown

Thrust fault—Dotted where concealed; sawteeth on upper plate

Small dikes and intrusive bodies

Min

general the anticlines and synclines trend north-northwest. The limbs have moderate to steep dips to the northeast and southwest. Locally, due to intense folding, the strikes and dips vary (Nelson and Ross, 1969a).

Within the White Knob thrust plate is the northeast-trending White Knob horst, which is composed of Mississippian sedimentary rocks cored by exposed middle Eocene plutonic rocks (Skipp and Hait, 1977; Skipp and Harding, 1985). The horst is bounded on the southeast by a high-angle fault that cuts rocks of the Eocene Challis Volcanic Group and on the northwest by a fault that has been mapped as buried by the same volcanic rocks (Nelson and Ross, 1968) (fig. 2). The latter fault may dip to the northwest at about 40° (L.G. Snider, Idaho State University, oral commun., May 1990). Uplift of the horst has been considered to be coincident with emplacement of the intrusion (Nelson and Ross, 1968; Skipp and Harding, 1985); however, recent studies of the volcanic rocks show that uplift may postdate intrusion (L.G. Snider, oral commun., May 1990).

Regional aeromagnetic data (Worl and others, 1989) indicate that the main body of the Mackay stock continues southwest of the exposed outcrop. The stock is elongate and dips under the Copper Basin mine area to the southwest.

Prevalent northeast-striking Eocene structures, including the horst, faults, intrusions, and dike swarms, reflect the influence of regional northwest-southeast extension. The same trend is noted in the Trans-Challis fault system to the northwest and in the Hailey gold belt to the west (Kiilsgaard and others, 1986; Worl and Johnson, this volume).

Numerous northwest-striking faults of probable Eocene age are present in rocks of the Challis Volcanic Group on either side of the White Knob horst. Geologists from Idaho State University have identified a major fault zone in the Challis Volcanic Group that appears to cross the horst at the northeastern end of the main pluton (L.G. Snider, oral commun., May 1990). The quartz monzonite and leucogranite porphyry, as well as most of the major ore deposits near Mackay, are confined to this zone. After discussions with L.G. Snider, we agree that these intrusions followed a well-defined fault zone and that the ore may be preferentially located at the intersections of northeast- and northwest-trending structures.

MINERAL DEPOSITS

COPPER SKARN DEPOSITS

Known mineral deposits in the Mackay and Copper Basin mineralized areas can be classified as copper skarn deposits. The Empire mine is typical of the inner zone of copper skarns, and the Blue Bird, Champion, Horseshoe, and White Knob mines represent the outer zone. There is, of course, some overlap between zones and their boundaries are gradational, but for the purpose of this paper the zones are discussed separately. The Copper Basin mine displays characteristics of both the inner and outer zones.

INNER ZONE

Description.—The most prominent and prolific ore producers in the Mackay and Copper Basin mineralized areas are copper skarn deposits. The deposits are along the contacts of the Mackay Granite and leucogranite porphyry with the surrounding White Knob Limestone. The skarn, a product of metamorphism and metasomatism, is the ore host. It is developed primarily in limestone (exoskarn) blocks that are included in the intrusive rocks and along parts of the leucogranite porphyry-limestone contact. The ore is massive to disseminated within the skarn. Locally, abundant iron minerals are suggestive of iron skarn, but iron minerals also are found in copper skarns (Einaudi and Burt, 1982). Copper, lead, zinc, silver, gold, and minor tungsten have also been produced. Gold-bearing skarns were the exploration targets in 1990.

Type example.—Empire mine, Alder Creek district

Other example.—Copper Basin mine, Copper Basin district

Orebodies.—Deposits are present where reactive carbonate strata were replaced by calcium-iron-magnesium-manganese silicate minerals and ore minerals within, or close to, the Mackay Granite and the leucogranite porphyry. Contacts between the intrusive and carbonate rocks, local bedding planes, faults, joints, and breccia zones control location of the orebodies. Most orebodies are pipelike, chimney or pod shaped, irregular, and small. Orebodies are at the contact between skarn and limestone, entirely within the skarn, and at the contact of skarn with intrusive rocks (Nelson and Ross, 1968, p. A25). Several orebodies are cut, and possibly were displaced, by rhyolite porphyry dikes that were emplaced after mineralization (Farwell and Full, 1944, p. 14).

Most of the deposits are confined to a northwest-trending, narrow, arcuate band, approximately 4 mi long, at the northeast end of the Mackay stock (fig. 2), and the largest skarn deposits are adjacent to the leucogranite porphyry. Concentrated near the middle of this zone are orebodies of the Empire mine. Individual orebodies are commonly 15–200 ft long by 5–55 ft wide, and the largest extend vertically for about 600 ft.

Texture.—In the skarn, ore minerals are interstitial to gangue. Copper ore minerals may be massive to disseminated, but other ore minerals are disseminated in widely varying concentrations throughout the skarn. Magnetite and hematite ore minerals are commonly, but not exclusively, massive. Farwell and Full (1944, p. 11) described the oxidized ore as calc-silicate minerals "carrying disseminations and spongelike aggregates of metallic minerals."

Calc-silicate skarn rocks range from fine to very coarse grained and granular. The ore-bearing skarn is coarser grained and contains more calcite than barren skarn. Pure white marble beds have a sugary texture. Blue-gray limestone is moderately crystalline and, locally, is interlayered with bands of the white marble.

Mineralogy.—At the Empire mine, the principal ore minerals are chrysocolla (copper silicate), malachite, and azurite (copper carbonates), tenorite (copper oxide), and sparse copper sulfate minerals. Locally, the deposits are almost entirely composed of magnetite and hematite, and siderite is abundant peripheral to these zones (such as at the Grand Prize mine, fig. 2). Sulfide minerals are rare, although chalcopyrite, pyrite, pyrrhotite, molybdenite, sphalerite, and, infrequently, bornite are present with associated calcite, quartz, magnetite, fluorite, scheelite, and specularite.

Similarly, most ore at Copper Basin is oxidized except where it is in black shaly limestone. Oxidized copper minerals are malachite, azurite, chrysocolla, cuprite, and tenorite. Magnetite and siderite are abundant. Sulfide minerals are rare, although bornite, chalcopyrite, chalcocite, and pyrite have been found, especially in the shaly layers.

Much of the material on the mine dumps consists of iron hydroxide in siliceous material with secondary hematite, magnetite, quartz, garnet, and diopside and stains of copper oxides. Well-developed extensively oxidized gossan is present in brecciated quartzite (probably part of the White Knob Limestone) at the Copper Basin mine.

Geochemical signature.—The geochemical signature commonly associated with copper skarns is copper-gold-silver in the proximal (inner) zone grading outward to gold-silver to lead-zinc-silver in the distal (outer) zone. Some copper skarns are anomalous in copper, arsenic, antimony, and bismuth (Cox and Theodore, 1986, p. 86). For comparison, lead-zinc skarn deposits are anomalous in Zn, Pb, Mn, Cu, Co, Au, Ag, As, W, Sn, F, and Be (Cox, 1986a, p. 90), and iron skarns tend to be anomalous in Fe, Cu, Co, Au, and Sn (Cox, 1986b, p. 94).

Thirty-four rock and two stream-sediment samples from the Mackay and Copper Basin mineralized areas were semiquantitatively analyzed for 35 elements by a direct-current arc emission spectrographic method (Bullock and others, 1990). Table 2 gives the lower limit of detection, minimum and maximum values, mean, standard deviation, and number of unqualified values for each element for the 34 rock samples. Of the 35 elements analyzed, Ca, Fe, Mg, Ti, Ag, Ba, Cr, Cu, Ga, Mn, Ni, Pb, V, Zn, and Zr were detected in at least half the samples. P, Au, Ge, Nb, Sc, and Th had insufficient detectable values to compute mean and standard deviation. The complete data set is given in Bullock and others (1990).

Statistical analysis of the geochemical data shows correlations of arsenic-cobalt-bismuth, silver-lead-zinc, and beryllium-tungsten. The first two associations are typical of copper skarns, but alone they are not conclusive evidence that the system near Mackay is exclusively a copper skarn. The same elements, beryllium, and tungsten, are present, for instance, in lead-zinc skarns. The same associations resulted when the data were reevaluated excluding the samples specifically collected from the copper skarn at the Empire mine. The data set was too small to make statistically valid correlations between specific rock types, ores, and alteration.

Geophysical signature.—Aeromagnetic data outline the Mackay stock and suggest that the stock is more extensive to the west in the subsurface than in surface exposure (Worl and others, 1989). The Mackay Granite probably underlies the Copper Basin mine area. No geophysical signatures specific to skarn are indicated in the Mackay and Copper Basin mineralized areas. Detailed electrical and magnetic surveys might reveal additional smaller sulfide-bearing orebodies, depending on the mineralogy and composition of the host rock.

Genesis and ore controls.—Mineralized rock is at the intersection of the northeast-trending intrusive rocks and the White Knob Limestone. Ore is in a narrow, north-trending arcuate zone in contact with metamorphosed limestone or in xenoliths or roof pendants in granitic rock and related porphyry close to the limestone. This zone is the only one in the horst that has prominent northwest-trending structures.

Ore stopes are along crosscuts that are radial to the northeast-trending haulageways (Nelson and Ross, 1968, p. A24). The crosscuts follow zones of shearing in both the igneous rock and altered limestone (Ross, 1930, p. 15; Nelson and Ross, 1968, p. A24). The shear zones are ore guides. Irregular porphyry dikes that cut both granite and limestone are approximately parallel with these shear zones. Ore is also present along bedding in the limestone.

Table 2. Summary of geochemical data for 34 rock samples from the Mackay and Copper Basin mineralized areas, White Knob Mountains, Idaho.

[Complete data set is given in Bullock and others (1990). Ca, Fe, Mg, Na, P, and Ti in percent; all other values in parts per million. Limit is the detection or determination limit. Mean and standard deviation were computed using the lower detection limit for all N and L values and the maximum value for all G values. Mean and standard deviation were not computed for P, Au, Ge, Nb, Sc, and Th; these elements either had no unqualified values or the highest value was the same as the lower detection limit. Valid indicates number of samples without qualified values (N, L, or G): N indicates not detected; L indicates detected but below limit of determination; G indicates greater than the uppermost detection limit]

Element	Limit	Minimum value detected	Maximum value detected	Mean	Standard Deviation	Valid
Ca	0.05	N	20	4.73	5.83	33
Fe	0.05	L	20	7.16	7.15	33
Mg	0.02	0.02	5	1.03	1.20	34
Na	0.2	N	3	0.685	0.866	10
P	0.2	N	L	--	--	0
Ti	0.002	N	.3	0.067	0.082	33
Ag	0.5	N	200	40.6	59.5	19
As	200	N	10,000G	629	1,732	6
Au	10	N	L	--	--	0
B	10	N	100	12.8	15.4	7
Ba	20	N	5,000	592	1,111	26
Be	1	N	7	1.40	1.09	12
Bi	10	N	200	27.7	43.4	9
Cd	20	N	500	39.7	8.45	4
Co	10	N	700	41.6	120	11
Cr	10	N	100	23.2	24.0	17
Cu	5	L	20,000G	284	5,209	30
Ga	5	N	50	19.1	13.4	26
Ge	10	N	N	--	--	0
La	50	N	150	54.1	17.6	5
Mn	10	L	5,000G	1,297	1,820	32
Mo	5	N	300	22.4	52.8	15
Nb	20	N	20	--	--	1
Ni	5	N	100	16.9	19.5	26
Pb	10	N	20,000G	497	8,053	28
Sb	100	N	200	103	17.2	1
Sc	5	N	5	--	--	3
Sn	10	N	500	45.2	92.5	12
Sr	100	N	500	140	87.7	11
Th	100	N	N	--	--	0
V	10	N	100	28.2	25.1	27
W	20	N	1,000	60.6	169	10
Y	10	N	20	10.4	1.89	4
Zn	200	N	10,000G	2,844	3,804	18
Zr	10	N	150	33.1	37.0	19

The Mackay stock and its related intrusions probably provided the heat and fluids that caused metasomatism and the formation of skarn. Presence of jasperoid and alteration of the intrusive and carbonate rocks are indications of hydrothermal activity.

Deposits at the Empire mine are representative of the skarn-forming process. The White Knob Limestone, with its calcic and dolomitic variations, was invaded by the intrusive rocks of the Eocene Mackay stock. Contact metamorphism caused the marbleization of the limestone beds, and minor skarn may have developed as hornfels in the more clastic layers. Metasomatism and continued skarn formation took place only where porosity and permeability permitted fluids to enter; thus, the ore-forming solutions penetrated the skarn rocks chiefly along preexisting fractures (Ross, 1930, p. 16; Worl and others, 1989, p. 60). These fractures probably formed coincident with the leucogranite porphyry, as did most of the larger skarn bodies. It should be emphasized that the ore formed by replacement, not by filling of open cavities.

Exploration guides.—Additional, undiscovered copper skarn deposits are likely to be found near the Empire and

Copper Basin mines. Elsewhere near the margins of the Mackay stock, potential for additional discoveries of skarn with economic value is moderate.

The orebodies are scattered along the intrusive rock-limestone contact in skarn, and exploration should obviously concentrate on this zone; however, fully engulfed blocks of skarn in the intrusive bodies and contacts not exposed at the surface or in underground workings also are potential sites for ore. In addition, further research should focus on locating the intersections of major northeast- and northwest-trending structures. It is likely that the entire Empire mine area may be at such an intersection.

Geochemical studies on the composition of the border phases of the intrusive rock bodies could help delineate locations where thicker skarn bodies might be present because most of the larger deposits are related to intrusions of leucogranite porphyry. Detailed studies of the spatial relationships of oxide and sulfide ores could be used to restrict the areas of search to those having the greatest potential. Most of the Empire ore was relatively shallow and, therefore, oxidized. Perhaps this zone is underlain by sulfide-bearing ore. Finally, there is not a clear understanding of the compositional variation in the White Knob Limestone, especially the calcite-dolomite distribution. Such an understanding might lead to delineation of calcium-rich areas more susceptible to sulfide deposition and magnesium-rich areas containing iron-oxide-rich skarn.

Future studies should include detailed geochemical sampling and analysis for gold and silver, both of which commonly are present in anomalous quantities in the inner and outer zones of typical copper skarns (Cox and Theodore, 1986, p. 86). Areas of the skarn that are enriched in precious metals could possibly be exploited in the future.

OUTER ZONE

Description.—Ore deposits of lead, zinc, and silver and accompanying copper and gold, such as at the Blue Bird, Champion, Horseshoe and White Knob mines, are in veins in the Paleozoic sedimentary rocks. Because a close spatial and genetic relationship exists between these deposits and the skarn deposits of the inner zone, discussed previously, the geologic characteristics are similar.

Host rock is the White Knob Limestone and intercalated quartzite and, locally, adjacent intrusive rocks, particularly dikes. Unlike the inner zone, the limestone is generally unaltered, and contact metamorphism is limited to a narrow zone of marble and some calc-silicate minerals such as wollastonite, scapolite, diopside, grossularite, and tremolite (Nelson and Ross, 1968, p. A20). Veins tend to parallel high-angle, northeast-striking faults and northwest-trending shear zones. Ore may be concentrated at the intersections of these northeast- and northwest-trending structures.

Lead, zinc, and silver are the main commodities mined from small deposits of the outer zone near the Mackay stock. These deposits are privately owned, and very little detailed information regarding them is published. Although these deposits contributed to the economic development of the region, they were insignificant from a commercial standpoint. Most of the ore in the Mackay area came from copper deposits in the inner zone.

Examples.—Blue Bird, Champion, Horseshoe, and White Knob mines in the Alder Creek district; Copper Basin mine in Copper Basin district (fig. 1)

Orebodies.—Deposits of lead, zinc, and silver are mainly replacements along fractures and bedding planes in the White Knob Limestone. Most are intimately associated with, and overlap, the inner zone skarn deposits that are so prominent in the region.

Individual orebodies are generally tabular and vary greatly in size. Exact dimensions are unknown, but these are considered to be small mines. The orebodies were probably several hundreds of feet by several tens of feet and not more than 30 ft thick.

At the Copper Basin mine, the principal orebodies were near aplite and porphyry dikes in the "limonite zone" of the quartzite conglomerate (H.H. Doelling and K.C. Thompson, written commun., 1960, USGS files, Spokane Field Office), in northeast- and northwest-trending fissures, and in crushed ground at the intersections of these fissures (E.R. Zalinsky, written commun., 1917, USGS files, Spokane Field Office). Replacement ore, generally striking N. 20° E. and dipping gently to the southeast, was also bedded in the limestone and quartzite units in the upper part of the White Knob Limestone. These replacement bodies were 6 in.–6 ft thick in the quartzite (Ross, 1930, p. 16), 6–20 ft thick in limestone (Anonymous, (probably C.P. Heiner) written commun. to OME Field Officer, April 18, 1969, USGS files, Spokane Field Office), and as much as 30 ft thick in shale (E.R. Zalinsky, written commun., 1917, p. 7, USGS files, Spokane Field Office).

Also at the Copper Basin mine, dikes of granite, andesite porphyry, and aplite are mineralized locally (Ross, 1930, p. 16). This local mineralized rock is always associated with introduced vein quartz or silicification (E.A. Baxter, written commun., November 1966, p. 4, USGS files, Spokane Field Office). A mineralized breccia zone containing remnants of leached chrysocolla in vuggy limonite was reported at depth (Anonymous (probably C.P. Heiner), April 18, 1969, USGS files, Spokane Field Office).

The ore zones and main workings at the Horseshoe mine trend N. 30° W. in marbleized limestone (Ross, 1930, p. 16; Nelson and Ross, 1968, p. A27). At least several of the ore zones in the Horseshoe are along limestone–intrusive rock contacts (G.M. Fowler, written commun., 1926; F.W. Anderson, written commun., 1941; both from the Anaconda Geological Documents Collection).

Most ore at the Champion mine is in brecciated and sheared limestone, and the ore shoots are generally parallel to each other. The main orebody strikes north and dips 60°–70° W.; it is 1,400 ft long and 15–30 ft wide and extends 250 ft downdip (T.H. Kiilsgaard, written commun., April 23, 1952, p. 3, USGS files, Spokane Field Office).

Ore shoots at the Blue Bird mine are small pipes that dip steeply to the northeast. Their locations probably are controlled by north-striking fissures (D.C. Gilbert, written commun., 1935, Anaconda Geological Documents Collection).

Mineralogy and texture.—The deposits in the outer zone were mined primarily for lead and silver; trace to byproduct amounts of zinc, copper, and gold contributed to their economic value.

The ore minerals are primarily cerussite, argentiferous galena, pyrite, chalcopyrite, and sphalerite. Lesser quantities of cerargyrite, smithsonite, and oxide minerals of zinc, manganese, and iron are common and are present as disseminations, clots, and lenses. Gangue is not abundant and is mostly quartz and lesser calcite, except at the Grand Prize mine where siderite dominates the dumps.

Cerussite is the primary ore mined at the Champion, Horseshoe, and White Knob mines and probably at the Blue Bird as well (Ross, 1930, p. 15; Farwell and Full, 1944, p. 12; T.H. Kiilsgaard written commun., April 23, 1952, p. 4, USGS files, Spokane Field Office). At the Champion mine it is present with complex secondary lead, silver, and zinc minerals, remnant fragments of galena (T.H. Kiilsgaard, written commun., April 1952, USGS files, Spokane Field Office), and rare chalcopyrite (Farwell and Full, 1944, p. 12). At the Horseshoe mine the ore, which is principally cerussite, also contains gold, silver, and copper. There are also coexisting bodies of sulfides, including pyrite, iron-rich sphalerite (marmatite), galena, chalcopyrite, (Ross, 1930, p. 15), and pyrrhotite (Farwell and Full, 1944, p. 12). At the Blue Bird mine the oxidized lead ore had high iron content (D.C. Gilbert, written commun., 1935, Anaconda Geological Documents Collection).

Ore in veins at the Grand Prize mine is argentiferous galena in siderite. Coarse-grained siderite is abundant on the surface at the mine site. Umpleby (1917, p. 101) described the ore as "sand carbonate in limonite gangue," probably referring to cerussite as sand carbonate, a popular miner's term (T.H. Kiilsgaard, written commun., January 1991).

Geochemical signature.—The geochemistry of the outer zone is discussed in the section on the inner zone. The Copper Basin, Blue Bird, Champion, Horseshoe, and White Knob deposits are classified as the outer zone of the copper skarn, even though most contain little copper, based on the element associations arsenic-cobalt-bismuth and silver-lead-zinc. Previous summaries of local ore deposits (Worl and Johnson, 1989; Worl and others, 1989) did not include detailed geochemistry, statistical analysis of the data, or a broader, more regional perspective and therefore classified these mines as polymetallic replacement veins. Such deposits would be expected to have anomalous concentrations of Cu, Pb, Ag, Zn, Mn, Au, As, Sb, Bi, and Ba (Morris, 1986, p. 99). In rocks of the outer zone most of these elements are present but not necessarily in anomalous quantities.

Geophysical signature.—Detailed electrical survey methods may detect sulfide-bearing veins that are typical of outer zone skarn deposits, but no such data are available for this area.

Genesis and ore controls.—Hydrothermal systems active during the late stages of intrusion and dike emplacement probably formed the deposits in this area. These replacement deposits in the outer zone tend to follow the northeast- and northwest-trending dike swarms or are at their intersections peripheral to main skarn bodies.

At the Copper Basin mine, ore deposits in the outer skarn zone have a wide variety of controls. The principal orebodies are near dikes, in crushed ground at the intersection of almost vertical northeast- and northwest-trending fissures (E.R. Zalinsky, written commun., August 1917, USGS files, Spokane Field Office; H.H. Doelling and K.C. Thompson, written commun., September 1960, USGS files, Spokane Field Office). Replacement ore is also bedded in the sedimentary units, generally striking N. 20° E. and dipping gently to the southeast (Anonymous (probably C.P. Heiner), written commun., April 1969, USGS files, Spokane Field Office).

Exploration guides.—Much of the lead and zinc production in the Alder Creek district was from relatively small replacement veins in limestone. Hidden orebodies of this type are difficult to detect; however, deposits similar to those at the Empire and Copper Basin mines may exist in the Mackay and Copper Basin mineralized areas. Exploration efforts should concentrate on the northwest-trending shear zones that cross the horst block, particularly at intersections with northeast-trending faults. Jasperoid may be an indicator of these deposits; it is present near most of the known mines in this region.

REFERENCES CITED

Bullock, J.H., Jr., Wilson, A.B., and Soulliere, S.J., 1990, Analytical results and sample locality maps of rock samples from the northwestern Idaho Falls and the northeastern Hailey 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 90–645, 20 p.

Cox, D.P., 1986a, Descriptive model of Zn-Pb skarn deposits, *in* Cox, D.P., and Singer, D.A., eds., Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 90.

———1986b, Descriptive model of Fe skarn deposits, *in* Cox, D.P., and Singer, D.A., eds., Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 94.

Cox, D.P., and Theodore, T.G., 1986, Descriptive model of Cu skarn deposits, *in* Cox, D.P., and Singer, D.A., eds., Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 86.

Dover, J.H., 1981, Geology of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–A, p. 15–75.

Doyle, M.A., 1989, Geology, petrology, and geochemistry of the Eocene intrusive suite of the Mackay Peak area, south-central Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 86–90.

Einaudi, M.T., and Burt, D.M., 1982, Introduction—Terminology, classification, and composition of skarn deposits: Economic Geology, v. 77, p. 745–754.

Farwell, F.W., and Full, R.P., 1944, Geology of the Empire Copper Mine near Mackay, Idaho: U.S. Geological Survey Open-File Report 370, 22 p.

Kemp, J.F., and Gunther, C.G., 1907, The White Knob copper deposits, Mackay, Idaho: Transactions of the American Institute of Mining Engineers, v. 38, p. 269–296.

Kiilsgaard, T.H., Fisher, F.S., and Bennett, E.H., 1986, The trans-Challis fault system and associated precious metal deposits: Economic Geology, v. 81, p. 721–724.

Leland, G.R., 1957, General geology and mineralization of the Mackay Stock area: Moscow, University of Idaho, M.S. thesis, 71 p.

Link, P.K., Skipp, Betty, Hait, M.H., Janecke, Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 87–98.

Morris, H.T., Descriptive model of polymetallic replacement deposits, *in* Cox, D.P., and Singer, D.A., eds., 1986, Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 99.

Moye, F.J., Hackett, W.R., Snider, L.G., and Mertzman, S.A., in press, Stratigraphy, chronology, and petrology of the southeastern Challis volcanic field, central Idaho: U.S. Geological Survey Bulletin 2064–EE.

Moye, F.J., Hackett, W.R., Blakley, J.D., and Snider, L.G., 1988, Regional geologic setting and volcanic stratigraphy of the Challis volcanic field, central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 87–98.

Neely, K.W., and Isaacson, P.E., 1982, Stratigraphy and depositional paleoenvironments of the White Knob Formation (late Mississippian), south-central Idaho: Geological Society of America Abstracts with Programs, v. 14, no. 6, p. 343.

Nelson, W.H., and Ross, C.P., 1968, Geology of part of the Alder Creek mining district, Custer County, Idaho: U.S. Geological Survey Bulletin 1252–A, 30 p.

———1969a, Geologic map of the Mackay quadrangle, south-central Idaho: U.S. Geological Survey Miscellaneous Geologic Investigations Series Map I–580, scale 1:125,000.

———1969b, Geology of the Mackay 30-minute quadrangle, Idaho: U.S. Geological Survey Open-File Report 70–242, 161 p.

Nilsen, T.H., 1977, Paleogeography of Mississippian turbidites in south-central Idaho, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 1, p. 275–299.

Ross, C.P., 1930, Geology and ore deposits of the Seafoam, Alder Creek, Little Smoky Creek, and Willow Creek mining districts, Custer and Camas Counties, Idaho: Idaho Bureau of Mines and Geology Pamphlet 33, p. 7–18.

Sanford, R.F., Wooden, J.L., Doe, B.R., and Worl, R.G., 1989, Origin of ore deposits in the central Idaho black shale belt—Implications from lead isotopes, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 39–52.

Skipp, Betty, 1987, Basement thrust sheets in the Clearwater orogenic zone, central Idaho and western Montana: Geology, v. 15, p. 220–224.

Skipp, Betty, and Hait, M.H., Jr., 1977, Allochthons along the northeast margin of the Snake River Plain, Idaho: Wyoming Geological Association Annual Field Conference, 29th, Guidebook, p. 499–515.

Skipp, Betty, and Harding, S.T., 1985, Preliminary report on geology of Borah Peak area, Idaho, including interpretation of seismic and gravity data, *in* Stein, R.S., and Bucknam, R.C., eds., Proceedings of Workshop 28 on the Borah Peak, Idaho, earthquake: U.S. Geological Survey Open-File Report 85–290–A, p. 657–671.

Skipp, Betty, Sando, W.J., and Hall, W.E., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States–Idaho: U.S. Geological Survey Professional Paper 1110–AA, p. AA1–AA42.

Soulliere, S.J., Wilson, A.B., Moye, F.J., and Rhea, K.P., 1988, Analytical results and sample locality map of jasperoid samples from Mackay Idaho: U.S. Geological Survey Open-File Report 88–589, 28 p.

Southworth, C.S., 1988, Preliminary lineament map of the Hailey, Idaho, and west half of the Idaho Falls, Idaho, 1°×2° quadrangles, with inset of regional image mosaic and geologic structures: U.S. Geological Survey Open-File Report 88–640, scale 1:250,000.

Umpleby, J.B., 1917, Geology and ore deposits of the Mackay region, Idaho: U.S. Geological Survey Professional Paper 97, scale 1:375,000, 129 p.

Wells, M.W., 1983, Gold camps and silver cities (2nd ed.): Idaho Bureau of Mines and Geology Bulletin 22, 165 p.

Wilson, A.B., and Skipp, Betty, 1994, Geologic map of the eastern part of the Challis National Forest and vicinity, Idaho: U.S. Geological Survey Miscellaneous Investigations Map I–2395, scale 1:250,000.

Wilson, A.B., Soulliere, S.J., Skipp, Betty, Worl, R.G., and Rhea, K.P., 1988, Geology and geochemistry of jasperoid near Mackay, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 183–192.

Worl, R.G., and Johnson, K.M., 1989, Geologic terranes and mineral deposit types in the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 3–16.

Worl, R.G., Wilson, A.B., Smith, C.L., Kleinkopf, M.D., and Sykes, R.C., 1989, Mineral resource potential and geology of the Challis National Forest, Idaho: U.S. Geological Survey Bulletin 1873, 101 p.

Jasperoid-Associated Precious-Metal Deposits in the Northwestern Part of the Idaho Falls 1°×2° Quadrangle, South-Central Idaho

By Sandra J. Soulliere, Anna B. Wilson, *and* Betty Skipp

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–J

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLE

Jasperoid-Associated Precious-Metal Deposits in the Northwestern Part of the Idaho Falls 1°×2° Quadrangle, South-Central Idaho

By Sandra J. Soulliere,[1] Anna B. Wilson,[1] *and* Betty Skipp[1]

ABSTRACT

Jasperoid bodies in the northwestern part of the Idaho Falls 1°×2° quadrangle in south-central Idaho have geologic and geochemical characteristics similar to those associated with known sedimentary-rock-hosted, disseminated precious-metal deposits elsewhere in the western United States. The jasperoid bodies are mainly replacements of limestone, commonly along high-angle fault systems, and many outcrops show evidence of more than one period of brecciation and silicification. Geochemical analyses of 117 jasperoid samples show that the suite of elements Ag, Au, As, Hg, Sb, and Tl is consistently present, commonly in anomalous amounts. The results of the study were used to assess the potential for undiscovered precious-metal deposits associated with jasperoid in this region.

INTRODUCTION

Recent exploration interest in precious-metal deposits associated with jasperoid prompted a reconnaissance study of jasperoid bodies in the northwestern part of the Idaho Falls 1°×2° quadrangle in south-central Idaho. This study was undertaken in order to assess the potential for the occurrence of this type of deposit within the quadrangle (Worl and Johnson, 1989). The study area is near the town of Mackay, west of U.S. Highway 93 and north of the Pioneer Mountains (fig. 1). Jasperoid, as used in this paper, refers to "an epigenetic siliceous replacement of a previously lithified host rock" (Lovering, 1972).

Jasperoid bodies in this area have geologic and geochemical characteristics similar to jasperoid associated with known sedimentary-rock-hosted, disseminated precious-metal deposits elsewhere in the Western United States (Wilson and others, 1988a). No gold deposits of this type have been discovered as yet near Mackay; however, claims have been staked around most of the jasperoid outcrops. Recent exploration activity has consisted mainly of detailed mapping, geochemical sampling, and some local drilling near jasperoid bodies.

Jasperoid bodies were located using recent geologic mapping by Skipp (1988, 1989), Skipp and others (1990), Skipp and Bollman (1992), earlier mapping by Nelson and Ross (1969a, b), and data supplied by Keith Rhea of Ampro Resources, Missoula, Montana. One hundred and seventeen jasperoid samples were collected for geochemical and petrographic analyses. The results are briefly summarized here; details are given in Soulliere and others (1988), Wilson and others (1988a, b), and Soulliere (1992).

GEOLOGIC SETTING

In the northwestern part of the Idaho Falls 1°×2° quadrangle, Paleozoic sedimentary rocks were intruded by Tertiary granitic stocks and hypabyssal bodies and covered by rocks of the Eocene Challis Volcanic Group (fig. 2). Northeast- and northwest-striking high-angle faults cut all rock types in the area. Extensive areas of jasperoid are along these faults.

Paleozoic sedimentary rocks that crop out in the area are Early Mississippian to Early Permian in age and consist of, in approximate ascending order, the Copper Basin and McGowan Creek Formations, White Knob Limestone, and Middle Canyon, Scott Peak, South Creek, Surrett Canyon, Bluebird Mountain, and Snaky Canyon Formations (fig. 2).

JASPEROID

Jasperoid bodies in the region are principally in Paleozoic limestone near high-angle faults or volcanic rocks; a few are in sandstone, mudstone, or conglomerate, or in volcanic rocks of the Challis Volcanic Group. Jasperoid outcrops range from massive, gray-pink, dense cryptocrystalline

[1]U.S. Geological Survey, Denver Federal Center, Denver, Colorado 80225.

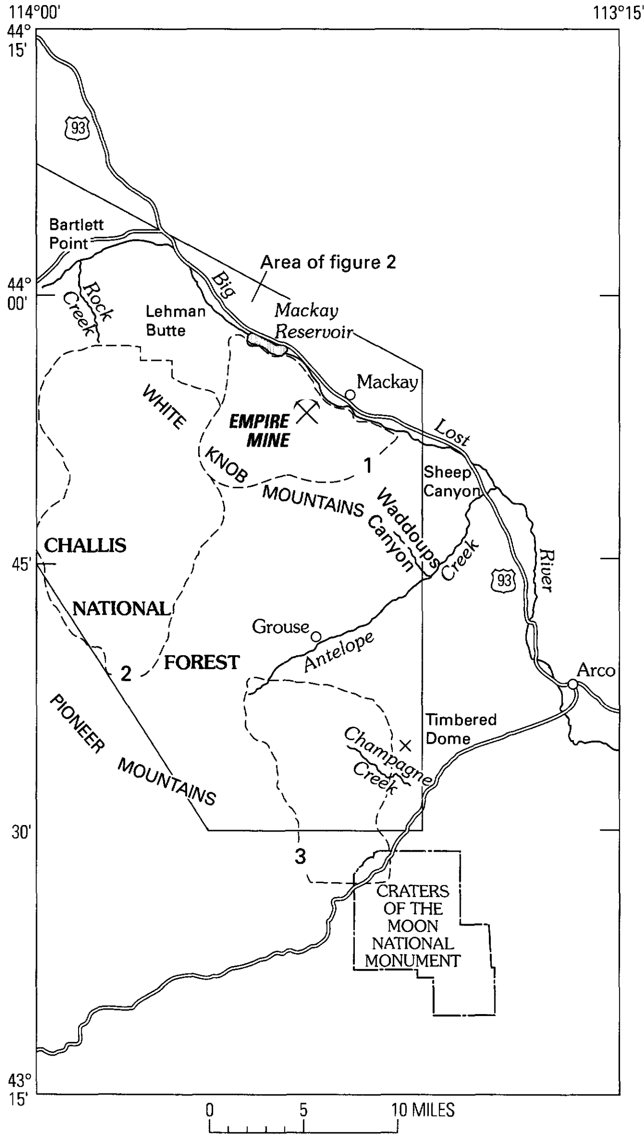


Figure 1. Map showing location of the study area in the northwestern part of the Idaho Falls 1°×2° quadrangle, Idaho. Dashed lines outline mining districts: 1, Alder Creek; 2, Copper Basin; 3, Lava Creek.

quartz bodies in which original host-rock textures are preserved to very brecciated, quartz-healed, gossan-stained siliceous bodies showing no sign of relict lithology (Wilson and others, 1988a). Some variations in grain size and texture are evident in thin section. Microtextures range from characteristic jigsaw-puzzle patterns (Lovering, 1972) that have sutured grain boundaries to granular textures. Petrographic analyses reveal a complex history of replacement followed by multiple periods of brecciation and cementation, oxidation, and weathering. Quartz and calcite veins are common, and veinlets of barite, fluorite, stibiconite, and jarosite are present locally.

One hundred and seventeen rock-chip samples were collected from jasperoid outcrops and analyzed by atomic absorption and inductively coupled plasma emission spectrographic methods. Ag, As, Au, Cu, Hg, Mo, Pb, Sb, and Zn were detected in more than 85 percent of the samples and Tl in almost half of the samples. A summary of the data set is shown in table 1. The highest gold value in this reconnaissance sampling was 0.16 ppm, from an outcrop sample near Bartlett Point (fig. 1) (Soulliere and others, 1988). The highest gold values reported from recent drilling in the Bartlett Point area are 0.4 ppm and 0.23 ppm in adjacent 5-foot intervals in pervasively silicified limestone and 0.3 ppm in partially silicified limestone (Keith Rhea, written commun., 1990).

Several theories have been put forth regarding the origin of the jasperoid bodies in the region. Nelson and Ross (1968) suggested that some jasperoid formed from fluids migrating downward from the Challis Volcanic Group into underlying limestone. Skipp suggested that some of the jasperoid may have been formed by meteoric water in karst zones, as well as in fault zones (Wilson and others, 1988a). Recent geochemical and petrographic studies by us indicate that the formation of some of the jasperoid is consistent with deposition from hydrothermal fluids (Wilson and others, 1988a).

Geochemical analyses of jasperoid show that the suite of elements Ag, Au, As, Hg, Sb, and Tl is consistently present, commonly in anomalous amounts (Soulliere and others, 1988; Wilson and others, 1988a). These elements are present in hot-springs-type mineral deposits (Berger and Silberman, 1985), geothermal systems (Silberman and Berger, 1985), and sedimentary-rock-hosted disseminated precious-metal deposits (Bagby and Berger, 1985). The presence of these elements in jasperoid samples from the Mackay area suggests that some, if not most, jasperoid formed by the replacement of sedimentary rocks with silica and associated elements that were carried in hydrothermal solutions (Wilson and others, 1988a).

Moye and others (1989) identified hot-springs deposits in volcanic rocks near Bartlett Point and Lehman Butte and at the Champagne Creek mine near Timbered Dome (fig. 1). Jasperoid bodies in sedimentary rocks at these locations may be related to these hot-springs systems. Fluids rising along structurally controlled breccia zones formed siliceous sinter in the volcanic rocks (Moye and others, 1989) and replaced sedimentary rocks with jasperoid at the surface. If the jasperoid was deposited in a shallow hot-springs environment, it is likely that fluid movement within the system was episodic.

Figure 2 (facing page). Map showing geology of the study area in the northwestern part of the Idaho Falls 1°×2° quadrangle, Idaho. Modified from Wilson and others (1988a) and Wilson and Skipp (1994).

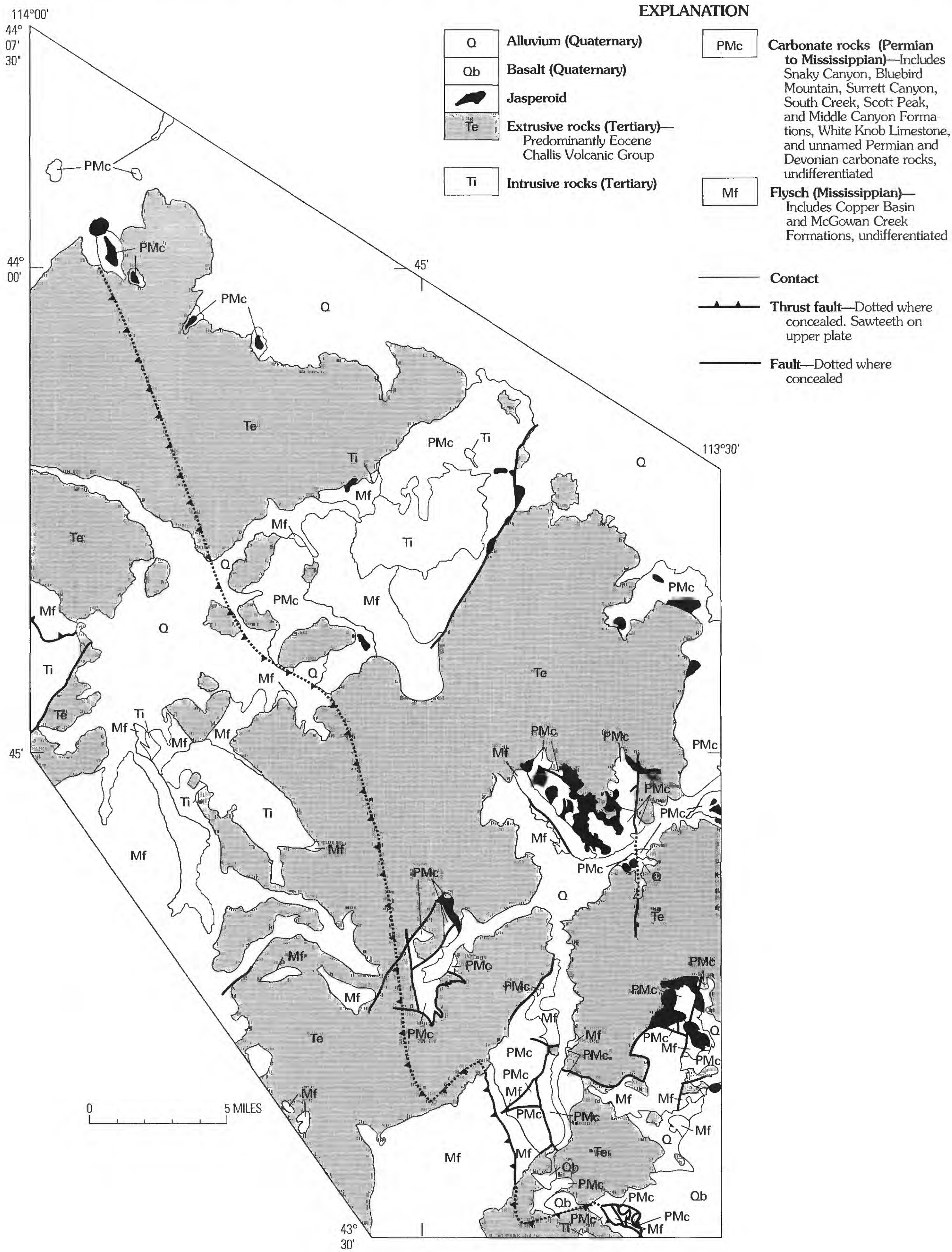

EXPLANATION
Q Alluvium (Quaternary)
Qb Basalt (Quaternary)
Jasperoid
Te Extrusive rocks (Tertiary)—Predominantly Eocene Challis Volcanic Group
Ti Intrusive rocks (Tertiary)
PMc Carbonate rocks (Permian to Mississippian)—Includes Snaky Canyon, Bluebird Mountain, Surrett Canyon, South Creek, Scott Peak, and Middle Canyon Formations, White Knob Limestone, and unnamed Permian and Devonian carbonate rocks, undifferentiated
Mf Flysch (Mississippian)—Includes Copper Basin and McGowan Creek Formations, undifferentiated
Contact
Thrust fault—Dotted where concealed. Sawteeth on upper plate
Fault—Dotted where concealed
114°00'
44° 07' 30"
44° 00'
45'
113°30'
43° 30'
0
5 MILES

Table 1. Summary of analytical results of samples from jasperoid outcrops in the northwestern part of the Idaho Falls 1°×2° quadrangle, Idaho.

[In parts per million. 117 samples were analyzed for Ag, As, Au, Hg, and Sb; 104 samples were analyzed for Cu, Mo, Pb, Tl, Zn, Bi, Cd, Ga, Pd, Pt, Pt, Se, Sn, and Te. Compiled from Soulliere and others (1988)]

	Minimum value detected	Maximum value detected	Mean	Deviation	Number of samples having detectable values
Ag	0.013	22.70	0.60	2.63	112
As	2.17	2,064	151.1	255.18	117
Au	<0.0005	0.16	0.01	0.02	106
Cu	1.24	127.00	11.36	16.38	104
Hg	0.04	20.60	1.33	2.17	102
Mo	0.34	121.00	11.46	17.74	104
Pb	0.62	5,708	74.47	560.38	104
Sb	<0.25	470.0	15.86	45.73	111
Tl	<0.44	17.70	1.36	2.22	64
Zn	<0.89	582.00	32.82	66.89	100
Bi	<0.22	18.60	0.54	1.83	22
Cd	<0.22	3.68	0.39	0.43	32
Ga	<0.44	3.52	0.86	0.69	49
Pd	<0.09	0.25	0.11	0.05	1
Pt	<0.22	0.50	0.27	0.08	9
Se	<0.87	15.90	1.86	1.91	28
Sn	<0.44	4.14	0.67	0.59	22
Te	<0.44	7.70	0.58	0.72	7

Modern hot springs associated with mineral deposition tend to deposit metallic sulfides in irregular episodic pulses (Dickson and Tunnell, 1968). Assuming that ancient hot springs acted similarly, this could account for the slight regional variations in the concentrations of elements in Mackay jasperoids. Distance from the source, pressure-temperature evolution of the fluids, lithology and chemistry of the host rocks, and structural setting also effect the geochemical associations in each area.

Oxygen-isotope studies by Snyder and Moye (1989) indicate that the hydrothermal alteration in the Lehman Basin area is related to large-scale convective movement of meteoric waters that were most likely heated by shallow intrusions or to high heat flow related to extensional tectonism. Some of the metals may have been remobilized by circulating meteoric fluids from Paleozoic sedimentary country rocks. Jasperoid bodies that are not proximal to hot springs may have formed as a result of mixing of heated and cold meteoric fluids. Research by Hofstra and others (1991) on the sedimentary-rock-hosted disseminated gold deposit at Jerritt Canyon, Nevada, indicates that heated, highly evolved meteoric fluids were responsible for mineral deposition. When these fluids reacted with the host rocks, the fluids were cooled, diluted, and oxidized, and gold and other metals precipitated. It is possible that some of the jasperoid in the study area formed in the same manner as jasperoid at Jerritt Canyon. Variation in the concentrations and diversity of elements at each geographic location may be the result of regional and local variations in temperature, fluid composition, host-rock composition, extent of fluid mixing, and extent of wallrock reaction. Detailed studies of the geochemistry, alteration, lithology, structure, and isotopic composition of the jasperoid and country rocks are needed to determine the origin of each jasperoid body.

RECENT EXPLORATION

Claims have been staked around most of the jasperoid outcrops in the Mackay area. Recent exploration activity consists mainly of detailed mapping, geochemical sampling, and some local drilling near jasperoid bodies. In 1986 and 1987, Homestake Mining Company collected samples from 2,695 ft of reverse-circulation drilling in eight holes near jasperoid outcrops at Bartlett Point (Keith Rhea, written commun., 1990).

Ore deposits at the Champagne mine near the jasperoid bodies at Timbered Dome are shallow-epithermal or hot-springs deposits in volcanic rocks localized in north-trending structures as siliceous veins and breccia zones (Moye and others, 1989). In 1988, Bema Gold Corporation of Canada began a pilot heap-leach test at the Champagne Creek mine and reported that an oxide orebody containing 2.4 million tons of minable reserves at 0.038 of gold per short ton equivalent had been defined. Production of gold and silver at the mine began in June 1989, and further exploration is continuing (Engineering and Mining Journal, 1989).

A heap-leach operation has been set up (as of 1989) about 5 mi southwest of the Champagne Creek mine along Dry Fork Creek. Jasperoid may be present at the site but could not be verified because permission to examine the property was denied.

REFERENCES CITED

Bagby, W.C., and Berger, B.R., 1985, Geologic characteristics of sediment-hosted, disseminated precious-metal deposits in the Western United States, *in* Berger, B.R., and Bethke, P.M., eds., Geology and geochemistry of epithermal systems: Society of Economic Geologists Reviews in Economic Geology, v. 2, p. 169–202.

Berger, B.R., and Silberman, M.L., 1985, Relationships of trace-element patterns to geology in hot-spring-type precious-metal deposits, *in* Berger, B.R., and Bethke, P.M., eds., Geology and geochemistry of epithermal systems: Society of Economic Geologists Reviews in Economic Geology, v. 2, p. 233–247.

Dickson, F.W., and Tunnell, G., 1968, Mercury and antimony deposits associated with active hot springs in the western United States, *in* Ridge, J.D., ed., Ore deposits of the United States, 1933–1967 (Graton-Sales volume): American Institute of Mining, Metallurgical, and Petroleum Engineers, p. 1673–1701.

Engineering and Mining Journal, 1989, A toast to the Champagne mine: Volume 190, no. 8, p. 15.

Hofstra, A.H., Leventhal, J.S., Northrup, H.R., Landis, G.P., Rye, R.O., Birak, D.J., and Dahl, A.R., 1991, Genesis of sediment-hosted disseminated-gold deposits by fluid mixing and sulfidization—Chemical-reaction-path modeling of ore-depositional processes documented in the Jerritt Canyon district, Nevada: Geology, v. 19, p. 36–40.

Lovering, T.G., 1972, Jasperoid in the United States—Its characteristics, origin, and economic significance: U.S. Geological Survey Professional Paper 710, 164 p.

Moye, F.J., Sanford, R.F., Erdman, J.A., and Johnson, K.M., 1989, Volcanic-hosted and intrusion-related mineral deposits in the Hailey and western Idaho Falls quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 70–71.

Nelson, W.H., and Ross, C.P., 1968, Geology of part of the Alder Creek mining district, Custer County, Idaho: U.S. Geological Survey Bulletin 1252–A, 130 p.

———1969a, Geology of the Mackay 30-minute quadrangle, Idaho: U.S. Geological Survey Open-File Report 70–242, 161 p.

———1969b, Geologic map of the Mackay quadrangle south-central Idaho: U.S. Geological Survey Miscellaneous Geologic Investigations Series Map I–580, scale 1:125,000.

Silberman, M.L., and Berger, B.R., 1985, Relationship of trace-element patterns to alteration and morphology in epithermal precious-metal deposits, *in* Berger, B.R., and Bethke, P.M., eds., Geology and geochemistry of epithermal systems: Society of Economic Geologists Reviews in Economic Geology, v. 2, p. 203–232.

Skipp, Betty, 1988, Geologic map of Mackay 4 (Grouse) NE quadrangle, Butte and Custer Counties, Idaho: U.S. Geological Survey Open-File Report 88–423, scale 1:24,000.

———1989, Geologic map of Mackay 4 (Grouse) NW quadrangle, Butte and Custer Counties, Idaho: U.S. Geological Survey Open-File Report 89–142, scale 1:24,000.

Skipp, Betty, and Bollmann, D.D., 1992, Geologic map of Blizzard Mountain North quadrangle, Blaine and Butte Counties, Idaho: U.S. Geological Survey Open-File Report 92–280, scale 1:24,000.

Skipp, Betty, Kuntz, M.A., and Morgan, L.A., 1990, Geologic map of Mackay 4 (Grouse) SE quadrangle, Butte Custer County, Idaho: U.S. Geological Survey Open-File Report 89–431, scale 1:24,000.

Snyder, L.G., and Moye, F.J., 1989, Regional stratigraphy, physical volcanology, and geochemistry of the southeastern Challis volcanic field, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 122–127.

Soulliere, S.J., 1992, Geochemistry and description of jasperoid near Grouse, south-central Idaho: Golden, Colorado School of Mines, M.S. thesis, 103 p.

Soulliere, S.J., Wilson, A.B., Moye, F.J., and Rhea, K.P., 1988, Analytical results and sample locality map of jasperoid samples from Mackay, Idaho: U.S. Geological Survey Open-File Report 88–589, 28 p.

Wilson, A.B., and Skipp, Betty, 1994, Geologic map of the eastern part of the Challis National Forest, Idaho: U.S. Geological Survey Miscellaneous Investigations Series Map I–2395, scale 1:250,000.

Wilson, A.B., Soulliere, S.J., Skipp, Betty, Worl, R.G., and Rhea, K.P., 1988a, Geology and geochemistry of jasperoid near Mackay, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 183–192.

———1988b, Geology and geochemistry of jasperoid from Mackay, Idaho: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 474.

Worl, R.G., and Johnson, K.M., 1989, Geologic terranes and mineral deposit types in the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 3–16.

Tertiary Sedimentary Rocks of the Payette Formation (Miocene) and the Lower Part (Miocene and Pliocene) of the Idaho Group, Hailey 1°×2° Quadrangle and Vicinity, West-Central Idaho

By Anthony B. Gibbons

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND
IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–K

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Tertiary Sedimentary Rocks of the Payette Formation (Miocene) and the Lower Part (Miocene and Pliocene) of the Idaho Group, Hailey 1°×2° Quadrangle and Vicinity, West-Central Idaho

By Anthony B. Gibbons[1]

ABSTRACT

Tertiary sedimentary rocks of the Hailey 1°×2° quadrangle and vicinity (west-central Idaho) consist of several thousands of feet of continental strata belonging to the Payette Formation (Miocene) and the lower part (Miocene and Pliocene) of the Idaho Group (Miocene to Pleistocene). Both the Payette Formation and the Idaho Group include conglomerate, sandstone, shale, and diatomite and are generally unmineralized; in addition, the Payette contains low-grade coal. A very few hydrothermal mineral deposits are present in rocks of the Payette Formation, and apparently at least one deposit is in sedimentary rocks of the Idaho Group.

Sandstone of the Payette Formation is highly arkosic and has heavy-mineral suites consistent with derivation from granitic rocks of the Idaho batholith. Three of eight panned concentrates of Payette sandstone analyzed contained unusual amounts of gold. The possibility of a paleoplacer gold resource in sandstone of the Payette Formation or in similar sandstone of the Idaho Group should be investigated by further sampling and chemical analysis.

INTRODUCTION

Tertiary sedimentary rocks in the Hailey 1°×2° quadrangle of west-central Idaho crop out in only small areas and are not known to contain significant mineral deposits. Information on these rocks, mostly from the literature, was compiled as part of the Hailey 1°×2° U.S. Geological Survey Conterminous United States Mineral Assessment Program (CUSMAP), mainly to insure that the possibility of undiscovered mineral resources in them was considered.

Tertiary sedimentary rocks in the Hailey 1°×2° quadrangle consist of the Payette Formation (Miocene) in the northwestern part of the quadrangle and the Idaho Group (Miocene-Pleistocene) in the southwestern part of the quadrangle (Worl and others, 1991). Of the two units, the Payette is of more interest with respect to mineral resources because it is present in the mineralized part of the Hailey 1°×2° quadrangle and directly overlies mineralized terranes (mainly gold mineralized) of the Idaho batholith. For this reason, a reconnaissance field study was made of the Payette Formation during three weeks in the summer of 1988 to supplement information from published sources. Published information on the chemistry of the Payette Formation is almost nonexistent, and therefore a main emphasis of the field study was sampling of the Payette at widely separated points both within and outside the Hailey 1°×2° quadrangle area (fig. 1, table 1) for chemical analysis, particularly for gold.

Acknowledgments.—Thanks are due to C.L. Smith, W.H. Raymond, T.H. Kiilsgaard, R.W. Leinz, S.K. McDanal, and G.A. Desborough for making their expertise available to the author at various stages of this study and to D.K. Allerton and D.E. Wilson for preparation of heavy-mineral concentrates of rock and sediment samples.

GEOLOGY OF THE PAYETTE FORMATION AND IDAHO GROUP

The study area is in west-central Idaho on the north side of the Snake River Plain and on the east side of the Columbia Plateau (fig. 1). The area of the Hailey quadrangle itself is largely mountainous country, much of it underlain by granitic rocks of the Idaho batholith.

Tertiary sedimentary rocks, generally interbedded with flows of basalt, crop out locally on the south and west sides of the Idaho batholith. Similar rocks may also be present in the subsurface in that part of Hailey 1°×2° quadrangle that is in the Snake River Plain (fig. 1). To the west, major areas of Tertiary crop out well beyond the quadrangle border, but small areas of Tertiary rock are present within the quadrangle near Idaho City and Atlanta. Tertiary sedimentary rocks

[1]U.S. Geological Survey, Denver Federal Center, Denver, Colorado 80225.

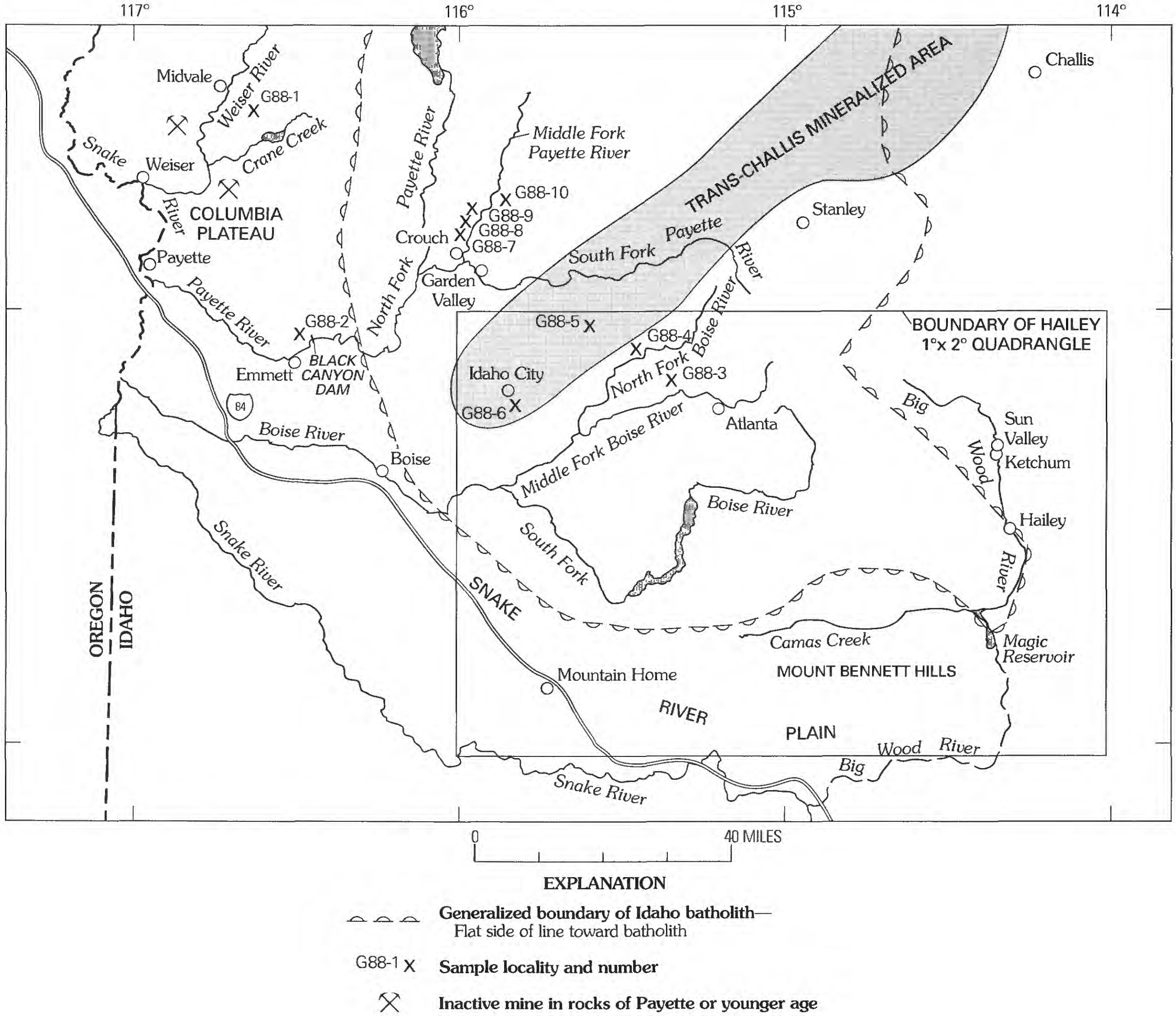


Figure 1. Map showing sample localities, Hailey 1°×2° quadrangle and vicinity, Idaho. Sample information is given in table 1. Trans-Challis mineralized area is from Kiilsgaard and others (1989).

of the study area belong to the mid-Miocene Payette Formation and to the lower part (Miocene and Pliocene) of the Idaho Group of Miocene to Pleistocene age (fig. 2).

Tertiary sedimentary rocks belonging to the lower part of the Idaho Group crop out on the south side of the Mount Bennett Hills in the southwestern part of Hailey quadrangle. These rocks comprise several hundred feet of mostly brownish sandstone and pebble gravel in lenticular channel deposits and light-colored silt, clay, and diatomite in lake deposits (Malde and others, 1963). They form a generally medial unit in the Banbury Formation (fig. 2), which consists mostly of basalt. Southward into the Snake River Plain, the Tertiary rocks, including the Banbury Formation sedimentary strata, dip beneath Quaternary deposits associated with later stages of the filling of the Snake River downwarp.

To the west of the Hailey 1°×2° quadrangle, sedimentary rocks of the Idaho Group form a sequence at least 1,800 ft thick (Kirkham, 1931) that rests with angular unconformity on a Miocene section that includes sedimentary rocks of the Payette Formation. The Payette, at least 1,180 ft thick, consists of sedimentary strata contemporaneous, and commonly interbedded, with the lavas of the Columbia River Basalt Group.

In general, the Payette Formation is lithologically similar to the sedimentary rocks of the Idaho Group as described above. Outcrops of Payette Formation along the Middle Fork of the Payette River near Crouch (fig. 1) include arkosic sandstone, conglomerate, thin-bedded dark-gray to black shale, diatomite, and tuff (Fisher and others, 1983). Thin, lenticular beds of carbonaceous shale and low-rank

Table 1. Sample information for the study.
[Sample localities are shown in figure 1]

Locality no.	Latitude (north)	Longitude (west)	Topographic quadrangle (scale 1:24,000)	Lithology sampled	Formation sampled	Stratigraphic assignment
G88–1	44°23'51"	116°35'33"	Hog Creek Butte, 1986	Sandstone	Payette Formation	Fitzgerald (1982)
G88–2	43°55' 49"	116°26'35"	Northeast Emmett, 1985	Sandstone	Payette Formation	Fitzgerald (1982)
G88–3	43°52'46"	115°23'00"	Bear River, 1972	Sandstone	Payette Formation	Kiilsgaard[1]
G88–4	43°54'42"	115°24'41"	Bear River, 1972	Gravelly sand	Stream sediment	—
G88–5	43°58'22"	115°33'36"	Big Owl Creek, 1972	Gravelly sand	Stream sediment	—
G88–6	43°49'05"	115°48'32"	Idaho City, 1976	Sandstone	Payette Formation	Lindgren (1898)
G88–7	44°09'04"	115°57'52"	Pyle Creek, 1988	Sandstone	Payette Formation	Fisher and others (1983)
G88–8	44°11'39"	115°56'20"	Pyle Creek, 1988	Coal, sandstone	Payette Formation	Fisher and others (1983)
G88–9	44°13'02"	115°55'50"	Pyle Creek, 1988	Sandstone	Payette Formation	Fisher and others (1983)
G88–10	44°16'03"	115°52'44"	Sixmile Point, 1988	Gravelly sand	Stream sediment	—

[1]U.S. Geological Survey, oral commun., March 2, 1988.

coal are also present; these are characteristic of the Payette Formation but not of younger sedimentary sequences, and thus the coaly layers constitute a lithologic marker for the Payette. Moreover, because such layers commonly contain well-preserved plant fossils, they help to date the Payette Formation, as well as the time-equivalent Columbia River Basalt Group (Knowlton, 1898; Lindgren, 1898; Smiley and others, 1975).

Despite their lithologic similarities, the Payette Formation and the Idaho Group probably represent different environments of deposition. According to Kirkham (1931), the Payette was deposited in a series of separate north-trending intramontane basins prior to development of the east-trending Snake River downwarp, whereas the Idaho Group represents deposition in one or more large lakes that formed in the downwarp concurrently with its development. Malde and Powers (1962), Kimmel (1982), and Swirydczuk and others (1982) related the stratigraphy of the Idaho Group to stages in the development of the Snake River downwarp.

Prior to and contemporaneous with downwarping of the Snake River Plain, high-angle faulting along north-south and northwest lines occurred in the tectonically positive area to the north (Malde and Powers, 1962). The faults, which have displacements of as much as several thousand feet, mostly determine the present outcrop pattern of the Tertiary rocks. Sizable outliers of Payette Formation in generally granitic terranes at Crouch and Idaho City (fig. 1) are preserved on blocks dropped down along late Cenozoic faults (Jones, 1917; Fisher and others, 1983).

Late Cenozoic faulting is commonly associated with hot-springs activity but not with hydrothermal mineralization, and the Payette Formation and Idaho Group are, with only a few local exceptions, unmineralized. Mercury deposits on an economic scale are present in the Payette Formation east of Weiser (fig. 1). Ross (1956) reported that between 1939 and 1942 deposits in this area produced almost 4,000 flasks of quicksilver from almost 53,000 tons of ore. About 10 mi north of Weiser, the Monroe Creek (Weiser) gold district consists of subeconomic occurrences of gold and mercury in and adjacent to small mafic bodies intrusive into Tertiary sedimentary rock (Varley and others, 1919; Ross, 1956). Available geologic mapping (Fitzgerald, 1982), suggests that rocks of both the Payette Formation and Idaho Group are mineralized.

Other commodities in the Tertiary sedimentary rocks of southwestern Idaho include diatomite deposits in both the Payette Formation and the Idaho Group (Powers, 1947).

PETROLOGY OF THE PAYETTE FORMATION

PETROGRAPHY

Estimated abundances of mineral species in bulk (unpanned) material for 10 disaggregated Payette rock samples and 1 stream-sediment sample are given in table 2. The grain-size fraction 35–170 mesh (0.5–0.088 mm), representing medium to very fine sand, of these samples was studied in a medium of refractive index n=1.5402, which facilitates recognition of alkali feldspar grains. Prior to study, successive heavy-liquid separations were made in bromoform (specific gravity about 2.85) and methylene iodide (specific gravity about 3.32), and strongly magnetic grains were removed using a hand magnet. The resulting nonmagnetic grain-density fractions—less than 2.85, between 2.85 and 3.32, and greater than 3.32—were examined separately. Mineral species abundances in each density fraction (table 2) were assigned according to the following scale.

Numerical designation	Semiquantitative characterization	Approximate grain percent
1	Rare	≤1
2	Not common	2–5
3	Common	5–19
4	Abundant	20–49
5	Dominant	≥50

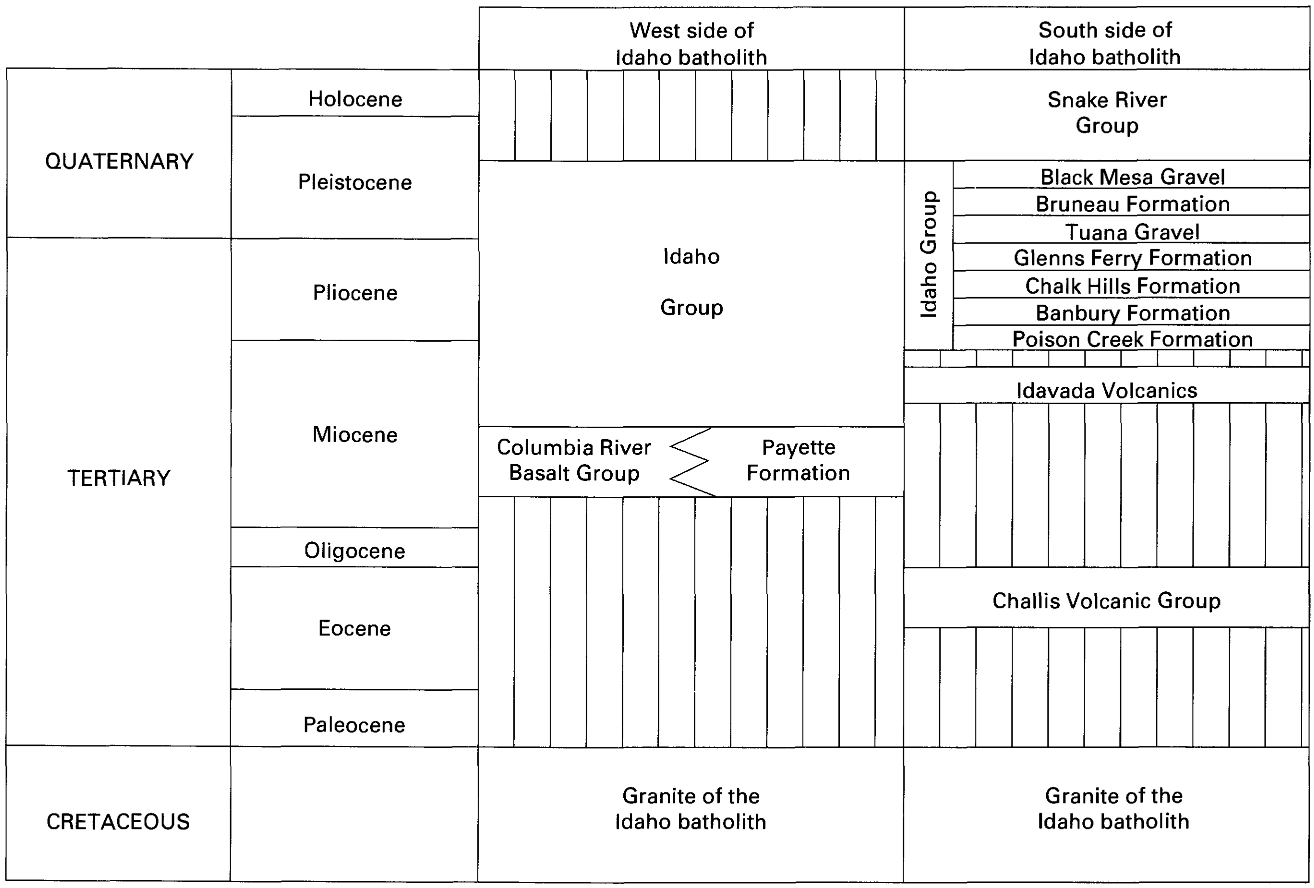


Figure 2. Generalized stratigraphy on the south and west sides of the Idaho batholith, Idaho. Subdivision of Idaho Group is from Malde and Powers (1962) and Malde and others (1963).

In the light-mineral fraction, the most striking feature is the great abundance of feldspar, particularly plagioclase, in the sandstone and sand samples (table 2). A petrographic study of the Payette Formation in the Weiser area (Nakai, 1979) indicates only a small proportion of feldspar in the sandstone. In material examined for the present study, most of the plagioclase is probably in the calcic oligoclase range. Alkali feldspar has large $2V$ and is mostly without grid twinning. Both plagioclase and alkali feldspar show much sericitic alteration in most samples.

In the moderately heavy mineral fraction, biotite is generally the most abundant mineral, then muscovite (table 2). Hornblende, of the common green variety, is negligible in most samples, but it is abundant or even dominant in some samples.

In the heaviest mineral fraction, opaque grains are very abundant, and garnet and epidote—a grain category here including some zoisite and clinozoisite—are abundant among the translucent, nonaggregate grains (table 2).

Aside from the great abundance of feldspar noted above, sand from the Payette Formation is characterized by angular to subangular grain shapes and, on average, a large proportion of heavy-mineral grains, about 0.6 weight percent of total rock. Recognizable volcanic constituents included only a few grains of probable augite and one grain of basaltic hornblende, a remarkably small amount considering the close association in space and time between the Payette Formation and the Columbia River Basalt Group. Basalt flows form part of the areal stratal succession at localities G88–1, G88–2, and G88–6 (fig. 1).

CHEMISTRY

Chemical analysis was carried out on all 10 samples—8 sandstone and 2 coal—from the Payette Formation and on 3 samples of modern stream-sediment alluvium from the area in which the Payette is present. Panning, followed by heavy-liquid and magnetic separation, was used on a portion of the

Table 2. Approximate abundance of mineral species in density fractions of nonmagnetic sand-size material from samples of the Payette Formation and from representative stream sediment derived from a granitic area, Hailey 1°×2° quadrangle and vicinity, Idaho.

[Sample information is given in table 1. Approximate mineral abundance: 1, rare; 2, not common; 3, common; 4, abundant; 5, dominant; see text for discussion]

Light mineral grains; density less than 2.85

Field no.	Sedimentary unit	Lithology	Quartz	Plagioclase	Alkali feldspar	Biotite	Muscovite	Opaques	Aggregates	Gypsum	Other
G88–1–1	Payette Formation	Sandstone	4	3	2	3	3		4		2
G88–2–1	Payette Formation	Sandstone	3	4	4	2	3		3		
G88–3–1	Payette Formation	Sandstone	4	4	3	2	2		2		
G88–3–2	Payette Formation	Sandstone	3	4	4	3	3		2		2
G88–6–1	Payette Formation	Sandstone	4	4	4	2	2		4		
G88–7–1	Payette Formation	Sandstone	4	4	2	4	3		2		
G88–8–3	Payette Formation	Sandstone	4	4	3	4	3		3		1
G88–9–1	Payette Formation	Sandstone	4	3	3	3	3	3	2		2
G88–8–1	Payette Formation	Coal	3	3	2	3	3	4	3		1
G88–8–2	Payette Formation	Coal	3	3	2	2	2		5	3	
G88–10–1	Stream sediment	Sand	4	4	3	2	2		2		2

Moderately heavy mineral grains; density between 2.85 and 3.32

Field no.	Sedimentary unit	Lithology	Biotite	Muscovite	Hornblende	Quartz	Plagioclase	Alkali feldspar	Apatite	Opaques	Aggregates	Other
G88–1–1	Payette Formation	Sandstone	4	1	5							2
G88–2–1	Payette Formation	Sandstone	5	3		2					3	2
G88–3–1	Payette Formation	Sandstone	5	4							2	2
G88–3–2	Payette Formation	Sandstone	4	4	2						3	2
G88–6–1	Payette Formation	Sandstone	5	3		2					3	
G88–7–1	Payette Formation	Sandstone	4	2	4				2		2	2
G88–8–3	Payette Formation	Sandstone	5	4		3	2	2				2
G88–9–1	Payette Formation	Sandstone	4	4		2	2	2		4		1
G88–8–1	Payette Formation	Coal	2	5						4		2
G88–8–2	Payette Formation	Coal	2	5						2		2
G88–10–1	Stream sediment	Sand	4	4		3	3	2	2			3

Heaviest mineral grains; density greater than 3.32

Field no.	Sedimentary unit	Lithology	Opaques, black	Opaques, nonblack	Aggregates	Garnet	Pale isotropics	Zircon	Sphene	Epidote	Biotite	Chlorite	Hornblende	Gypsum	Other
G88–1–1	Payette Formation	Sandstone	4	3				2	3	4	1	1	4		2
G88–2–1	Payette Formation	Sandstone	3	4		3			3	4	3	2			2
G88–3–1	Payette Formation	Sandstone	5			3	2	2	3	4					
G88–3–2	Payette Formation	Sandstone	4	4		2	2	1	3	4			1		1
G88–6–1	Payette Formation	Sandstone	4	4		3		4							2
G88–7–1	Payette Formation	Sandstone	3	4		3			3	4	1				2
G88–8–3	Payette Formation	Sandstone		4		3		2	2	2	1	4			
G88–9–1	Payette Formation	Sandstone	5		2		4	2	3						2
G88–8–1	Payette Formation	Coal	4			3		5	2		2		2		2
G88–8–2	Payette Formation	Coal	4		4	3		3						4	
G88–10–1	Stream sediment	Sand	3	4		4		2	2	4					2

Table 3. Results of instrumental neutron activation analysis of coal from the Payette Formation, Hailey 1°×2° quadrangle, and vicinity, Idaho.

[Analyst: J.S. Mee. In parts per million unless otherwise indicated. Error limits (in percent) are one standard deviation based on counting statistics alone. Sample information is given in table 1]

Laboratory number / Field number	D–323476 / G88–8–1	D–323477 / G88–8–2
Na (percent)	0.488±3	0.140±3
K (percent)	1.07±8	0.189±14
Fe (percent)	1.34±3	0.736±3
Sc	9.07±2	7.32±3
Cr	22.1±6	5.6±15
Co	13.3±3	14.8±3
Ni	<40	<40
Zn	58±23	57±6
As	4.1±8	3.44±8
Se	<2	<2
Br	3.1±12	4.2±8
Rb	<140	<80
Sr	370±11	390±13
Sb	2.14±5	1.65±5
Cs	3.63±4	0.62±11
Ba	<500	700±18
La	39.4±3	18.1±4
Ce	53.7±3	28.8±6
Nd	<60	<50
Sm	4.94±3	2.70±3
Eu	1.22±4	0.68±5
Tb	0.620±5	0.387±6
Yb	2.96±5	1.98±6
Lu	0.37±36	0.33±27
Hf	4.65±4	2.78±4
Ta	2.54±5	0.54±7
W	<1	<2
Au	<0.019	<0.012
Th	13.6±3	4.36±4
U	6.9±9	7.1±9

sandstone and alluvium samples to provide nonmagnetic heavy-mineral concentrates for analysis. (These panned concentrates were not duplicates of any of the grain-density fractions of unpanned material prepared for petrographic study.) All of the concentrates were analyzed for 37 elements by direct current-arc emission spectrography (Myers and others, 1961; Grimes and Marranzino, 1968). Four bulk sandstone and alluvium samples whose concentrates had high values for gold were analyzed for 10 elements by inductively coupled plasma-atomic emission spectrometry (ICP–AES) following acid digestion and solvent extraction (Motooka, 1988). All of the concentrates, as well as the four bulk samples selected for ICP–AES analysis, were also analyzed for gold by flameless atomic absorption spectrophotometry (Meier, 1980). The two coals were analyzed for 30 elements by instrumental neutron activation analysis (INAA) (Baedecker and McKown, 1987).

The chemical data are presented in tables 3, 4, and 5. In the discussion that follows, the significance of the analytical values for the bulk sandstone and alluvium samples and the coal samples is evaluated mainly by comparing these values with published values for average abundances of the elements (Mason, 1958; Hawkes and Webb, 1962; Levinson, 1974). The significance of the data for the concentrates can be similarly evaluated if one considers that concentrations for some elements may be enhanced by as much as three orders of magnitude relative to bulk sample data.

If viewed as the product of a granitic source area, the Payette Formation has few apparent chemical peculiarities. The two samples of Payette coal (table 3) are substantially enriched in arsenic and antimony with respect to crustal abundances, though not with respect to average shale, and are moderately enriched in uranium with respect to both standards. As a class, the sandstone samples from the Payette (tables 4, 5) are not enriched in any of the chalcophile elements. An exception is arsenic in samples G88–2–1 and G88–6–1 (table 5), which is at levels moderately above crustal abundance. With reference to the crustal standard, the Payette sandstones as a group are enriched to some degree in manganese, barium, and niobium, in the rare earth elements lanthanum and scandium, and, most significantly, in gold.

PLACER GOLD

The analyses obtained for the present study indicate that there is gold in the Payette Formation but do not indicate how much. Unusually large amounts of gold in the panned concentrates of three of the eight samples of Payette sandstone were indicated by flameless atomic absorption analysis (table 4, second column of gold values); however, bulk material from the same samples (table 5) showed no gold, either by flameless atomic absorption analysis or by solvent extraction ICP–AES. Taken together, the high gold values for concentrates and the low gold values for bulk material suggest that gold is present in rare particles separated by large volumes of barren material. Because this kind of distribution produces mostly nonrepresentative individual samples (Clifton and others, 1969), gold values obtained for the present study do not give a good idea of the overall gold content of even the Payette sandstone beds that were sampled.

The exact form in which gold is present in the Payette sandstones is not known. It probably is fairly coarse, as suggested by its ability to impart high gold values to the particular samples or sample splits in which it is present. It must also be dense because it can be concentrated by panning. No free gold was identified in any of the samples at any stage of processing. Lack of chemical or other evidence for hydrothermal mineralization in the sampled parts of the Payette implies that the gold values represent placer deposition.

Table 4. Results of semiquantitative emission spectrographic analysis of heavy-mineral concentrates of sandstones of the Payette Formation and of modern stream sediments, Hailey 1°×2° quadrangle and vicinity, Idaho.

[Sample location information in table 1. In parts per million unless otherwise indicated. Six-step semiquantitative spectrographic analysis by D.E. Detra and M. Malcolm. Results are reported to the nearest number in the series 1, 0.7, 0.5, 0.3, 0.2, 0.15, 0.1, and so on, which represent approximate midpoints of grouped data on a geometric scale. N indicates not detected at limit of detection or at value shown; < indicates detected but below limit of determination or below value shown. Second value reported for gold is by flameless atomic absorption; analysts, R.M. O'Leary and P. Roushey. Asterisk (*) is before field number of stream-sediment sample]

Lab no.	Field no.	Ca(%)	Fe(%)	Mg(%)	Na(%)	P(%)	Ti(%)	Ag	As	Au	Au
D–323478	G88–1–1	7	1	0.2	<0.5	2	>2	N1	N500	N20	351
D–323479	G88–2–1	10	1	<0.05	<0.5	3	>2	20	N500	300	135
D–323480	G88–3–1	1.5	5	0.2	0.5	2	>2	N1	N500	N20	N1.66
D–323481	G88–3–2	5	5	0.3	0.5	3	>2	N1	N500	N20	N1.176
D–323484	G88–6–1	0.2	2	0.2	<0.5	1.5	>2	N1	N500	N20	264
D–323485	G88–7–1	20	0.5	0.07	<0.5	10	>2	N1	N500	N20	N0.005
D–323486	G88–8–3	1.5	2	0.15	<0.5	7	>2	N1	N500	N20	N0.161
D–323487	G88–9–1	0.5	5	0.15	0.5	1.5	>2	N1	N500	N20	N0.869
D–323482	*G88–4–1	20	1.5	0.3	0.5	7	>2	N1	N500	N20	N0.104
D–323483	*G88–5–1	20	0.7	0.3	<0.5	10	>2	N1	N500	N20	72.6
D–323488	*G88–10–1	15	2	0.5	0.5	10	2	70	N500	1,000	0.869

Lab no.	Field no.	B	Ba	Be	Bi	Cd	Co	Cr	Cu	Ga	Ge
D–323478	G88–1–1	30	300	N2	N20	N50	N20	150	<10	20	N20
D–323479	G88–2–1	N20	200	N2	N20	N50	N20	150	<10	<10	N20
D–323480	G88–3–1	<20	500	3	N20	N50	N20	100	<10	30	N20
D–323481	G88–3–2	<20	500	5	N20	N50	N20	70	10	30	N20
D–323484	G88–6–1	N20	1,000	N2	N20	N50	N20	1,500	<10	<10	N20
D–323485	G88–7–1	N20	70	N2	N20	N50	N20	30	<10	N10	N20
D–323486	G88–8–3	N20	1,500	3	N20	N50	N20	<20	<10	10	N20
D–323487	G88–9–1	N20	>10,000	3	N20	N50	N20	<20	70	15	N20
D–323482	*G88–4–1	N20	300	2	N20	N50	N20	100	100	10	N20
D–323483	*G88–5–1	N20	150	3	N20	N50	N20	200	<10	<10	N20
D–323488	*G88–10–1	N20	300	N2	N20	N50	N20	70	<10	20	N20

Lab no.	Field no.	La	Mn	Mo	Nb	Ni	Pb	Sb	Sc	Sn	Sr
D–323478	G88–1–1	1,500	1,000	<10	100	N10	<20	N200	70	100	200
D–323479	G88–2–1	1,500	2,000	20	500	<10	20	N200	70	100	200
D–323480	G88–3–1	>2,000	7,000	N10	1,500	<10	150	N200	200	1,000	700
D–323481	G88–3–2	>2,000	7,000	<10	1,500	<10	100	N200	200	70	1,000
D–323484	G88–6–1	>2,000	5,000	N10	200	10	30	N200	150	30	N200
D–323485	G88–7–1	500	1,000	<10	200	<10	N20	N200	20	100	200
D–323486	G88–8–3	>2,000	5,000	N10	300	N10	100	N200	150	20	200
D–323487	G88–9–1	>2,000	5,000	15	500	N10	100	N200	150	1,500	300
D–323482	*G88–4–1	2,000	2,000	<10	700	<10	15,000	1,000	100	700	500
D–323483	*G88–5–1	1,500	1,500	N10	500	<10	1,000	N200	70	70	300
D–323488	*G88–10–1	>2,000	7,000	N10	300	<10	100	N200	200	N20	300

Lab no.	Field no.	Th	V	W	Y	Zn	Zr	Pd	Pt
D–323478	G88–1–1	200	500	N50	500	N500	>2,000	N5	N20
D–323479	G88–2–1	200	700	N50	700	N500	>2,000	N5	N20
D–323480	G88–3–1	3,000	300	N50	1,500	N500	>2,000	N5	N20
D–323481	G88–3–2	3,000	300	N50	1,500	N500	>2,000	N5	N20
D–323484	G88–6–1	3,000	150	N50	1,000	N500	>2,000	N5	N20
D–323485	G88–7–1	N200	300	N50	500	N500	>2,000	N5	N20
D–323486	G88–8–3	700	150	N50	700	N500	>2,000	N5	N20
D–323487	G88–9–1	1,000	150	N50	700	N500	>2,000	N5	N20
D–323482	*G88–4–1	700	150	200	1,000	N500	>2,000	N5	N20
D–323483	*G88–5–1	500	150	100	700	N500	>2,000	N5	N20
D–323488	*G88–10–1	>5,000	70	N50	3,000	N500	>2,000	N5	N20

Table 5. Results of inductively coupled plasma-atomic emission spectrographic (ICP-AES) analysis of sandstones of the Payette Formation and modern stream sediments, Hailey 1°×2° quadrangle, and vicinity, Idaho.

[Sample location information in table 2. In parts per million. Samples analyzed by inductively coupled plasma–atomic emission spectrography using diisobutyl ketone extraction by P.H. Briggs and J.M. Motooka. N indicates not detected at limit of detection or at value shown. Second value shown for gold is by flameless atomic absorption (analysts, R.M. O'Leary and P. Roushey). Asterisk (*) is before field number of stream-sediment sample]

Lab no.	Field no.	Ag	As	Au	Au	Bi	Cd	Cu	Mo	Pb	Sb	Zn
D–341248	G88–1–1	0.075	N0.60	N0.15	N0.002	N0.60	0.030	4.5	N0.090	5.0	N0.60	22
D–341249	G88–2–1	0.045	9.3	N0.15	N0.002	N0.60	N0.030	1.0	0.28	2.4	N0.60	7.7
D–341250	G88–6–1	0.057	10	N0.15	N0.002	N0.60	N0.030	4.0	1.7	6.7	N0.60	13
D–341251	*G88–10–1	N0.045	N0.60	N0.15	N0.002	N0.60	N0.030	0.51	N0.090	1.4	N0.60	11

POTENTIAL FOR GOLD RESOURCES OF THE PAYETTE FORMATION AND IDAHO GROUP

The chemical data resulting from reconnaissance sampling of the Payette Formation indicate the possibility of a gold rs ource in the Payette. The three sandstone samples that have high gold values in the concentrates are from widely separated locations (fig. 1, localities G88–1, G88–2, and G88–6). In addition, the samples are from widely separated levels above granitic or other pre-Payette, pre-Columbia River basement terranes. Inferred vertical distance to basement is on the order of 3,500 and 1,500 ft, respectively, for samples G88–1–1 and G88–2–1, on the basis of data in Fitzgerald (1982), and on the order of 600 ft for sample G88–6–1, on the basis of data in Jones (1917).

In the analytical data for the Payette Formation, high gold values have a high incidence. This is true if incidence is taken at the level of three occurrences of high values out of eight concentrates, on the basis of atomic absorption analysis, or even at the level of one such occurrence out of eight samples, on the basis of spectrographic analysis.

A comparison with the Payette data is furnished by the incidence of detectable gold in the semiquantitative spectrographic analyses of a select group of 74 concentrates of modern stream-sediment samples. The sediment samples were collected from that part of the drainage basin of the South Fork of the Payette River within the Trans-Challis mineralized area of Kiilsgaard and others (1989) (fig. 1). Of these alluvium samples, taken from streams draining terrain marked by many mines and prospects for gold, lead, and silver, only one contained detectable gold (McDanal and others, 1984).

The analyses of the alluvial concentrate samples cited above and the analyses of the Payette concentrate samples collected for this study are directly comparable in most respects. The two groups of samples were analyzed in the same laboratory and received exactly the same laboratory processing; however, field panning was by different operators, and the Payette samples were panned to a volume of perhaps a fifth of what is usual. Thus, a possible concentration of gold in samples of the Payette Formation relative to the alluvial samples must be taken into account. The relative enhancement of gold values in the Payette would probably be somewhat more than fivefold. Even so, these values seem very significant, and the incidence of high gold values in sandstone of the Payette Formation compares more than favorably with the incidence of high gold values in alluvium selected to be gold rich.

Overall, the results of chemical analysis of the Payette Formation are distinctly positive for gold. Unusually high gold values are indicated to be present in the Payette. Moreover, they apparently are widely distributed and of common occurrence.

SUGGESTIONS FOR FURTHER STUDY

The possibility of significant placer gold in the Payette was considered by Lindgren (1898, p. 668–669). Although Lindgren felt that there might be placer gold in the very basal part of the Payette Formation at some locations, his evaluation was essentially negative. His evaluation reflects a want of essential information, the presence of gold in nonbasal parts of the Payette, as shown by the present study. His evaluation also reflects the presence of competing resources. The gold lode mines and Quaternary placers of southwest-central Idaho were readily amenable to the exploration and extraction technology of Lindgren's day and were then still rich. In the very different economic and technical context of today, re-examination of the placer gold potential of the Payette may be in order. Sedimentary rocks of the Idaho Group are similar in lithology and apparent provenance to those of the Payette Formation and should be included in any program of further sampling for placer gold in the Tertiary sedimentary rocks of the Hailey 1°×2° quadrangle and vicinity.

REFERENCES CITED

Baedecker, P.A., and McKown, D.M., 1987, Instrumental neutron activation analysis of geochemical samples, *in* Baedecker, P.A., ed., Methods for geochemical analysis: U.S. Geological Survey Bulletin 1770, p. H1–H14.

Clifton, H.E., Hunter, R.E., Swanson, F.J., and Phillips, R.L., 1969, Sample size and meaningful gold analysis: U.S. Geological Survey Professional Paper 625–C, p. C1–C17.

Fisher, F.S., McIntyre, D.H., and Johnson, K.M., 1983, Geologic map of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 83–523, scale 1:250,000, 41 p.

Fitzgerald, J.F., 1982, Geology and basalt stratigraphy of the Weiser embayment, west-central Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 103–128.

Grimes, D.J., and Marranzino, A.P., 1968, Direct-current arc and alternating-current spark emission spectrographic field methods for the semiquantitative analysis of geologic materials: U.S. Geological Survey Circular 591, 6 p.

Hawkes, H.E., and Webb, J.S., 1962, Geochemistry in mineral exploration: New York, Harper and Row, 415 p.

Jones, E.L., Jr., 1917, Lode mining in the Quartzburg and Grimes Pass porphyry belt, Boise Basin, Idaho: U.S. Geological Survey Bulletin 640–E, p. 83–111.

Kiilsgaard, T.H., Fisher, F.S., and Bennett, E.H., 1989, Gold-silver deposits associated with the trans-Challis fault system, Idaho, *in* Shawe, D.R., and Ashley, R.P., eds., United States gold terranes—Part I: U.S. Geological Survey Bulletin 1857–B, p. B22–B44.

Kimmel, P.G., 1982, Stratigraphy, age, and tectonic setting of the Miocene-Pliocene lacustrine sediments of the western Snake River Plain, Oregon and Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 559–578.

Kirkham, V.R.D., 1931, Revision of the Payette and Idaho Formations: Journal of Geology, p. 39, no. 3, p. 193–239.

Knowlton, F.H., 1898, The fossil plants of the Payette Formation: U.S. Geological Survey 18th Annual Report, pt. 3, p. 721–744.

Levinson, A.A., 1974, Introduction to exploration geochemistry: Calgary, Canada, Applied Publishing, 612 p.

Lindgren, W., 1898, The mining districts of the Idaho Basin and the Boise Ridge, Idaho: U.S. Geological Survey 18th Annual Report, pt. 3, p. 625–719.

Malde, H.E., and Powers, H.A., 1962, Upper Cenozoic stratigraphy of the western Snake River Plain, Idaho: Geological Society of America Bulletin, v. 73, p. 1197–1220.

Malde, H.E., Powers, H.A., and Marshall, C.H., 1963, Reconnaissance geologic map of west-central Snake River Plain, Idaho: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–373, scale 1:125,000.

Mason, Brian, 1958, Principles of geochemistry: New York, John Wiley and Sons, 310 p.

McDanal, S.K., Cooley, E.F., and Callahan, J.E., 1984, Analytical results of stream-sediment and non-magnetic heavy-mineral-concentrate samples with locality map from portions of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 84–634, 287 p.

Meier, A.L., 1980, Flameless atomic absorption determination of gold in geological materials: Journal of Geochemical Exploration, v. 13, p. 77–85.

Motooka, J.M., 1988, An exploration geochemical technique for the determination of preconcentrated organometallic halides by ICP–AES: Applied Spectroscopy, 42, p. 1293–1296.

Myers, A.T., Havens, R.G., and Dunton, P.J., 1961, A spectrochemical method for the semiquantitative analysis of rocks, minerals, and ores: U.S. Geological Survey Bulletin 1084–I, p. I1–I229.

Nakai, T.S., 1979, Stratigraphy of the Payette Formation, Washington County, Idaho: Moscow, University of Idaho, M.S. thesis, 187 p.

Powers, H.A., 1947, Diatomite deposits of southwestern Idaho: Idaho Mineral Resources Report 4, 27 p.

Ross, C.P., 1956, Quicksilver deposits near Weiser, Washington County, Idaho: U.S. Geological Survey Bulletin 1042–D, p. 79–104.

Smiley, C.J., Shah, S.M., and Jones, R.W., 1975, Later Tertiary stratigraphy and paleobotany of the Weiser area, Idaho: Idaho Bureau of Mines and Geology Information Circular 28, 12 p.

Swirydczuk, K., Larson, G.P., and Smith, G.R., 1982, Volcanic ash beds as stratigraphic markers in the Glenns Ferry and Chalk Hills Formations from Adrian, Oregon, to Bruneau, Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 543–558.

Varley, Thomas, Wright, C.A., Soper, E.K., and Livingston, D.C., 1919, A preliminary report on the mining districts of Idaho: U.S. Bureau of Mines Bulletin 166, p. 1–89.

Worl, R.G., Kiilsgaard, T.H., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Interpretation of the Regional Geochemistry of the Hailey 1°×2° Quadrangle, South-Central Idaho

By Cole L. Smith

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–L

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Interpretation of the Regional Geochemistry of the Hailey 1°×2° Quadrangle, South-Central Idaho

By Cole L. Smith[1]

ABSTRACT

Results of 1,914 analyses of sediment samples were used to interpret the regional geochemistry of the Hailey 1°×2° quadrangle, south-central Idaho. Many of the samples, 1,240, were originally collected for the National Uranium Resource Evaluation (NURE) program. The low sampling density in this geochemical study makes it difficult to recognize individual mineral deposits. Only large geochemical features can be reliably distinguished by their characteristic suites of elements in this survey. The geochemistry was interpreted by using suites of elements rather than individual elements. The suites were chosen because they characterize either rocks associated with some types of mineral deposits or primary alteration halos associated with some types of mineral deposits.

Two geochemical suites, Li+Be and Li+Be+Nb+Th+Y, are associated with evolved rocks. The geochemical suite Li+Be+Nb+Th+Y is associated with outcrops of Eocene granite and with outcrops of some Miocene to Pliocene rhyolites. The element suite Li+Be tends to be associated with pegmatites and veins that formed near the margins of the Eocene granite.

Four element suites, Sb+As+Au, Ag+Pb, Zn+Cu, and Bi+Mo, are used to recognize primary geochemical halos formed by mineralizing events. The element suite Bi+Mo is commonly associated with plutonic margins, and the element suite Sb+As+Au is commonly distal from plutonic margins. The element suites Zn+Cu and Ag+Pb are commonly between the other two element suites. Zoning patterns effectively outline most known mineral deposits and prospects in the Hailey 1°×2° quadrangle. Some of the element suites are anomalous at sites not associated with significant known mining activity. If supported by geologic information, this geochemical information can be useful for both exploration and resource assessment.

[1]U.S. Geological Survey, W. 904 Riverside Avenue, Spokane, Washington 99201.

INTRODUCTION

This report is a description and interpretation of the regional geochemistry of the Hailey 1°×2° quadrangle, south-central Idaho (fig. 1). The Hailey quadrangle, which covers an area of about 6,800 mi^2, was sampled at a density of about one sample per 4.2 mi^2. Work was performed as part of the U.S. Geological Survey Conterminous United States Mineral Assessment Program (CUSMAP) study of the Hailey quadrangle, which was started in the fall of 1986; fieldwork was completed in the summer of 1989.

Previously published reports (table 1) that incorporate geochemistry discuss only small parts of the Hailey quadrangle. The only geochemical data set that covers the entire Hailey quadrangle with widely spaced sample sites is from the National Uranium Resource Evaluation (NURE) program. In order to limit the expense of sample collection, the U.S. Geological Survey (USGS) reanalyzed NURE samples collected by Savannah River Laboratories and analyzed by Oak Ridge Gaseous Diffusion Plant (ORDGP). Samples were collected for this study to fill gaps in NURE coverage. Personnel from the USGS, mostly volunteers, collected samples during the summer of 1987. Three areas of special interest were more densely sampled by the USGS in the summers of 1988 and 1989.

A major goal of this study was to provide an interpretation of how mineralized areas in the Hailey quadrangle fit into a regional geochemical framework. Another goal was to use multielement analytical methods to provide geochemical information about a wide range of mineral deposit types and their associated alteration halos. The emphasis of the study was on those mineral deposit types whose occurrence is plausible within the Hailey quadrangle (Worl and Johnson, this volume). Many deposit types described in Cox and Singer (1986) cannot be distinguished by their chemical characteristics alone. Nash (1988, p. 6), after studying the chemical analyses of more than 2,000 mineralized rock samples from the Tonopah 1°×2° quadrangle in Nevada, stated "that there are no simple chemical distinctions between the many ore types because element concentrations usually have overlapping ranges." Geological characteristics must be

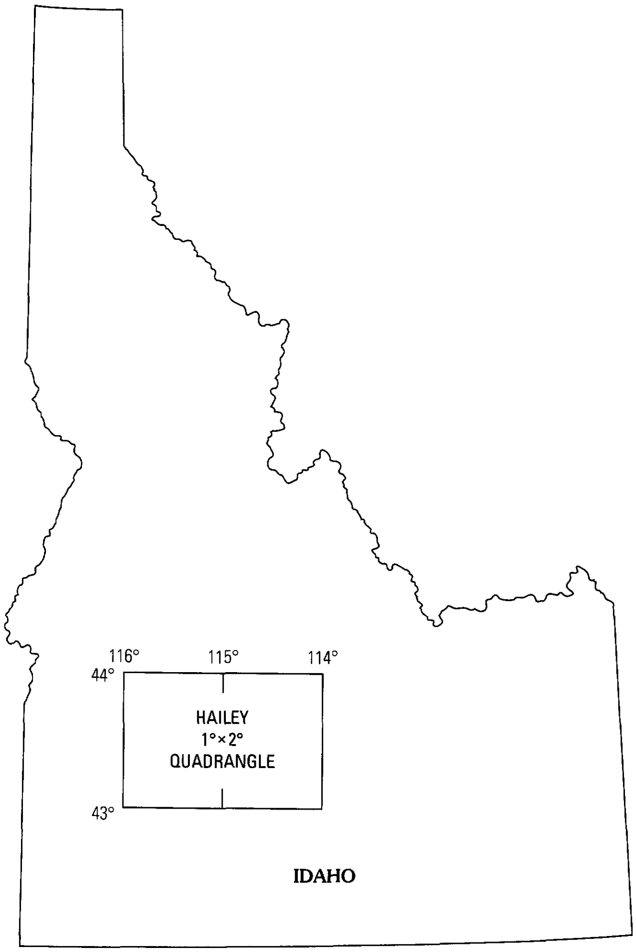


Figure 1. Index map showing location of the Hailey 1°×2° quadrangle, south-central Idaho.

considered to separate one deposit type from another; however, for many common metalliferous deposits, the vertical zoning sequence of primary geochemical halos is somewhat independent of geologic environment (Ovchinnikov and Grigoryan, 1971).

Acknowledgments.—Many people, both within and outside the USGS, contributed to this geochemical study. Samples were collected by many volunteers including Craig Fanshier, Craig Bogdon, Brian Nelson, and Keegan Schmidt from outside the USGS. Janice Nadeau, an employee of the USGS, collected samples for this study during her vacation. Jack Antweiler, a retired employee of the USGS, also collected samples. Employees of the USGS, those from other projects, who collected samples were Betty Bailey, Helen Whitney, Janet Jones, and Jerry Gaccetta. Chris McDougal both controlled sample flow of USGS samples and helped to locate and recover samples collected for the NURE program. Paul Briggs, Mollie Malcolm, and Rich O'Leary provided analyses of samples collected for the USGS. Jerry Motooka analyzed USGS samples and reanalyzed samples collected for the NURE program. Gary Selner and Dick Taylor designed and wrote a computer program to digitize sample locality information, as well as a program to display geochemical information. Joe Christie and Larry Jackson wrote computer programs to automate sample submittal for this project. Jim Hoffman and Wendy Speckman helped retrieve existing geochemical data from the NURE program. Ron Worl, Reed Lewis, Earl Bennett, Thor Kiilsgaard, and Kate Johnson provided helpful advice on geology. Thor Kiilsgaard and Ken Watts thoroughly reviewed the manuscript. I thank these and other people for their contributions to this project.

METHODS OF STUDY

SAMPLING AND CHEMICAL ANALYSIS

Samples of soil and samples of stream sediment from the channels of dry streams were collected during the NURE program; details of sampling procedures are given in Ferguson and others (1977). In 1987, the USGS collected and analyzed 337 samples of stream sediment from channels of both dry and flowing streams; details of sampling procedures are given in Malcolm and Smith (1992). In order to integrate NURE geochemical data with that of the USGS and to save sample collecting costs, the USGS reanalyzed 1,240 samples that Savannah River Laboratories had collected for the NURE program. Analytical methods for both groups of samples are summarized in table 2. Limits of determination for the methods of analysis are given in table 3. Analytical results for both the samples collected and analyzed by the USGS (337 samples) and the samples reanalyzed by the USGS (1,240 samples) were used to interpret the regional geochemistry of the Hailey quadrangle.

Because of the cost and time required to analyze 1,577 samples, only a limited number of analytical techniques could be used. Multielement analytical techniques were chosen because they determine the concentrations of a wide range of elements that are characteristic of many types of mineral deposits. An alternative to using multielement techniques would have been to select a few single-element analytical methods that have limits of detection at or near crustal abundance for the analyzed element; however, this would not have provided information about many types of deposits that are likely to be present in the Hailey quadrangle. Even using multielement analytical procedures, the geochemical sampling and analysis program was not designed to evaluate the Hailey quadrangle for the occurrence of industrial minerals.

In the summers of 1988 and 1989, USGS volunteers collected 106 samples in three areas of special interest. Both stream-sediment samples and their heavy-mineral fractions were analyzed. Sediment samples collected from the special study areas were analyzed by the USGS methods previously

Table 1. Previous geochemical studies in the Hailey 1°×2° quadrangle, south-central Idaho.

Study name	Size (mi^2)	Sample type	Number of samples	Reference
Boulder-Pioneer Wilderness	450	Stream sediment	1,151	Simons (1981)
		Soil	161	
		Rock	783	
White Cloud-Boulder Wilderness	78.6	Stream sediment	101	Fisher and May (1983).
		Heavy-mineral concentrate	73	
		Rock	172	
Eastern part of the Sawtooth National Recreation Area	820	Stream sediment	2,875	Tschanz and Kiilsgaard (1986).
		Heavy-mineral concentrate	255	
		Rock	1,506	
Sawtooth Primitive Area	314	Stream sediment	820	Kiilsgaard and others (1970).
		Soil	15	
		Heavy-mineral concentrate	78	
		Rock	359	
Ten Mile Roadless Area	134	Stream sediment	313	Kiilsgaard (1982).
		Rock	271	
King Hill Creek Wilderness	43.2	Stream sediment	43	Toth and others (1987).
		Heavy-mineral concentrate	28	
		Rock	70	
National Uranium Resource Evaluation (NURE) of the Hailey quadrangle	6,937	Stream sediment	610	Grimes (1982a).
		Soil	638	

Table 2. NURE and USGS sampling programs in the Hailey 1°×2° quadrangle, south-central Idaho.

	NURE		USGS
Sample media	Sediments from dry stream beds	Soils	Sediments from the beds of flowing streams
Number of samples	608	638	337
Method of analysis	Plasma source emission spectrometry, neutron activation (Grimes, 1982b) Plasma source emission spectrometry after partial extraction (Motooka, 1988)	Plasma source emission spectrometry, neutron activation (Grimes, 1982b) Plasma source emission spectrometry after partial extraction (Motooka, 1988)	Plasma source emission spectrometry (Crock and others, 1983)
Results reported	Grimes (1982a) Malcolm and Smith (1992)	Grimes (1982a) Malcolm and Smith (1992)	Malcolm and Smith (1992)

described. The nonmagnetic heavy-mineral fractions of the stream-sediment samples were analyzed for 36 elements by a semiquantitative emission spectrographic method. Results of the analyses and a description of the analytical method is given in Malcolm and Smith (1992).

DATA PROCESSING

A microcomputer and a combination of commercial computer programs and computer programs written by employees of the USGS were used to aid interpretation of the data generated by the Hailey CUSMAP project. Much of the data manipulation was done using commercial spreadsheet programs. The commercial computer programs SYSTAT and SYGRAPH (by SYSTAT Inc.) were used for statistical and some graphical applications. Illustrations were originally produced using a combination of USGS programs (GSPOST and GSMAP) and a commercial program (MASS11DRAW, by Microsystems Engineering Corp.).

The regional geochemistry of the Hailey quadrangle was interpreted using groups of elements (geochemical suites) that were chosen to reflect the vertical zonation (Tauson and others, 1971) of primary geochemical halos.

Table 3. Limits of detection for USGS and NURE methods of analysis
[In parts per million unless otherwise indicated. ICP indicates inductively coupled plasma; partial indicates partial extraction. Leaders (--) indicates element not analyzed for. Crustal abundances from Rose and others (1979, p. 30)]

Element	NURE ICP	USGS ICP	USGS ICP partial	Neutron activation	Crustal abundance
Ag	2	2	0.05	--	0.05
Al (percent)	0.05	0.05	--	--	8.1
As	--	10	1.0	--	2
Au	--	8	0.3	--	0.003
B	10	--	--	--	8
Ba	2	1	--	--	580
Be	1	1	--	--	2
Bi	--	10	1.0	--	0.1
Ca (percent)	0.05	0.05	--	--	3.3
Cd	--	2	0.075	--	0.1
Ce	10	4	--	--	81
Co	4	1	--	--	25
Cr	1	1	--	--	100
Cu	2	1	0.1	--	50
Eu	--	2	--	--	1.1
Fe (percent)	0.05	0.05	--	--	4.65
Ga	--	4	--	--	26
Hf	15	--	--	--	3
Ho	--	4	--	--	0.78
K (percent)	0.05	0.05	--	--	2.5
La	2	2	--	--	25
Li	1	2	--	--	30
Mg (percent)	0.05	0.005	--	--	1.7
Mn	4	4	--	--	1,000
Mo	4	2	0.25	--	1.5
Na (percent)	0.05	0.005	--	--	2.5
Nb	4	4	--	--	20
Nd	--	4	--	--	16
Ni	2	2	--	--	75
P (percent)	0.0005	0.005	--	--	0.09
Pb	10	4	1.5	--	10
Sb	--	--	1.5	--	0.1
Sc	1	2	--	--	13
Sn	--	10	--	--	2
Sr	1	2	--	--	300
Ta	--	40	--	--	2
Th	2	4	--	--	10
Ti (percent)	0.001	0.005	--	--	0.44
U	--	100	--	0.02	2.5
V	2	2	--	--	150
Y	1	2	--	--	20
Yb	--	1	--	--	2.2
Zn	2	2	0.1	--	80
Zr	2	--	--	--	150

Table 4. Rank sum example.
[Concentrations are in parts per million]

	Concentration		Rank		Rank Sum	Anomaly threshold		
Sample site	Ag	Pb	Ag	Pb	Ag+Pb	Ordered sample sites		Percentile
AA01	43.0	21	20	1.5	21.5	AA04	39	95
AA02	0.51	37	10	9.5	19.5	AA18	37	90
AA03	0.91	42	16	12	28	AA17	35	85
AA04	1.20	75	19	20	39	AA06	31	80
AA05	0.72	45	13	14	27	AA16	30	75
AA06	0.81	57	14	17	31	AA03	28	70
AA07	0.09	32	2	6.5	8.5	AA11	28	65
AA08	0.31	44	6	13	19	AA05	27	60
AA09	0.42	34	8	8	16	AA01	21.5	55
AA10	0.53	37	11	9.5	20.5	AA10	20.5	50
AA11	0.67	48	12	16	28	AA13	20	45
AA12	0.22	28	4	4	8	AA02	19.5	40
AA13	0.49	41	9	11	20	AA08	19	35
AA14	0.08	29	1	5	6	AA09	16	30
AA15	0.32	32	7	6.5	13.5	AA15	13.5	25
AA16	0.85	46	15	15	30	AA07	8.5	20
AA17	0.92	59	17	18	35	AA12	8	15
AA18	1.07	63	18	19	37	AA20	8	10
AA19	0.17	21	3	1.5	4.5	AA14	6	5
AA20	0.23	27	5	3	8	AA19	4.5	0

A nonparametric statistical method (Goldfarb and others, 1983) that combines information about chemical components having widely different concentration ranges was chosen to manipulate the geochemical data. This method was used to assign scores to each sample site for each of the groups of elements that were used to characterize primary geochemical halos. An unpublished computer program called "RANKS" was used to organize, in descending order, the concentrations of chemical components in the geochemical database and to replace concentration with rank. For equal concentrations the ranks were averaged, and concentrations below the limit of determination were assigned a rank of 0.

Table 4 gives an example of how geochemical suites were used to define anomaly thresholds for this study. In this example, the geochemical suite (Ag+Pb) is discussed. Data are shown for 20 samples sites. Silver concentrations range from 0.08 to 43 ppm and lead concentrations from 21 to 75 ppm. Anomaly thresholds were chosen at the 95th and 90th percentile. If a sample-site score is at the 95th percentile, it means that 95 percent of all sample-site scores are below that score and 5 percent of all sample-site scores are above that score. For 20 samples, only one sample site is anomalous at the 95th percentile, site AA04, with a rank sum of 39. Another sample site is anomalous at the 90th percentile, site AA18, with a rank sum of 37. Because a suite of elements, rather than a single element, is used to define anomaly thresholds, a sample site having a highly anomalous concentration of a single element of the suite might not have an anomalous rank sum value. For example, sample site AA01 has a very anomalous silver concentration of 43 ppm; all other silver concentrations are less than or equal to 1.2 ppm. Sample site AA01 is anomalous, however, only at the 55th percentile. By using suites of elements rather than single elements to interpret the regional geochemistry of the Hailey quadrangle, spurious analytical results involving single elements tend to be filtered out.

Concentrations of silver, copper, lead, molybdenum, and zinc were determined using more than one analytical method. Because of its lower limits of detection (table 3), the analytical results of the USGS partial-extraction ICP method (Motooka, 1988) were used to rank Ag, As, Au, Bi, Cu, Pb, Mo, Sb, and Zn.

Parts of the geochemical data set were affected by the following factors: (1) concentrations of some elements were determined by more than one method of analysis; (2) unmineralized samples of some rock types, such as black shale, have high background concentrations of some elements relative to other unmineralized rocks; and (3) three different sample media were collected and used to interpret the Hailey geochemistry. Not all elements and element suites were affected to the same degree by these factors. Where necessary, the Hailey geochemical data set was

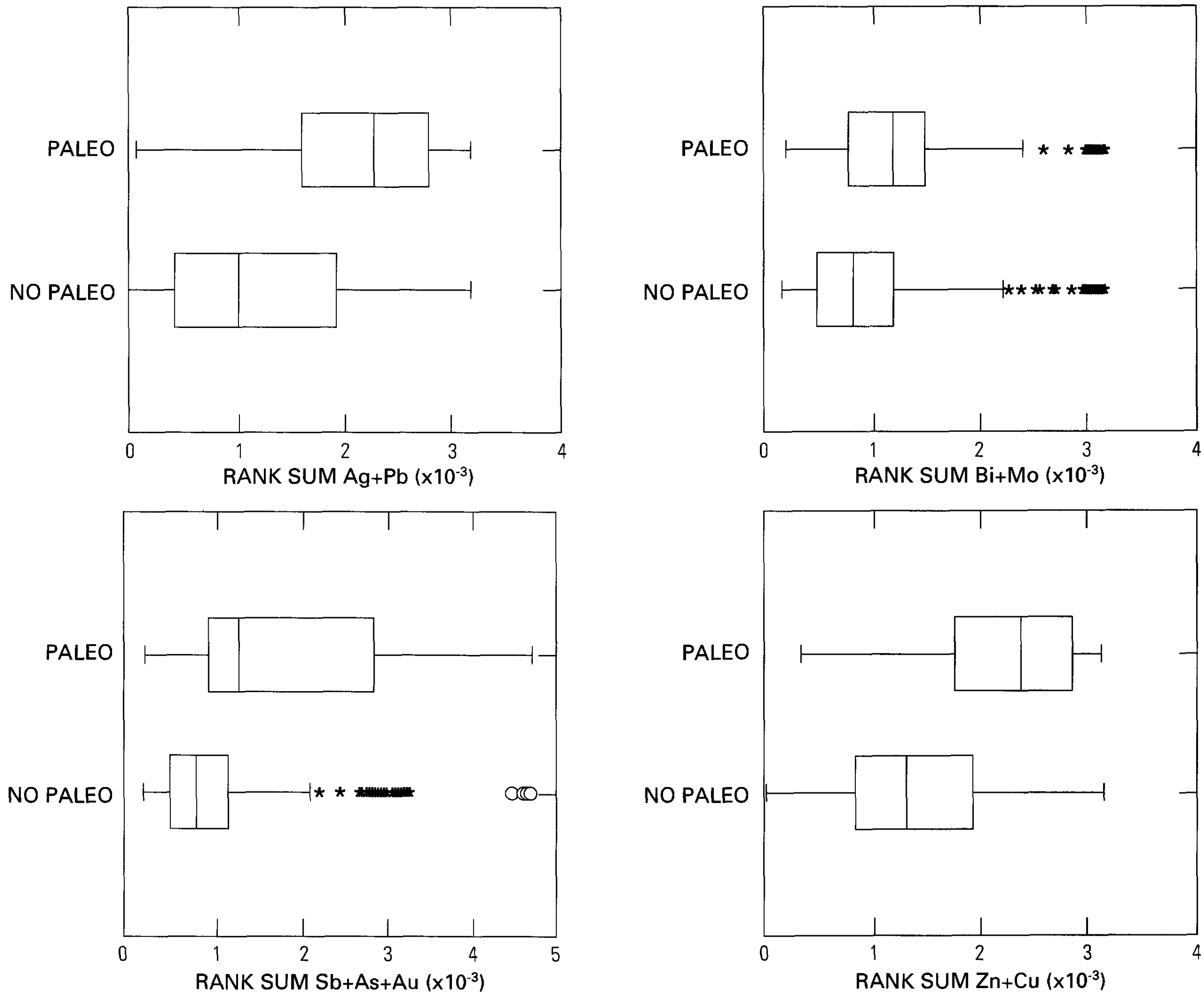


Figure 2. Box plots of sample-site scores within area of Paleozoic rock outcrops (PALEO) and outside of area of Paleozoic outcrops (NO PALEO), Hailey 1°×2° quadrangle, south-central Idaho. See text for discussion.

separated into subsets, and sample-site scores were calculated separately for each subset.

Concentrations of the elements beryllium, lithium, niobium, thorium, and yttrium were determined by different analytical methods at the USGS and ORDGP laboratories. For geochemical suites containing these elements, sample-site scores for samples analyzed by the USGS laboratory were calculated separately from those for samples analyzed by ORDGP.

Samples of black shale of Paleozoic age in the Hailey quadrangle have high background concentrations (Simons, 1981) of elements that are used in some of the geochemical suites. The southern and western extent of the Paleozoic rocks, including black shale, is known; however, the distribution of black shale within the area of Paleozoic rocks is not as well known. Because of this, subsets of the Hailey geochemical data were formed of sample sites associated with Paleozoic rocks and sample sites not associated with Paleozoic rocks. For geochemical suites containing Ag, As, Au, Bi, Cu, Mo, Pb, Sb, or Zn, sample-site scores of samples collected from the area where sample sites might be associated with Paleozoic black shale were calculated separately from sample-site scores of samples collected elsewhere.

Box plots graphically summarize data distribution (Chambers and others, 1983). The box plots in figure 2 show that median sample-site scores within the area where sample sites might be associated with Paleozoic black shale are higher than median sample-site scores outside this area. For several reasons not all sample sites within this area are associated with high background values: (1) rocks other than those of Paleozoic age crop out within the area of Paleozoic rock outcrop; (2) not all Paleozoic rocks are metal-rich black shale; and (3) no detailed studies have been made of metal distribution in all black-shale units, and

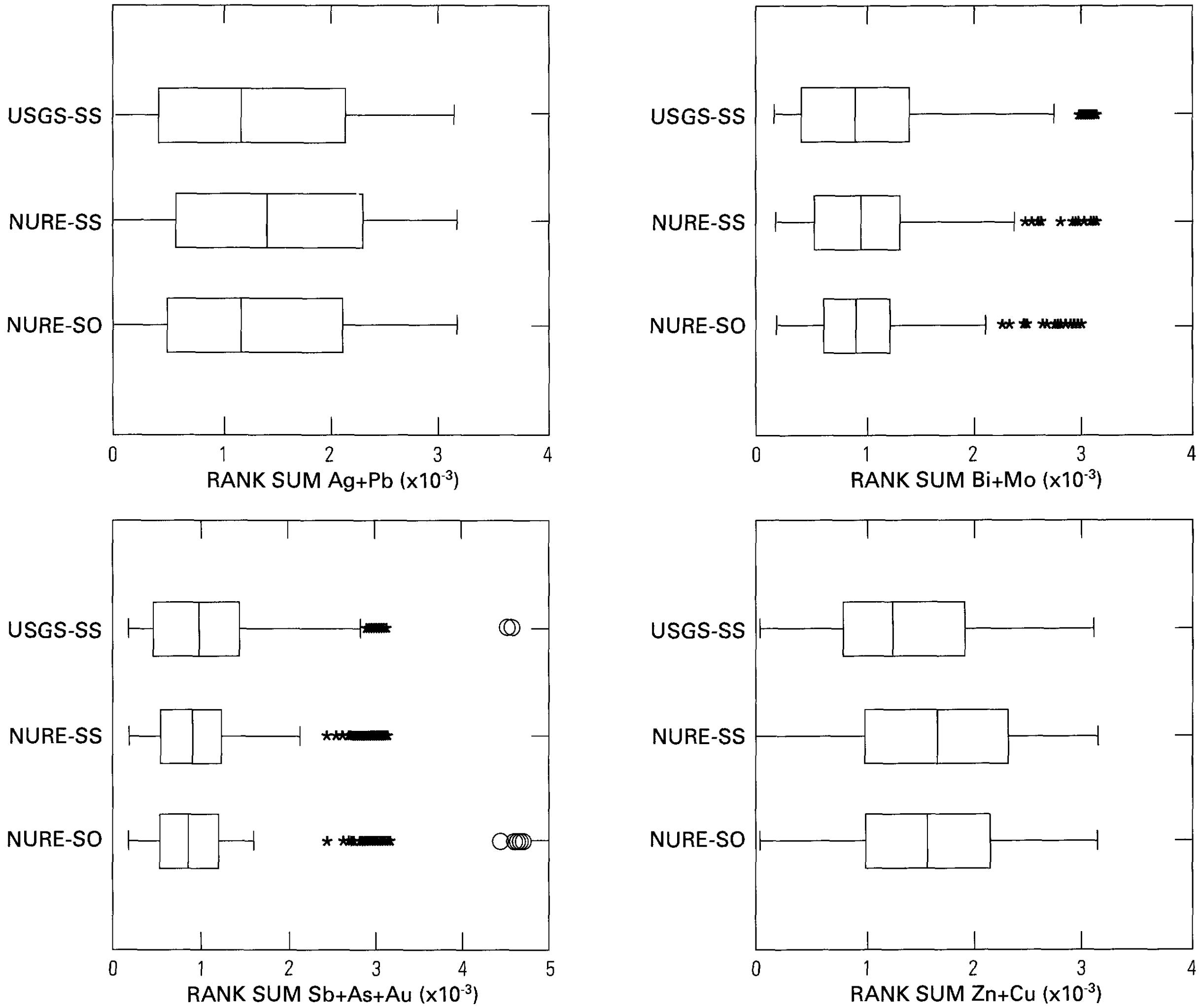


Figure 3. Box plots of sample-site scores of different sample media, Hailey 1°×2° quadrangle, south-central Idaho. USGS–SS indicates samples from beds of flowing streams; NURE–SS indicates samples from dry stream beds; NURE–SO indicates soil samples.

it is not certain that all Paleozoic black shale units in the Hailey quadrangle are metal rich.

Despite being analyzed by the same method, different sample media can produce significantly different ranges of sample-site scores. The Hailey geochemical data show, however, that different sample media have a significant overlap in the ranges of sample-site scores for the same geochemical suites (fig. 3). Because of this overlap, it was decided not to form sample-media subsets.

GEOLOGIC SETTING

Rocks in the Hailey quadrangle are Precambrian to Holocene in age. Thrust-faulted Paleozoic sedimentary and metasedimentary rocks crop out in the northeastern part of the quadrangle. These rocks are flanked by granitic rocks of the Idaho batholith on the west and locally overlain by volcanic rocks of the Challis Volcanic Group. The southern part of the Hailey quadrangle is covered by volcanic rocks of the Snake River Plain.

The oldest rocks in the Hailey quadrangle are in the northeastern part of the quadrangle. Precambrian granitic gneiss forms a part of the northwest-trending core of the Pioneer Mountains. Paleozoic sedimentary and metasedimentary rocks also are present in this part of the quadrangle and are commonly cut by major thrust faults. The Phi Kappa Formation is mainly black, locally silicified argillite and shale of Early Ordovician to Middle Silurian age, and the Trail Creek Formation is siliceous metasiltstone and very fine grained quartzite of Middle Silurian age (Dover and others, 1980). The Devonian Milligen Formation contains carbonaceous argillite, quartzite, and limestone. The Mississippian Copper Basin Formation contains limestone, sandstone, argillite, siltstone, and conglomerate. The Middle Pennsylvanian to Early Permian Wood River Formation is dominantly calcareous sandstone and calcarenite that contain layers of conglomerate, quartzite, limestone, and argillite.

Rocks of the Atlanta lobe of the Idaho batholith of Cretaceous age crop out in much of the northwestern and central parts of the Hailey quadrangle. The Idaho batholith is composed of plutonic rocks that range in composition from tonalite to granite; granodiorite is dominant. Parts of the batholith were intruded by Eocene plutons that range in composition from diorite to evolved granite. The granite has geochemical affinities with evolved (A-type) granite (Lewis and Kiilsgaard, 1991). Evolved granite is enriched in incompatible elements and depleted in compatible elements. Some of the Paleozoic sedimentary rocks in the northeastern part of the quadrangle also were intruded by Eocene quartz monzonite stocks that postdate metamorphism and regional thrust faulting. Eocene volcanic rocks of the Challis volcanic field are mainly andesitic in composition and are coeval with quartz monzonite stocks in the northeastern part of the quadrangle.

The northern edge of the Snake River Plain extends across the southern part of the Hailey quadrangle. Rocks of the Snake River Plain include the Idavada Volcanics of Miocene age and fluvial and lacustrine sediments and interbedded basalt flows that were deposited in the subsiding basin. The silicic Idavada Volcanics lie unconformably on granite and compose much of the Mount Bennett Hills. They were extruded during the formation of the Snake River Plain. The youngest rhyolite domes of the Magic Mountain eruptive center are of Pliocene age. Fluvial and lacustrine sediments and interbedded basalt were deposited on the Snake River Plain from Miocene to Pleistocene time and include the Banbury Basalt, the Glenns Ferry Formation, and the Bruneau Formation.

Just north of the northern edge of the Snake River Plain, in the south-central part of the Hailey quadrangle, is the Camas prairie, a downwarped block filled with Eocene and younger volcanic rocks and sediments.

Mineral deposits in the Hailey quadrangle include tungsten stockworks, Tertiary molybdenum stockworks, Tertiary polymetallic veins, epithermal precious-metal veins, Tertiary polymetallic replacements, copper skarns, Cretaceous molybdenum stockworks, polymetallic quartz veins and lodes, Cretaceous polymetallic veins, and Cretaceous polymetallic replacements (Johnson and Worl, 1989, in press; Worl and Johnson, 1989, this volume). Gold and radioactive black-sand placers, hot-springs deposits, antimony veins, gold-bearing skarns, sedimentary stratabound lead-zinc deposits, and stratabound barite are also present.

GEOCHEMISTRY

INTRODUCTION

One objective of this study was to develop methods with which to present regional geochemical data in a way that aids resource assessment. The low sampling density of many regional surveys hinders reliable detection of individual mineral deposit targets; however, two types of large geochemical features believed to be present in the Hailey quadrangle are appropriate targets for more detailed studies: (1) the zoned primary geochemical halos described by Ovchinnikov and Grigoryan (1971) and (2) evolved rocks, both Eocene granite and Miocene to Pliocene rhyolite.

Mineralizing events that form many types of sulfide-bearing mineral deposits also form zoned primary geochemical halos (Ovchinnikov and Grigoryan, 1971). These primary geochemical halos can affect areas large enough to be detected using the low-sampling density used in regional surveys. Tauson and others (1971) summarized the vertical zoning sequence of primary geochemical halos as follows, from top to bottom: Hg, Sb, As, Ba, Ag, Pb, Zn, Cu, Bi, W, Mo, Sn, Co, Ni, and Be. Tschanz and Kiilsgaard (1986) indicated that geochemical halos could have been used in their interpretation of the geochemistry of Sawtooth National Recreation Area. Primary halos suggested by the regional geochemical data are discussed in following sections of this report.

Lewis and Kiilsgaard (1991) described Eocene granite in south-central Idaho as having some geochemical characteristics of A-type (evolved) granitoids, and Bennett's (1980) description of the characteristics of the Eocene granite in the Idaho batholith matches some of the characteristics of evolved granite. Based on this and on geochemical evidence from this study, Eocene granite in the Hailey quadrangle is considered evolved granite. Relative to other types of granite and rhyolite, evolved rocks tend to be enriched in incompatible elements and depleted in compatible elements. Evolved rocks, both granite and rhyolite, are characterized by high concentrations of elements such as Be, Cs, F, K, Li, Nb, Rb, Sn, Th, U, and Y and low concentrations of elements such as Ba, Ca, Mg, and Sr. Evolved granite commonly comprises epizonal multiphase plutons that crop out over broad areas. Because of these characteristics, evolved rocks are good targets in regional geochemical surveys. This was demonstrated by Howarth and others (1981), who used NURE geochemical data to identify uranium-rich (evolved) granite in Colorado. Outside of Idaho, evolved rocks are associated with mineralized areas that are important sources of elements such as beryllium, lithium, uranium, and tin; however, in the Hailey quadrangle, evolved rocks have produced only minor amounts of beryl.

Regional geochemical studies can suggest areas where some types of mineral deposits might be present; however, with any geochemical survey a possibility of error exists in

Table 5. Element suites used for interpretation of regional geochemistry of the Hailey 1°×2° quadrangle, south-central Idaho.

Name	Characteristic elements	Comments
Outer zone	Sb+As+Au	Elements tend to be present in near-surface environments. Some of these elements may be volatile. Deposits may be precious metal enriched.
Outer middle zone	Ag+Pb	Deposits associated with this part of the primary geochemical halo may be precious metal enriched.
Inner middle zone	Zn+Cu	Deposits associated with this part of the primary geochemical halo tend to be base metal enriched.
Inner zone	Bi+Mo	This part of the primary geochemical halo suggests nearby plutonic margins. Gold may also be associated with these elements.
Pegmatite	Li+Be	Element suite is associated with margins of evolved plutons.
Evolved rock	Li+Be+ Nb+Th+Y	Elements is associated with outcrops of evolved rocks.

sample collecting, handling, and analysis. For this reason, not much importance is attached to an anomalous sample site that is isolated from other anomalous sample sites. More attention is given to an anomaly that is represented by a cluster of sample sites.

The multielement analytical methods used in this regional survey provide information about the concentrations of a variety of elements in stream-sediment and soil samples. This information can lead to identification of primary geochemical halos of the types that are associated with a wide range of sulfide-bearing mineral deposits (Ovchinnikov and Grigoryan, 1971). Single-element analytical methods having low detection limits are better for providing information about an element, such as gold, that is associated with specific deposit types, particularly deposit types having low anomaly-to-background contrasts and poorly developed primary geochemical halos.

GEOCHEMICAL SUITES

Geochemical suites were used to interpret the regional geochemistry of the Hailey quadrangle. Sample sites are defined as anomalous on the basis of scores that are the sums of the ranks of geochemical suites of elements rather than single elements. The distribution of geochemical anomalies defined by these suites is consistent with the distribution of prospects, alteration, and mining activity. Some of the mineralized areas defined by Worl and Johnson (this volume) are characterized by anomalous sample-site scores for more than one geochemical suite.

In this study, six element suites are used to discuss the regional geochemistry of the Hailey quadrangle (table 5). The first four element suites are related to the vertical zoning sequence of primary geochemical halos described by Tauson and others (1971). The first three elements in the top of the vertical zoning suite summarized by Tauson are mercury, antimony, and arsenic. Because samples in this study were not analyzed for mercury and in order to emphasize epithermal gold deposits, the outer zone geochemical suite in this study was defined as Sb+As+Au. Names given to the first four element suites are based on their vertical position in primary geochemical halos; those for the last two element suites reflect the geochemical signature associated with evolved rocks. The term "polymetallic" is used to describe mineralized areas in which no one or two suites of the first four geochemical suites of elements dominate.

Unless otherwise indicated, the mineralized areas referred to in this report are those delineated by Worl and Johnson (this volume). The boundaries of the mineralized areas delineated by Worl and Johnson are not necessarily the same as those that would be delineated by geochemistry. Most of these mineralized areas are characterized by or dominated by one or more of the element suites Sb+As+Au, Ag+Pb, Zn+Cu, and Bi+Mo (fig. 4).

A point system was used to characterize geochemical suites in each of the mineralized areas. The classification of the mineralized areas and a description of the point system used to classify the areas are given in table 6. The geochemistry of two mineralized areas, Neal (7) and Elk Creek (23), is not well characterized by the sample media and analytical methods used in this study, and these two areas are unclassified. Samples were not collected in the Bunker Hill (23) or Washington Basin (31) mineralized areas, and the areas are not discussed herein.

POLYMETALLIC AREAS

Most of the mineralized areas containing polymetallic deposits are wholly or partly within the area where Paleozoic rocks crop out (fig. 5). In these mineralized areas none of the geochemical suites are dominant. Polymetallic deposits

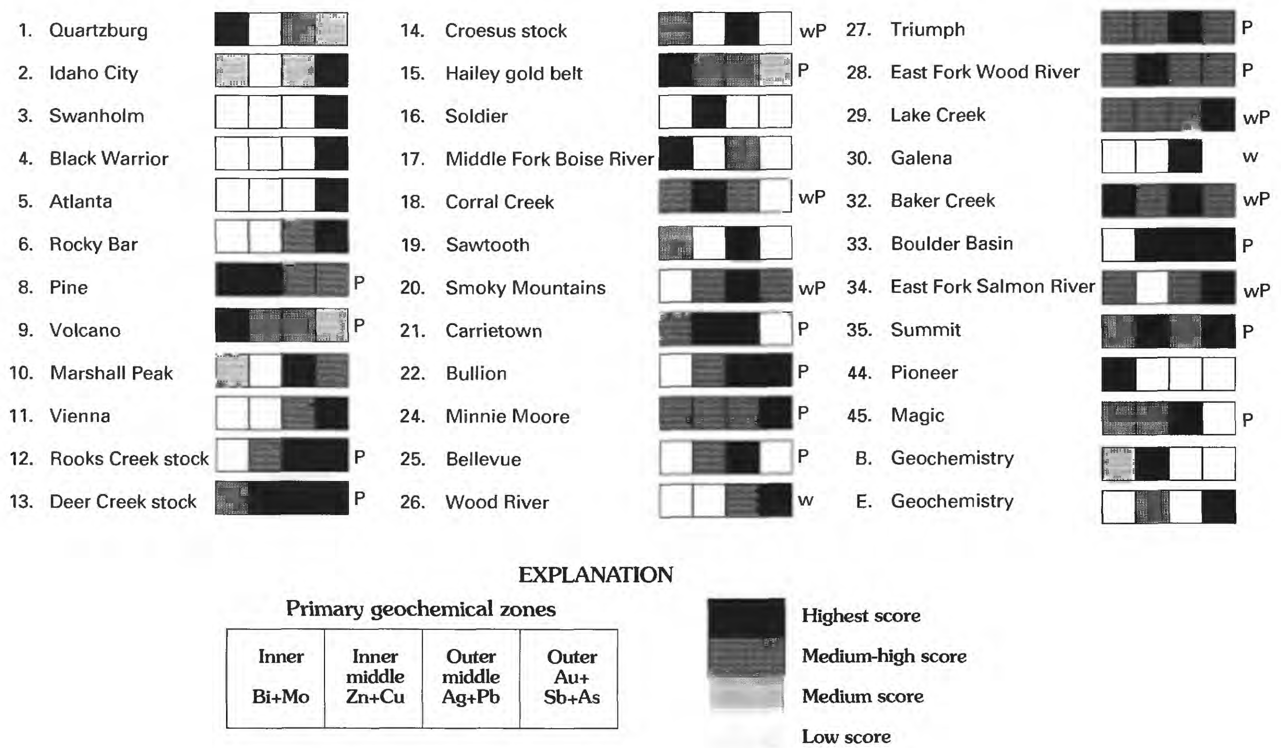


Figure 4. Summed sample-site scores of geochemical suites in mineralized areas of Worl and Johnson (this volume), Hailey 1°×2° quadrangle, south-central Idaho. P indicates area is polymetallic; w indicates weak geochemical signature. Mineralized areas are shown in figure 5.

characterize the following areas: Boulder Basin (33), Summit (35), East Fork Wood River (28), Triumph (27), Rooks Creek stock (12), Deer Creek stock (13), Bullion (22), Bellevue (25), Minnie Moore (24), Magic (45), Carrietown (21), and Hailey gold belt (15). The polymetallic areas at Pine (8), Volcano (9), and Magic (45) are outside of the area of Paleozoic rock outcrop.

OUTER ZONE SUITE

Figure 6 shows sample sites that contain anomalous amounts of elements that characterize the outer zone of the primary geochemical halo described by Tauson and others (1971). The elements antimony, arsenic, and gold define this suite. Samples were not collected and analyzed specifically for gold at levels near crustal abundance (about 3 ppb), and only 1 percent of the samples contained detectable amounts of gold. In this study the elements arsenic and antimony are the main pathfinder elements for precious-metal mineralization that might be associated with the outer zone halo. Most anomalous samples of the outer zone that are not in polymetallic-mineralized areas are in the northern third and western two-thirds of the quadrangle (fig. 6). The outer zone suite is the dominant geochemical signature in the Idaho City (2), Swanholm (3), Black Warrior (4), Atlanta (5), Vienna (11), and Rocky Bar (6) mineralized areas (fig. 5). With the exception of Swanholm, precious metals were produced from mines in these areas. The main commodity produced at Swanholm was antimony. The outer zone suite also is dominant in area E in the southeastern corner of the Hailey quadrangle (fig. 5), which was delineated on the basis of the regional geochemistry.

A few scattered samples in the southern part of the quadrangle contain anomalous amounts of the outer zone suite of elements. The anomalous samples in this area are single-sample anomalies associated with the Idavada Volcanics. Epithermal, low-sulfide, precious-metal deposits associated with volcanic terrane can have weak geochemical signatures (Nash, 1988), and the occurrence of hot-springs-type precious-metal deposits may be underestimated in the southern part of the Hailey quadrangle.

OUTER MIDDLE ZONE SUITE

Figure 7 shows sample sites that have anomalous scores for the outer middle zone suite of elements (Ag+Pb) (Tauson and others, 1971). This geochemical signature is the dominant suite in the Marshall Peak (10) and Sawtooth (19) mineralized areas (fig. 5). The Sawtooth mineralized area is associated with rocks of a Tertiary granite batholith, and the Marshall Peak area is just to the south in an area of regional shear zones in rocks of the Idaho batholith.

Table 6. Classification of mineralized areas in the Hailey 1°×2° quadrangle, south-central Idaho.

[Mineralized areas are defined in Worl and Johnson (this volume), and the location of the areas is shown in figure 5. No. is number of sample sites whose drainages are within boundary of mineralized area. Sample sites having a percentile ranking between the first and second number are given a score of 1; those having a ranking of greater than the second number are given a score of 2. I is the inner zone (Bi+Mo) score; I–M is the inner middle zone (Zn+Cu) score; O–M is the outer middle zone (Ag+Pb) score; and O is the outer zone (Sb+As+Au) score. The geochemical suite having the highest sample site score is shown in uppercase letters; geochemical suite(s) having score(s) greater than or equal to half of the dominant sample site score is shown in lowercase letters. If the dominant suite at the 90th and 95th percentiles has less than 15 percent of the maximum possible points for the mineralized area, then the suites were evaluated at the 75th and 90th percentiles. Geochemical suites chosen at the 75th and 90th percentiles are described by the adjective "weak." Polymetallic suites either have at least three of four zones whose scores area separated by three or less points or have at least three of four zones whose scores are greater than or equal to one-half that of the zone having the highest score. If a mineralized area has a total score of less than four for all zones when evaluated at the 75th and 90th percentiles, then the area was not classified]

	Mineralized area	No.	Percentile	I	I–M	O–M	O	Classification
1	Quartzburg	21	90–95	12	2	7	5	INNER with outer middle zones
2	Idaho City	35	90–95	6	5	8	19	OUTER zone
3	Swanholm	4	90–95	0	0	1	6	OUTER zone
4	Black Warrior	12	90–95	0	0	2	6	OUTER zone
5	Atlanta	12	90–95	2	0	3	16	OUTER zone
6	Rocky Bar	15	90–95	0	0	5	6	OUTER with outer middle zones
7	Neal	12	75–90	1	0	2	0	Not classified
8	Pine	37	90–95	12	12	10	6	POLYMETALLIC
9	Volcano	9	90–95	12	9	9	4	POLYMETALLIC
10	Marshall Peak	16	90–95	5	3	21	19	OUTER MIDDLE with outer zone
11	Vienna	8	90–95	2	2	7	13	OUTER with outer middle zone
12	Rooks Creek stock	6	90–95	0	2	3	3	POLYMETALLIC
13	Deer Creek stock	3	90–95	1	2	2	2	POLYMETALLIC
14	Croesus stock	4	90–95	2	2	4	1	Weak polymetallic
15	Hailey gold belt	12	90–95	7	6	6	4	POLYMETALLIC
16	Soldier	9	90–95	1	6	1	1	INNER MIDDLE zone
17	Middle Fork Boise River	45	90–95	26	5	23	5	INNER with outer middle zone
18	Corral Creek	21	75–90	4	7	4	0	Weak polymetallic
19	Sawtooth	17	90–95	10	2	13	2	OUTER MIDDLE with inner zone
20	Smoky Mountains	17	75–90	7	10	15	14	Weak polymetallic
21	Carrietown	14	90–95	3	4	4	0	POLYMETALLIC
22	Bullion	5	90–95	0	2	3	3	POLYMETALLIC
23	Bunker Hill	0						No samples
24	Minnie Moore	3	90–95	2	3	3	4	POLYMETALLIC
25	Bellevue	8	90–95	0	2	3	1	POLYMETALLIC
26	Wood River	24	75–90	0	4	9	15	Weak outer with outer middle zone
27	Triumph	4	90–95	3	3	6	4	POLYMETALLIC
28	East Fork Wood River	7	90–95	4	7	5	6	POLYMETALLIC
29	Lake Creek	16	75–90	9	10	9	11	Weak polymetallic
30	Galena	7	75–90	1	1	3	0	Weak polymetallic
31	White Clouds	0						No samples
32	Baker Creek	5	75–90	2	1	2	1	Weak polymetallic
33	Boulder Basin	1	90–95	0	2	2	2	POLYMETALLIC
34	East Fork Salmon River	14	75–90	5	2	4	6	Weak polymetallic
35	Summit	18	90–95	11	13	8	13	POLYMETALLIC
44	Pioneer	11	90–95	10	1	2	2	INNER zone
45	Magic	2	90–95	2	2	3	0	POLYMETALLIC
46	Elk Creek	3	75–90	0	2	0	0	Not classified
B	Defined by geochemistry	17	90–95	6	18	1	0	INNER MIDDLE zone
E	Defined by geochemistry	22	90–95	0	11	6	22	OUTER with inner middle zone

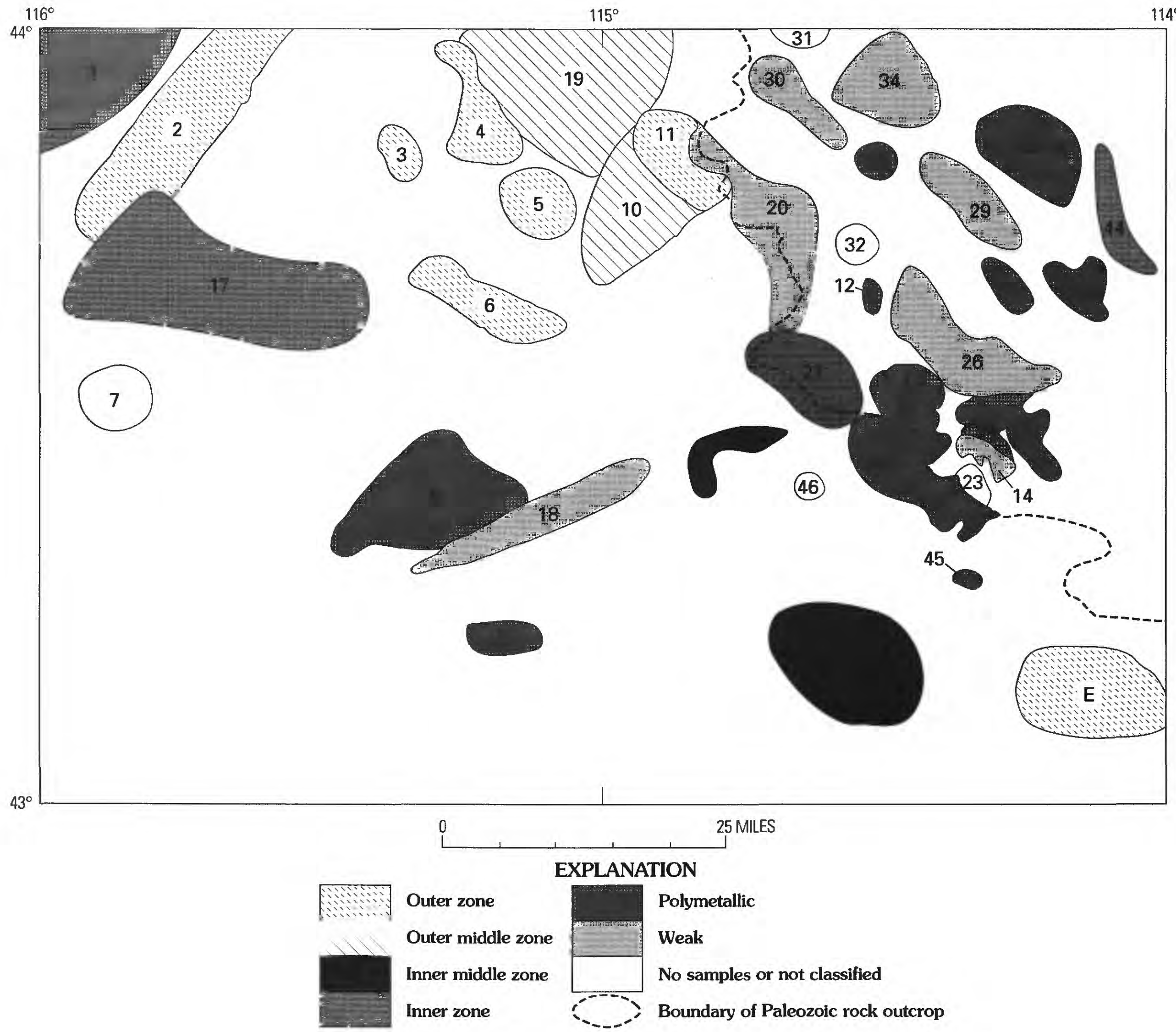


Figure 5. Dominant geochemical signatures of mineralized areas, Hailey 1°×2° quadrangle, south-central Idaho. Numbers of mineralized areas are from Worl and Johnson (this volume) and table 6 (this report).

INNER MIDDLE ZONE SUITE

Figure 8 shows sample sites that have high scores for the inner middle zone suite of elements (Zn+Cu) (Tauson and others, 1971). Most of these samples sites are associated with Paleozoic rocks in mineralized areas that have a polymetallic character; only two mineralized areas that are not associated with Paleozoic rocks have inner middle zone signatures as the dominant signature. One of the areas, Soldier (16), is a mineralized area defined by Worl and Johnson (this volume). The other area, area B (fig. 5), was delineated by the regional geochemistry.

INNER ZONE SUITE

The elements bismuth and molybdenum define the inner zone signature (Tauson and others, 1971). Only 4.1 percent of the sediment samples contain amounts of bismuth greater than 1.5 ppm, the limit of determination for bismuth in this study. Authors such as Nash (1988) found bismuth to be a good indicator of pluton margins and used it, together with tungsten, as a diagnostic element for skarn deposits. The inner zone suite of elements is the dominant element suite in the Quartzburg (1), Middle Fork Boise River (17), and Pioneer (44) mineralized areas (fig. 5). With the exception of Quartzburg, mineralized areas having inner zone signatures are associated with Eocene granite (fig. 9). The inner zone signature associated with the Middle Fork Boise River (17) mineralized area extends beyond the mineralized area delineated by Worl and Johnson (this volume). A weak inner zone signature is associated with rhyolite of the Idavada Volcanics in the southern part of the quadrangle. The inner zone signature in this area is

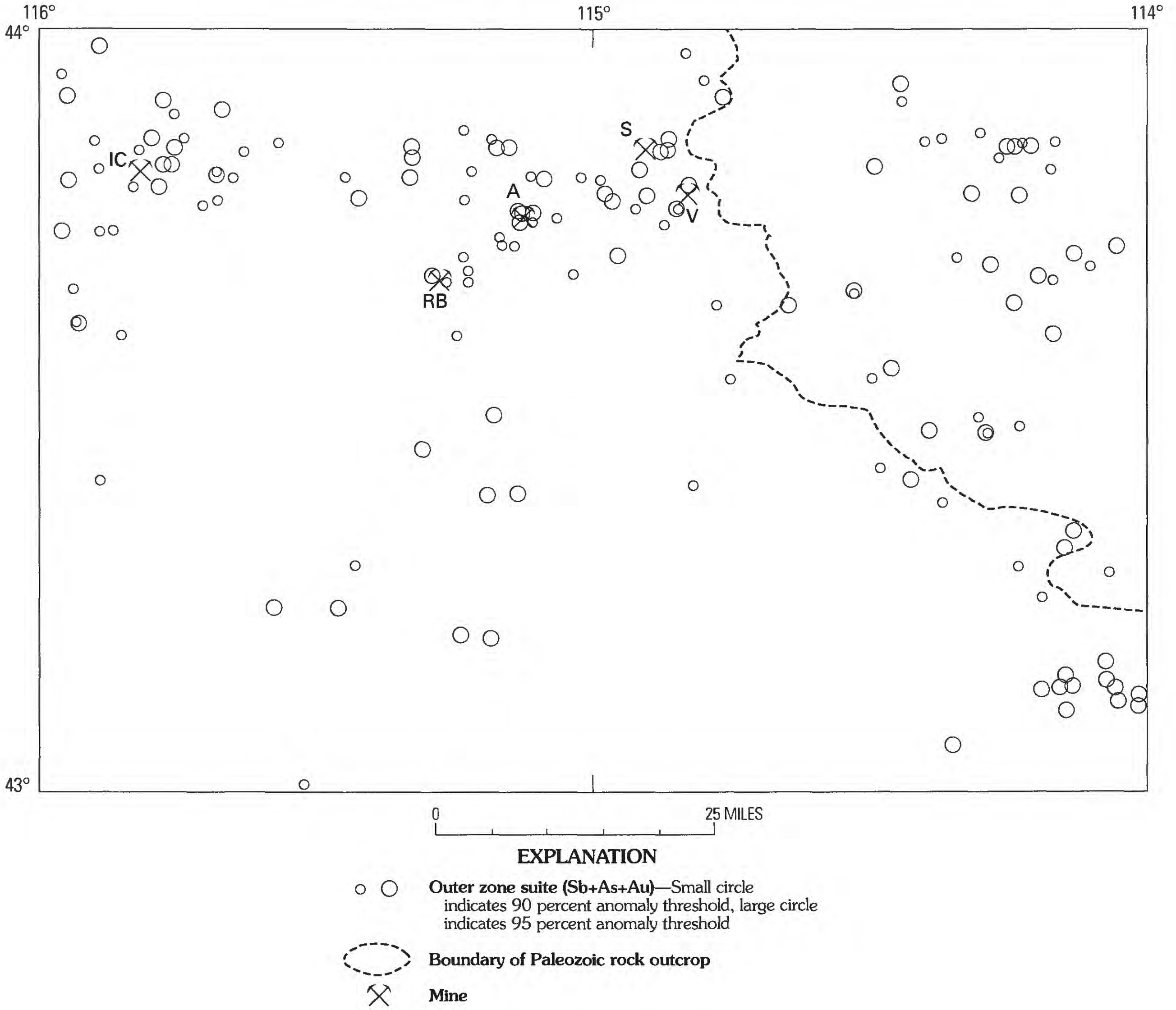


Figure 6. Sample sites having anomalous outer zone scores, Hailey 1°×2° quadrangle, south-central Idaho. Mineralized areas: IC, Idaho City; RB, Rocky Bar; S, Sawtooth; V, Vienna; A, Atlanta.

probably the result of high background concentrations of molybdenum in the Idavada rhyolites.

EVOLVED-ROCK SUITE

The elements beryllium, lithium, niobium, thorium, and yttrium characterize the evolved-rock geochemical suite in the Hailey quadrangle. The evolved-rock element suite is usually associated with either Eocene granite or Miocene to Pliocene rhyolite (fig. 10). Some Miocene to Pliocene rhyolite has been recognized as evolved rock. In the Cannonball Mountain area, Lewis (1990) found an evolved peralkaline rhyolite, a comendite, of Miocene age (about 10.2 Ma) and proposed that this unit be named "tuff of Cannonball Mountain." A single anomalous sample is associated wth rhyolite at Wedge Butte (fig. 10), where a 3-million-year-old rhyolite was described by Leeman (1982) as "evolved" as compared with most Snake River Plain rhyolites. Some scattered sample sites having high scores for the evolved-rock element suite do not have close associations with either Eocene granite or Miocene to Pliocene rhyolite (fig. 10); these sites are north and northwest of the Eocene granite around Sheep Mountain, within the Pine (8) mineralized area, and south-southeast of the Sawtooth batholith.

PEGMATITE SUITE

Pegmatite associated with evolved granite can be enriched in both beryllium and lithium, and pegmatite fields commonly emanate from the cupolas of parental granite (see, for example, London, 1990). Compared to the evolved-rock element suite, the pegmatite suite has fewer anomalous

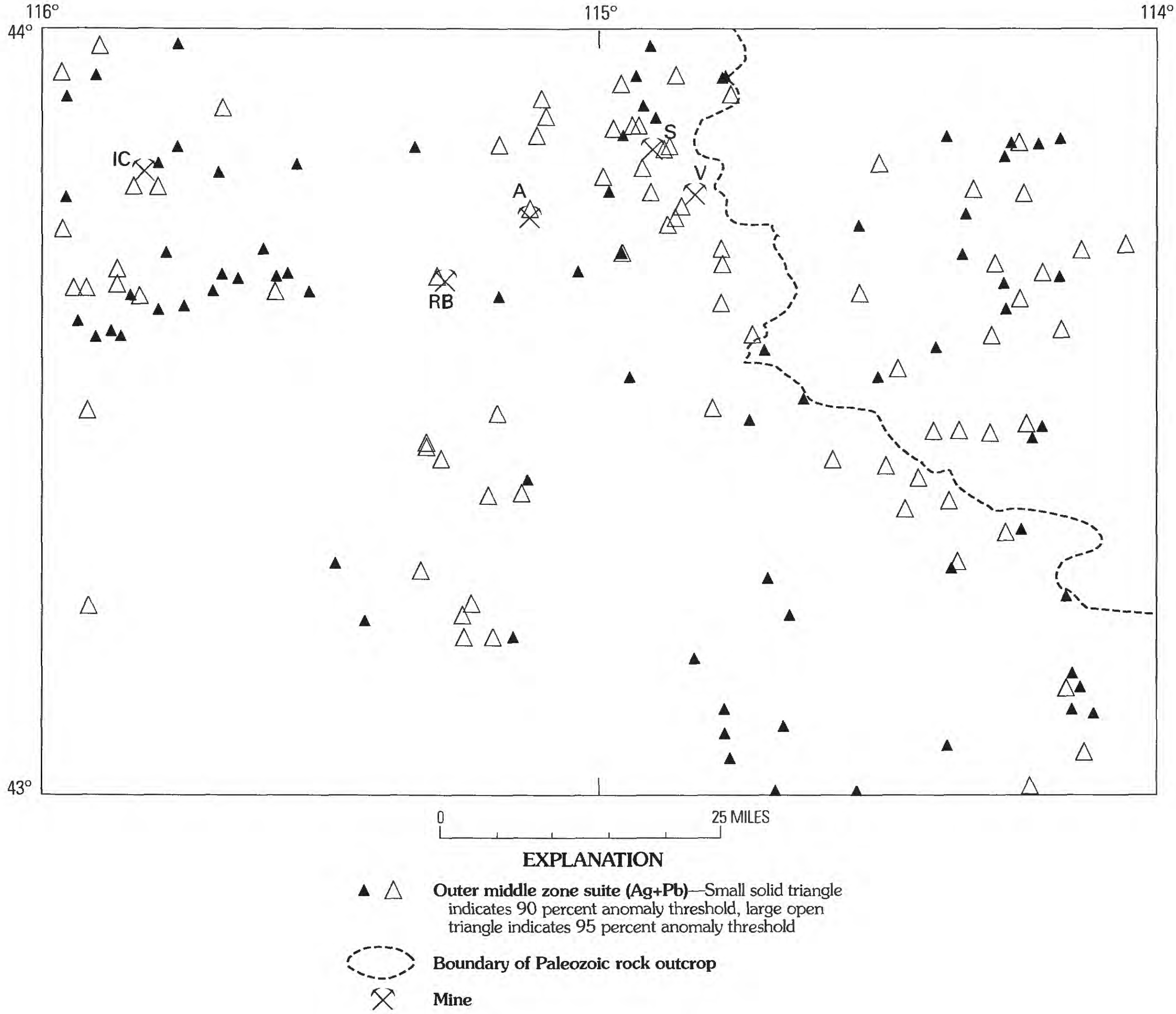


Figure 7. Sample sites having anomalous outer middle zone scores, Hailey 1°×2° quadrangle, south-central Idaho. Mineralized areas: IC, Idaho City; RB, Rocky Bar; S, Sawtooth; V, Vienna; A, Atlanta.

sample sites within the area where Eocene granite crops out. Also, unlike the evolved-rock element suite, the pegmatite suite is not strongly associated with Miocene rhyolite (fig. 11). Because of these contrasts with the evolved-rock suite, Li+Be was chosen as a separate element suite. Despite the name "pegmatite suite" given to this suite of elements, mineralized veins can be the source of anomalous amounts of beryllium in sediment samples. As much as 150 ppm Be is in mineralized veins near Mattingly Creek and La Moyne and Leggitt Creeks (fig. 12) (Kiilsgaard and others, 1970). These veins also contain anomalous amounts of other elements, including silver and gold.

Clusters of sample sites containing anomalous amounts of the pegmatite suite of elements are in the following areas (fig. 11): south of the Sawtooth batholith, in the Pioneer (44) mineralized area, east of the Soldier Mountains, in a northeast-trending corridor indicated by outcrops of Eocene granite including those in the Soldier and Smoky Mountains, and within the outcrop area and north of the Eocene granite at Sheep Mountain. Many of the sediment samples containing anomalous amounts of the pegmatite suite of elements were not derived from areas where Eocene granite crops out.

DISCUSSION

Several authors have estimated the age or ages of mineralization in the Idaho batholith. For example, Bennett (1980) emphasized the close spatial relationship between Eocene epizonal granite and mineralized rock in the Idaho

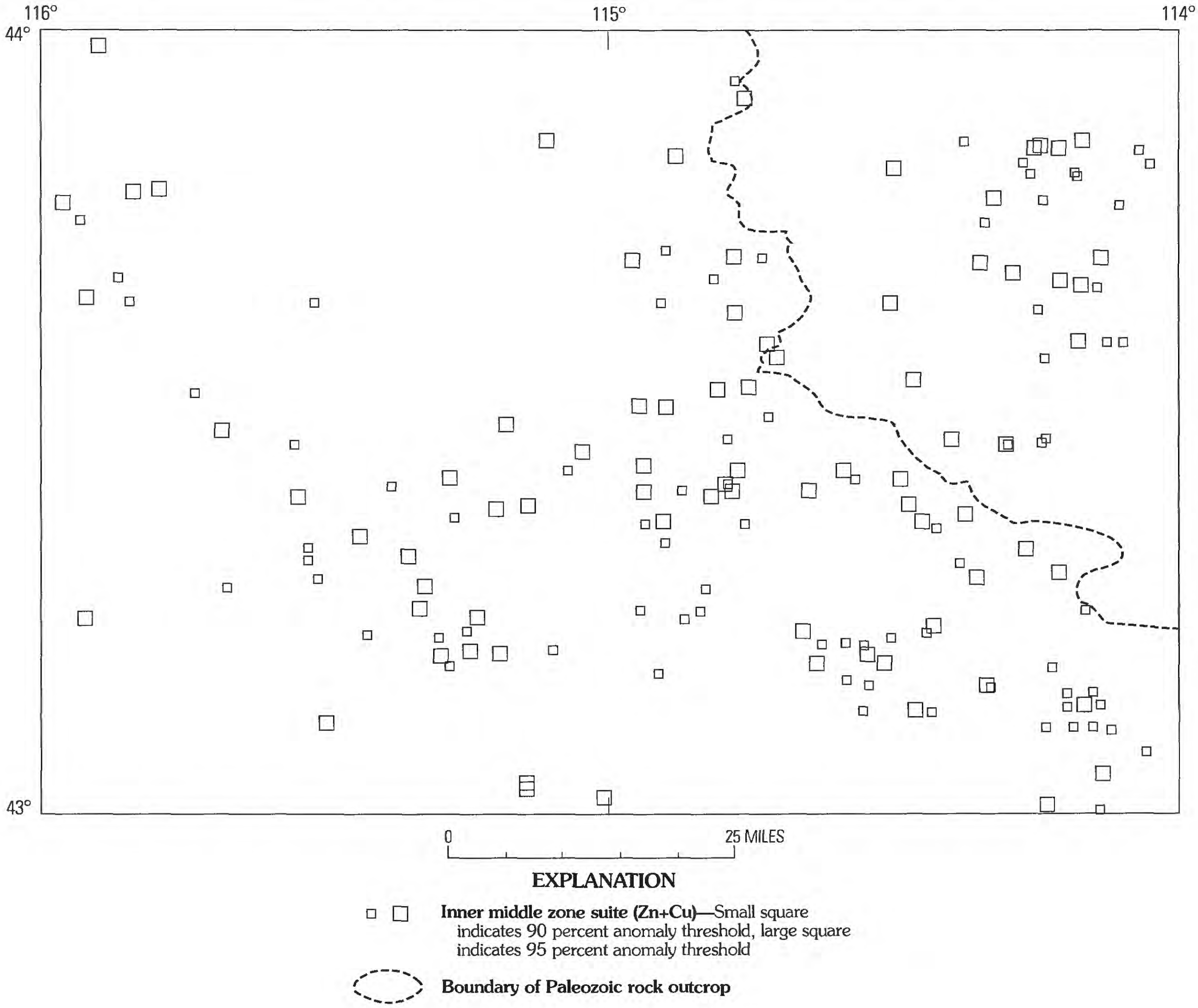


Figure 8. Sample sites having anomalous inner middle zone scores, Hailey 1°×2° quadrangle, south-central Idaho.

batholith. Anderson (1951) believed that all mineralization in the southern batholith is Tertiary in age. Ross (1931) believed that mineralization is both Cretaceous and Tertiary in age. Snee (1989), using $^{40}Ar/^{39}Ar$ thermochronology, found that some mineralization in the southern part of the Idaho batholith occurred between 78 and 57 Ma.

Results of regional geochemical surveys do not directly indicate age of mineralization. Most geochemical suites cannot be tied to geologic events that occurred in a particular time period; however, geochemical sediment samples containing large amounts of the elements that define two geochemical suites, the evolved-rock and pegmatite suites, are related to source rocks that are of Eocene age or younger. Source rocks of the sediment samples enriched in these suites of elements are probably the result of polyphase plutonism that caused incompatible elements to be enriched in Eocene granite and younger rhyolite.

KNOWN MINERALIZED AREAS

POLYMETALLIC MINERALIZATION

Most mineralized areas characterized by polymetallic mineralization are related to the area in which Paleozoic rocks crop out (fig. 5). Black shale usually is enriched in a range of elements (Vine and Tourtelot, 1970). Although Paleozoic black shale of the Hailey quadrangle is not enriched relative to other black shale (Simons, 1981), the unmineralized parts of the black shale units in the Hailey

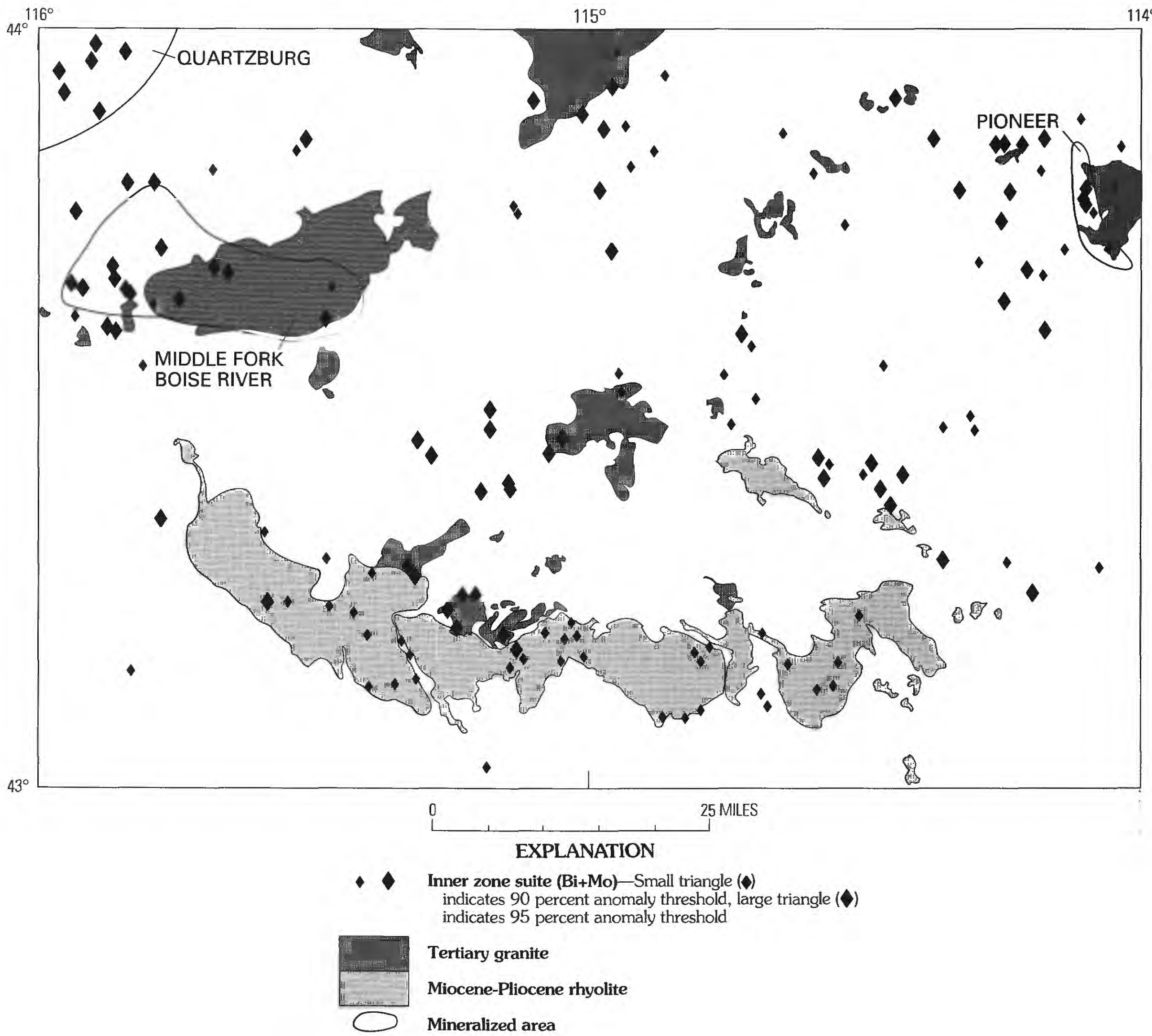


Figure 9. Sample sites having anomalous inner zone scores, Hailey 1°×2° quadrangle, south-central Idaho.

quadrangle are enriched in a range of elements relative to other unmineralized rocks in the quadrangle.

Black shale could be the source of the metals in at least some of the mineral deposits within the area where Paleozoic rocks crop out. Simons (1981) concluded that black shale could be the source of some of his geochemically anomalous stream-sediment samples. Hall and others (1978) proposed that metals in vein deposits were leached from Paleozoic black shale and deposited in favorable structures related to the Wood River thrust fault.

In areas containing black shale, mineral deposits may be telescoped. Epizonal intrusions could have provided heat to drive shallowly circulating hydrothermal cells that extracted metals from a nearby source, the black shale. As a result of rapidly changing pressure and temperature conditions between the shallow intrusion and the surface, the metals could have been precipitated in a relatively narrow zone to form the deposits. Telescoping of deposits is consistent with the polymetallic character of many mineralized areas associated with Paleozoic rock outcrops, and it is consistent with the suggestion of Sanford and others (1989) that at least the lead in some deposits could have been remobilized from syngenetic deposits in black shale.

OUTER ZONE MINERALIZED AREAS

As previously indicated, Bennett (1980) found a spatial relationship between Eocene granite and mineralization,

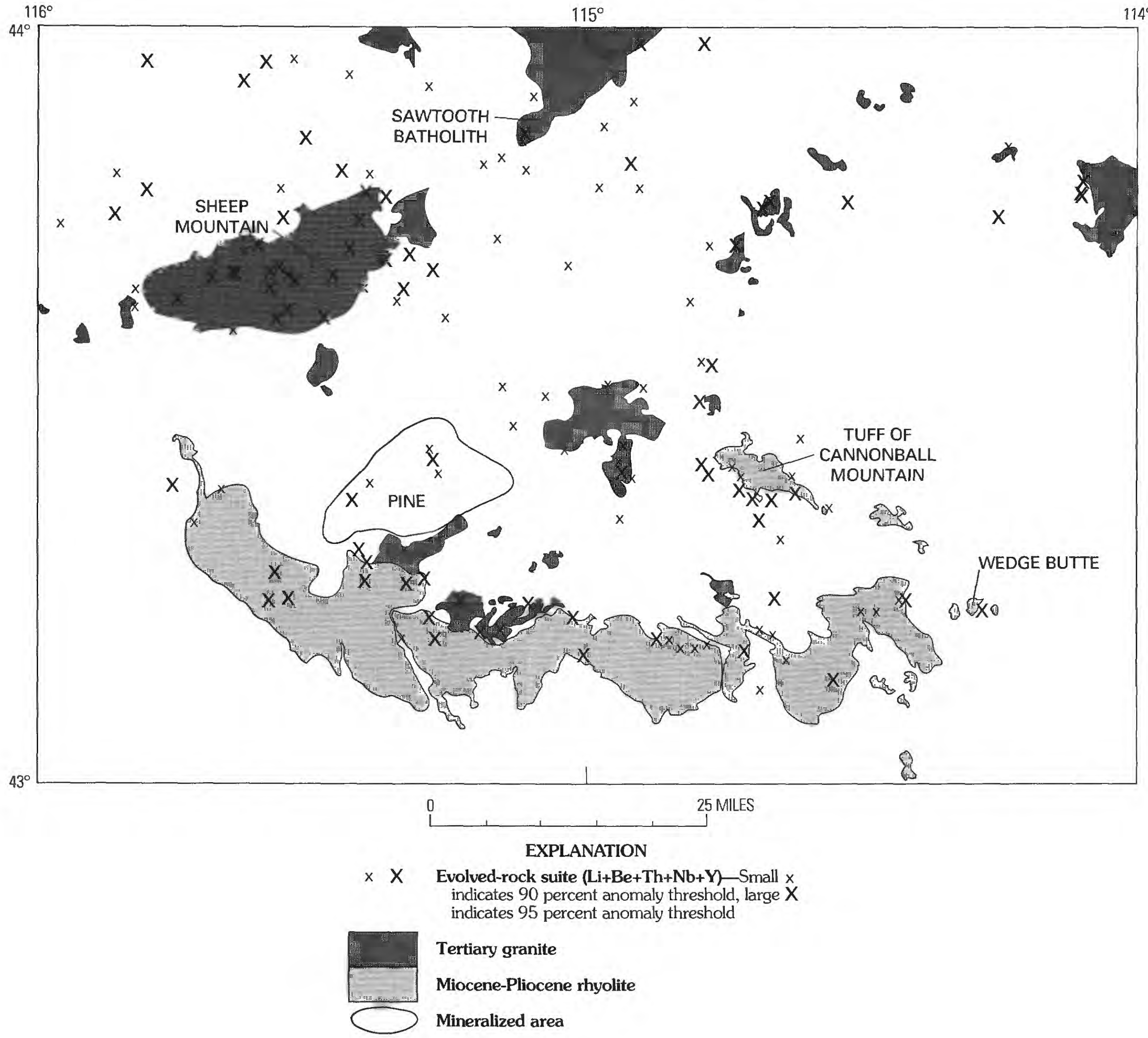


Figure 10. Sample sites having anomalous evolved-rock scores, Hailey 1°×2° quadrangle, south-central Idaho. Solid triangle shows the location of Sheep Mountain.

including epithermal mineralization. Eocene igneous activity is suggested by anomalous amounts of elements that characterize both the evolved-rock and pegmatite geochemical suites. The area containing mines at Atlanta, Sawtooth, Mattingly Creek, and La Moyne and Leggitt Creeks has a combined pegmatite and outer zone geochemical suite signature (fig. 12). The pegmatite signature in this area, and the veins found by Kiilsgaard and others (1970) that contain both beryllium and precious metals, suggests that near-surface, but unexposed, Eocene granite might have affected both mineralization associated with outer zone primary geochemical halos and the distribution of the pegmatite suite of elements. The implication that Eocene igneous activity played a role in some of the mineralization at Atlanta is difficult to reconcile with Snee's findings that the youngest mineralizing event at Atlanta is 57 Ma.

The outer zone suite of elements is enriched farther to the south and southeast of the Sawtooth batholith than is the pegmatite suite of elements (fig. 12). The mines at Vienna and Rocky Bar are characterized only by the outer zone suite of elements. The outer zone suite of elements is enriched between the Rocky Bar and Atlanta mineralized areas. This linear zone of enrichment could reflect a structure that controlled mineralization between these known mineralized areas.

MIDDLE ZONE MINERALIZED AREAS

Two mineralized areas are dominated solely by the outer middle zone geochemical signature: Marshall Peak

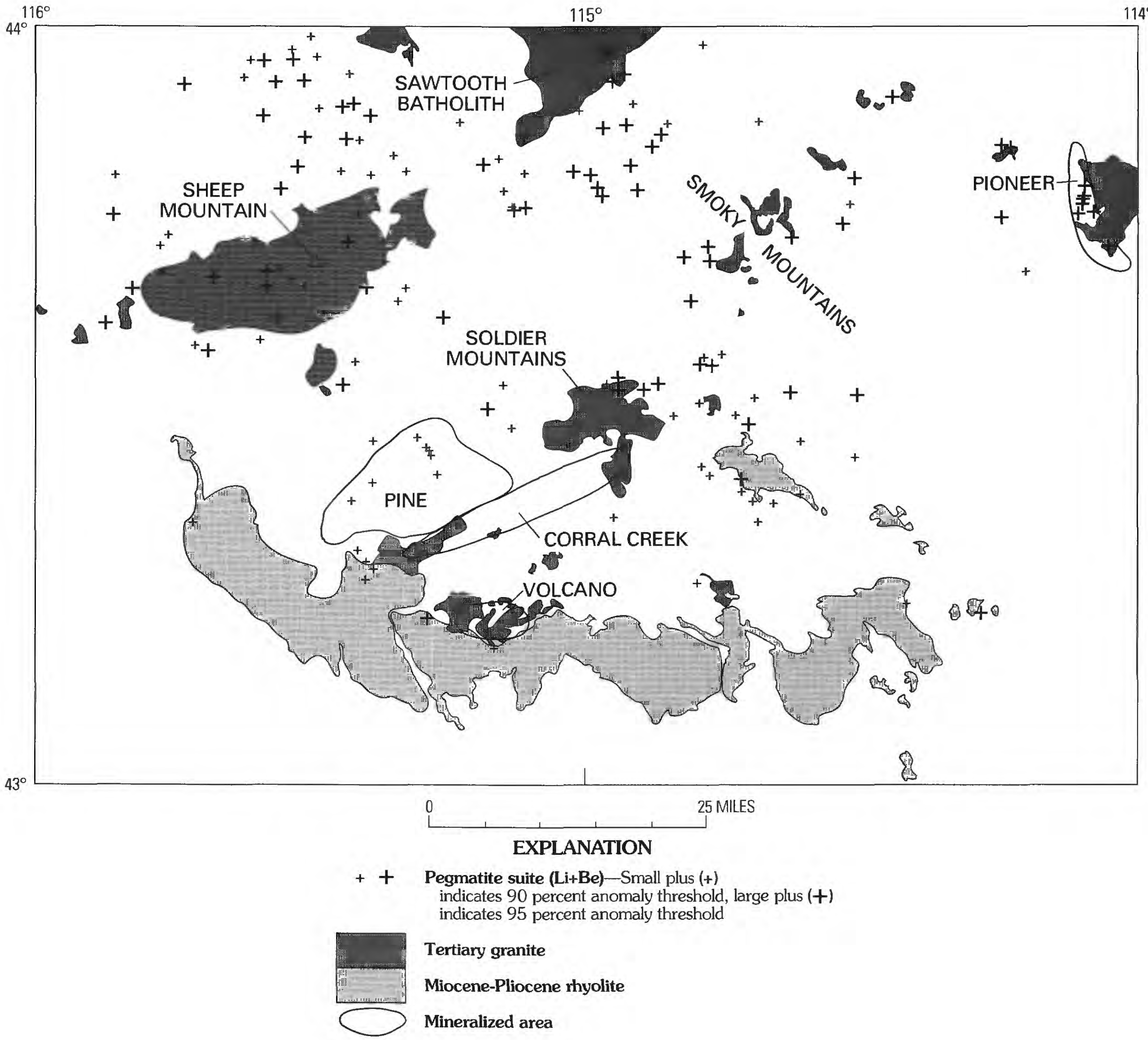


Figure 11. Samples sites having anomalous pegmatite scores, Hailey 1°×2° quadrangle, south-central Idaho. Solid triangle shows the location of Sheep Mountain.

(10) and Sawtooth (19). The Middle Fork Boise River (17) mineralized area also has a significant outer middle zone signature. In all three areas the outer middle zone signature is associated with the inner zone signature (fig. 4). It is not known why these areas do not have a stronger inner middle zone signature.

Soldier (16) is the only known mineralized area that is characterized solely by an inner middle zone signature. Mineralization in the Soldier area is hosted by rocks of the Idaho batholith (Worl and Johnson, this volume), west of an evolved peralkaline rhyolite (fig. 13), a comendite, of Miocene age (Lewis, 1990). The chemistry of this rhyolite is unique in the Hailey quadrangle and unusual for Snake River Plain rhyolite (Lewis, 1990). Evolved rocks can have high concentrations of halogens such as fluorine and chlorine. Both fluorine and chlorine can be important in the transport of gangue and ore minerals from granitic magmas during the final stages of crystallization (see, for example, Burnham and Ohmoto, 1980); however, it is not certain if the emplacement of the comendite played a role in forming the deposits of the Soldier mineralized area.

INNER ZONE MINERALIZED AREAS

Mineralized areas characterized by an inner zone signature commonly are associated with Eocene granite outcrops or with geologic or geochemical indications of near-surface

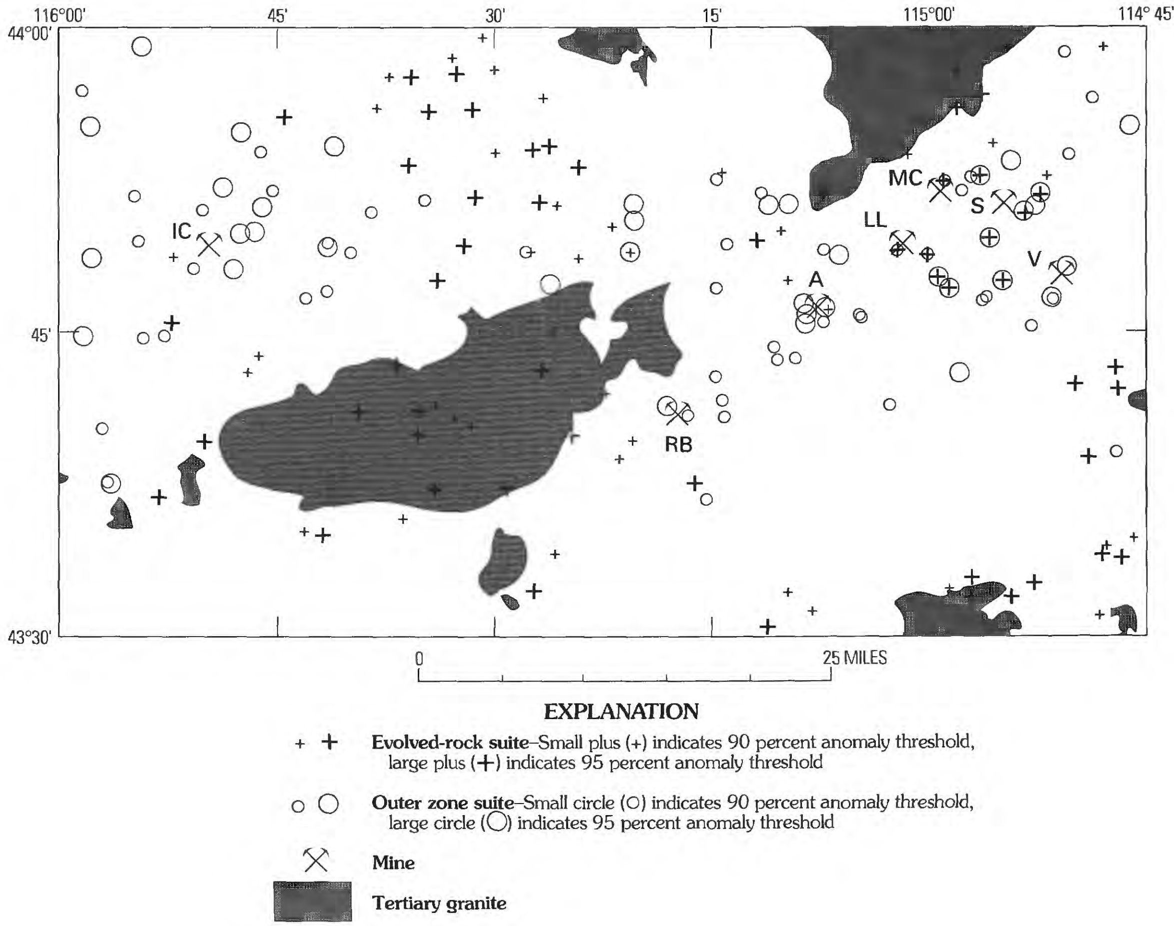


Figure 12. Sample sites in the northwest quadrant of the Hailey 1°×2° quadrangle, south-central Idaho, having anomalous evolved-rock and outer zone scores. Mineralized areas: IC, Idaho City; RB, Rocky Bar; S, Sawtooth; V, Vienna; A, Atlanta; LL, La Moyne Creek–Leggit Creek; MC, Mattingly Creek.

Eocene granite. Both the Pioneer (44) and the Middle Fork Boise River (17) mineralized areas are near exposures of Eocene granite (fig. 9). The Corral Creek (18), Pine (8), and Volcano (9) mineralized areas have polymetallic signatures and a significant inner zone component and are within a northeast-trending belt that is indicated both by outcrops of Eocene granite, such as in the Soldier and Smoky Mountains, and by a pegmatite geochemical signature (fig. 11). The Quartzburg (1) mineralized area (fig. 9) contains Tertiary dike swarms. Rhyolite dikes in this swarm are compositionally equivalent to evolved Eocene granite (Kiilsgaard and Hingley, 1989; Kiilsgaard and others, in press). In the Quartzburg area sample sites that are anomalous with respect to the evolved-rock signataure are not contiguous. The scattered character of the evolved-rock signature in the Quartzburg area is consistent with dike source rocks (fig. 10). Bennett (1980) suggested that Tertiary dike swarms in the Boise Basin are apophyses of Eocene granite plutons. The relationship between the inner zone signature and Eocene granite is important. Not only does this signature indicate a potential for tungsten and molybdenum mineralization, it also is associated with gold-mineralized rock in the Quartzburg, Volcano, Corral Creek, and Pioneer mineralized areas.

MINERALIZED AREAS HAVING WEAK GEOCHEMICAL SIGNATURES

Except for the Corral Creek (18) mineralized area, mineralized areas having weak geochemical signatures (table 6, fig. 5) are wholly or partly within the area where Paleozoic rocks might crop out. Anomaly thresholds for the area where Paleozoic rocks might crop out were calculated separately from the rest of the Hailey quadrangle. The area where Paleozoic rocks might crop out contains black shale as well as other rock types. Commonly, mineralized areas that are within the area of Paleozoic outcrop and that have weak geochemical signatures may not be closely associated with black shale. For example, most mines in the Galena (30)

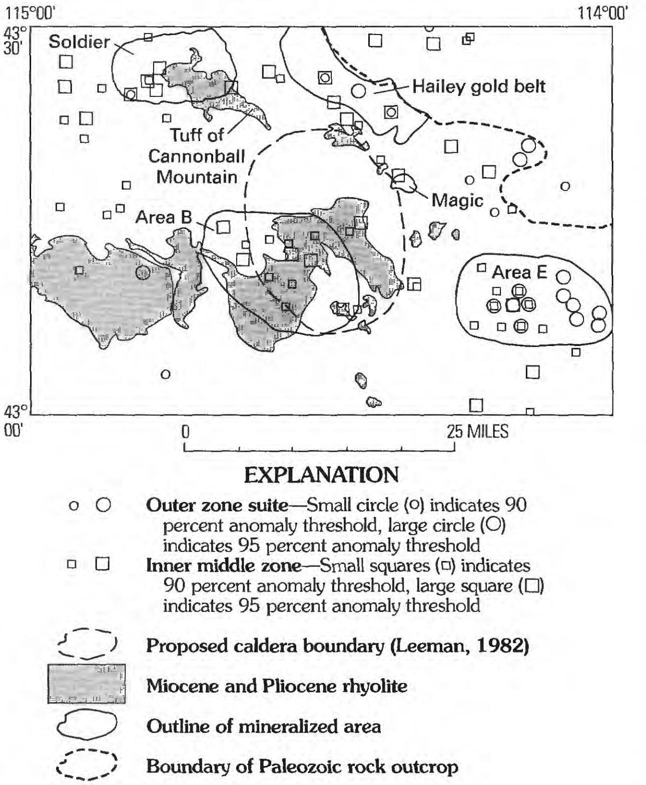


Figure 13. Sample sites in the southeast quadrant of the Hailey 1°×2° quadrangle, south-central Idaho, having anomalous outer and inner middle zone scores.

mineralized area are hosted in quartzite (Van Noy and others, 1986) of the upper part of the Wood River Formation (Worl and Johnson, this volume). Mineralized areas containing black shale commonly have high background concentrations of elements that define some of the element suites. Because of the high background produced by black shale, the occurrence of mineral deposits that are not closely associated with black shale might be underestimated in the area where Paleozoic rocks crop out.

SPECULATIVE TARGETS BASED ON GEOCHEMISTRY

Some areas in the Hailey quadrangle have strong geochemical signature but are not associated with known mining districts. Two areas in the southeastern part of the Hailey quadrangle, areas B and E (fig. 13), have strong geochemical signatures. Area B has an inner middle zone signature; area E has an outer zone signature. Area B is spatially associated with the buried caldera proposed by Leeman (1982) (fig. 13), who estimated that the main eruptive activity occurred 5–10 million years ago. The youngest (3 Ma) rhyolite in the area is the dome at Wedge Butte. Area E is associated with Quaternary basalt. The recent igneous activity in both areas B and E suggests the possibility of associated recent hydrothermal activity; however, the source or sources of the anomalous elements in these areas is not known. Further sediment and rock sampling is suggested for these two areas. Samples from area E should be analyzed for gold using an analytical technique that has a limit of detection close to the crustal abundance of gold, about 2 ppb.

Three areas (fig. 14) in the northwestern part of the Hailey quadrangle were sampled in more detail than the regional geochemical survey. Both stream-sediment samples and the heavy-mineral-fraction of stream sediments (concentrates) were collected in these areas, and the results of that sampling are reported in Kiilsgaard and Smith (in press).

In China and Canada, some black shale beds are characterized by enrichments in platinum-group elements, gold, arsenic, and zinc and molybdenum- and nickel-sulfide minerals (Coveney and Nansheng, 1991; Grauch and others, 1991). Grauch indicated that the mineralization is in a single stratiform, metal-rich sulfide layer. The sulfide layer is thin, from 2 to about 15 cm in thickness, and commonly overlies a phosphorite layer that is as thick as a meter. Some sediment samples associated with Paleozoic rocks in the Hailey quadrangle are enriched in molybdenum, nickel, arsenic, zinc, and phosphorus (Malcolm and Smith, 1992), and black shale in the Hailey quadrangle should be investigated for the occurrence of molybdenum-nickel sulfide minerals.

REFERENCES CITED

Anderson, A.L., 1951, Metallogenic epochs in Idaho: Economic Geology, v. 46, p. 592–607.

Bennett, E.H., 1980, Granitic rocks of Tertiary age in the Idaho batholith and their relation to mineralization: Economic Geology, v. 75, p. 278–288.

Burnham, C.W., and Ohmoto, H., 1980, Late-stage processes of felsic magmatism: Mining Geology Special Issue, v. 8, p. 1–11.

Chambers, J.M., Cleveland, W.S., Kleiner, B., and Tukey, P.A., 1983, Graphical methods for data analysis: Boston, Duxbury Press, 395 p.

Coveney, R.M., and Nansheng, C., 1991, Ni-Mo-PGE-Au-rich ores in Chinese black shales and speculations on possible analogues in the United States: Mineralium Deposita, v. 26, no. 2, p. 83–88.

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

Crock, J.G., Lichte, F.E., and Briggs, P.H., 1983, Determination of elements in National Bureau of Standards geological reference materials SRM 278 obsidian and SRM 688 basalt by inductively coupled argon plasma-atomic emission spectrometry: Geostandards Newsletter, v. 7, p. 334–340.

Dover, J.H., Berry, W.B.N., and Ross, R.J., Jr., 1980, Ordovician and Silurian Phi Kappa and Trail Creek Formations, Pioneer Mountains, central Idaho—Stratigraphic and structural revisions, and new data on graptolite faunas: U.S. Geological Survey Professional Paper 1090, 54 p.

Ferguson, R.B., Price, V., and Baucom, E.I., 1976, Field manual for stream sediment reconnaissance: U.S. Department of Energy Open-File Report GJBX–30(77), 55 p.

Fisher, F.S., and May, G.D., 1983, Geochemical characteristics of metalliferous Salmon River sequence, central Idaho: U.S. Geological Survey Open-File Report 83–670, 28 p.

Goldfarb, R.J., Leach, D.L., and Speckman, W.S., 1983, An application of rank-order statistics to exploration geochemistry [abs.]: Helsinki, Finland, Association of Exploration Geochemists, p. 30.

Grauch, R.I., Coveney, R.M., Jr., Murowchick, J.B., and Chen Nansheng, 1991, Black shales as hosts for unconventional platinum-group-element resources? Examples from southern China and the Yukon, Canada, and implications for U.S. resources, *in* Good, E., Slack, J., and Kotra, R., eds., USGS Research on Mineral Resources—1991, Program and Abstracts, Seventh Annual V.E. McKelvey Forum on Mineral and Energy Resources: U.S. Geological Survey Circular 1062, p. 33.

Grimes, J.G., 1982a, Hydrogeochemical and stream sediment reconnaissance basic data for Hailey and Challis quadrangles, Idaho: U.S. Department of Energy Open-File Report GJBX–55(82), 121 p.

———1982b, Supplement to hydrogeochemical and stream sediment reconnaissance basic data reports K/UR–405 and K/UR–408 through K/UR–443 [GJBX–52(82) through GJBX–88(82)]: U.S. Department of Energy Open-File Report GJBX–51(82), 25 p.

Hall, W.E., Rye, R.O., and Doe, B.R., 1978, Wood River mining district, Idaho—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, p. 579–592.

Howarth, R.J., Koch, G.S., Plant, J.A., and Lowry, R.K., 1981, Identification of uraniferous granitoids in the USA using stream sediment geochemical data: Mineralogical Magazine, v. 44, p. 455–470.

Johnson, K.M., and Worl, R.G., 1989, Preliminary assessment of mineral resources of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 17–22.

———in press, Assessment of the potential for undiscovered mineral resources in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, Idaho: U.S. Geologial Survey Bulletin 2064–S.

Kiilsgaard, T.H., 1982, Analytical determinations from samples taken in the Ten Mile West Roadless Area, Boise and Elmore Counties, Idaho: U.S. Geological Survey Open-File Report 82–1099, 32 p.

Kiilsgaard, T.H., Freeman, V.L., and Coffman, J.S., 1970, Mineral resources of the Sawtooth Primitive Area, Idaho: U.S. Geological Survey Bulletin 1319–D, p. D1–D174.

Kiilsgaard, T.H., and Hingley, Frank, 1989, Regional geology of the northwestern part of the Hailey 1°×2° quadrangle and ore deposits of the Atlanta district, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangle quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 35–36.

Kiilsgaard, T.H., Lewis, P.S., and Bennett, E.H., in press, Plutonic and hypabyssal rocks of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 2064–U.

Kiilsgaard, T.H., and Smith, C.L., in press, Geology, geochemistry, and mineral resources of the lower part of the Middle Fork Boise River drainage basin, Boise and Elmore Counties, Idaho: U.S. Geological Survey Bulletin 2064–AA.

Leeman, W.P., 1982, Geology of the Magic Reservoir area, Snake River Plain, Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 369–376.

Lewis, R.S., 1990, Geology, geochemistry and mineral potential of Cretaceous and Tertiary plutons in the eastern part of the Soldier Mountains, Idaho: Corvallis, Oregon State University, Ph.D. dissertation, 214 p.

Lewis, R.S., and Kiilsgaard, T.H., 1991, Eocene plutonic rocks in south central Idaho: Journal of Geophysical Research, v. 96, no. B8, p. 13295–13311.

London, D., 1990, Internal differentiation of rare-earth pegmatites—A synthesis of recent research, *in* Stein, H.J., and Hannah, J.L., eds., Ore-bearing granite systems—Petrogenesis and mineralizing processes: Geological Society of America Special Paper 246, p. 35–50.

Malcolm, M.J., and Smith, C.L., 1992, Analytical results and sample locality maps of USGS and NURE samples from the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 92–24, 267 p.

Motooka, J.M., 1988, An exploration geochemical technique for the determination of preconcentrated organometallic halides by ICP–AES: Applied Spectroscopy, v. 42, p. 1293–1296.

Nash, J.T., 1988, Interpretation of the regional geochemistry of the Tonopah 1°×2° quadrangle, Nevada, based on analytical results from stream-sediment and nonmagnetic heavy-mineral-concentrate samples: U.S. Geological Survey Bulletin 1849, 28 p.

Ovchinnikov, L.N., and Grigoryan, S.V., 1971, Primary halos and prospecting for sulfide deposits, *in* Geochemical exploration: International Geochemical Exploration Symposium, 3rd, Toronto, Canada, April 16–18, 1970; Canadian Institute of Mining and Metallurgy Special Volume 11, p. 375–380.

Rose, A.W., Hawkes, H.E., and Webb, J. S., 1979, Geochemistry in mineral exploration (2nd ed.): New York, Academic Press, 657 p.

Ross, C.P., 1931, A classification of the lode deposits of south-central Idaho: Economic Geology, v. 26, p. 169–185.

Sanford, R.F., Wooden, J.L., Doe, B.R., and Worl, R.G., 1989, Origin of ore deposits in the central Idaho black shale belt—Implications from lead isotopes, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 39–52.

Simons, F.S., 1981, A geological and geochemical evaluation of the mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–C, p. 85–180.

Snee, L.W., 1989, $^{40}Ar/^{39}Ar$ thermochronology of mineral deposits in the southern part of the Idaho batholith [abs.]: U.S. Geological Survey Open-File Report 89–639, p. 37–38.

Tauson, L.V., Ovchinnikov, L.N., Polikarpochkin, V.V., and Grigoryan, S.V., 1971, The use and development of geochemical prospecting methods for ore deposits in the U.S.S.R., *in* Geochemical exploration: International Geochemical Exploration Symposium, 3rd, Toronto, Canada, April 16–18, 1970; Canadian Institute of Mining and Metallurgy Special Volume 2, p. 42–43.

Toth, M.I., King, H.D., Kulik, D.M., and Leszcykowski, A.M., 1987, Mineral resources of the King Hill Creek Wilderness Study Area, Elmore County, Idaho: U.S. Geological Survey Bulletin 1721–B, 8 p.

Tschanz, C.M., and Kiilsgaard, T.H., 1986, Geologic appraisal of mineral resources in the eastern part of the Sawtooth National Recreation Area, Idaho, *in* Mineral resources of the eastern part of the Sawtooth National Recreation Area, Custer and Blaine Counties, Idaho: U.S. Geological Survey Bulletin 1545–C, p. 54–210.

Van Noy, R.M., Ridenour, James, Zilka, N.T., Federspiel, F.E., Evans, R.K., Tuchek, E. T., and McMahan, A.B., 1986, Economic appraisal of the eastern part of the Sawtooth National Recreation Area, Idaho: U.S. Geological Survey Bulletin 1545–E, p. 232–472.

Vine, J.D., and Tourtelot, E.B., 1970, Geochemistry of black shale deposits—A summary report: Economic Geology, v. 65, p. 253–272.

Worl, R.G., and Johnson, K.M., 1989, Geologic terranes and mineral deposit types in the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 3–16.

Lead-Isotope Characteristics of Ore Systems in Central Idaho

By Bruce R. Doe *and* Richard F. Sanford

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–M

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Lead-Isotope Characteristics of Ore Systems in Central Idaho

By Bruce R. Doe[1] *and* Richard F. Sanford[2]

ABSTRACT

More than 60 new lead-isotope analyses on ore minerals, igneous rocks, and shale were used to test the use of lead isotopes in mineral prospect evaluation of Cretaceous and Tertiary magmatothermal deposits of central Idaho in the Challis and Hailey 1°×2° quadrangles. The lead-isotope pattern is complex but separable into two groups: group A has many values of $^{206}Pb/^{204}Pb$ of 18.0 or less but with a continuum of values that exceed 21, whereas group B always has values of $^{206}Pb/^{204}Pb$ of 18.5 or more and most values are between 19 and 20. Although samples of groups A and B have similar values of $^{206}Pb/^{204}Pb$, values of $^{208}Pb/^{204}Pb$ are always greater for group A than for group B. Data for group A are similar to those for Early Proterozoic crystalline-rock terranes of the Rocky Mountains, whereas data for group B are similar to those for deposits in the Phanerozoic sedimentary rock terrane of the Great Basin. Large fluorine-poor molybdenum-porphyry deposits generally have lower values of $^{206}Pb/^{204}Pb$ (most notably Thompson Creek), although values are in both isotopic groups (for example, group A, Thompson Creek; group B, White Cloud). The Clayton Silver mine—the largest silver deposit in Idaho outside the Coeur d'Alene district—is in group A but has a value of $^{206}Pb/^{204}Pb$ between 18 and 19. Tin granites can also be in either group A or B and cannot be distinguished from molybdenum deposits by lead isotopes.

The area of interest is mostly within the central Idaho gravity low, which may represent extensions of the Idaho batholith in the subsurface. Although some exposures of Early to Middle Proterozoic sedimentary rocks equivalent to the Belt Supergroup are present around the margins of the area of interest, Precambrian crystalline rocks are present only in the eastern part of the Challis quadrangle. Two major areas containing group A lead isotopes in the Challis quadrangle are in areas of Bouguer gravity highs within the regional low and possibly indicate the presence of structural highs in the crystalline Precambrian basement. Thus, the group A lead isotopes may reflect a buried terrane of geographically restricted surface exposure. Two major areas in the Challis quadrangle containing group B lead isotopes have magnetic anomalies but not Bouguer gravity anomalies. Both of these areas are within a Basin and Range terrane south of the Trans-Challis fault system that trends southwest-northeast across the Challis 1°×2° quadrangle. Lead isotopes do not reflect a Rocky Mountain style of Proterozoic metamorphosed crystalline rocks at depth. Precambrian basement, if present, is deeper than the horizon at which isotopic signatures were acquired by magmas and ore fluids.

Central Idaho differs from the normal situation in which $\delta^{34}S$ is $-5\geq\delta^{34}S\leq+3$ in the best magmatothermal deposits because $\delta^{34}S$ tends to be high in the best deposits (for example, Thompson Creek $\geq+8$). Consideration thus should be given to prospects having nonradiogenic lead isotopes and heavy sulfur isotopes. Further confirmation of this unusual hypothesis is needed.

Both lead-isotope signatures and sulfur-isotope data of igneous rocks and ores are representative of crustal rocks and do not indicate mantle influences. Molybdenum and tin also are enriched in the crust. The granites therefore are most likely crustal melts or, if mantle melts, were extensively modified by reactions within the crust. Eocene tin granite has the same lead-isotope signature as Cretaceous and Eocene molybdenum granites, whether the lead-isotope signature is in group A or B. This similarity is surprising because molybdenum granites are classified as I-type (igneous rock parent) and tin granites as S-type (sedimentary rock parent). The Eocene tin granites could have formed, however, by remelting sulfide-mineralized Cretaceous granite or by extensively reacting tin granite melts with mineralized Cretaceous granite. The lead-isotope signatures would then be inherited from the Cretaceous granite; the inherited sulfur would keep the tin granite melts reduced, and the observed stannite mineralization would result rather than cassiterite mineralization. If the reduced tin granite magmas were produced from reactions with carbon-rich material, the material must have had a

[1]U.S. Geological Survey, National Center, Reston, Virginia 22092.
[2]U.S. Geological Survey, Federal Center, Denver, Colorado 80225.

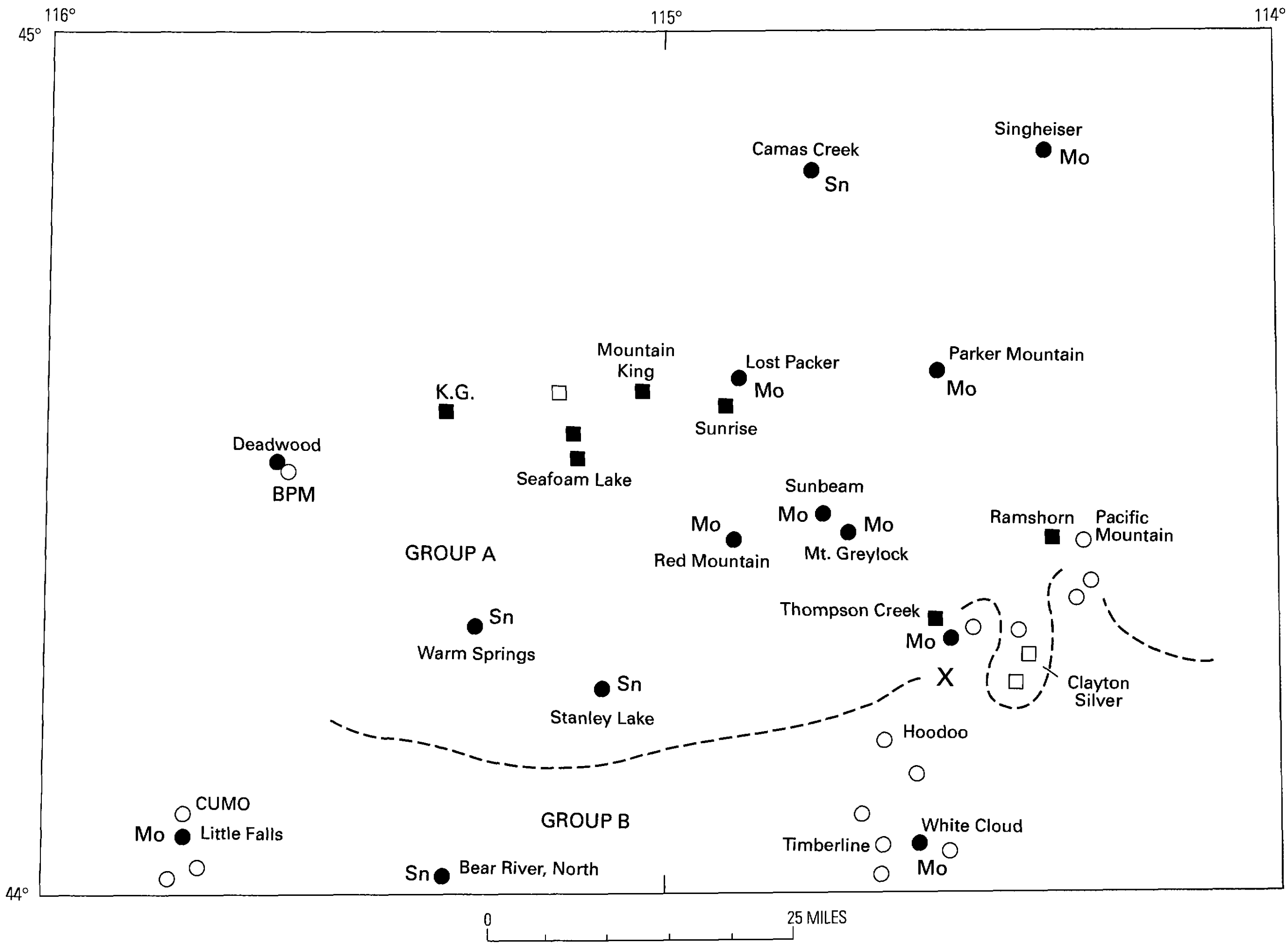


Figure 1. Map showing sample localities, Challis 1°×2° quadrangle, Idaho. Names are included for selected localities including all granite and rhyolite and all mineral deposits having $^{206}Pb/^{204}Pb$ of less than 18.0. The dashed line separates lead-isotope compositions of group A and group B. Symbols: open circle, ore having $^{206}Pb/^{204}Pb$ greater than 19.1; open square, ore having $^{206}Pb/^{204}Pb$ 18.1–19.1; solid square, ore having $^{206}Pb/^{204}Pb$ less than 18.0; solid circle, granite- or rhyolite-type (Sn, tin; Mo, molybdenum; BPM, base- and precious-metals); X, shale.

low lead content, such as petroleum, natural gas, or perhaps coal. Eocene molybdenum granite and rhyolite then could have evolved without extensive reaction with mineralized Cretaceous intrusive rocks.

INTRODUCTION

New lead-isotope analyses for 38 samples of ore minerals, 9 whole-rock and potassium feldspar samples from molybdenum porphyry units, 7 whole-rock samples of tin granite, and 2 samples of shale from the Challis and Hailey 1°×2° quadrangles, together with 12 previously published analyses (Hall and others, 1978; Davis, 1978), were used to test the use of lead isotopes in mineral prospect evaluation of Cretaceous and Tertiary magmatothermal deposits (figs. 1, 2). The northwestern part of the Challis quadrangle and the southwestern part of the Hailey quadrangle were not sampled. This study documents the conclusions of Doe and Delevaux (1985) that the lead-isotope ratios form two groups: group A, similar to lead-isotope ratios in the Rocky Mountain provinces, and group B, similar to those in the Basin and Range provinces. These data, data by Small (1968), and 59 additional analyses were used to further subdivide the major isotopic groupings into six subgroups by Sanford and others (1989) and Sanford and Wooden (this volume). The geology of the Challis quadrangle is described by Fisher and others (1983), McIntyre (1985), and Fisher and Johnson (1987b) and the geology of the Hailey quadrangle by Worl and others (1991).

Molybdenum granite and tin granite are thought to form by quite different processes; notably, tin granite differentiates under more highly reduced conditions than molybdenum granite (Ishihara, 1977). The molybdenum porphyry granite

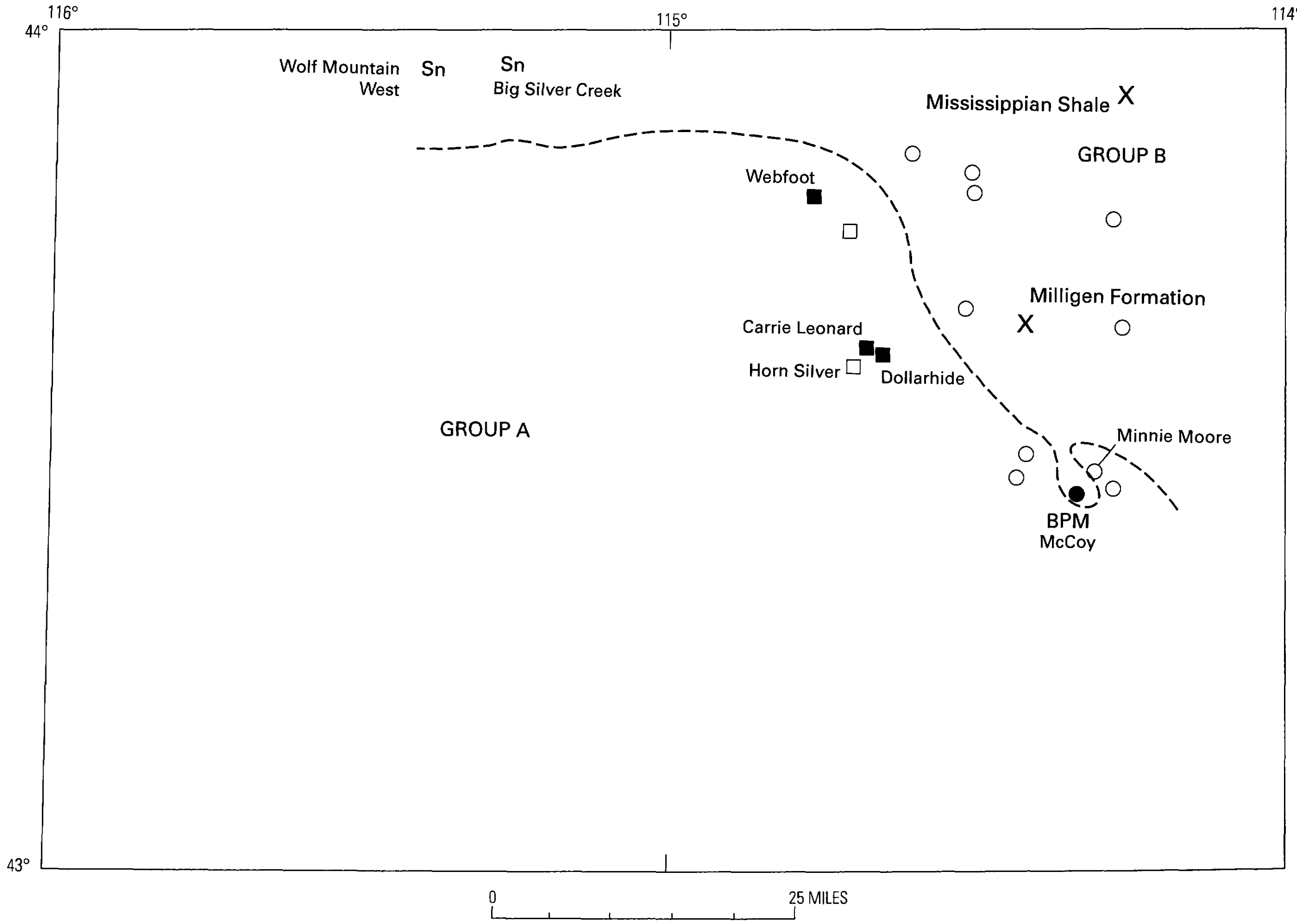


Figure 2. Map showing sample localities, Hailey 1°×2° quadrangle, Idaho. Names are included for selected localities including all granite and rhyolite and all mineral deposits having $^{206}Pb/^{204}Pb$ of less than 18.0. The dashed line separates lead-isotope compositions of group A and group B. Symbols: open circle, ore having $^{206}Pb/^{204}Pb$ greater than 19.1; open square, ore having $^{206}Pb/^{204}Pb$ 18.1–19.1; solid square, ore having $^{206}Pb/^{204}Pb$ less than 18.0; solid circle, granite- or rhyolite-type (Sn, tin; BPM, base- and precious-metals).

of central Idaho is included in a belt of subduction-related, I-type, molybdenum-enriched plutons that extends from east-central California to Northwest Territories (Westra and Keith, 1981; Hall and others, 1984). Molybdenum porphyry granite of central Idaho is low in fluorine (model 21b, Theodore, 1986), generally 0.0X percent, is believed to be free of tin minerals (Westra and Keith, 1981), and differs from molybdenum porphyry granite of the more famous Climax-type fluorine-rich ores (as much as 2–3 percent F) in Colorado and northern New Mexico (model 16, Ludington, 1986), which can contain tin minerals (Westra and Keith, 1981). Lead-isotope compositions of Climax-type molybdenum porphyry granite are given by Stein (1985); Church and others (1986) give lead-isotope data for a low-fluorine porphyry granite in Washington. Tin has almost no production in the United States, and, because tin-mineralized rocks (tin as stannite) and bodies of S-type granite, tin granite, are known in central Idaho, tin was selected as a second commodity for emphasis. Because the main tin mineralization is stannite in polymetallic veins in black shale, no mineral deposit model completely fits the Idaho setting; however, the mineralization was said to be of the Bolivian type by Hall (1985), and perhaps the porphyry tin (model 20a, Reed, 1986) and tin-polymetallic veins (model 20b, Togashi, 1986) models may be of interest. A variety of samples from silver-producing mines were also analyzed including a sample from the Clayton Silver mine, the largest producer of silver in Idaho outside the Coeur d'Alene district.

The goals of the study are to (1) determine the use of lead isotopes in central Idaho for mineral prospect evaluation using the methods outlined by Doe and Stacey (1974) and Doe (1979); (2) ascertain if coupling of sulfur isotopes and lead isotopes might refine this evaluation, as was done by Doe and others (1986) for west-central Montana; (3) examine if there is any information in the lead isotopes to show why molybdenum and tin are so common in central Idaho (for example, different or unique source rocks); and (4) refine the plumbotectonic classification of Zartman (1974) for central Idaho. Gravity (Bankey and Kleinkopf, 1988) and magnetic (Mabey and Webring, 1985; Webring and Mabey, 1987) geophysical data are utilized to strengthen the interpretations based on isotopic data.

Acknowledgments.—Most of the lead-isotope analyses of ore minerals were made under the supervision of M.H. Delevaux by Stephen Kish (now at Florida State University), Åke Johansson (now at Geologiska Institutionen, Stockholm, Sweden), Anne LeHuray (now at Lamont-Doherty Geological Observatory), and Holly Stein (now at the U.S. Geological Survey). Discussions with D. Mabey, M. Webring, and W. Hanna of the U.S. Geological Survey on geophysical measurements and their possible meaning were very helpful. Samples were provided by and we benefitted greatly from discussion with our colleagues, the late Wayne Hall, Frederick S. Fisher, Charles M. Tschanz, Thor Kiilsgaard, and S. Warren Hobbs. This paper was originally to be written by Wayne Hall, but he died before he was able to do so; we here recognize his enormous contributions to the study of Idaho geology.

THE THEORY

The theory of lead isotopes has been discussed by a number of authors. In order that this report may be self contained, we excerpt the following from a readily accessible source (Doe and Stacey, 1974, p. 758–759).

The variations in lead isotopic composition are the result of radioactive decay of ^{238}U and ^{235}U to the radiogenic isotopes ^{206}Pb and ^{207}Pb (uranogenic leads) respectively and ^{232}Th to the radiogenic isotope ^{208}Pb (thorogenic lead). The fourth stable lead isotope, ^{204}Pb, has no long-lived radioactive parent. The physicochemical effects that are so all important on the stable light isotopes are very minor in the case of heavy elements such as lead; variations of $^{238}U/^{235}U$ in nature are only 0.07 percent (an empirical justification for this statement is given in Doe [and others, 1966]). The generalized relationships of primordial and radioactivity produced (i.e., radiogenic) stable lead isotopes are shown in [fig. 3 of this report]. Because it is reasonable to believe that the earth began with some primeval lead (primordial lead) just as it began with some silver and other elements, the primeval lead isotope values are adopted from the least radiogenic leads ever found, i.e., the values in the troilite phase of meteorites * * * [fig. 3 of this report] shows that the abundance of ^{206}Pb has about doubled since the earth formed, ^{207}Pb has increased by about one-half, and ^{208}Pb by about one-third. Unlike U-Th-Pb dating where the emphasis is placed on the lead produced by radioactive decay of uranium and thorium present in a phase, in common lead studies emphasis is placed on interpreting the significance of the initial lead isotopic composition of a phase, i.e., the lead that cannot be accounted for by the in situ decay of uranium and thorium in the phase being analyzed.

The most popular way of representing lead isotopic compositions is through ratios of a radiogenic isotope to the nonradiogenic isotope whose abundance does not change with time—$^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$—because such representation are subject to the simplest mathematical interpretation. A present-day lead isotope ratio evolves from an initial lead isotope ratio plus radiogenic additions by radioactive decay involving the product of a time term with the ratio of the radioactive parent isotope to the nonradiogenic isotope of the daughter element ($^{238}U/^{204}Pb$, $^{235}U/^{204}Pb$, and $^{208}Pb/^{204}Pb$ which are always normalized to the present day to cancel variable effects of radioactive decay) * * * Although the lead isotope ratios are not subject to physicochemical fractionation, the ratios of the parent radioactive isotope to the nonradiogenic isotope of the daughter element are. Changes in these ratios eventually result in differences in lead isotope ratios. For example, the present-day value of $^{206}Pb/^{204}Pb$ in a Precambrian mineral such as zircon or monazite will be very large (perhaps several thousand) because $^{238}U/^{204}Pb$ is very large (perhaps tens of thousands). In contrast, the present-day values of $^{206}Pb/^{204}Pb$ in a Precambrian mineral such as galena or feldspar will be relatively small (perhaps less than 20) because $^{238}U/^{204}Pb$ is near zero. Variations in the values of parent/daughter for rocks are more subtle—for example, a subalkaline basalt may have a value of $^{238}U/^{204}Pb$ of about 6, whereas a silicic volcanic might be more like 20—but usable variations in lead isotope ratios still result. Customarily lead isotopes have evolved through a sequence of stages which will be discussed below. A number of ores and igneous rocks have leads that evolved under conditions that closely approach single-stage conditions, that is, where the only changes in the value of the parent/daughter ratio are due to the radioactive decay of the parent. These are the model leads and the model lead ages approach the time of formation of the deposit or igneous rocks. Many ores and igneous rocks have leads that clearly did not form under such simple conditions and are anomalous. A number of these anomalous leads approximate two-stage conditions, that is, the values of the parent/daughter ratio were altered sometime after the earth was formed, usually to a variety of values, so that the present-day lead isotope ratio evolved under two stages of parent/daughter values. Leads that underwent two-stage evolution are called anomalous because model-lead "formation" ages calculated from them have a demonstrated tendency to be grossly incorrect. Secondary Pb-Pb isochron ages may often be calculated, making use of the near-constant value of $^{238}U/^{235}U$, for either the formation-age or the age of the source material from which the lead was derived (source-ages), if one of the ages can be independently determined. Because the earth has had a complex history, often three-stage systems must be involved in the evolution of lead in many ores and igneous rocks. Three-stages systems are difficult to interpret, but several special cases that have some characteristics of two-stage systems [can be solved].

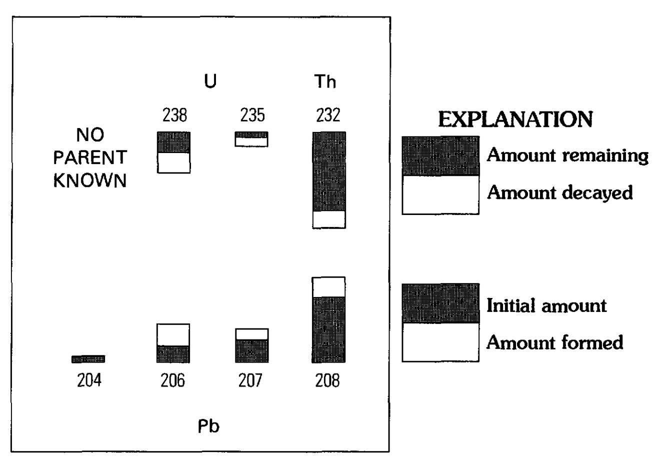


Figure 3. Isotopic relationships between uranium, thorium, and lead in the Earth's crust. Modified from Cannon and others (1961, fig. 1). Reproduced with permission, as modified, from Economic Geology Publishing Company.

Table 1. Maximum $^{206}Pb/^{204}Pb$ for Mesozoic and Cenozoic magmatothermal ore deposits in various dollar-value production groups. [Dollar amounts not adjusted for inflation. Modified from Doe (1979)]

Production (dollars)	Maximum observed $^{206}Pb/^{204}Pb$	Largest districts or deposits in production group
≥$1,500,000,000	18.0	Butte, Montana; Bingham, Utah; Climax and Urad-Henderson, Colorado; Globe-Miami, Arizona; Santa Rita-Hanover and Tyrone, New Mexico.
$150,000,000–$1,500,000,000	19.1	Tintic and Park City, Utah; pene- and post-caldera ores, San Juan Mountains and Leadville-Gilman, Colorado; Bisbee, Arizona.
$15,000,000–$150,000,000	22	Selected deposits include Wood River, Idaho; Ophir and Milford, Utah; Alma and Central City-Silver Plume, Colorado.
$1,500,000–$15,000,000	20	Selected deposits include Star and Big Cotton Wood, Utah; Hilltop, Colorado.
≤$1,500,000	25	Santaquin (Mt. Nebo), Utah; Yarmony and precaldera ores (Embargo district), San Juan Mountains, Colorado.

THE MODEL

The model is well illustrated in table 1, modified from Doe (1979). In the 15 years since the table was first published, there has been little change in dividing isotopic compositions between the economic groupings. Some supergiant deposits have been found that have associated galenas with values of $^{206}Pb/^{204}Pb$ of 18 or more, and a bounding value of 18.1 might work a little better than the value of 18.0 given in table 1; however, some grayness in the dividing isotope-ratio values is to be expected. Doe (1979, p. 229) pointed out that, although all medium-sized to supergiant magmatothermal deposits known at the time had values of $^{206}Pb/^{204}Pb$ of 19.1 or less and ratios similar to plutonic leads, not all prospects having the appropriate lead-isotope ratios may be expected to be of medium to supergiant in size.

Of course, the lead isotope data are only pointing to trends of economic potential, and many valueless prospects are to be expected with $^{206}Pb/^{204}Pb$ values less than 19.1. Conversely, not all districts currently inactive have been mined out, and much more may remain to be found. A case in point is the Leadville/Gilman grouping which, according to Ogden Tweto (oral commun., 1977) might move into the supergiant category and be a lead-zinc skarn-type deposit to rival porphyry coppers, if the excessively thick cover did not hamper prospecting.

This cautionary note should not be ignored, and the study of the Hailey and Challis quadrangles forms an interesting test of the statement.

We still know of no operating Cretaceous or Tertiary magmatothermal deposit that has a value of $^{206}Pb/^{204}Pb$ of 20 or more. The observation that the more radiogenic the lead isotopes, the less likely one is to find an economic ore deposit is probably a reliable indicator. If exceptions are found, they are likely to involve the precious metals—gold, silver, and platinum—that respond to monetary factors as well as to utilitarian values. Lateral-secretion deposits can have more radiogenic values, such as those in southeast Missouri with values of $^{206}Pb/^{204}Pb$ of 20.0–22.5, and a judgment must be made as to just what class of ore deposit to which the prospect belongs. This judgment is also not always easy. Other kinds of ore deposits may also be involved in a regional study, such as bedded ores in black shale of essentially syngenetic origin. These usually have model lead ages in approximate agreement with the sedimentary or submarine volcanic host rocks by both the $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ models of Stacey and Kramers (1975). In the Cretaceous-Tertiary magmatogenic ore deposits, $^{208}Pb/^{204}Pb$ provides little information of value in plumbic prospecting but commonly helps to distinguish the Cretaceous-Tertiary deposits because agreement between the $^{206}Pb/^{204}Pb$ age and the $^{208}Pb/^{204}Pb$ age is poor.

SULFUR ISOTOPES

Doe and others (1986) presented a first attempt to combine sulfur isotopes and lead isotopes to refine the plumbic prospecting evaluation, and Rye and others (1980) used sulfur isotopes to classify ore deposit types. We expand on both studies herein.

Unlike lead-isotope ratios, which are controlled by radioactive decay and mixing, sulfur-isotopic variations are a consequence of physico-chemical effects and, perhaps, mixing. Data are usually expressed relative to a standard,

$$\delta = \left[\frac{(^{34}S/^{32}S)_{sample}}{(^{34}S/^{32}S)_{standard}} - 1 \right] \times 1{,}000$$

with the values reported in per mil. Although sulfur-isotope ratios are difficult to utilize by themselves in mineral-prospect evaluation, especially on single samples, they help us refine the lead-isotope results.

Figure 4 shows values of $\delta^{34}S$ for a variety of hydrothermal ore deposits. Magmatothermal deposits such as porphyry copper deposits generally have values of $\delta^{34}S$ for pyrite of between –3 and +1 (Ohmoto and Rye, 1979). Because we are analyzing galena, and because sulfur

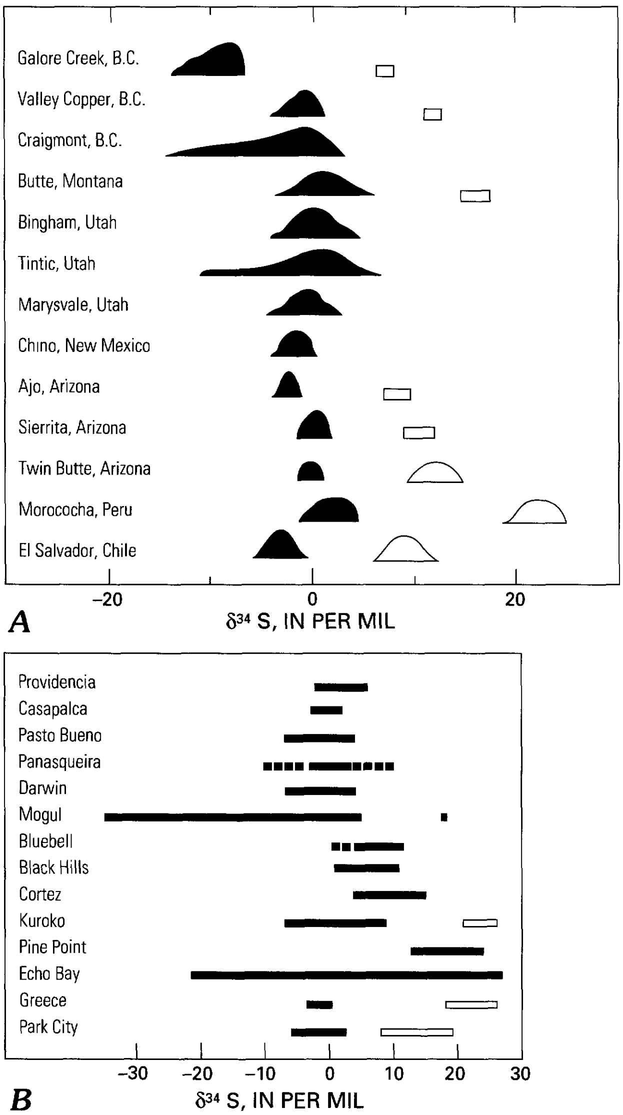


Figure 4. Sulfur-isotope data for some magmatothermal ore deposits. Solid line or area indicates sulfide mineral; open line or area indicates sulfate mineral; dashed line indicates late-stage mineral. *A*, Porphyry copper deposits. Modified from Ohmoto and Rye (1979, fig. 10.12). Reproduced with permission from John Wiley and Sons. *B*, Various hydrothermal deposits. Modified from Rye and Ohmoto (1974, fig. 1). Reproduced with permission, as modified, from Economic Geology Publishing Company.

isotopes in galena are usually somewhat isotopically lighter than sulfur isotopes in pyrite, we use a range for igneous rocks of –5 to +3. There are a few exceptions; values for Galore Creek are around –10 and values for scattered samples from some other deposits are close to –10 (Craigmont and Tintic). No subaerial magmatothermal deposits are known to have values of $\delta^{34}S$ greater than +6 (the extreme value for the Henderson molybdenum porphyry of Colorado; Stein, 1985). Normally, however, heavy sulfur-isotope values are indicative of an interaction with evaporites, as for Mississippi Valley deposits (such as Pine Point), and some light sulfur-isotopic values are in ores associated with organic material (such as Mogul) or where sulfur in the ore fluids is mainly in the form of sulfates (anhydrite, barite, alunite, and so on). Echo Bay is an unusual Precambrian uranium-nickel-silver-copper deposit.

Although we emphasize subaerial magmatothermal deposits in this study, a number of other kinds of ore deposits must be taken into account. Sulfur-isotope values for submarine volcanogenic massive-sulfide deposits (such as kuroko-type deposits) can range widely, and the sulfur isotopes can be heavy as a result of the role of seawater in the ore fluid. As seen in figure 4, however, the more values of $\delta^{34}S$ for galena are lighter than –5, the less likely that one is dealing with a major magmatothermal ore deposit. In fact, few major ore deposits of any origin contain such light sulfur isotopes, although some notable exceptions are known among shale-hosted ores, such as the Kupferschiefer in Germany and Udokan in the Soviet Union. The same predictive statement could be said for values greater than +3 for magmatothermal ores; however, a number of important ore deposits of other origins—especially of lateral-secretion (Mississippi Valley) and sedimentary exhalative (SEDEX) origin—exist and must be considered in the present study.

By combining lead and sulfur isotopes, the plumbic-prospecting evaluation can be refined (table 2). In many cases, whether one is looking at a magmatothermal deposit can be determined just from looking at the lead and sulfur isotopes; however, incorporation of as much other geologic information as possible improves the isotopic assessment. For example, volcanogenic and shale-hosted deposits should be of an age similar to the host rocks, lateral-secretion deposits normally form at temperatures lower than economic magmatothermal deposits, a crosscutting ore is not stratiform, and so forth.

GEOLOGY

The stratigraphy, structure, and geologic history of the study area have been extensively described as a result of the Conterminous United States Mineral Assessment Program (CUSMAP) of the U.S. Geological Survey and associated research groups (McIntyre, 1985; Fisher and Johnson, 1987a; Link and Hackett, 1988; Winkler and others, 1989; this volume). We briefly review the geology of the parts of the Challis and Hailey quadrangles that were sampled for lead isotopes.

Although Middle Proterozoic metasedimentary rocks of the Belt Supergroup are widely exposed north of the study area, the only Precambrian rocks exposed in the study area are in the structural core of the Pioneer Mountains (Dover, 1981). Highly deformed Middle Proterozoic or older gneiss is overlain by less deformed, medium- to high-grade, younger Proterozoic to Ordovician metasedimentary rocks (Dover, 1981). Preexisting and (or) underlying Archean

Table 2. Hypothesis for lead and sulfur isotopes in mineral prospect evaluation.
[Modified from Doe and others (1986). For examples of magmatothermal ore deposits, see table 1]

Ore type and largest deposits	$^{206}Pb/^{204}Pb$	$\delta^{34}S$
Lateral secretion: southeast Missouri, Tri-State, east and central Tennesee[1,2]	≥19.1	≥0
Magmatothermal, very small	≥19.1	−3 to +3
Magmatothermal, very small, oxidized	≥19.1	≤−3
Lateral secretion: Pine Point	≥18.1–≤19.1	≥0
Magmatothermal, large[3]	≥18.1–≤19.1	−3 to+3
Submarine-volcanogenic or shale hosted: Bathhurst, Canada; Rammelsberg, West Germany (Paleozoic)[3]	≥18.1–≤19.1	≥0
Shale hosted: Kupferschiefer, West Germany (Paleozoic)	≥18.1–≤19.1	≤−5
Magmatothermal, medium, oxidized: Pasto Bueno, Peru	≥18.1–≤19.1	≤−3
Magmatothermal, supergiant, rejuvenated cratons (Cretaceous-Tertiary)[3]	≤18.1	−3 to +3
Submarine-volcanogenic or shale hosted: Mt. Isa, Australia (Proterozoic)	≤18.1	≥0
Submarine-volcanogenic: Kidd Creek, Canada (Archean)	≤18.1	~0

[1]Lateral-secretion deposits usually plot on or below the evolution curve of Stacey and Kramers (1975) for $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$.

[2]Unusual samples from Mississippi Valley deposits have $\delta^{34}S$ ≤+3. The most notable exception is the Tri-State district where $\delta^{34}S$ is 0 to −10.

[3]Magmatothermal ores usually plot above the curve of Stacey and Kramers (1975) for $^{208}Pb/^{204}PbPb$ versus $^{206}Pb/^{204}Pb$ in rejuvenated cratons; however, Creede, Colorado, does not. Submarine volcanogenic ores usually plot on or below the $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ curve, but the Kuroko ores of Japan plot above it. Some molybdenum porphyry deposits in the Challis 1°×2° quadrangle have values of $\delta^{34}S$ ≥+3 (Howe and Hall, 1985; this paper). Because these deposits may be of supergiant size (such as Thompson Creek), an additional refinement would be to survey the sulfur isotopes in a region to establish the possibility of an anomalous character.

crust to the south is indicated by xenoliths in lavas of the Snake River Plain that suggest Archean crust was present prior to major thermal events dated at about 3.4–3.1 Ga and 2.8 Ga (Leeman and others, 1985).

Paleozoic sedimentary rocks underlie much of the study area and are host to numerous ore deposits, many of which are represented in this study. These rocks span all time periods from Cambrian to Permian. A continental shelf environment during the Paleozoic in the eastern part of the study area is characterized by predominantly shallow water carbonate rocks, whereas continental slope and basin environments in the western part are characterized by fine-grained, commonly carbonaceous, argillaceous rocks of the "black shale belt" (Skipp and Hall, 1975; Skipp and Sandberg, 1975; Poole and others, 1977; Skipp and Hait, 1977; Hall and others, 1978; Hobbs, 1987; McFaddan and others, 1988).

Rocks representing the carbonate shelf include the Cambrian or Ordovician Bayhorse Dolomite, Ordovician Ella Dolomite, Ordovician Saturday Mountain Formation, Devonian Jefferson Formation, and undifferentiated Proterozoic and Paleozoic limestone in roof pendants of the Idaho batholith. Ore in these carbonate rocks typically formed by replacement, although vein- and breccia-filling are locally important (Hobbs, 1987; Hobbs and others, 1987). The Bayhorse Dolomite hosts ore deposits of lead, zinc, silver, and fluorite at the Pacific mine and lead, silver, and zinc at the Riverview mine (Davis, 1978; Hobbs, 1987). The Ella Dolomite hosts lead, silver, and zinc deposits of the Clayton Silver mine (Hobbs, 1987). The Saturday Mountain Formation hosts vein and replacement deposits of lead, copper, and zinc in the Dryden mine (Mitchell and others, 1986; Hobbs, 1987). The Jefferson Formation hosts deposits of lead, silver, zinc, and gold at the Hilltop mine (Mitchell and others, 1981). A total of $24.6 million (values uncorrected for inflation) of ore has been produced from carbonate-hosted deposits in the Challis quadrangle. Mineralization probably took place in the Late Cretaceous (Hobbs and others, 1987).

The black shale belt of the slope-basin environment includes the Ordovician Ramshorn Slate, Devonian Milligen Formation, Paleozoic Salmon River assemblage, Mississippian Copper Basin Formation, and Pennsylvanian and Permian Wood River and Dollarhide Formations (Paull and others, 1972; Hall and others, 1974; Sandberg and others, 1975; Hall, 1985; Wavra and others, 1986; Hall and Hobbs, 1987; Turner and Otto, 1988). Black-shale-hosted deposits are predominantly in veins that yielded mainly lead, silver, zinc, antimony, and tin (Hall, 1987a). The veins are typically beneath major thrust faults or near regional unconformities and are typically associated with Late Cretaceous or Eocene intrusive rocks (Hall, 1987a). Some of these deposits (Livingston, Hoodoo, and Triumph mines) have textures suggesting at least partial syngenetic deposition of ore and sediments. Deposits of interest for their tin concentrations include the sulfide-bearing quartz and siliceous vein deposits of the Timberline and Patti Flynn mines, among others, in the Fourth of July, Galena, and Boulder Basin mineralized areas (Tschanz and others, 1986; Tschanz and Kiilsgaard, 1986; Hall, 1987a).

Late Cretaceous and Tertiary igneous rocks of the Idaho batholith and Challis Volcanic Group underlie the largest part of the study area (McIntyre and others, 1982; Bennett and Knowles, 1985; Ekren, 1985; Kiilsgaard and Lewis, 1985; Fisher and Johnson, 1987a; Johnson and others, 1988; Moye and others, 1988). Igneous rocks include Cretaceous intrusive rocks, Tertiary intrusive rocks, and Tertiary extrusive rocks (Challis Volcanic Group). Biotite granodiorite is the most widespread of the Cretaceous intrusive rocks; hornblende-biotite granodiorite, porphyritic granodiorite, tonalite, quartz diorite, muscovite-biotite granite, and leucocratic granite are also present (Johnson and others, 1988; Lewis, 1989). The granodiorite and granite further belong to distinct potassic and sodic suites (Lewis, 1989). Tertiary intrusive rocks are compositionally bimodal, including both granite and quartz monzodiorite (Johnson and others, 1988). Tertiary extrusive rocks of the Challis Volcanic Group include basaltic, intermediate, and rhyolitic types (McIntyre and others, 1982). Numerous dikes and hypabyssal rocks were feeders for the extrusive rocks, and these are compositionally and texturally transitional between intrusive and extrusive varieties. Virtually all of the Tertiary rocks were emplaced during the Eocene between 51 and 40 Ma.

Cretaceous and Tertiary igneous rocks are associated with characteristic types of ore deposits (Hardyman, 1985; Hardyman and Fisher, 1985; Kiilsgaard and Bennett, 1985, 1987a, b; McIntyre and Johnson, 1985; Kiilsgaard and Hingley, 1989; Lewis, 1989; Moye and others, 1989). Stockwork molybdenum deposits and mixed precious- and base-metal veins are thought to be genetically related to Cretaceous intrusions. Stockwork molybdenum deposits of probable Cretaceous age include the Little Boulder Creek deposit (White Cloud) (>100 million tons averaging 0.15 percent MoS_2 and 0.02 percent WO_3) and the productive silver-bearing Thompson Creek open-pit mine (200 million tons averaging 0.18 percent MoS_2) (Hall, 1987b). Mineralized rock at the Thompson Creek mine is in the stock but near the contact with the Salmon River assemblage; mineralized rock at the White Cloud deposit is in the Wood River Formation, but the ore grade is greatest near the stock. The molybdenum-bearing copper-silver veins of the Lost Packer mine and the zinc-lead-silver veins of the Deadwood, Greyhound, K.G., Seafoam, Mountain King, Sunrise, and Comeback mines also are thought to be related to Cretaceous intrusions (Kiilsgaard and Bennett, 1987c).

Deposits of probable Tertiary age include vein stockwork and disseminated molybdenum deposits, precious-metal veins, and high-level precious-metal veins (Fisher and others, 1987; Johnson and Fisher, 1987; Kiilsgaard and Bennett, 1987b, d). Significant Tertiary molybdenum mineralization is in the CUMO (1,000 million tons at an estimated 0.1 percent MoS_2) and Little Falls (a multimillion ton deposit at 0.05 percent MoS_2) deposits and other smaller deposits such as that at Red Mountain (Kiilsgaard and Bennett, 1987d). Molybdenum typically is in subparallel quartz veins that are in or near intensely altered host rocks that are cut by Eocene rhyolite dikes. Precious-metal veins include the Comeback, Singheiser, and Parker mines (Fisher and others, 1987a). Gold and silver are the main commodities, but copper, lead, and zinc are present in variable amounts. The veins are epithermal and are in altered host rocks typically cut by Eocene rhyolite dikes. Rhyolite-hosted precious-metal deposits include the Sunbeam, Parker Mountain, and Singheiser mines (Johnson and Fisher, 1987). Ore is present as irregular stockwork quartz veins and disseminations in hydrothermally altered host rocks, typically rhyolite.

The three types of Tertiary deposits discussed here may be genetically related. At many locations elsewhere, stockwork molybdenum deposits pass upward and outward into precious-metal vein deposits, which in turn grade into stockwork and disseminated precious-metal deposits. In the study area, all three deposit types are associated with hydrothermally altered hypabyssal rhyolitic rocks and are concentrated along the Trans-Challis fault system. Precious-metal veins and high-level precious-metal deposits in the study area are commonly enriched in molybdenum. Thus, for this study, samples from these molybdenum-enriched deposits and their associated intrusive rocks are classified with molybdenum systems.

Although minable tin deposits are absent, much Tertiary granite in the study area has geochemical characteristics of tin-bearing granite elsewhere. Most Tertiary intrusive rocks in the area are evolved or anorogenic (A-type), high-potassium, slightly peraluminous, probably alkali-calcic to calc-alkaline granite (Bennett and Knowles, 1985). They typically have high tin, tungsten, uranium, thorium, molybdenum, niobium, zirconium, certain rare earth elements, lithium, beryllium, and fluorine contents and low strontium and barium contents (Bennett and Knowles, 1985; Cole Smith, written commun., 1990). Thus, their major- and trace-element abundances are similar to those of tin granites worldwide (for example, Tischendorf, 1977).

ANALYTICAL TECHNIQUES

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

Lead, uranium, and thorium concentrations were determined by stable-isotope dilution as described in Doe and Delevaux (1980).

Sulfur-isotope compositions were determined by the U.S. Geological Survey laboratories in Menlo Park, Calif., or by Geochron Laboratories Division of Kruger Enterprises, Inc. Analyses from Geochron Laboratories have a standard error of the mean of 0.3 per mil for $\delta^{34}S$. Although this laboratory uses Trenton City troilite as a reference sample on which they reported values of 0.18 and 0.20 per mil, their data are systematically 0.4 per mil greater than those obtained on a common sample that was analyzed several times in both Geochron Laboratories and the U.S. Geological Survey laboratories in Denver. In order to facilitate direct comparison of data from Geochron Laboratories with that from the U.S. Geological Survey, 0.4 per mil arbitrarily has been subtracted from analyses done by Geochron Laboratory. The difference, however, is not considered to be significant compared to the isotopic differences between samples.

DATA

LEAD ISOTOPES

The lead-isotope data (tables 3–5) form a complex pattern (figs. 5, 6) that is easily separable into two groups on a plot of $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ (Doe and Delevaux, 1985). The two groups were originally defined on the basis of ore leads. Group A forms a diffuse cluster having relatively high $^{208}Pb/^{204}Pb$ for its $^{206}Pb/^{204}Pb$, whereas group B forms a fairly tight cluster having relatively low $^{208}Pb/^{204}Pb$ for its $^{206}Pb/^{204}Pb$. Igneous rocks of group A lie close to group A ore leads, and igneous rocks of group B lie close to group B ore leads. Ore samples of group A lie on or above the regression line for igneous rocks of group A in figure 6. Many samples in group A have values of $^{206}Pb/^{204}Pb$ of 18.0 or less, but some values are as high as 21.0 or more. In contrast, all samples of group B have values of $^{206}Pb/^{204}Pb$ of 18.5 or more, and most values are between 19.0 and 20.0. Where $^{206}Pb/^{204}Pb$ values overlap, group A shows a trend of $^{208}Pb/^{204}Pb$ values greater than group B. The isotopic distinction is so clear that there is confusion in classification of only one ore mineral in a group of more than 50 ore minerals and (or) igneous rocks. Groups A and B can be further subdivided (Sanford and others, 1989; and Sanford and Wooden, this volume), but these subdivisions are not discussed here.

Data for group A are similar to those from the Rocky Mountains or from area I in the classification of Zartman (1974) (see fig. 7), although there is some tendency for $^{208}Pb/^{204}Pb$ values in central Idaho to be greater for given values of $^{206}Pb/^{204}Pb$. For example, the average values of the two samples from the Thompson Creek mine in central Idaho for $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ are 17.597, 15.588, 40.945, respectively, whereas the Climax-type (fluorine-rich) molybdenum deposit having the highest values of $^{208}Pb/^{204}Pb$ in the Front Range of the Rocky Mountains of Colorado is the Henderson mine (342 million metric tons, averaging 0.297 percent Mo; David Singer, oral commun., July 11, 1991), which has ratios of 17.696, 15.508, and 39.375, respectively (Stein, 1985). The Henderson mine also has the highest $\delta^{34}S$, as much as +6 (Stein, 1985), for a molybdenum porphyry, aside from that for Thompson Creek. For two samples from the low-fluorine Cannivan Gulch molybdenum porphyry deposit of Montana (203 million metric tons of 0.096 percent Mo; David Singer, oral commun., July 11, 1991), the lead-isotope ratios are 17.602, 15.552, and 38.378, respectively (Doe and others, 1986). The similarity of $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ for all three deposits is remarkable, close to analytical uncertainties.

In contrast, group B values in central Idaho are similar to those in the Great Basin or area II in Zartman's classification (see fig. 7). For example, a feldspar and a galena from the White Cloud molybdenum porphyry system have average values of 19.304, 15.705, and 39.310 for $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$, respectively, whereas these ratios for the low-fluorine Mt. Tolman molybdenum porphyry deposit (1,180 million metric tons of 0.055 percent Mo; Donald Singer, oral commun., July 11, 1991) in the Zartman area II of Washington are 19.377, 15.672, and 39.191 (Church and others, 1986). Again, the differences in isotope ratios between these two deposits are close to analytical uncertainties.

Large deposits of molybdenum porphyry generally have lower values of $^{206}Pb/^{204}Pb$ but are in either isotopic grouping (for example, group A, Thompson Creek; group B, White Cloud and CUMO). The data for the largest and most productive deposit in group A (Thompson Creek) follow the apparent rule observed by Doe (1979) of $^{206}Pb/^{204}Pb$ values of 18 or less for supergiant and many giant magmatothermal ore deposits of area I of Zartman.

Previously (Doe, 1979), consideration was given only to deposits in area Ib of Zartman (oldest rocks are Proterozoic basement to the south) with the exception of Butte in area Ia of Zartman (oldest rocks are Archean in the north). Data for Thompson Creek add additional evidence that the same lead-isotope rule of $^{206}Pb/^{204}Pb$ less than 18 for supergiant deposits can be applied to area Ia as well as to area Ib. The lead in potassium feldspar from the stock at Thompson Creek is highly radiogenic ($^{206}Pb/^{204}Pb \geq 20$), however, and, because the data were the last obtained, we were unable to determine whether this radiogenic nature is a primary feature of the stock or is the result of additions of radiogenic lead to the feldspar subsequent to crystallization of the stock.

Mineral-deposit evaluation through use of lead isotopes has not yet been assessed for the Great Basin area or area II of Zartman. Galenas from both the White Cloud and CUMO molybdenum porphyry bodies are toward the less radiogenic

Table 3. Sample information, Challis and Hailey 1°×2° quadrangles, Idaho.
[Two-letter designation following sample number indicates sample type: GN, galena; BO, boulangerite; BA, barite; JA, jamesonite; KF, potassium feldspar; OR, ore; WR, whole rock. Latitude in degrees, minutes, and seconds north, longitude in degrees, minutes, and seconds west]

Group	Sample number	Name or location	Age and host rock type	Relative age	Deposit type	Commodities	District	County	Latitude	Longitude
				Ore samples						
A	Buckskin? GN	Thompson Creek–Buckskin Drill Road	Cretaceous Idaho batholith	Epithermal	Vein	Pb, precious metals?	Bayhorse	Custer	44 19 05	114 33 43
A	S76-895 GN	Thompson Creek–Buckskin Drill Road?	Cretaceous Idaho batholith	Epithermal	Porphyry	Mo	Bayhorse	Custer	44 19 05	114 33 43
B	T 49M GN	Bulldozer cut/dump	Pennsylvanian-Permian Wood River Formation?	Epithermal			Boyle Mountain	Blaine	43 39 58	114 30 55
B	F112 GN	Carlson Gulch–Grimes Creek	Cretaceous Idaho batholith	Epithermal			Grimes Pass	Boise	44 01 56	115 45 18
A	T 68M GN	Carrie Leonard	Lower Permian upper member of the Dollarhide Formation	Epithermal	Vein		Carrietown	Camas	43 36 08	114 41 41
A	SWH 78-65 GN	Clayton Silver	Ordovician Ella Dolomite	Epithermal	Vein-replacement	Pb, Ag, Zn	Bayhorse	Custer	44 16 48	114 24 44
A	80WH-1 GN	Clayton Silver	Ordovician Ella Dolomite	Epithermal	Vein-replacement	Pb, Ag, Zn	Bayhorse	Custer	44 16 48	114 24 44
B	79T565M GN	Combination? New vein	Pennsylvanian-Permian Wood River Formation, limy sandstone	Epithermal	Vein		Galena	Blaine	43 51 55	114 36 12
B	F107 GN	Comeback	Cretaceous Idaho batholith	Epithermal	Vein	Precious metals	Grimes Pass	Boise	44 00 24	115 48 32
B	T597M GN	Crest Lake?–Trail Creek?	Devonian Milligen Formation(?)?	Epithermal			Warm Spring	Blaine	43 47 01	114 17 52
B	79T 5M GN	Crown Point	Pennsylvanian-Permian Wood River Formation, limy sandstone	Epithermal			Boulder Basin	Blaine	43 49 41	114 30 25
A	584-13 GN	CUMO	Tertiary Idaho batholith	Epithermal	Porphyry	Mo	Pioneerville	Boise	44 05 00	115 46 00
B	KR880 GN	Deadwood	Cretaceous Idaho batholith	Epithermal	Vein	Base- and precious-metals	Dixie	Elmore	44 28 20	115 34 25
A	76T 23M GN	Dollarhide	Pennsylvanian Dollarhide Formation	Epithermal	Vein		Carrietown	Camas	43 36 13	114 40 58
A	SWH154-76a GN	Dryden	Ordovician Saturday Mount Formation	Epithermal	Vein	Pb, Ag, Zn, Sb, Sn	Bayhorse	Custer	44 14 43	114 25 56
A	SWH154-76b GN	Dryden	Ordovician Saturday Mount Formation	Epithermal	Vein	Pb, Ag, Zn, Sb, Sn	Bayhorse	Custer	44 14 43	114 25 56

B	76T 32M GN	Emperium	Pennsylvanian-Permian Wood River Formation, limy sandstone	Epithermal	Vein?		Germania Basin	Custer	43 59 32	114 38 34
B	R-366 OR	Empire	Paleozoic Salmon River assemblage, carbonaceous argillite	Epithermal			Washington Basin	Custer	44 00 31	114 39 13
B	WH70-13B GN	Eureka	Lower Permian upper member of the Dollarhide Formation	Epithermal			Mineral Hill	Blaine	43 29 58	114 24 27
A	76T 31 GN	GAF, Big Smokey Creek	Tertiary Idaho batholith	Epithermal			Vienna	Blaine	43 46 20	114 42 27
B	79T580M GN	Golden Glow	Pennsylvanian-Permian Wood River Formation, limy sandstone	Epithermal			Boulder Basin	Blaine	43 50 24	114 30 24
A	GHM-1 GN	Greyhound	Cretaceous Idaho batholith	Epithermal	Vein	Base- and precious-metals	Seafoam	Custer	44 35 01	115 07 28
A	80R-1 GN?	Hilltop	Devonian Jefferson Formation	Epithermal			Gilmore (Texas)	Lemhi	44 28 49	113 19 08
A	80R-1 GN?	Hilltop	Devonian Jefferson Formation	Epithermal			Gilmore (Texas)	Lemhi	44 28 49	113 19 08
B	S-16 GN	Hoodoo	Paleozoic Salmon River assemblage, argillite and limestone	Syngenetic-epithermal	Bed-vein	Pb, Ag, Zn, Sb, Sn	Boulder Creek-Slate	Custer	44 10 11	114 38 25
B	R-13 BA	Hoodoo	Paleozoic Salmon River assemblage, argillite	Syngenetic	Bed-vein	Pb, Ag, Zn, Sb, Sn	Boulder Creek-Slate	Custer	44 10 11	114 38 25
A	T 81M GN	Horn Silver	Pennsylvanian Dollarhide Formation	Epithermal			Carrietown	Camas	43 35 22	114 43 24
A	R-151 GN	K.G.	Cretaceous Idaho batholith	Epithermal			Seafoam-Deadwood	Valley	44 33 32	115 20 20
B	A176 GN	Liberty Gem	Pennsylvanian Dollarhide Formation	Epithermal			Mineral Hill	Blaine	43 27 15	114 26 14
B	80T-2 GN	Livingston	Paleozoic Salmon River assemblage, carbonaceous argillite	Syngenetic remobilized	Bed-vein	Zn, Pb, Ag, Sb, Sn	Boulder	Custer	44 08 16	114 35 12
B	80T-1 GN-JA	Livingston	Paleozoic Salmon River assemblage, carbonaceous limestone	Syngenetic remobilized	Bed-vein	Zn, Pb, Ag, Sb, Sn	Boulder	Custer	44 08 16	114 35 12
A	R-154 GN	Mountain King	Paleozoic limestone in Idaho batholith	Epithermal	Base- and precious-metal		Sheep Mountain	Custer	44 36 17	115 01 45

Table 3. Sample information, Challis and Hailey 1°×2° quadrangles, Idaho—Continued.

Group	Sample number	Name or location	Age and host rock type	Relative age	Deposit type	Commodities	District	County	Latitude	Longitude
				Ore samples—Continued						
A	K52-1 GN	Mountain King	Paleozoic limestone in Idaho batholith	Epithermal	Base- and precious-metal		Sheep Mtn.–Vienna	Custer	44 36 18	115 01 42
B	WH74-2B GN	North Star	Devonian Milligen Formation(?)	Syngenetic?-epithermal	Bed-vein		Mineral Hill	Blaine	43 39 01	114 16 47
A	D6S3 GN?	Pacific Mountain	Ordovician Bayhorse Dolomite	Epithermal			Bayhorse	Custer	44 24 41	114 19 23
A	KD74-P9-5B GN	Pacific Mountain	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Zn, Ag, F	Bayhorse	Custer	44 24 41	114 19 23
A	KD74-P9-5B GN	Pacific Mountain	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Zn, Ag, F	Bayhorse	Custer	44 24 41	114 19 23
A	KD72-9-124 GN	Pacific Mountain, above main ore	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Zn, Ag, F	Bayhorse	Custer	44 24 41	114 19 23
A	KD68 GN	Pacific Mountain, dump	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Zn, Ag, F	Bayhorse	Custer	44 24 41	114 19 23
A	KD69 GN	Pacific Mountain, main	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Zn, Ag, F	Bayhorse	Custer	44 24 41	114 19 23
A	KD73-4-161 GN	Pacific Mountain, top, main ore	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Zn, Ag, F	Bayhorse	Custer	44 24 41	114 19 23
A	GN?	Pacific ?	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Zn, Ag, F	Bayhorse	Custer	44 24 41	114 19 23
B	T576M BO-JA	Patti Flynn	Paleozoic Salmon River assemblage, argillite	Epithermal	Tin vein		Fourth of July	Custer	44 04 22	114 40 04
A	79T 23M GN	Ramshorn	Ordovician Ramshorn Slate	Epithermal	Vein	Galena, cuprite, tetrahedrite, siderite	Bayhorse	Custer	44 24 36	114 21 42
B	SWH 52-67 GN	Rattlesnake Prospect	Ordovician Bayhorse Dolomite	Epithermal			Bayhorse	Custer	44 20 50	114 19 53
A	1885 Ag GN?	Riverview	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Ag, Zn	Bayhorse	Custer	44 21 56	114 18 22
A	KD-7 GN	Riverview	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Ag, Zn	Bayhorse	Custer	44 21 56	114 18 22
A	GN?	Riverview ?	Ordovician Bayhorse Dolomite	Epithermal	Replacement	Pb, Ag, Zn	Bayhorse	Custer	44 21 56	114 18 22
A?	SWH120-79 GN	Rob Roy	Ordovician Ella Dolomite	Epithermal	Vein	Pb, Ag, Zn, Sb, Sn	Bayhorse	Custer	44 17 56	114 25 05

A	F154 GN	Seafoam Lake, West	Cretaceous Idaho batholith	Epithermal			Seafoam	Custer	44 30 30	115 08 00
A	R-155 GN	Silver Bell	Cretaceous Idaho batholith	Epithermal			Seafoam	Custer	44 32 17	115 08 13
B	WH70-16 GN	Silver Star Queen	Devonian Milligen Formation	Epithermal			Mineral Hill	Blaine	43 27 58	114 17 16
A	SR-1 GN	Sunrise	Cretaceous Idaho batholith	Epithermal	Base- and precious-metal		Yankee Fork	Custer	44 34 11	114 53 48
B'	T572M JA	Timberline	Paleozoic Salmon River assemblage, carbonaceous argillite	Epithermal	Tin vein	Pb, Ag, Zn, Sb, Sn	Fourth of July	Custer	44 03 24	114 39 12
B	SWH GN	Twin Apex	Paleozoic Salmon River assemblage, carbonaceous shale	Epithermal			Bayhorse	Custer	44 18 26	114 30 11
A	T302M GN	Webfoot	Cretaceous Idaho batholith	Epithermal			Vienna	Blaine	43 48 13	114 50 30
B	W312A GN	White Cloud-East	Pennsylvanian-Permian Wood River Formation, limy sandstone	Epithermal	Porphyry-vein	Mo, Pb, Ag, Zn, Sn	East Fork-Lower Boulder	Custer	44 02 40	114 32 40

Group	Sample number	Name or location	Age and host rock type	Association	Deposit type	Commodities	District	County	Latitude	Longitude
				Granite and rhyolite						
B	KR 56T WR	Bear River, North	Tertiary granite	(Sn)	Tin granite		East Fork?	Custer	44 00 55	115 21 29
B	KR842T WR	Big Silver Creek	Tertiary granite	(Sn)	Tin granite		Gambrinus	Boise	43 59 03	115 19 18
A	KR493C WR	Camas Creek	Tertiary granite	(Sn)	Tin granite		Gravel Range	Lemhi	44 50 25	114 30 35
B	KR743 KF	Deadwood	Cretaceous Idaho batholith	Zn-Pb-Ag	Vein	Base- and precious-metals	Dixie	Elmore	44 28 15	115 34 55
B	LF-1 WR	Little Falls	Tertiary Challis Volcanic Group, rhyolite	Mo	Porphyry	Mo	Bear Creek	Boise	44 04 22	115 46 12
A	LP-1 WR	Lost Packer	Cretaceous granite	Cu-Zn-(Mo)	Porphyry	Mo, base- and precious-metals	Loon Creek	Custer	44 35 55	114 52 55
B	A230 KF	McCoy	Cretaceous granite	Au-Ag-Pb-Zn			Mineral Hill	Blaine	43 27 33	114 21 28
A	Grey-1 WR	Mt. Greylock	Tertiary Challis Volcanic Group, rhyolite	Au-Ag-(Mo)	Porphyry	Mo	Yankee Fork	Custer	44 25 10	114 41 49
A	GM-31 WR	Parker Mountain	Tertiary Challis Volcanic Group, rhyolite	Au-Ag-(Mo)	Porphyry	Mo, precious metals	Parker Mountain	Lemhi	44 35 30	114 32 55
A	RH-122 WR	Red Mountain Dome	Tertiary Challis Volcanic Group, rhyolite	Au-Ag-(Mo)	Porphyry	Mo, precious metal	Yankee Fork	Custer	44 24 20	114 51 42
A	GM-365 WR	Singheiser	Tertiary Challis Volcanic Group, rhyolite	Au-Ag-(Mo)	Porphyry	Mo, precious metals	Gravel Range	Lemhi	44 51 31	114 23 17

Table 3. Sample information, Challis and Hailey 1°×2° quadrangles, Idaho—Continued.

Group	Sample number	Name or location	Age and host rock type	Association	Deposit type	Commodities	District	County	Latitude	Longitude
				Granite and rhyolite						
A	KR804T WR	Stanley Lake Creek	Tertiary granite	(Sn)	Tin granite	Sn	Seafoam	Custer	44 43 52	115 06 08
A	SB215120 WR	Sunbeam	Tertiary Challis Volcanic Group, rhyolite	Au-Ag-(Mo)	Porphyry	Mo	Yankee Fork	Custer	44 26 12	114 44 00
B	KR529T WR	Wolf Mountain	Tertiary granite	(Sn)	Tin granite		Big Creek	Boise	43 59 14	115 24 11
A	KR803T WR	Warm Springs Creek	Tertiary granite	(Sn)	Tin granite		Stanley?	Boise	44 18 30	115 17 50
B	W188 KF	White Cloud-West	Cretaceous granite porphyry	Mo	Porphyry	Mo	East Fork	Custer	44 03 00	114 35 30
A	W320 KF	Thompson Creek	Cretaceous granite	Mo	Porphyry	Mo	Bayhorse	Custer	44 19 00	114 32 15
				Sedimentary rocks						
B	WH2112 WR	"Sawmill Gulch"	Devonian Milligen Formation, argillite	Ag-Pb-Zn			Warm Springs	Blaine	43 40 14	114 24 50
B	T-270 WR	Holman Creek-West	Paleozoic Salmon River assemblage, argillite	Ag-Pb-Zn	Porphyry?	Mo?	Bayhorse	Custer	44 15 13	114 32 27
B	T-602 WR	Urexco	Mississippian Copper Basin Formation, argillite	U			Alto	Custer	43 56 47	114 14 20

end of the assemblage and are similar in lead-isotopic composition to the potassium feldspar from the White Cloud stock. The observation of Stacey and others (1968), Doe and Stacey (1974), and Doe (1979) that the largest magmatothermal deposits are most similar to igneous rocks in lead-isotopic composition may also apply to the Great Basin environment.

Ores from the largest Cretaceous and Tertiary magmatothermal gold and silver deposits in the Rocky Mountain region have values of $^{206}Pb/^{204}Pb$ of 19.1 or less (Doe, 1979). The largest production for these kinds of deposits is from the Sunnyside gold mine ($^{206}Pb/^{204}Pb$ 18.2 –18.6) and the Creede silver mine ($^{206}Pb/^{204}Pb$ 18.8–19.1) of the San Juan Mountains of Colorado (Doe and others, 1979) and the Clayton Silver mine of central Idaho ($^{206}Pb/^{204}Pb$ 18.9–19.0) (table 4). There is some indication (Doe, 1979), however, that these deposits have more radiogenic values of $^{206}Pb/^{204}Pb$ than ores from the largest Cretaceous and Tertiary magmatothermal base-metal deposits, such as Butte, Montana, Bingham Canyon, Utah, and Climax, Colorado, which have values of $^{206}Pb/^{204}Pb$ of less than 18.1 (table 2). Although at one time the Sunnyside mine was the second largest gold producer in the United States (after the Precambrian deposit of the Homestake mine, South Dakota) and the Clayton Silver mine was the largest silver producer in Idaho (after the Precambrian deposit of the Coeur d'Alene mine), the dollar production of both the Sunnyside mine and the Clayton Silver mine is modest in comparison to the largest base-metal mines and gold or silver mines. Although the new data from the Clayton Silver mine validates the previous suggestion that Cretaceous and Tertiary magmatothermal gold and silver deposits tend to have more radiogenic values of $^{206}Pb/^{204}Pb$ than the largest Cretaceous and Tertiary magmatothermal base-metal deposits, data from gold and silver deposits of equivalent production to the largest base-metal deposits are not available for comparison. Therefore, the suggestion remains tentative that, among Cretaceous and Tertiary magmatothermal deposits in the Rocky Mountain area, gold and silver deposits will have somewhat more radiogenic values of $^{206}Pb/^{204}Pb$ than the largest base-metal deposits.

Tin is currently classified as a strategic mineral but has little production in the United States. Finding a good tin deposit is complicated by the requirement that the tin be in the form of cassiterite because of smelting considerations. Although some cassiterite has been found in the study area, stannite is more abundant (Hall, 1985, 1987a). The best occurrence of tin is at the Timberline prospect, but stannite also is present at the Patti Flynn prospect and elsewhere. Both Timberline ($^{206}Pb/^{204}Pb$ of 20.10) and Patti Flynn ($^{206}Pb/^{204}Pb$ of 19.33), which are vein deposits in black shale, are in group B. The Rob Roy mine has the greatest value of $^{206}Pb/^{204}Pb$ but is of uncertain isotopic classification. In contrast, samples of tin granite (all but one having values of $^{206}Pb/^{204}Pb$ less than 19.0) are in either isotopic group and are isotopically indistinguishable from molybdenum porphyry. No granite has the lead-isotope composition of the lead in the tin veins. The lead in the tin veins most likely was derived from a Precambrian source at depth and a lower Paleozoic source in the upper crust (Sanford and Wooden, this volume). Thus, the relationship between tin granite and tin-bearing veins is not simple.

SULFUR ISOTOPES

No economic subaerial magmatothermal deposit previously was known that has $\delta^{34}S$ values greater than +6 per mil (the extreme value for the Henderson molybdenum porphyry of Colorado; Stein, 1985). The "normal" igneous range is –5 to +3 per mil, as discussed above. Central Idaho, however, is an astounding exception to this previous observation (table 4) (see also Howe and Hall, 1985). Of 25 samples from different mines, 17 have $\delta^{34}S$ greater than +3 per mil. Indeed, the very biggest mineral deposits of Thompson Creek and White Cloud have exceptionally heavy sulfur isotopes (+8 per mil or more and +4 per mil or more, respectively). Some deposits have galenas of the appropriate $\delta^{34}S$ values for the igneous range, but they are not among the best producers. Among prospects having values of $^{206}Pb/^{204}Pb$ of 18.0 or less, only Webfoot ($\delta^{34}S$ of +3.6) is near the normal igneous range for sulfur isotopes. For $^{206}Pb/^{204}Pb$ values between 18.0 and 19.1, galenas from the Carrie Leonard mine, GAF prospect, Horn Silver mine, and perhaps the Dryden and Comeback mines are in or near the igneous range for $\delta^{34}S$. Among radiogenic group B deposits, only the mine at Crest Lake–Travel Creek is in the normal igneous range for $\delta^{34}S$.

Unlike all other areas for which data are available, the sulfur isotope rule in the Challis and Hailey quadrangles may be that the biggest deposits contain heavy sulfur isotopes. If this is true, then the areas near the prospect at Seafoam Lake ($^{206}Pb/^{204}Pb$=17.75, $\delta^{34}S$=+7.0) and near the Ramshorn mine ($^{206}Pb/^{204}Pb$=17.75, $\delta^{34}S$=+10.3) would have exceptional potential.

DISCUSSION

NATURE OF BASEMENT SOURCE LEAD

Deposits of group A such as the Ramshorn mine have lead-isotope signatures representative of metamorphosed Proterozoic crystalline rocks from the Rocky Mountains. Such an isotopic signature is derived either directly from Precambrian crystalline rocks at depth or indirectly from Mesozoic or Cenozoic igneous intrusions that derived their lead from such a source. Deposits of group A are associated with structural highs in Precambrian basement or with Cretaceous intrusive complexes. The two main examples are the Bayhorse anticline and the Casto structure. The Ramshorn, Pacific, Riverview, Rattlesnake Creek, Rob Roy, Clayton Silver, Dryden, Thompson Creek, and Twin Apex mines are associated with the Bayhorse anticline. The Seafoam Lake, KG, Silver Bell, Mountain King, Sunrise, Deadwood, Sunbeam, Parker Mountain, Greyhound, Lost Packer, and Singheiser mines are associated with the Casto structure.

The Ramshorn mine is in the Bayhorse anticline and is within 1 km of the exposed Cretaceous granodiorite and quartz monzonite of Juliette Creek (Hobbs and others, 1975; McIntyre and others, 1976; Hobbs, 1985) to the south. Although the exposures are less than 2.5 km^2 in areal extent, they are within the Bayhorse gravity high and adjacent to a magnetic anomaly that has closure at 2,000 gammas (relative). Closure at 1,800 gammas (relative) is shaped rather like an Idaho potato (fig. 8), about 16 km long and 7.25 km wide. Webring and Mabey (1987, p. 78) (also see Mabey and Webring, 1985) said of this feature:

> The gravity high may reflect a topographic high on the Precambrian crystalline basement or an underlying intrusive basement *** The rocks in the Bayhorse anticline may be an autochthon or an allochthon overlying a basement high. The other possibility, that a large intrusive complex may underlie the Bayhorse high, is suggested by the extensive magnetic high apparently at least in part produced by the 98 m.y. Juliette stock. McIntyre and others (1976) consider this stock to be a satellite of the Idaho batholith, but the large mass of rock underlying the Bayhorse high is both more dense and more strongly magnetized than the main mass of the Idaho batholith. The gabbro dikes and sills [that occur in the Bayhorse high] suggest that a mafic intrusive may underlie the area.

The intrusion may be a thin, batholith-sized body, and the Ramshorn mine may be within 1.5 km of the cupola. A number of other deposits may overlie deeper parts of the igneous mass or be adjacent to it. These include deposits of the Pacific and Riverview mines to the east, Rattlesnake Creek mine in the center, and Rob Roy, Clayton Silver, and Dryden mines to the southwest.

The Bayhorse gravity anomaly has a lobe that extends considerably to the west of the Bayhorse magnetic anomaly and includes the area of the Thompson Creek molybdenum porphyry and Twin Apex mine. The area of the gravity anomaly, which possibly represents elevated basement, is one of the areas containing nonradiogenic group A lead and heavy sulfur isotopes. Although some deposits in this area are in group A, the rest have values of $^{206}Pb/^{204}Pb$ greater than 18.0. Deposits in group A include the Dryden mine, Clayton Silver mine, Rattlesnake River prospect, and Pacific Mountain mine. Potassium feldspar from the Thompson Creek stock is also in this radiogenic assemblage. Some deposits, however, are in group B, including the Riverview and Twin Apex mines. The Rob Roy mine may be a mixture of group A and group B leads.

The Casto structure, as used here, includes the Casto pluton and associated gravity and magnetic highs. In the area of the Casto structure, all deposits except the Deadwood mine are in group A (Seafoam Lake prospect, K.G. prospect, Silver Bell prospect, Mountain King mine, Sunrise prospect). Also in group A are, from southwest to northeast, the Eocene tin

Table 4. Lead and sulfur isotopic compositions, Challis and Hailey 1°×2° quadrangles, Idaho.
[Sample information is given in table 3. Two-letter designation following sample number indicates sample type: GN, galena; BO, boulangerite; BA, barite; JA, jamesonite; KF, potassium feldspar; OR, ore; WR, whole rock. Analysts: 1, M.H. Delevaux; 2, Å. Johansson; 3, S.A. Kish; 4, A.P. LeHuray; 5, H.J. Stein; 6, K.E. Davis; 7, Geochron Laboratory; 8, S.S. Howe]

Sample no.	Mine or location	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Analyst	$\delta^{34}S$	Analyst
			Ore samples				
Buck? GN	Thompson Creek-Buckskin Drill Road	17.577	15.583	40.875	3	8.7	8
S76–895 GN	Thompson Creek-Buckskin Drill Road?	17.617	15.594	41.014	3	8.7	8
T 49M GN	Bulldozer cut	19.692	15.808	39.383	3	5.0	7
F112 GN	Carlson Gulch	19.144	15.672	39.087	2	7.1, 7.1	7
T 68M GN	Carrie Leonard	18.094	15.591	39.975	3	-0.1	7
SWH 78–65 GN	Clayton Silver	18.965	15.718	39.948	2	–	–
80WH–1 GN	Clayton Silver	18.944	15.709	39.990	4	–	–
79T565M GN	Combination?-new	19.735	15.800	39.367	3	7.6, 7.9	7
F107 GN	Comeback	19.097	15.678	39.083	2	0.8, 0.8	7, 8
T597M GN	Crest Lake?	19.952	15.797	39.619	3	0.8	7
79T 5M GN	Crown Point	19.624	15.800	39.274	2	4.5	7
584–13 GN	CUMO	19.039	15.656	38.965	5	–	–
KR880 GN	Deadwood	19.341	15.754	39.527	1	–	–
76T 23M GN	Dollarhide	18.081	15.572	39.918	2	-0.4	
SWH154–76a GN	Dryden	18.586	15.706	40.608	2	4.6	7
SWH154–76b GN	Dryden	18.584	15.690	40.549	2	5.4	
76T 32M GN	Emperium	19.785	15.788	39.311	2	–	–
R–366 OR	Empire	19.831	15.673	39.111	3	–	–
WH70–13B GN	Eureka	20.220	15.910	40.904	1	6.3	7
76T 31 GN	GAF, Big Smokey Creek	18.311	15.622	39.883	2	0.2	7
79T580M GN	Golden Glow	19.767	15.844	39.677	3	6.8	7
GHM–1 GN	Greyhound	18.110	15.646	40.565	5	–	–
80R–1 GN?	Hilltop	18.376	15.573	38.955	1	–	–
80R–1 GN?	Hilltop	18.407	15.606	39.039	4	–	–
S–16 GN	Hoodoo	19.502	15.859	39.177	3	–	–
R–13 BA	Hoodoo	19.810	15.795	39.192	3	–	–
T 81M GN	Horn Silver	18.441	15.652	39.735	2	1.2	7
R–151 GN	K.G.	17.990	15.578	39.286	1	–	–
A176 GN	Liberty	20.013	15.917	40.619	1	2.2	
80T–2 GN	Livingston	19.779	15.786	39.499	2	11.1	8
80T–1 GN, JA	Livingston	20.030	15.833	39.673	3	–	–
R–154 GN	Mountain King	17.943	15.606	40.299	1	–	–
K52–1 GN	Mountain King	17.810	15.593	40.290	5	–	–
WH74–2B GN	North Star	19.688	15.906	39.438	1	7.2	
D6S3 GN?	Pacific	21.452	15.840	41.145	1	10.1	8
KD74–P9–5B GN	Pacific Mountain	21.355	15.777	40.897	6	–	–
KD74–P9–5B GN	Pacific Mountain	21.375	15.807	41.004	6	–	–
KD72–9–124 GN	Pacific Mountain above	21.323	15.765	40.796	6	–	–
KD68 GN	Pacific Mountain, dump	21.332	15.771	40.948	6	–	–
KD69 GN	Pacific Mountain, main	21.431	15.767	40.899	6	–	–
KD73–4–161 GN	Pacific Mountain, top	21.347	15.801	40.966	6	–	–
T576M BO, A	Patti Flynn	19.333	15.809	39.026	3	–	–
79T 23M GN	Ramshorn	17.776	15.578	39.861	3	10.1, 10.1	7, 8
SWH 52–67 GN	Rattlesnake Peak	20.479	15.784	40.874	2	4.2	7
1885 Ag GN?	Riverview	21.073	15.804	40.808	1	–	–

Table 4. Lead and sulfur isotopic compositions, Challis and Hailey 1°x2° quadrangles, Idaho—Continued.
[Sample information is given in table 3. Two-letter designation following sample number indicates sample type: GN, galena; BO, boulangerite; BA, barite; JA, jamesonite; KF, potassium feldspar; OR, ore; WR, whole rock. Analysts: 1, M.H. Delevaux; 2, Å. Johansson; 3, S.A. Kish; 4, A.P. LeHuray; 5, H.J. Stein; 6, K.E. Davis; 7, Geochron Laboratory; 8, S.S. Howe]

Sample no.	Mine or location	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Analyst	$\delta^{34}S$	Analyst
	Ore samples—Continued						
KD–7 GN	Riverview	19.704	15.806	39.281	6	–	–
SWH120–79 GN	Rob Roy	20.474	15.870	40.143	2	6.6	8
F154 GN	Seafoam Lake, West	17.747	15.604	39.765	2	7.0	7
R–155 GN	Silver Bell	17.596	15.574	39.785	1	–	–
WH70–16	Silver Star Queen	20.421	15.962	41.301	1	10.7	
SR–1 GN	Sunrise	17.473	15.562	40.348	5	–	–
T572M JA	Timberline	20.099	15.772	39.460	3	–	–
SWH GN	Twin Apex	19.357	15.753	39.341	2	–	–
T302M GN	Webfoot	17.785	15.587	39.895	3	4.2	7
W312A GN	White Cloud, East	19.404	15.732	39.358	1	4.0	
	Granite and rhyolite						
KR 56T WR	Bear River, North	18.842	15.637	39.094	1	–	–
KR842T WR	Big Silver Creek	18.826	15.632	39.108	1	–	–
KR493C WR	Camas Creek	18.762	15.645	39.755	1	–	–
KR743 WR	Deadwood	19.488	15.716	39.320	1	–	–
LF–1 WR	Little Falls	19.149	15.653	39.186	1	–	–
LP–1 WR	Lost Packer	19.768	15.742	40.078	1	–	–
A230 KF	McCoy (Croesus stock)	19.572	15.762	39.309	1	–	–
Grey–1 WR	Mt. Greylock	17.536	15.525	39.272	1	–	–
GM–31 WR	Parker Mountain	17.869	15.526	39.634	1	–	–
RH–122 WR	Red Mountain Dome	17.493	15.522	39.459	1	–	–
GM–365 WR	Singheiser	18.885	15.645	39.818	1	–	–
KR804T WR	Stanley Lake Creek	17.551	15.549	39.330	1	–	–
SB215120 WR	Sunbeam	17.803	15.529	39.497	1	–	–
KR529T WR	Wolf Mountain	18.637	15.612	39.055	1	–	–
KR803T WR	Warm Springs Creek	17.580	15.552	39.219	1	–	–
W188 KF	White Cloud, West	19.204	15.694	39.262	1	–	–
W320 KF	Thompson Creek	20.508	15.811	40.329	1	–	–
	Sedimentary rocks						
WH2112 WR	"Sawmill Gulch"	19.308	15.673	39.526	1	–	–
T–270 WR	Holman Creek, West	24.073	15.954	39.145	1	–	–
T–602 WR	Urexco	29.384	16.238	38.881	1	–	–

granites of Stanley Lake Creek and Warm Springs Creek, the molybdenum prospects in rhyolitic intrusive rocks of volcanic assemblages of Red Mountain dome, Mt. Greylock barren dome, and Sunbeam mine, and the Eocene rhyolite of the Parker Mountain mine. Samples in group A that have values of $^{206}Pb/^{204}Pb$ of 18.0 or more are galena from the Greyhound mine, igneous rocks from the granite of the Lost Packer mine, which have a base-metal ore association, tin granite of Camas Creek in the Casto pluton geophysical anomaly of both magnetic and gravity highs, and Singheiser mine molyrhyolite (table 3), to the east of which Precambrian rock crops out. The Casto structure contains major geologic and structural complexities that defy simple synthesis, although the western part of the area is one generally of positive residual gravity anomaly (Mabey and Webring, 1985) that may reflect Precambrian crystalline rocks at some depth. D.R. Mabey (written commun., February 1980) said about the northeastern part of this region, "A large area of relative high magnetic intensity and high Bouguer gravity values lies between the Thunder Mountain Caldera and the Twin Peaks area. The gravity high may reflect a general high in the Precambrian basement. The magnetic anomalies appear to reflect a combination of basement, intrusive, and volcanic rocks."

Alternatively, Webring and Mabey (1987) interpreted the Casto structure as being underlain by units of the Casto pluton that are more dense than the Idaho batholith because Precambrian rocks are not known at the surface. In this case, the higher density indicates a more mafic composition. Thus, as with the Bayhorse geophysical anomaly, the Casto anomaly is permissive of some combination of more mafic

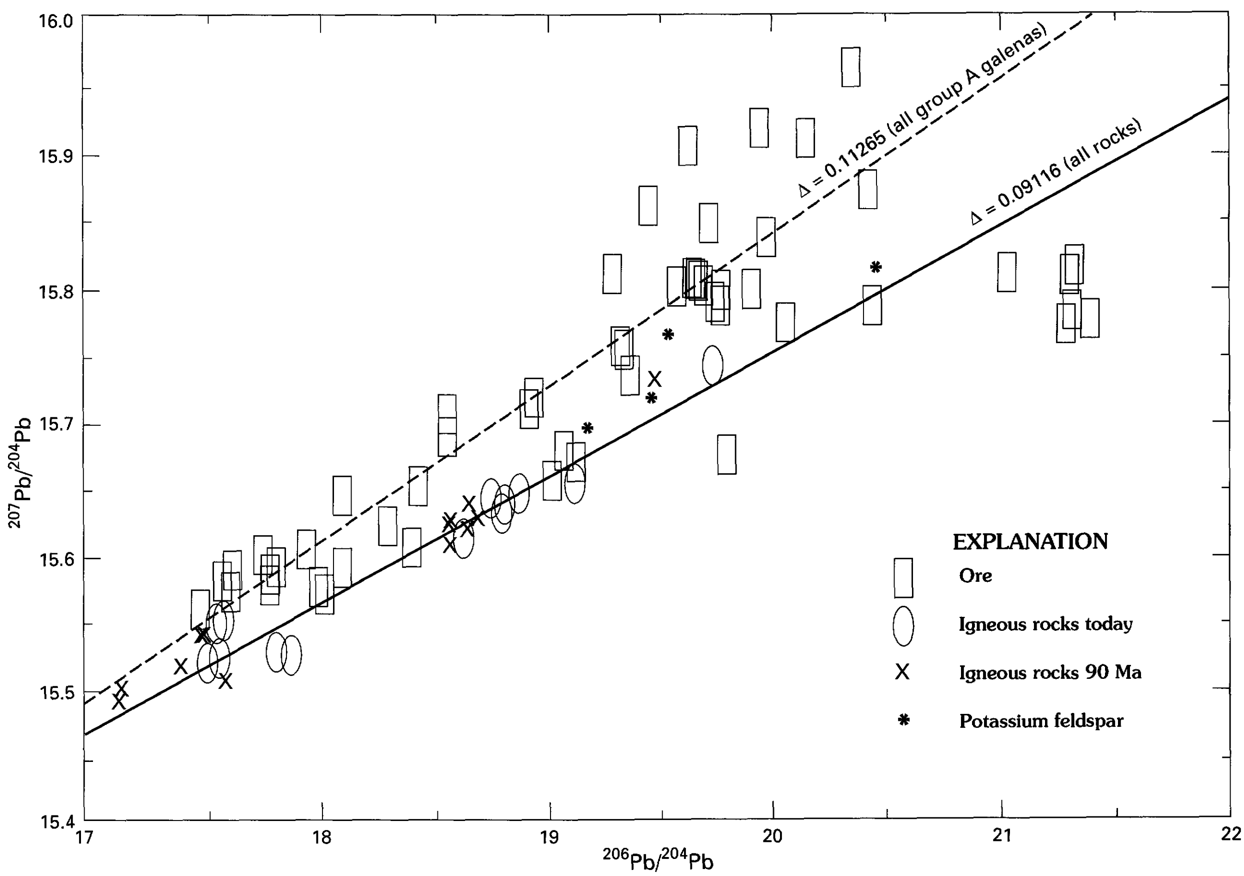


Figure 5. $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ for ore minerals and igneous rocks of the Challis and Hailey 1°×2° quadrangles, Idaho. Symbol sizes are approximately two standard deviations. Solid and dashed sloping lines are regressions as labeled.

intrusive rocks and a general high in the Precambrian crystalline basement. Areas of gravity lows can be attributed to near-surface, less dense volcanic rocks (for example, the Twin Peaks caldera) or granitic intrusive rocks (for example, the Soldier Lakes magnetic high) that can make positive gravity anomalies at depth.

Thus, the two areas of samples of group A, the Bayhorse anticline and Casto structures, include major geophysically defined areas of positive gravity anomalies that possibly indicate structural highs in the crystalline Precambrian basement. Almost no samples of group B are in the area of the Casto structure.

Although the igneous-rock and ore samples have, in general, similar values of $^{206}Pb/^{204}Pb$, the ore samples generally have values of $^{208}Pb/^{204}Pb$ greater than the igneous rocks (except for the K.G. deposit). Corrections to $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ in the igneous rocks for in situ decay of uranium and thorium since the igneous rocks formed 90–45 Ma, though small, will only increase the differences in $^{208}Pb/^{204}Pb$ between ore and igneous rock. Thus, derivation of some ore lead from outside the stocks is likely. The geographic distribution of ore and igneous rocks sampled is similar; however, the sampling scheme did not involve possibly related igneous rock–ore pairs in group A (except perhaps for Thompson Creek, the pair that has a major difference in isotopic composition). It may be, therefore, that stocks exist that have values of $^{208}Pb/^{204}Pb$ greater for a given value of $^{206}Pb/^{204}Pb$ than we found during this study. Perhaps more likely is the existence at depth of high-rank metamorphic Precambrian basement. High values of Th/U are common in granulite facies metamorphic rocks (Doe and Zartman, 1979).

The two major areas studied for which samples are exclusively in group B are along the southern boundary of the Challis 1°×2° quadrangle and are associated with magnetic anomalies without gravity anomalies (D.R. Mabey, written commun., February 1980; also see Mabey and Webring, 1987). The White Cloud Peaks–Little Boulder Creek area includes the White Cloud Peaks magnetic high (granite) and Little Boulder Creek magnetic low (volcanic rocks), the White Cloud stockwork molybdenum prospect, the Patti Flynn and Timberline prospects, the Empire vein, and the Livingston and Hoodoo mines and adjacent magnetic anomalies to the north. The Grimes area includes the magnetic anomaly of Grimes, the area of the CUMO and Comeback deposits, Carlson Gulch adit, and the prospect of the Little Falls intrusion. The magnetic anomaly may represent

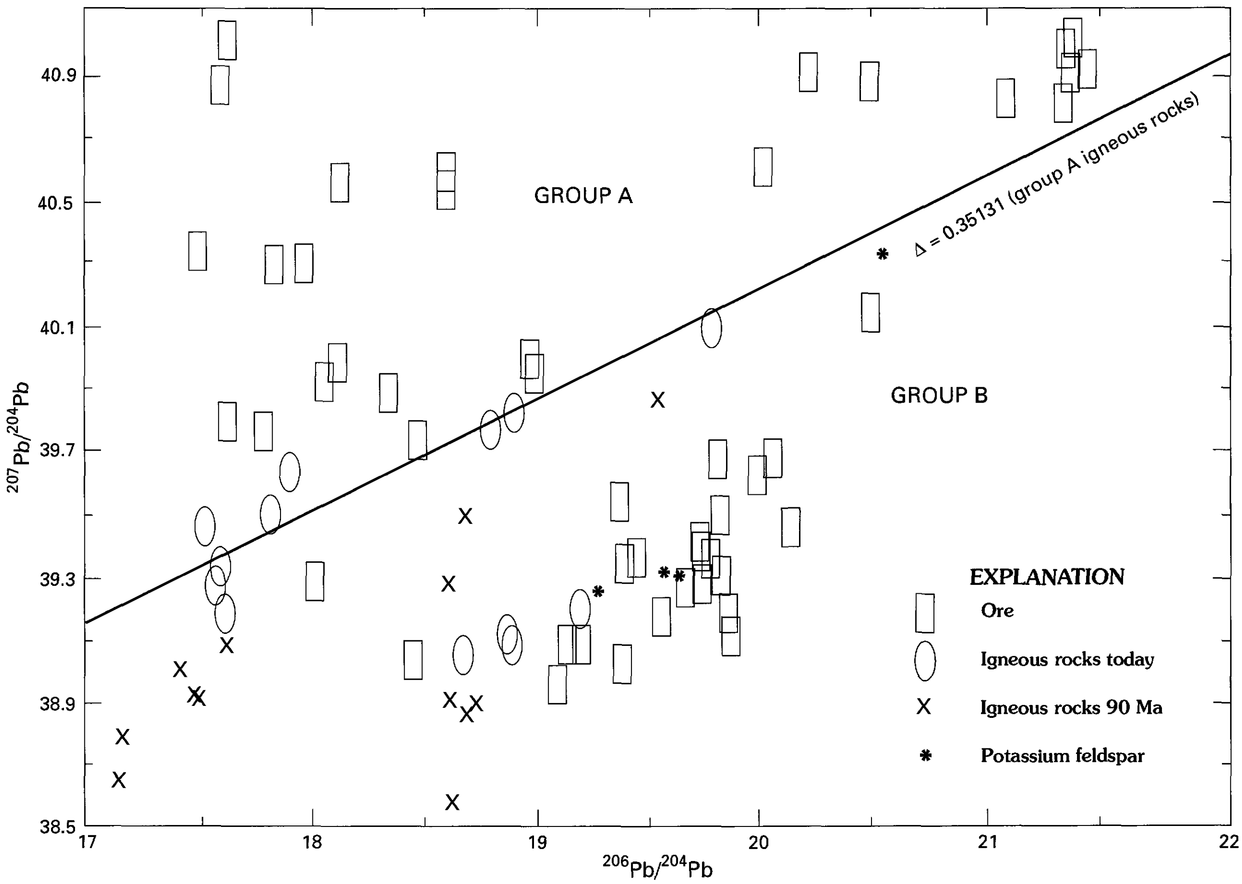


Figure 6. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ for ore minerals and igneous rocks of the Challis and Hailey 1°×2° quadrangles, Idaho. Symbol sizes are approximately two standard deviations. Solid line is regression as labeled.

a large diorite stock underlying the area, with some surface exposure. The Deadwood ore–igneous rock pair is the most northerly location of group B. Because it is associated with gravity and magnetic anomalies and a large Eocene diorite stock, it may be related to the Grimes deposits, which are associated with a magnetic anomaly and an Eocene dike swarm. The intrusion of Bear River on the southern boundary of the Challis quadrangle is also on a slight magnetic anomaly within the Sawtooth magnetic anomaly without a gravity anomaly and includes the tin granites of Wolf Mountain and Big Silver Creek just south of the boundary with the Hailey 1°×2° quadrangle.

The lack of gravity anomalies associated with these group B deposits suggests that Precambrian crystalline basement, if present at all, is deeper than in group A areas such that the isotopic signatures of the group B magmas were imprinted on both igneous rocks and veins above any Precambrian basement.

Although we do not have equivalent geophysical information for the Hailey quadrangle, one might speculate that the isotopic signatures from the area including the Webfoot mine to the Carrie Leonard mine are from Precambrian crystalline basement at unknown depth. Deposits to the east of these deposits again have group B isotopic signatures of Basin and Range rocks. Precambrian crystalline basement, if present at all, must be deeper than the horizon where the isotopic signatures were acquired.

The two areas of group B deposits and rocks along the boundary between the Challis and Hailey 1°×2° quadrangles suggest that a strip of Great Basin (group B) lead in igneous rocks and ores is present along the length of the boundary. More data are needed to fill in the gaps.

CONSIDERATION OF OTHER LEAD SOURCES

No rock has been analyzed from the study area that has $^{208}Pb/^{204}Pb$ values as high as those for the ore minerals of group A. Mineral deposits in the black shale belt of central Idaho are discussed at length by Sanford and others (1989) and Sanford and Wooden (this volume), who conclude that

Table 5. U, Th, and Pb concentrations and calculated initial lead ratios for igneous and sedimentary rocks, Challis and Hailey 1°×2° quadrangles, Idaho.
[Sample information is given in table 3. U, Th, Pb concentrations are in parts per million]

Group	Name	U	Th	Pb	Age (Ma)	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
	GRANITE AND RHYOLITE							
	Molybdenum stockwork							
A	Sunbeam	6.63	24.2	8.55	90	17.11	15.50	38.66
A	Red Mountain dome	4.45	24.2	10.7	90	17.12	15.50	38.79
A	Mt. Greylock	5.46	27.9	31.5	90	17.38	15.52	39.01
A	Parker Mountain	4.23	24.4	13.0	90	17.58	15.51	39.08
A	Singheiser	3.14	16.3	9.06	90	18.57	15.63	39.28
B	Little Falls	5.12	16.3	8.14	90	18.58	15.63	38.59
B	White Cloud-West[1]	–	–	–	90	19.20	15.69	39.26
B	Lost Packer	1.27	3.11	4.61	90	19.51	15.73	39.87
A	Thompson Creek[1]	–	–	–	90	20.51	15.81	40.33
	Base- and precious-metal veins							
B	Deadwood[1]	–	–	–	90	19.49	15.72	39.32
B	McCoy[1]	–	–	–	90	19.57	15.76	39.31
	Tin granite							
A	Stanley Lake Creek	2.28	24.9	18.0	90	17.44	15.54	38.93
A	Warm Springs Creek	2.43	17.3	16.7	90	17.45	15.55	38.92
B	Wolf Mountain	2.32	14.4	29.4	90	18.57	15.61	38.91
B	Big Silver Creek	3.83	14.7	18.4	90	18.64	15.62	38.87
A	Camas Creek	1.60	12.1	13.6	90	18.65	15.64	39.49
B	Bear River, North	3.44	12.7	20.3	90	18.69	15.63	38.91
	SEDIMENTARY ROCKS							
	"Sawmill Gulch"	1.05	4.41	4.12	90	19.06	15.66	39.20
	Holman Creek-West	91.10	10.90	42.10	500	11.98	15.26	38.68
	Urexco	13.80	2.99	15.00	500	23.98	15.92	28.50

[1]Potassium feldspar.

few, if any, of the vein leads in either group A or B were derived solely from the shale in the section. One might ask whether mineralized samples in either group A or B might contain significant components (for example, lead in detritus) of either Paleozoic sedimentary rocks or rocks of the Belt Supergroup. Three samples of Paleozoic shale or argillite from central Idaho that were analyzed in this study (tables 2, 4) are plotted in figure 9, along with four samples of limestone from the Dillon, Montana, 1°×2° quadrangle (Doe and others, 1986) and one sample of limestone from north-central Nevada (Rye and others, 1974). Although the samples of Paleozoic limestone are not from central Idaho, little difference in isotopic ratios is seen between them and the local shaly rocks. We have no reason to suspect that the lead-isotope data are not reasonably representative of a very large area. The significant point is that, although the overall trend in $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ for the Paleozoic sedimentary rocks is much flatter than that for ore minerals, one sample of shale and two of limestone have lead-isotopic compositions similar to ore lead of group B. Thus a component of lead from sediments in the group B assemblage cannot be ruled out.

Rocks of the Belt Supergroup are exposed in the extreme northern part of the Challis 1°×2° quadrangle. Data for ore minerals of group B are shown in figure 10 together with data for ore minerals from Cretaceous and Tertiary veins in the Wallace 1°×2° quadrangle to the north in the Belt Supergroup. Although there is some overlap in the lead-isotopic compositions for the two groups of ore minerals, samples from the Belt Supergroup have a tendency to be less radiogenic than samples from the Challis and Hailey quadrangles. Any rocks of the Belt Supergroup underlying the Paleozoic section seem to be excluded as possible source rocks.

MODEL FOR LEAD DERIVATION

In the Challis and Hailey 1°×2° quadrangles, there are two large areas of group A lead-isotope assemblages—one in the central and northern parts of the Challis quadrangle and one in the central part of the Hailey quadrangle. These areas are separated by a strip of group B lead-isotope assemblages along the length of the boundary between the two

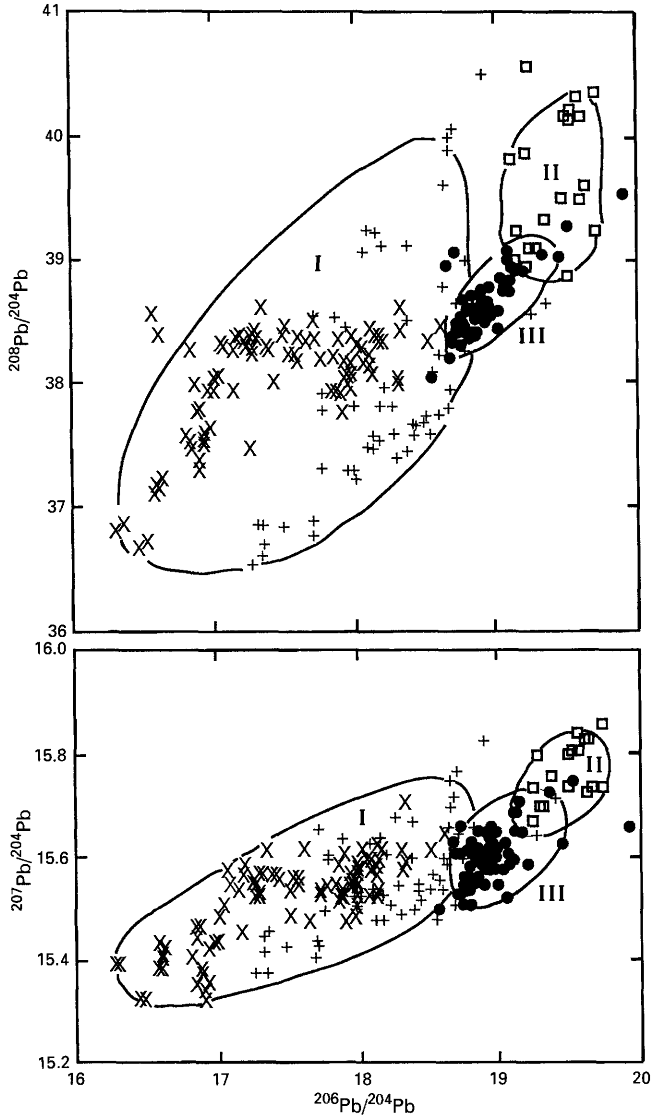


Figure 7. Lead-isotope areas of Zartman (1974). Area I is the area of rejuvenated cratons: Ia (+) includes Archean basement rocks and is the area north of the Colorado-Wyoming State line; Ib (X) includes only Proterozoic basement rocks. Area II (open boxes) is the Basin and Range miogeosynclinal area mostly in Nevada. Area III (solid circles) is the eugeosynclinal continental margin mostly in California, Oregon, and Washington.

quadrangles that perhaps widens to the east. Because the geophysical anomalies tend to have north-south trends, our sampling leaves the possibility of patches of group A and group B leads.

Group A lead is deduced to come from areas where magmas and ore fluids were generated in or interacted with Precambrian crystalline rocks metamorphosed to high rank during the Precambrian. Because exposures of Precambrian crystalline rocks are sparse on the surface and restricted to the far eastern parts of the Challis and Hailey quadrangles, the source of these igneous rocks and ore minerals must be deeper than rocks now observed on the surface. These crystalline source rocks must have high and variable values of Th/U as reflected by high values of $^{208}Pb/^{204}Pb$ relative to the values of $^{206}Pb/^{204}Pb$ in our samples. Because some of the ore minerals have values of $^{208}Pb/^{204}Pb$ greater than any of the igneous rocks, we conclude that the lead of these ore minerals represents a larger component of partially extracted lead from phases having very high values of Th/U (such as monazite) in the Precambrian crystalline basement, whereas lead of the igneous rocks probably more closely represents whole-rock extraction. That is, lead of the mineral deposits may have components derived either from Cretaceous or Tertiary igneous rocks and from partially extracted lead from Precambrian crystalline wallrocks. The possibility remains that appropriate igneous rocks were not sampled.

There is little distinction between lead-isotope ratios of igneous rocks of group B and mineral deposits of group B, although many ore minerals of group B have values of $^{207}Pb/^{204}Pb$ greater than any of the igneous rocks. Some Paleozoic sedimentary rocks at the surface have lead-isotope ratios similar to those for ore and igneous rock samples of group B. Thus, igneous rocks and ores of group B may be derived, at least in part, from sources such as now at the surface. Mineral deposits of group B may be adjacent to igneous rocks of group A, but no mineral deposits of group A are known to be adjacent to igneous rocks of group B, except for the Wood River district (for example, Minnie Moore). These relationships lead us to propose a conceptual model whereby (1) the deep crust is composed of Precambrian crystalline rocks, at least in part, (2) units of the Great Basin are present in patches, particularly along the boundary between the Hailey and Challis quadrangles, to establish the observed lead-isotope signatures in some igneous rocks and ores, either through their generation or through interactions during transport or emplacement, and (3) lead isotopes are mainly derived from beneath the exposed section through interactions of magmas and attendant ore fluids with country rocks.

COMPARISON WITH MAJOR TIN-PRODUCING AREAS

In addition to tin granite in the Hailey and Challis 1°×2° quadrangles, ore deposits in the black shale belt commonly contain stannite and may contain minor cassiterite (Hall, 1985, 1987a). Samples of mineralized rock in the black shale belt can contain more than 1 percent tin[3] (in the Boulder Basin at lat 43°50′ N., long 114°31′ W.) (Hall, 1987a) in a

[3]Earlier analyses by quantitative X-ray fluorescence in Tschanz and others (1974) and Tschanz and Kiilsgaard (1986) gave much higher values for tin, as much as 6 percent. These high values may be due to loss of sulfur and other volatiles in the fusion step used to form a bead for X-ray fluorescence.

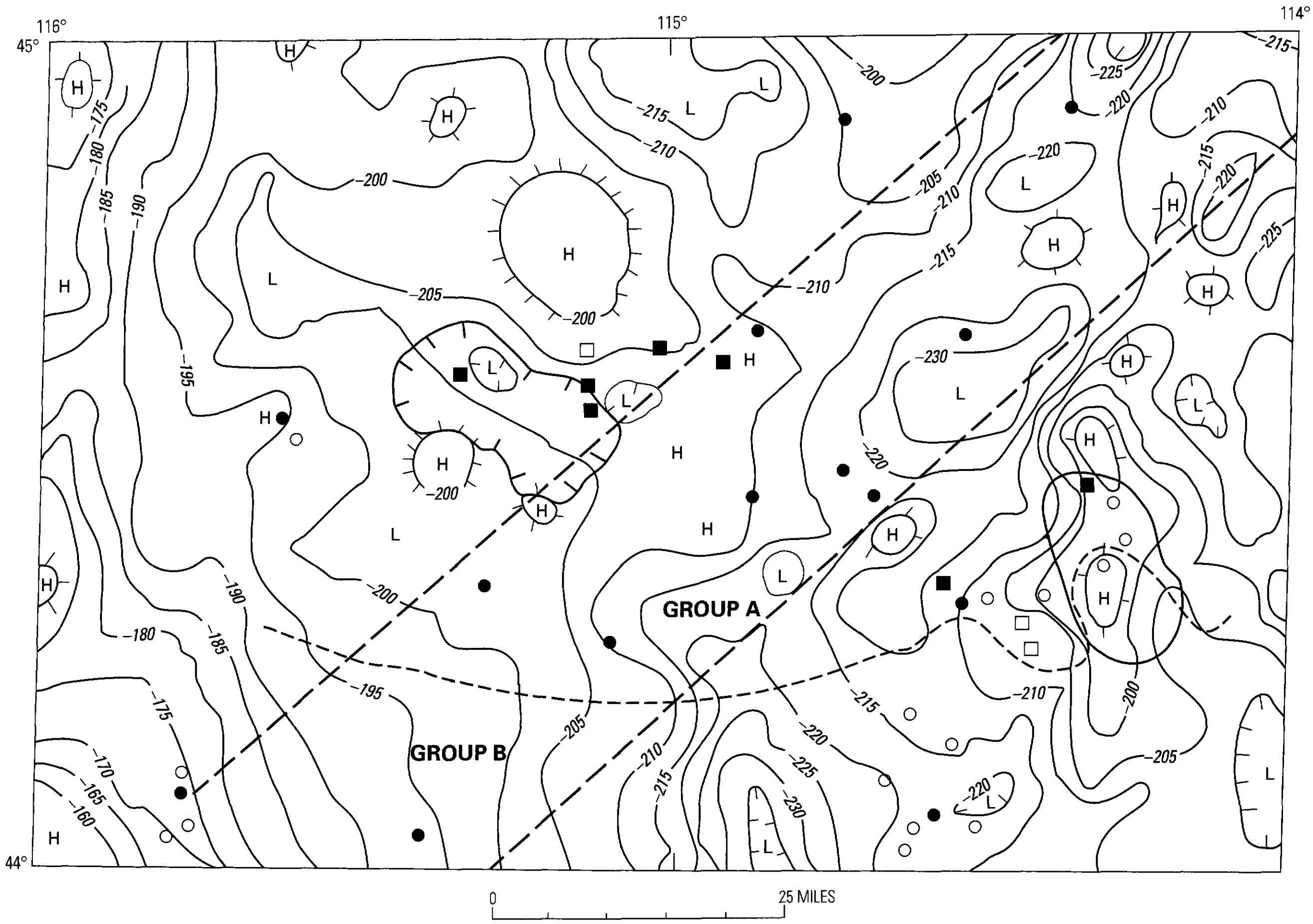


Figure 8. Map showing lead-isotope data and simplified Bouguer gravity anomalies, Challis 1°×2° quadrangle, Idaho. Contour interval 10 gammas; H, gravity high; L, gravity low. Heavy solid lines delineate aeromagnetic high at 1,800 gammas (relative) closure in the Bayhorse anticline area and magnetic low west of the Trans-Challis fault complex. Dashed line separates group A and group B isotopic areas. Straight heavy dashed lines trending northeast-southwest indicate margins of Trans-Challis fault complex. Symbols: open circle, ore having $^{206}Pb/^{204}Pb$ greater than 19.1; open square, ore having $^{206}Pb/^{204}Pb$ 18.1–19.1; solid square, ore having $^{206}Pb/^{204}Pb$ less than 18.0; solid circle, granite- or rhyolite-type. Geophysical data from D.R. Mabey (written commun., February 1980).

mineralogical and chemical assemblage similar to Bolivian tin deposits (Hall, 1985; Tschanz and Kiilsgaard, 1986). Although they are half a world apart, the two major tin mining areas of the world, Malaysia and Bolivia, both have unusual but similar lead-isotope ratios for ore deposits, typical of upper crust (Doe, 1992). Though the sample size is small, the values of $^{208}Pb/^{204}Pb$ and, especially, $^{207}Pb/^{204}Pb$ are unusually great for galenas in both tin-producing areas.

Lead-isotopic compositions for galena and tin granite from the Challis and Hailey quadrangles are shown in figure 11, along with data from the tin provinces of Malaysia (Doe, 1992) and Bolivia (Tilton and others, 1981). A few galenas of group A from the Challis and Hailey quadrangles have values of $^{207}Pb/^{204}Pb$ that are similarly as high as those from the tin provinces, but their $^{208}Pb/^{204}Pb$ values are even higher. Galenas of group B from the Grimes Pass district (and Gilmore district) compare more favorably with data from the tin-producing areas, and an especially good fit is for tin granite along the boundary between the Challis and Hailey quadrangles. Perhaps surprisingly, data from the best stannite-bearing areas of the Challis (Timberline and Patti Flynn prospects) and the Hailey (Golden Glow and Crown point deposits in Boulder Basin and Combination deposit in Galena Basin) quadrangles show values of $^{206}Pb/^{204}Pb$ higher than the tin producers; however, all the tin-bearing deposits have high values of $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$, typical of the upper crust. Although not too much weight should be placed on this "fingerprinting" involving such widely spaced localities, further interest in the Challis and Hailey 1°×2° quadrangles for tin resources is warranted, especially in those areas of the Boulder Basin that are also silver rich such that tin might be produced as a byproduct (Hall, 1987a).

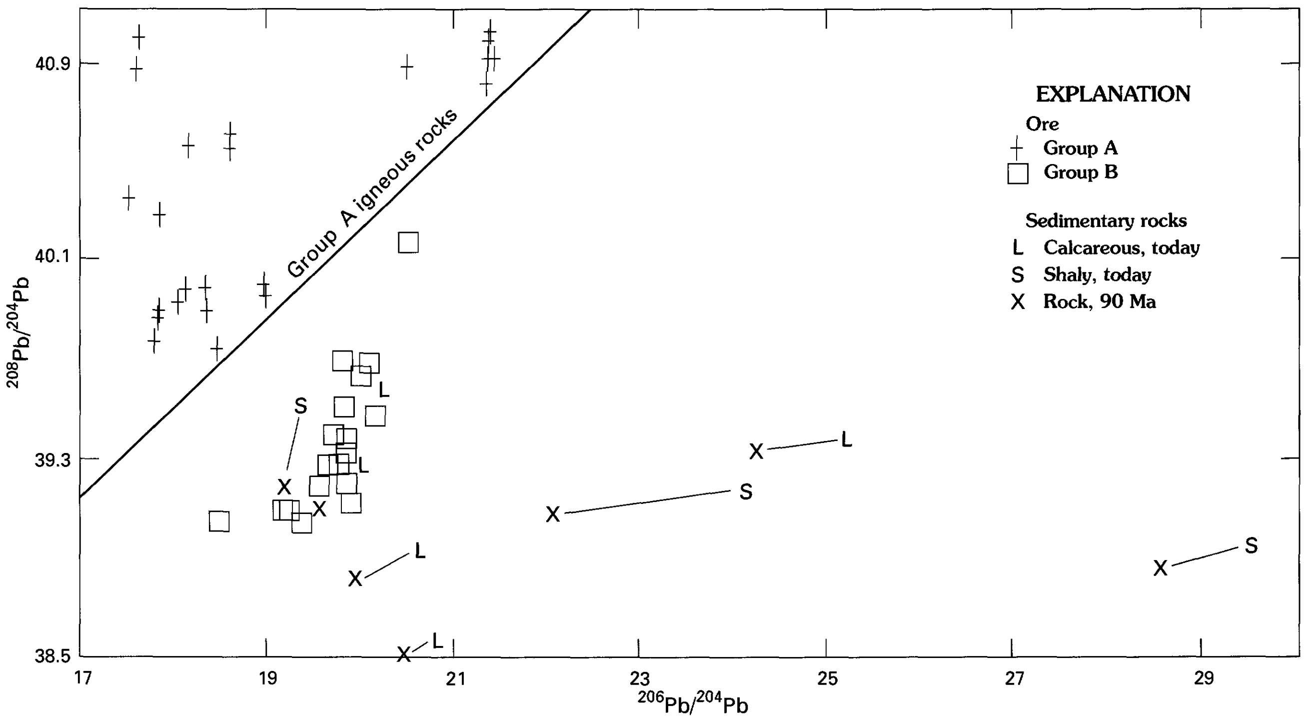


Figure 9. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ for ore minerals and sedimentary rocks (ore minerals and shaly rocks are from central Idaho, limestone is from Montana). Solid line is regression for group A igneous rocks (fig. 6).

ORIGIN OF GRANITE

There were three significant surprises in this study. The first is the discovery of two lead-isotope terranes: one terrane similar to Proterozoic and Precambrian crystalline-rock terranes of the Rocky Mountains, in which rock units have non-radiogenic uranogenic leads that have high values of $^{232}Th/^{238}U$, and the other terrane similar to the Basin and Range, in which normal to moderately radiogenic lead is present. The second surprise is that stockwork molybdenum porphyry and tin granite are present in both terranes. The third surprise is that the stockwork molybdenum and tin granite within each isotopic terrane have similar lead-isotopic compositions (table 5).

All of the lead-isotope data in central Idaho have crustal signatures, and there is no evidence of an oceanic mantle component. In fact, the values of $^{207}Pb/^{204}Pb$ tend to be high for their values of $^{206}Pb/^{204}Pb$, a characteristic of the upper crust (Doe and Zartman, 1979). Norman and Leeman (1989) presented lead-isotope data for four samples of Idaho batholith (Cretaceous) and five samples of Challis Volcanic Group (Eocene) from central and southwestern Idaho, as well as data for younger igneous rocks. They also presented data on strontium and neodymium isotopes and a number of trace elements. Although they do not use the group A and B classification of Doe and Delevaux (1985) or the areas of Zartman (1974), their lead-isotope data for the Cretaceous Idaho batholith fit in group B and their data for the Eocene Challis Volcanic Group fit in group A. Utilizing trace element and other isotopic data, they interpreted the group B leads as being mixtures of subduction-related magmas (asthenospheric mantle, oceanic lithosphere, and continental lithospheric mantle), as represented by basalts of the Cascade Range (included in area III of Zartman) and Precambrian continental crust. They interpreted group A leads as coming from continental lithospheric sources (subcontinental lithospheric mantle, unradiogenic lower crust, and more radiogenic upper crust) but said there is a strong case for intracrustal melting.

Norman and Leeman (1989) related the isotopic differences between leads of group A and group B to the change in tectonic environments, probably from compression during the Cretaceous to strike-slip movement during the Eocene. We find, however, that leads of group A and group B are present in either the Cretaceous or Tertiary igneous rocks, and, thus, the isotopic groupings probably are not time related. We are in substantial agreement with Norman and Leeman as to the sources of group A leads, but their interpretation of the sources of group B leads is an alternative interpretation to ours and that of Church and others (1986) for the Mt. Tolman molybdenum porphyry in eastern Washington; that is, derivation from a Basin and Range geologic section or area II of Zartman. Unfortunately, neodymium and strontium isotope data are not available for our samples in order

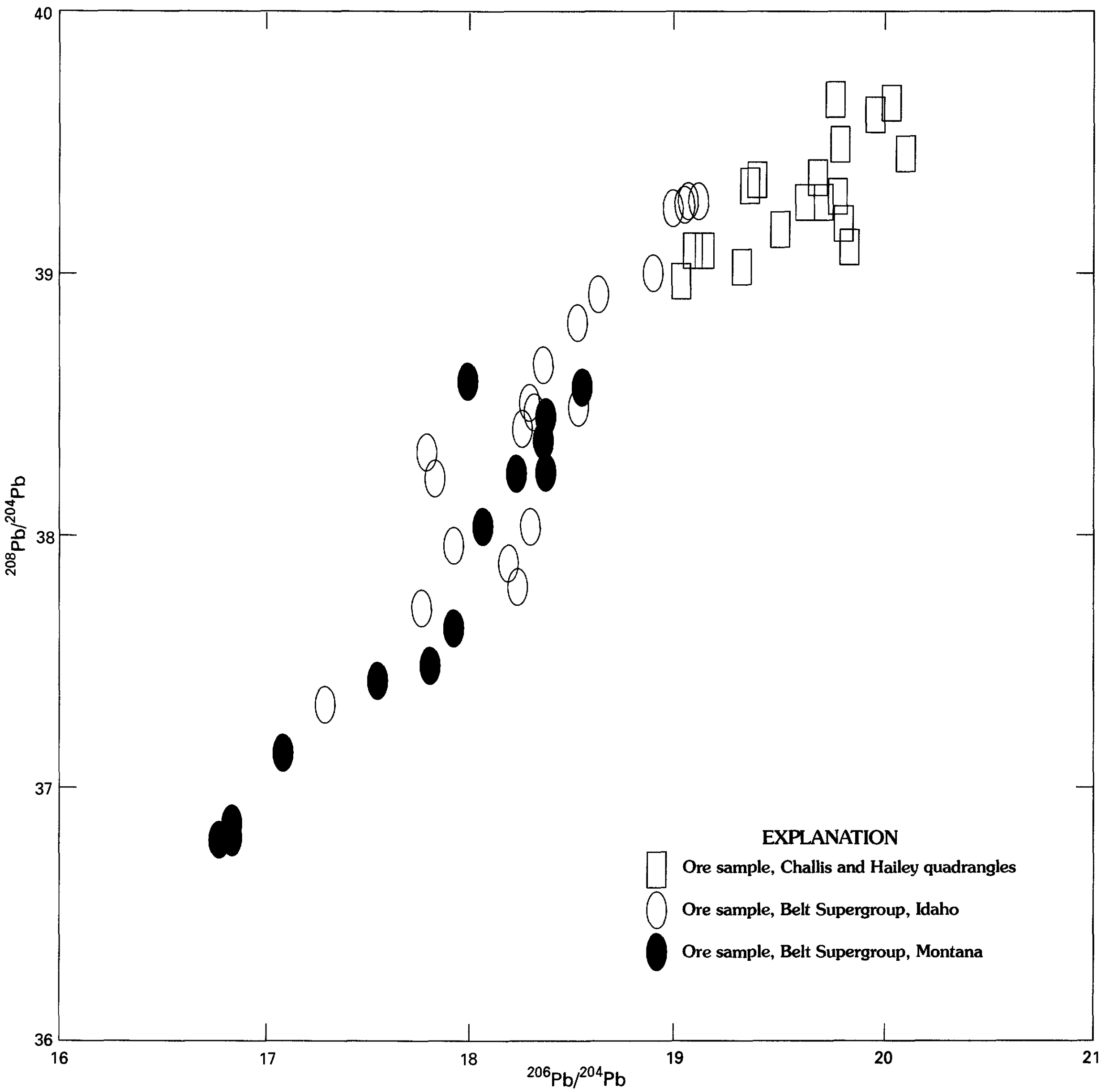


Figure 10. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ for ore minerals of the Challis and Hailey 1°×2° quadrangles, Idaho, and Mesozoic and Cenozoic ore minerals of the Belt Supergroup in Idaho and Montana. Data for Belt Supergroup from Zartman and Stacey (1971), Marvin and others (1983), and Marvin and Zartman (1984).

that we might further evaluate these possibilities. The important point, however, is that consideration of expanded geochemical data indicates derivation of the igneous rocks from two quite different source environments. Both molybdenum porphyry and tin granite formed, however, in both source environments.

Although the molybdenum porphyries of central Idaho are classed as low-fluorine porphyries, samples from the Thompson Creek mine, a group A isotopic deposit, have nonradiogenic values of $^{206}Pb/^{204}Pb$ and high values of $^{208}Pb/^{204}Pb$ (table 4), as do samples from the Henderson Climax-type of molybdenum porphyry in the Colorado Front Range (Stein, 1985); such values indicate derivation from high-rank metamorphic rocks (Doe and Zartman, 1979). The sulfur-isotope data are also characteristic of crustal sources, particularly sedimentary sources (Howe and Hall, 1985). Although their $\delta^{34}S$ values are not as heavy as those for the stockwork molybdenum deposit of Thompson Creek in Idaho, molybdenum porphyries of the Climax type in the Rocky Mountain provinces tend to have high values of $\delta^{34}S$ relative to the mantle. Stein (1985) gave values of $\delta^{34}S$ for molybdenum porphyries at Henderson and Mount Emmons,

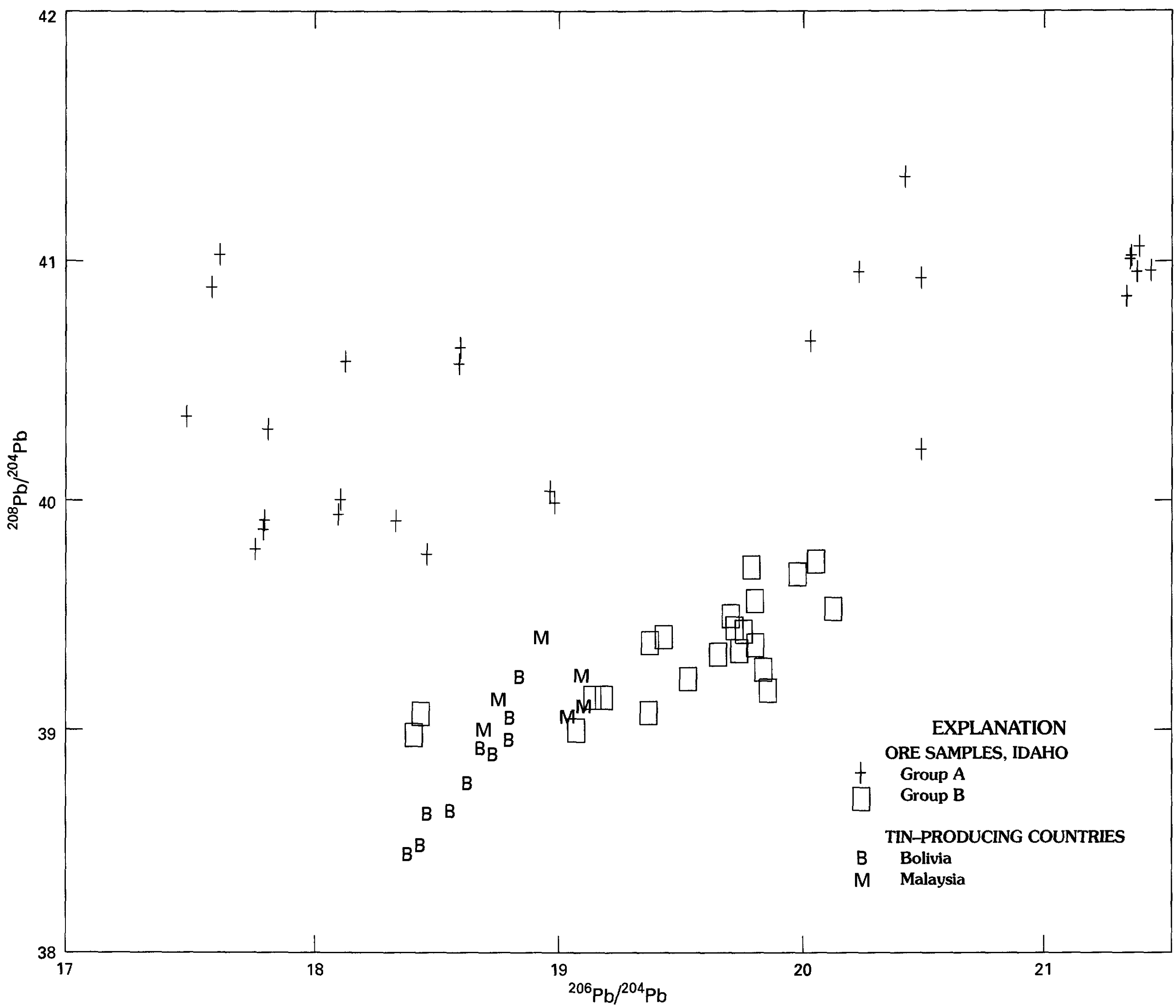


Figure 11. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ for ore minerals of the Challis and Hailey 1°×2° quadrangles, Idaho, and ore minerals from the largest tin-producing countries, Malaysia and Bolivia (Tilton and others, 1981; Doe, 1992).

Colorado, of about +4 to +6 and at Climax, Colorado, of about +2 to +4, whereas $\delta^{34}S$ values at Questa, New Mexico, are close to mantle values of zero.

The mantle is generally highly depleted in molybdenum and moderately depleted in tin, as compared to the bulk crust (primitive mantle is depleted in molybdenum by more than a factor of 10 and in tin by a factor of 2; Lehman, 1990), and is a poor source for these metals. In fact, Lehman (1990) concluded that the Eastern tin belt of Malaysia is less tin rich than the Main tin belt because the magmas of the Eastern belt may have included a mantle component, as deduced from strontium and neodymium isotope data. Molybdenum and tin are enriched in the upper crust as compared to primitive mantle (molybdenum by more than a factor of 20 and tin by a factor of 5), whereas copper is present in about equal amounts (Lehman, 1990). It therefore is likely that the molybdenum and tin granites of Idaho were derived from within the crust or at least from differentiating mantle magmas that reacted extensively with intruded crust.

Tin granite is expected to have a lead-isotope signature different from that of stockwork molybdenum granite because tin enrichment by differentiation requires that the magma be highly reduced, whereas molybdenum-enriched magma is more oxidizing (Ishihara and others, 1979; Candela and Holland, 1986; Lehman, 1990). This reduced nature is commonly presumed to arise through reaction of magma with carbon-rich rocks such as pelitic schist or black shale; however, sulfur-rich sources also may reduce a magma (Lehman, 1990). Reduction by sulfur generally has less appeal than reduction by carbon because the tin mineral of interest is the tin oxide cassiterite. If tin granite magmas have interacted more with carbonaceous sediments than have molybdenum granite, a difference in lead-isotope signatures might be expected because shaly rocks have lead contents similar to those of the granites analyzed, and old shale generally has more radiogenic lead by Eocene time (fig. 9) than does the granite, especially more than the granite of group A. But such an isotopic difference between molybdenum and tin granites is not the case.

A model is needed whereby the molybdenum and tin granite magmas are derived from the same sources and follow the same reaction paths with wallrocks of normal lead contents (which would give similar lead-isotope signatures) but follow different differentiation trends. In one possible model, tin granite magmas reacted with lead-poor, carbon-rich material such as petroleum, natural gas, or perhaps coal, whereas molybdenum granite magmas did not. Because the major stockwork molybdenum-bearing granites in the study area are about 45 m.y. older than the tin granites, the carbon-rich material could have been absent during the Cretaceous but present during the Eocene as a result of thrusting and fluid migration. Eocene molybdenum-enriched magmas then would not have intersected the carbonaceous lead-poor material. In a second, and perhaps more likely, model, tin granite magmas reacted with older sulfide-mineralized Cretaceous igneous rocks and mineralized country rocks. Some radiogenic lead would have formed in the Cretaceous igneous rocks between the time of their origin about 90 Ma and the time of formation of the tin granite about 45 Ma. Because Cretaceous granites were not isotopically uniform when they formed, the radiogenic lead evolution may be lost in the noise of the isotopic signatures. Also, we have corrected all whole-rock samples back to a common time of 90 Ma (table 5) to remove any isotopic effects associated with the age difference.

In the second model, the tin granite magmas were reduced by sulfur associated with the Cretaceous intrusive rocks. That the predominant tin mineral is stannite, a copper-iron-tin sulfide, rather than cassiterite, a tin oxide, supports this hypothesis. The Cretaceous intrusive rocks may also be a source of tin, even though the low-fluorine molybdenum granites do not contain tin minerals, because the tin may be in minerals such as sphene, magnetite, ilmenite, epidote, and hornblende. Eocene molybdenum granites and rhyolites then would have evolved without extensive reactions with Cretaceous mineralized intrusive rocks.

REFERENCES CITED

Bankey, Viki, and Kleinkopf, M.D., 1988, Bouguer gravity anomaly map and four derivative maps of Idaho: U.S. Geological Survey Geophysical Investigations Map GP–978, 3 sheets.

Bennett, E.H., and Knowles, C.R., 1985, Tertiary plutons and related rocks in central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–F, p. 81–95.

Candela, P.A., and Holland, H.H., 1986, A mass transfer model for copper and molybdenum in magmatic hydrothermal systems—The origin of porphyry-type ore deposits: Economic Geology, v. 81, p. 1–17.

Cannon, R.S., Pierce, A.P., Antweiler, J.C., and Buck, K.L., 1961, The data of lead isotope geology related to problems of ore genesis: Economic Geology, v. 56, p. 1–38.

Church, S.E., LeHuray, A.P., Grant, A.R., Delevaux, M.H., and Gray, J.E., 1986, Lead-isotopic data from sulfide minerals from the Cascade Range, Oregon and Washington: Geochimica et Cosmochimica Acta, v. 50, p. 317–328.

Davis, K.E., 1978, Lead isotope ratios of galena from the Bayhorse mining district, Custer County, Idaho: Isochron/West, no. 22, p. 15–16.

Doe, B.R., 1979, The application of lead isotopes to mineral prospect evaluation of Cretaceous-Tertiary magmatothermal ore deposits in the Western United States, *in* Watterson, J.R., and Theobald, P.K., eds., Geochemical exploration 1978: International Geochemical Exploration Symposium, 7th, Golden, Colorado, Proceedings; Association of Exploration Geochemists, p. 227–232.

———1992, Lead isotopic composition of galena from Malaysia, an S-type granite terrane, *in* Informal notes of the workshop on the application of isotope systems to geological problems: U.S. Geological Survey Open-File Report 92–525, p. 28–62.

Doe, B.R., Berger, B.R., and Elliott, J.E., 1986, Lead-isotope evaluation of selected ores and mineral prospects in the Butte and Dillon 1°×2° quadrangles, Montana: U.S. Geological Survey Open-File Report 86–111, 40 p.

Doe, B.R., and Delevaux, M.H., 1973, Variations in lead isotopic compositions in Mesozoic granitic rocks of California—A preliminary investigation: Geological Society of America Bulletin, v. 84, p. 3513–3526.

———1980, Lead isotope investigations in the Minnesota River valley—I, Late- and post-tectonic granites: Geological Society of America Special Paper 182, p. 105–112.

———1985, Lead-isotope characteristics of ore systems in central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–O, p. 181.

Doe, B.R., Hedge, C.E., and White, D.E., 1966, Preliminary investigation of the source of lead and strontium in deep geothermal brines underlying the Salton Sea geothermal area: Economic Geology, v. 61, p. 462–483.

Doe, B.R., and Stacey, J.S., 1974, The application of lead isotopes to the problems of ore genesis and ore prospect evaluation—A review: Economic Geology, v. 69, p. 757–776.

Doe, B.R., Steven, T.A., Delevaux, M.H., Stacey, J.S., Lipman, P.W., and Fisher, F.S., 1979, Genesis of ore deposits in the San Juan volcanic field, southwestern Colorado—Lead isotope evidence: Economic Geology, v. 74, p. 1–26.

Doe, B.R., and Zartman, R.E., 1979, Plumbotectonics, The Phanerozoic, *in* Barnes, H., ed., Geochemistry of hydrothermal ore deposits: New York, Wiley Interscience Publications, p. 22–70.

Dover, J.H., 1981, Geology of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho, *in* Mineral resources of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–A, p. 15–75.

Ekren, E.B., 1985, Eocene cauldron-related volcanic events in the Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–C, p. 43–58.

Fisher, F.S., and Johnson, K.M., eds., 1987a, Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, 251 p.

———1987b, Challis rock terrane, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 51–54.

Fisher, F.S., Kiilsgaard, T.H., Johnson, K.M., and Bennett, E.H., 1987, Precious-metal veins, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 117–126.

Fisher, F.S., McIntyre, D.H., and Johnson, K.M., 1983, Geologic map of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 83–523, 1:250,000 scale, 41 p.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

———1987a, Lead-silver-zinc-antimony-tin vein deposits, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 133–140.

———1987b, Cretaceous molybdenum stockworks, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 152–158.

Hall, W.E., Batchelder, J.N., and Douglass, R.D., 1974, Stratigraphic section of the Wood River Formation, Blaine County, Idaho: U.S. Geological Survey Journal of Research, v. 2, no. 1, p. 89–95.

Hall, W.E., and Hobbs, S.W., 1987, Black shale terrane, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 29–42.

Hall, W.E., Rye, R.O., and Doe, B.R., 1978, Wood River mining district, Idaho—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, p. 579–592.

Hall, W.E., Schmidt, E.A., Howe, S.S., and Broch, M.J., 1984, The Thompson Creek, Idaho, porphyry molybdenum deposit—An example of a fluorine deficient molybdenum granodiorite system: Proceedings of the Sixth Quadrennial IAGOD Symposium, v. 6, p. 349–357.

Hardyman, R.F., 1985, The Twin peaks caldera and associated ore deposits, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–G, p. 97–105.

Hardyman, R.F., and Fisher, F.S., 1985, Rhyolite intrusions and associated mineral deposits in the Challis volcanic field, Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–N, p. 167–179.

Hobbs, S.W., 1985, Structural and stratigraphic controls of ore deposits in the Bayhorse area, Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–K, p. 133–140.

———1987, Carbonate rock terrane, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 20–29.

Hobbs, S.W., Fisher, F.S., and Johnson, K.M., 1987, Irregular replacement deposits of base and precious metals, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 175–179.

Hobbs, S.W., Hays, W.H., and McIntyre, D.H., 1975, Geologic map of the Clayton quadrangle, Custer County, Idaho: U.S. Geological Survey Open-File Report 75–76, scale 1:62,500, 23 p.

Howe, S.S., and Hall, W.E., 1985, Light-stable-isotope characteristics of ore systems in central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–P, p. 183–192.

Ishihara, S., 1977, The magnetite-series and ilmenite-series granitic rocks: Mining Geology (Japan), v. 27, p. 293–305.

Ishihara, S., Sawata, H., Arpornsuwan, S., Busaracome, P., and Bungbrakearti, N., 1979, The magnetite-series and ilmenite-series granitoids and their bearing on tin mineralization, particularly of the Malay Peninsula region: Geological Society of Malaysia Bulletin, v. 11, p. 103–110.

Johnson, K.M., and Fisher, F.S., 1987, High-level rhyolite-hosted precious-metal deposits, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 163–166.

Johnson, K.M., Lewis, R.S., Bennett, E.H., and Kiilsgaard, T.H., 1988, Cretaceous and Tertiary intrusive rocks of south-central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 55–86.

Kiilsgaard, T.H., and Bennett, E.H., 1985, Mineral deposits in the southern part of the Atlanta lobe of the Idaho batholith and their genetic relation to Tertiary intrusive rocks and to faults, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–M, p. 153–165.

———1987a, Idaho batholith terrane, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 43–51.

———1987b, Eocene plutonic rock terrane, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2°

quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 54–60.

———1987c, Mixed base- and precious-metal veins, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 126–133.

———1987d, Tertiary molybdenum stockworks, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 158–162.

Kiilsgaard, T.H., and Hingley, Frank, 1989, Regional geology of the northwestern part of the Hailey 1°×2° quadrangle and ore deposits of the Atlanta district, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangle quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 35–36.

Kiilsgaard, T.H., and Lewis, R.S., 1985, Plutonic rocks of Cretaceous age and faults in the Atlanta lobe of the Idaho batholith, Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–B, p. 29–44.

Leeman, W.P., Menzies, M.A., Matty, D.J., and Embree, G.F., 1985, Strontium, neodymium and lead isotopic compositions of deep crustal xenoliths from the Snake River Plain—Evidence for Archean basement: Earth and Planetary Science Letters, v. 75, p. 354–368.

Lehman, Bernd, 1990, Metallogeny of tin, *in* Bhattacharji, S., and others, eds., Lecture notes in earth sciences: New York, Springer Verlag, v. 32, 211 p.

Lewis, R.S., 1989, Plutonic rocks in the southeastern part of the Idaho batholith and their relation to mineralization, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles: U.S. Geological Survey Open-File Report 89–639, p. 33–34.

Link, P.K., and Hackett, W.R., eds., 1988, Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, 319 p.

Ludington, S.D., 1986, Descriptive model of Climax Mo deposits, *in* Cox, D.P., and Singer, D.A., eds., Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 73.

Mabey, D.R., and Webring, M.W., 1985, Regional geophysical studies in the Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–E, p. 69–79.

Marvin, R.F., and Zartman, R.E., 1984, A tabulation of lead isotopic ratios for lead minerals from mines and prospects in the Belt-Purcell Basin in the United States and Canada: U.S. Geological Survey Open-File Report 84–210, 15 p.

Marvin, R.F., Zartman, R.E., Obradovich, J.D., and Harrison, J.E., 1983, Geochronometric and lead isotope data on samples from the Wallace 1°×2° quadrangle, Montana and Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF–1354–G.

McFadden, M.D., Measures, E.A., and Isaacson, P.E., 1988, Early Paleozoic continental margin development, central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 129–152.

McIntyre, D.H., ed., 1985, Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658, 227 p.

McIntyre, D.H., Ekren, E.B., and Hardyman, R.F., 1982, Stratigraphic and structural framework of the Challis volcanics in the eastern half of the Challis 1°×2° quadrangle, Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 3–22.

McIntyre, D.H., Hobbs, S.W., Marvin, R.F., and Mehnert, H.H., 1976, Late Cretaceous and Eocene ages for hydrothermal alteration and mineralization, Bayhorse District and vicinity, Custer County, Idaho: Isochron/West, no. 16, p. 11.

McIntyre, D.H., and Johnson, K.M., 1985, Epithermal gold-silver mineralization related to volcanic subsidence in the Custer graben, Custer County, Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–I, p. 109–115.

Mitchell, V.E., Strowd, W.B., Hustedde, G.S., and Bennett, E.H., 1981, Mines and prospects of the Dubois quadrangle, Idaho: Idaho Bureau of Mines and Geology, Mines and Prospects Map Series, scale 1:250,000, 22 p.

———1986, Mines and prospects of the Challis quadrangle, Idaho (2nd ed.): Idaho Bureau of Mines and Geology, Mines and Prospects Map Series, scale 1:250,000, 56 p.

Moye, F.J., Hackett, W.R., Blakley, J.D., and Snider, L.G., 1988, Regional geologic setting and volcanic stratigraphy of the Challis volcanic field, central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 87–97.

Moye, F.J., Sanford, R.F., Erdman, J.A., and Johnson, K.M., 1989, Volcanic-hosted and intrusion-related mineral deposits in the Hailey and western Idaho Falls quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 70–72.

Norman, M.D., and Leeman, W.P., 1989, Geochemical evolution of Cenozoic-Cretaceous magmatism and its relation to tectonic setting, southwestern Idaho, U.S.A.: Earth and Planetary Science Letters, v. 94, p. 78–96.

Ohmoto, Hiroshi, and Rye, R.O., 1979, Isotopes of sulfur and carbon, *in* Barnes, H.L., ed., Geochemistry of hydrothermal ore deposits (2nd ed.): New York, Wiley-Interscience, p. 509–567.

Paull, R.A., Wolbrink, M.A., Volkmann, R.G., and Grover, R.L., 1972, Stratigraphy of the Copper Basin Group, Pioneer Mountains, south-central Idaho: American Association of Petroleum Geologists Bulletin, v. 56, p. 1370–1401.

Poole, F.G., Sandberg, C.A., and Boucot, A.J., 1977, Silurian and Devonian paleogeography of the western United States, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the Western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 1, p. 39–65.

Reed, B.L., 1986, Descriptive model of porphyry Sn, *in* Cox, D.P., and Singer, D.A., eds., Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 108.

Rye, R.O., Doe, B.R., and Wells, J.D., 1974, Stable isotope and lead isotope study of the Cortez, Nevada, gold deposit and surrounding area: U.S. Geological Survey Journal of Research, v. 2, p. 13–23.

Rye, R.O., and Ohmoto, Hiroshi, 1974, Sulfur and carbon isotopes and ore genesis—A review: Economic Geology, v. 69, p. 826–842.

Rye, R.O., Roberts, R.J., Whelan, J.F., and Mawad, M.M., 1980, Reconnaissance sulfur isotope investigations of mineral deposits in the Precambrian shield, Kingdom of Saudi Arabia: U.S. Geological Survey Saudi Arabian Mission Technical Record 7 (Interagency Report 346), 77 p.

Sandberg, C.A., Hall, W.E., Batchelder, J.N., and Axelson, Claus, 1975, Stratigraphy, conodont dating, and paleotectonic interpretation of the type Milligen Formation (Devonian), Wood River area, Idaho: U.S. Geological Survey Journal of Research, v. 3, p. 707–720.

Sanford, R.F., Wooden, J.L., Doe, B.R., and Worl, R.G., 1989, Origin of ore deposits in the central Idaho black shale belt—Implications from lead isotopes, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 39–52.

Skipp, Betty, and Hait, M.H., Jr., 1977, Allochthons along the northeast margin of the Snake River Plain, Idaho: Wyoming Geological Association Guidebook, Twenty-Ninth Annual Field Conference, p. 499–515.

Skipp, Betty, and Hall, W.E., 1975, Structure and Paleozoic stratigraphy of a complex of thrust plates in the Fish Creek Reservoir area, south-central Idaho: U.S. Geological Survey Journal of Research, v. 3, p. 671–689.

Skipp, Betty, and Sandberg, C.A., 1975, Silurian and Devonian miogeosynclinal and transitional rocks of the Fish Creek Reservoir window, central Idaho: U.S. Geological Survey Journal of Research, v. 3, p. 691–706.

Small, W.D., 1968, Cordilleran geochronology deduced from hydrothermal leads: Vancouver, University of British Columbia, Ph.D. thesis, 96 p.

Stacey, J.S., and Kramers, J.D., 1975, Approximation of terrestrial lead isotope evolution by a two-stage model: Earth and Planetary Science Letters, v. 26, p. 207–221.

Stacey, J.S., Zartman, R.E., and NKomo, I.T., 1968, A lead isotope study of galenas and selected feldspars from mining districts in Utah: Economic Geology, v. 63, p. 796–814.

Stein, H.J., 1985, A lead, strontium, and sulfur isotope study of Laramide-Tertiary intrusions and mineralization in the Colorado mineral belt with emphasis on Climax type porphyry molybdenum systems plus a summary of other newly acquired isotopic and rare earth element data: Chapel Hill, University of North Carolina, Ph.D. thesis, 500 p.

Theodore, T.G., 1986, Descriptive model of porphyry Mo, low-F, *in* Cox, D.P., and Singer, D.A., eds., Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 120.

Tilton, G.R., Pollak, R.J., Clark, A.H., and Robertson, R.C.R., 1981, Isotopic composition of Pb in central Andean ore deposits: Geological Society of America Memoir 154, p. 791–816.

Tischendorf, G., 1977, Geochemical and petrographic characteristics of silicic magmatic rocks associated with rare-element mineralization, *in* Stemprok, M., and others, eds., Metallization associated with acid magmatism: Czechoslovakia Geological Survey, v. 2, p. 41–96.

Togashi, Y., 1986, Descriptive model of Sn-polymetallic veins, *in* Cox, D.P., and Singer, D.A., eds., Mineral deposit models: U.S. Geological Survey Bulletin 1693, p. 109.

Tschanz, C.M., and Kiilsgaard, T.H., 1986, Geologic appraisal of mineral resources of the eastern part of the Sawtooth National Recreation Area, Idaho, *in* Mineral resources of the eastern part of the Sawtooth National Recreation Area, Custer and Blaine Counties, Idaho: U.S. Geological Survey Bulletin 1545–C, p. 54–210.

Tschanz, C.M., Kiilsgaard, T.H., and Seeland, D.A., 1986, Geology of the eastern part of the Sawtooth National Recreation Area, Idaho, *in* Mineral resources of the eastern part of the Sawtooth National Recreation Area, Custer and Blaine Counties, Idaho: U.S. Geological Survey Bulletin 1545–A, p. 17–43.

Tschanz, C.M., Kiilsgaard, T.H., Seeland, D.A., Van Noy, R.M., Ridenour, James, Zilka, N.T., Federspiel, F.E., Evans, R.K., Tuchek, E.T., and McMahan, A.B., 1974, Mineral resources of the eastern part of the Sawtooth National Recreation Area, Custer and Blaine Counties, Idaho, *with sections on* Aeromagnetic survey and tentative interpretation, by D.R. Mabey and C.M. Tschanz, *and* Electromagnetic surveys, by F.C. Frischknecht: U.S. Geological Survey Open-File Report 74–1100, 2 v., 648 p.

Turner, R.J.W., and Otto, B.R., 1988, Stratigraphy and structure of the Milligen Formation, Sun Valley area, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 153–167.

Wavra, C.S., Isaacson, P.E., and Hall, W.E., 1986, Studies of the Idaho black shale belt—Stratigraphy, depositional environment, and economic geology of the Permian Dollarhide Formation: Geological Society of America Bulletin, v. 97, p. 1504–1511.

Webring, M.W., and Mabey, D.R., 1987, Geophysical anomalies in the Challis quadrangle, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 66–80.

Westra, G., and Keith, S.B., 1981, Classification and genesis of stockwork molybdenum deposits: Economic Geology, v. 76, p. 844–873.

Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., 1989, Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, 140 p.

Worl, R.G., Kiilsgaard, T.K., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Zartman, R.E., 1974, Lead isotopic provinces in the Cordillera of the Western United States and their geologic significance: Economic Geology, v. 69, p. 792–805.

Zartman, R.E., and Stacey, J.S., 1971, Lead isotopes and mineralization ages in Belt Supergroup rocks, northwestern Montana and northern Idaho: Economic Geology, v. 66, p. 849–860.

Sources of Lead in Ore Deposits of Central Idaho

By Richard F. Sanford *and* Joseph L. Wooden

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND
IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–N

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Sources of Lead in Ore Deposits of Central Idaho

By Richard F. Sanford[1] *and* Joseph L. Wooden[2]

ABSTRACT

Ore and rock samples in central Idaho are divided into six types on the basis of distinctive lead-isotope signatures. Lead previously identified as group A lead is subdivided into the Carrietown-, Minnie Moore-, Pacific-, and Lava Creek-types, which are characterized by discrete ranges in uranogenic ($^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$) and thorogenic ($^{208}Pb/^{204}Pb$) lead ratios. Carrietown-type lead, which is low in uranogenic lead and high in thorogenic lead, is probably derived from Precambrian upper crustal rocks beneath the Cretaceous Idaho batholith. Minnie Moore- and Pacific-type leads, which are high in both uranogenic and thorogenic lead, have no obvious source, but recycled Precambrian upper crustal rocks are likely candidates. Lava Creek-type lead, which is low in both uranogenic and thorogenic lead, is probably derived from Precambrian upper crustal rocks having lower $^{208}Pb/^{204}Pb$ than rocks beneath the Idaho batholith.

Lead previously identified as group B lead is subdivided into the Triumph- and black shale-types, both of which are high in uranogenic lead and low in thorogenic lead. Triumph-type lead has higher $^{207}Pb/^{204}Pb$ than black shale-type lead. Black shale-type lead probably was derived from recycled Proterozoic upper crustal rocks eroded from the craton to the east and northeast, whereas Triumph-type lead probably was derived from recycled Archean and Proterozoic rocks in the middle or lower crust. The variety of lead-isotope types that originated in Precambrian basement suggests that the Precambrian basement is quite heterogeneous.

The six types of lead suggest three crustal zones: a western crustal zone characterized by Carrietown-type lead (low uranogenic lead ratios and high thorogenic lead ratio); a central zone having four lead-isotope types, Minnie Moore, Pacific, black shale and Triumph, all of which are high in $^{207}Pb/^{204}Pb$; and an eastern zone characterized by Lava Creek-type lead, which is low in both uranogenic and thorogenic lead. These crustal zones correspond closely to five of eight geologic terranes previously recognized in the central Idaho area. The broad western crustal zone corresponds to the Cretaceous intrusive rock terrane and small parts of the black shale terrane; the narrow arcuate central zone corresponds to almost all of the black shale terrane; and the broad eastern zone corresponds to the flysch, shelf carbonate, and Precambrian terranes.

Ore deposit types typically have characteristic lead-isotope signatures. Carrietown-type lead is in veins hosted by the Idaho batholith and adjacent, locally metamorphosed sedimentary rocks. The association of intrusive rocks, metamorphic rocks, and veins suggests a common and contemporaneous plumbing system that extends down to the Precambrian basement. The Cretaceous intrusions probably provided the heat for hydrothermal circulation, but the metals apparently came from the underlying basement, not from the intrusions.

Triumph-type lead is typical of the lead in syngenetic, massive sulfide, "black smoker" deposits that formed in a syndepositional rift basin. This lead probably came from a Precambrian source at depth and was transported upward by hydrothermal fluids and deposited at submarine vents. During transport, this lead may have mixed with a component of lead from argillite, but argillite could not have been the sole source of lead in these deposits. Triumph-type lead in epigenetic veins represents either the contemporaneous feeder system for the syngenetic deposits or syngenetic sulfides from which lead was later remobilized. Most likely, veins in Devonian argillite having Triumph-type lead were part of the feeder system for contemporaneous syngenetic deposits, whereas those in Pennsylvanian-Permian argillite consist of remobilized Devonian syngenetic sulfides.

Black shale-type lead is typical of argillite as well as Cretaceous intrusive rocks. Very few vein deposits have a black shale-type lead signature similar to that of the host argillite. Therefore, these ore deposits probably did not form by lateral secretion. Most argillite-hosted veins probably derived their lead ultimately from Precambrian basement, either directly or by remobilization of syngenetic lead, but not from the detrital component of the argillite.

[1]U.S. Geological Survey, Denver Federal Center, Denver, Colorado 80225.

[2]U.S. Geological Survey, 345 Middlefield Road, Menlo Park, California 94025.

Epigenetic sulfides in central Idaho thus have lead-isotope types that differ with crustal zone. Epigenetic sulfides in the western crustal zone have mostly Carrietown-type lead, those in the central crustal zone have Minnie Moore-, Pacific-, black shale-, and Triumph-type lead, and those in the eastern crustal zone have mainly Lava Creek-type lead. Lava Creek-type lead probably is unique to, and characteristic of, Eocene veins. Syngenetic sulfides are present only in the black shale terrane and have only Triumph-type lead.

INTRODUCTION

The black shale terrane of central Idaho (figs. 1, 2) has been a significant area of silver-lead-zinc production since the late 1800's and is the subject of several modern ore-deposit studies (Hall and others, 1978; Hall, 1985). Lead isotopes in galena from the Minnie Moore mine in the Big Wood River valley, central Idaho, were noted as unusually radiogenic for deposits of this size in the Western United States (Doe, 1978), and metals for the Minnie Moore mine were thought to be derived from the enclosing host rocks by lateral secretion (Hall and others, 1978). Regression of $^{207}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ data yields isochron ages of 2,500 and 1,600–1,400 Ma for the primary source material (Small, 1968; Doe and Delevaux, 1985). Sulfur isotope data suggest that sulfur was derived from crustal sources such as syngenetic sulfide minerals in sedimentary rocks (Howe and Hall, 1985). Hydrogen- and oxygen-isotope studies suggest that the hydrothermal fluid that transported the metals was meteoric water driven by heat from igneous intrusions (Criss and others, 1984, 1985; Howe and Hall, 1985). Although these studies suggest some of the mineralizing processes involved, they fail to account for the complexity of different ore deposit types and ages that have been identified by recent mineral deposit studies. For example, a classification of deposit types in the Challis quadrangle includes 23 types of known hydrothermal deposits (Fisher and Johnson, 1987). Further, dating of hydrothermal minerals shows at least five periods of mineralization in Late Cretaceous to Eocene time alone (Snee and Kunk, 1989). This complexity requires that present and future studies focus in detail on the major types of deposits and geologic settings.

The goal of the present study is to extend earlier isotopic studies to include much of central Idaho and to investigate in more detail the sources of lead as a function of mineral deposit type, host-rock lithology, and geologic terrane. We present new analyses for 59 samples of vein material, much of which contains galena, and metalliferous argillitic whole rock and discuss the implications of these data in combination with previously reported results of Doe and Delevaux (1985), Small (1968), Davis (1978), Hall and others (1978), Sanford and others (1989), and Doe and Sanford (this volume). These data show that lead-isotope groups A and B of Doe and Delevaux (1985) and Doe and Sanford (this volume) can be divided into six subgroups that are related to distinct source-rock terranes and ore-deposit types. (Group B has no relation to the B type of early workers (Holmes, 1946; Houtermans, 1946; Russell and Farquhar, 1960)). Further, the lateral-secretion model for shale-hosted ore deposits (Hall and others, 1978) is revised in light of the new data.

The geology and ore deposits of the study area are described elsewhere in this volume, but a brief review is given here. The study area is underlain by eight discrete geologic terranes (fig. 1), all of which, except the Miocene volcanic rock terrane, contain significant mineral deposits (Fisher and Johnson, 1987; Worl and Johnson, 1989, this volume). The Paleozoic and Mesozoic terranes generally have parallel northwest trends. Precambrian and Paleozoic shelf carbonate terranes are dominant in the northeast. The next terrane to the southwest is the Mississippian flysch terrane representing foreland-basin deposits. The next terrane to the southwest, the Paleozoic black shale terrane, represents deeper water basinal deposits. The western third of the study area is underlain by the Cretaceous intrusive rock terrane of the Idaho batholith. Tertiary igneous rock terranes are distributed throughout the area. The Eocene extrusive rock terrane forms an irregular north-trending belt in the center of the area, and most, but not all, of the Eocene intrusive rock terrane is west of this belt. The Miocene volcanic rock terrane is restricted to the southwestern part of the study area.

The ore deposit types studied here are mainly epigenetic polymetallic veins in various rock types, replacement polymetallic sulfide deposits in carbonate rocks, and syngenetic massive sulfide deposits in black shale and argillite. As we will show, the relationships between ore deposit types and lead isotopes are complex but systematic. Initially, field criteria were used to group samples, but the lead-isotope results indicated further refinements in classification. For example, vein sulfides originally were distinguished from bedded, possibly syngenetic, sulfides based on macroscopic textural features. Later, we found that the vein sulfides required further subdivision based on systematic isotopic variations according to geographic area and geologic terrane. Similarly, we found that bedded sulfides include both epigenetic and syngenetic material. Our final classification of six lead-isotope types reflects both field criteria and isotopic characteristics.

Table 1 provides a summary of the results and a guide through the subsequent presentation of data, particularly the relationships among lead-isotope types, lithostratigraphic units, and geologic terranes. Divisions based on lead-isotope

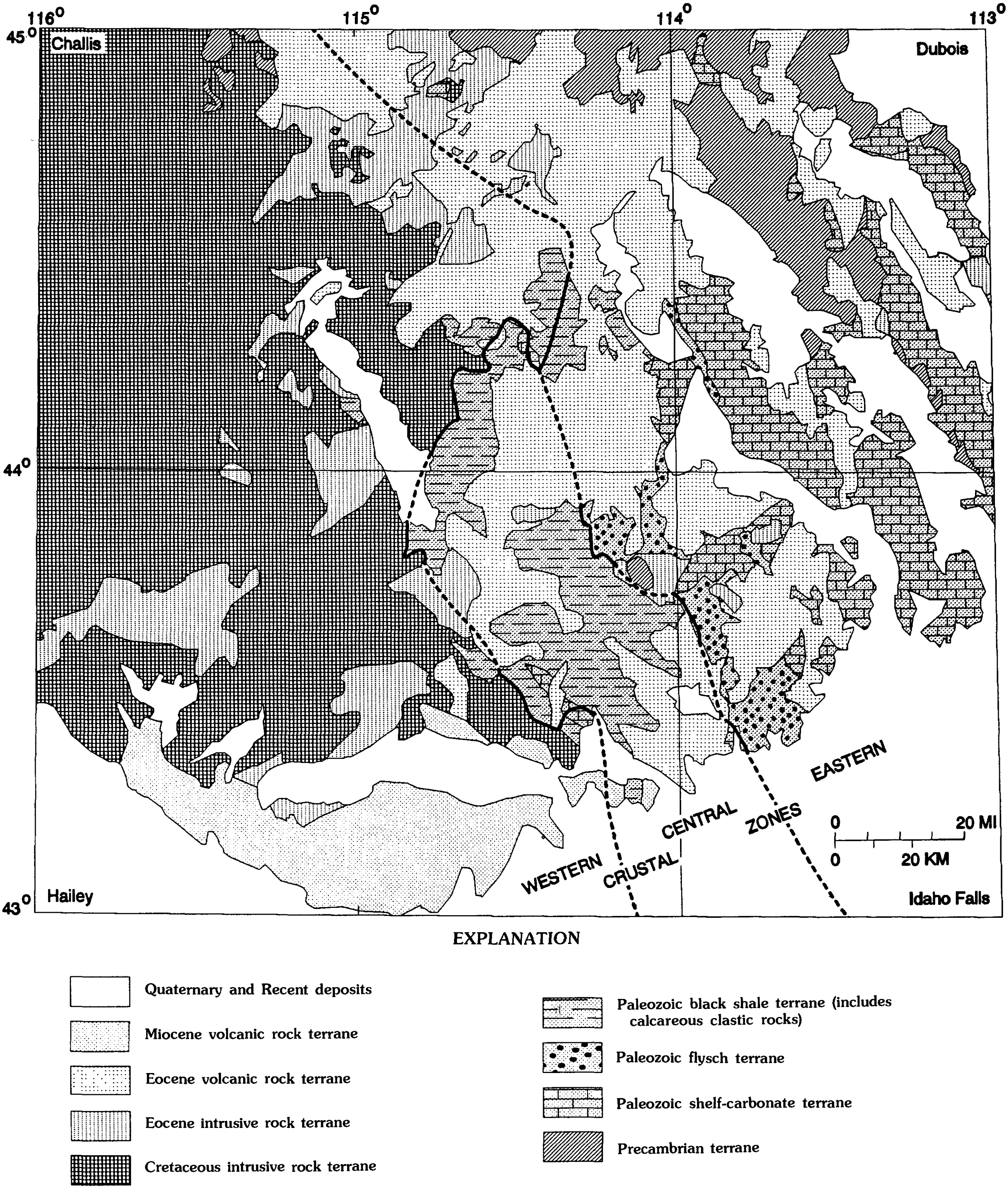


Figure 1. Map showing geologic terranes in the area of the Hailey and Challis 1°×2° quadrangles and the western parts of the Dubois and Idaho Falls 1°×2° quadrangles, central Idaho. Heavy solid and dashed lines delineate western, central, and eastern crustal zones. Modified from Rember and Bennett (1979); Worl and Johnson (1989); A.B. Wilson (unpublished compilation, 1991).

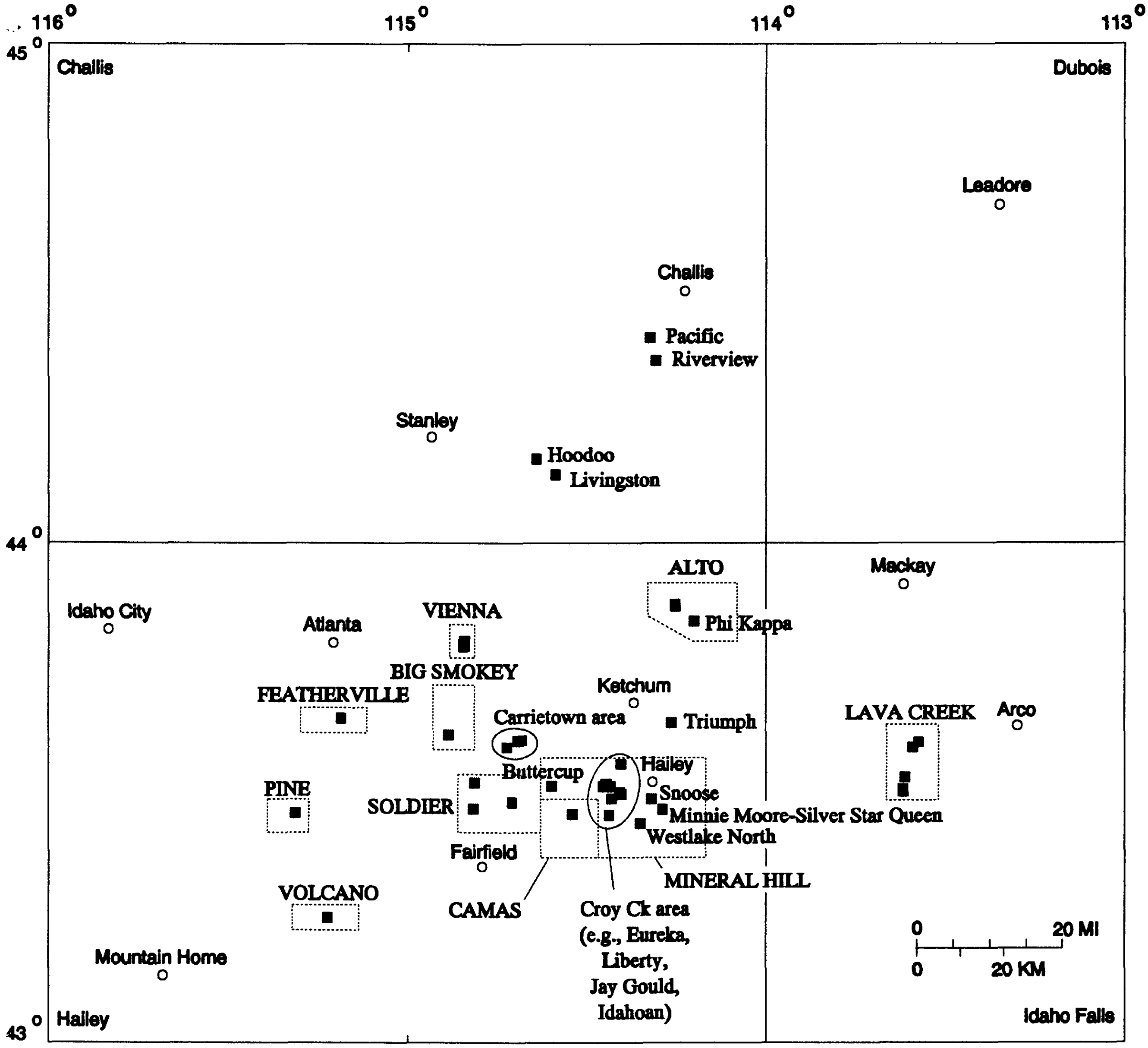


Figure 2. Map showing locations cited in text, central Idaho. Mining districts are shown by dotted lines; names are in capital letters. Selected mines (solid squares) are shown.

ratios are shown across the top, and divisions based on geology are shown down the left side. Group A and B leads are divided into six types. Group A lead is divided into Carrietown-, Minnie Moore-, Pacific-, and Lava Creek-types, and Group B lead is divided into black shale- and Triumph-types. Except for the black shale type, the names correspond to important mines or districts that exhibit the particular type of lead in sulfides. The black shale type is named for the argillitic rocks, which generally exhibit a characteristic lead-isotope signature.

Table 1 shows the isotopic characteristics of rocks (whole rock and potassium feldspar), epigenetic vein and bedded sulfides, and possibly syngenetic sulfides. Sulfides are listed according to the enclosing host rock. Almost all rocks (Paleozoic argillite and Cretaceous intrusive rocks) have black shale-type lead, except for Eocene rocks, which exhibit a range of lead-isotope types. Epigenetic sulfides have lead-isotope types that differ with crustal zone. Epigenetic sulfides in the western crustal zone have mostly Carrietown-type lead, those in the central crustal zone have Minnie Moore-, Pacific-, black shale-, and Triumph-types of lead, and those in the eastern crustal zone have mainly Lava Creek-type lead. Syngenetic sulfides are present only in the black shale terrane and have only Triumph-type lead.

Table 1. Summary of relationships among lead-isotope types, crustal zones, geologic terranes, and lithostratigraphic units in central Idaho.

[Abbreviations: R, r, whole rock and potassium feldspar; E, e, epigenetic vein and bedded sulfide; S, syngenetic sulfide. Capital letter indicates that interpretation is based on a larger number of samples]

	Group A			Group B		Group A
Lead-isotope type	Carrie-town	Minnie Moore	Pacific	Triumph	Black shale	Lava Creek
Western crustal zone						
Eocene volcanic rock terrane: Eocene volcanic rock, central Idaho batholith	R					r
Eocene intrusive rock terrane						
Eocene granite, central Idaho batholith	R, e					R
Eocene granite, southeast Idaho batholith	r					
Cretaceous intrusive rock terrane						
Cretaceous intrusive rocks, central Idaho batholith	E		r	e	R	E, r
Cretaceous intrusive rocks, Vienna district, Idaho batholith	E					
Cretaceous intrusive rocks, southeast Idaho batholith	E				r	e
Cretaceous intrusive rocks, south-central Idaho batholith	E					
Paleozoic carbonate roof pendants in Idaho batholith	E					
Black shale terrane						
Pennsylvanian-Permian argillite, Dollarhide Formation, Carrietown-Westlake area	E				R	
Ordovician carbonate rocks, Ella Dolostone group	E	e				
Central crustal zone						
Black shale terrane						
Pennsylvanian-Permian argillite, Dollarhide Formation, Croy Creek area		E			R	
Pennsylvanian-Permian argillite, Grand Prize Formation				e		
Pennsylvanian-Permian argillite, Wood River Formation				E	e	
Devonian argillite, part of Paleozoic Salmon River assemblage				E, S	E, r	
Devonian argillite, Milligen Formation, (except Triumph block)		E		E, S	R, e	
Devonian argillite, Milligen Formation, Triumph block		e		E, S	r	
Cambrian Ordovician carbonate rocks and phyllite, Bayhorse Dolostone group		e	E	e		
Eastern crustal zone						
Eocene extrusive rock terrane: Eocene volcanic rock, Lava Creek district						E
Flysch terrane: Mississippian argillite, Copper Basin Formation					r	E
Shelf carbonate terrane: Devonian carbonate rocks, Jefferson Dolostone						E
Precambrian terrane: Middle Proterozoic Swauger Formation						E

The five pre-Tertiary terranes closely correspond to the lead-isotope types identified here (table 1, fig. 1). The broad western crustal zone consists of the Cretaceous intrusive rock terrane and a small part of the black shale terrane. The narrow central crustal zone corresponds to almost all of the black shale terrane. The broad eastern crustal zone corresponds to the flysch, shelf carbonate, and Precambrian terranes. The three Tertiary rock terranes are either unmineralized or exhibit several lead-isotope types in sulfides. Isotopically, the Tertiary terranes are in crustal zones, and apparently the major control on lead-isotope types was the preexisting, pre-Tertiary crust beneath the Tertiary volcanic cover.

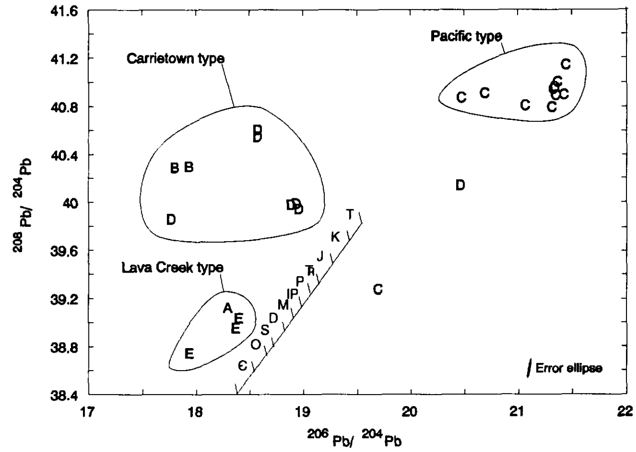
Acknowledgments.—We thank Reed Lewis, Paul Link, Falma Moye, Sandy Soulliere, Craig Wavra, and Anna Wilson for collecting some of the samples analyzed in this study. Constructive comments by Brian Mahoney, Betty Skipp, Gary Winkler, and Ron Worl improved the manuscript. Anna Wilson's compilations were invaluable for constructing figure 1. Larry Dee, U.S. Bureau of Land Management, provided much assistance in the field. Stan Church, Anna Wilson, and Paul Link gave particularly thoughtful reviews of the manuscript.

METHODS

Sample material includes galena from epigenetic crosscutting and bedding-conformable veins, one sample each of siderite vein and copper sulfide vein without visible galena, and several whole-rock samples of pyrite-rich argillite. Sulfide samples were prepared by handpicking a few milligrams of sulfide material and dissolving it in 1–2 mL of concentrated HNO_3. After 12 hours, this solution was diluted to about 15 mL total volume with distilled water and heated at 50°–80°C for another 12 hours. Approximately 10 µL of this final solution was loaded directly on a bed of silica gel-H_3PO_4 that had been partly dried on a single rhenium filament. Whole-rock shale samples were treated with concentrated HNO_3 to dissolve sulfides and then given a normal HF dissolution treatment for the silicate residue. The total sample was centrifuged to remove graphite, and then lead was separated by anion exchange using HBr as the medium. Loading of all the samples for mass spectrometry followed the silica gel-H_3PO_4 procedure described above. All samples were analyzed for lead-isotope composition by simultaneous collection of all four isotopes (^{204}Pb, ^{206}Pb, ^{207}Pb, and ^{208}Pb) on multicollector Finnigan-MAT mass spectrometers. Isotopic ratios were corrected for thermal fractionation by 0.11 percent per mass unit based on average values of numerous analyses of the standards, NBS–981 and –982, determined at the same operating conditions as samples. Precision at the 95 percent confidence level is approximately 0.04 percent per mass unit difference (that is, 0.08 percent for ^{206}Pb/^{204}Pb, 0.12 percent for ^{207}Pb/^{204}Pb, and 0.16 percent for ^{208}Pb/^{204}Pb), and the correlation of errors is about 0.97. Regressions were performed according to the method of York (1969) using the program ISOPLOT (Ludwig, 1990). Our new data, as well as previously published data, corrected for different standards where necessary, are listed in table 2.

Results are presented by age of lithostratigraphic unit starting with the Precambrian (table 2, figs. 3–8). Veins are discussed according to the host rock in which they are found. The clustering of lead-isotope ratios into recurring characteristic types will become apparent during this analysis. The data presentation concludes with a summary of the six lead-isotope types. In the discussion, the lead-isotope types are related to geologic terranes, the three crustal zones are identified, and the lead-isotope types are related to ore deposit types.

The variation in ^{208}Pb/^{204}Pb is moderately independent of ^{206}Pb/^{204}Pb in sulfide and whole-rock samples and has been successfully used to discriminate between source terranes (see, for example, Zartman, 1974). The ^{207}Pb/^{204}Pb ratio better correlates with ^{206}Pb/^{204}Pb than with ^{208}Pb/^{204}Pb, but it can also be used to discriminate between source terranes. Because the ^{208}Pb/^{204}Pb-^{206}Pb/^{204}Pb diagrams provide more resolution in discriminating between lead-isotope type than the ^{207}Pb/^{204}Pb-^{206}Pb/^{204}Pb diagrams, the data are first presented on a series of ^{208}Pb/^{204}Pb-^{206}Pb/^{204}Pb diagrams (figs. 3–8). Further subdivision of types is accomplished using ^{207}Pb/^{204}Pb-^{206}Pb/^{204}Pb diagrams (figs. 9, 10). (The terms "low," "high," "intermediate," and other similar terms for isotopic ratios in this paper are used to describe *relative* values within the range of ratios observed in this data set.)



EXPLANATION

A Middle Proterozoic Swauger Formation quartzite host rock, sulfide sample

B Paleozoic carbonate roof pendants in Idaho batholith, sulfide samples

C Cambrian or Ordovician carbonate host rock, Bayhorse Dolostone group, sulfide samples

D Ordovician carbonate and phyllite host rock, Ella Dolostone group, sulfide samples

E Devonian carbonate host rock, Jefferson Dolostone, sulfide samples

Figure 3. Lead-isotope ratios for samples of galena in veins hosted by Precambrian and lower Paleozoic quartzite and carbonate rocks, central Idaho. Average shale growth curve of Godwin and Sinclair (1982) is also shown.

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

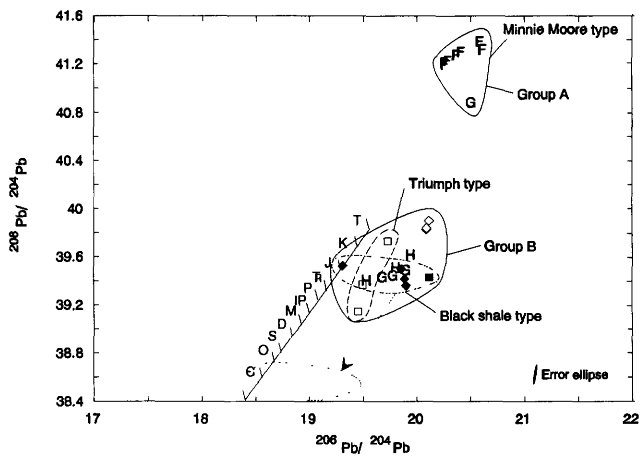
PRECAMBRIAN AND LOWER PALEOZOIC QUARTZITE AND CARBONATE ROCKS

Lead-isotope ratios for galena in an epithermal vein in Precambrian quartzite and in several replacement deposits in lower Paleozoic carbonate rocks show three main clusters that belong to Lava Creek-, Carrietown-, and Pacific-types (fig. 3).

Lead-isotope ratios for galena in a vein in the Middle Proterozoic Swauger Formation (A) and in replacement deposits in the Upper Devonian Jefferson Dolostone (E) form a cluster that has ^{206}Pb/^{204}Pb=17.9–18.4 and ^{208}Pb/^{204}Pb=38.7–39.1. This cluster partly defines the Lava Creek type (fig. 3). It has ^{206}Pb/^{204}Pb ratios similar to Carrietown-type lead but lower ^{208}Pb/^{204}Pb ratios.

Replacement and vein galena in Paleozoic carbonate roof pendants in the Idaho batholith (B) and in the Ella dolostone group (D), which we define to include the Ordovician Ramshorn slate, Middle Ordovician Ella Dolostone, and Middle Ordovician to Lower Silurian Saturday Mountain Formation, forms a cluster of low ^{206}Pb/^{204}Pb and high ^{208}Pb/^{204}Pb ratios, ^{206}Pb/^{204}Pb=17.8–19.0 and ^{208}Pb/^{204}Pb=39.9–40.6, and belongs to the Carrietown type.

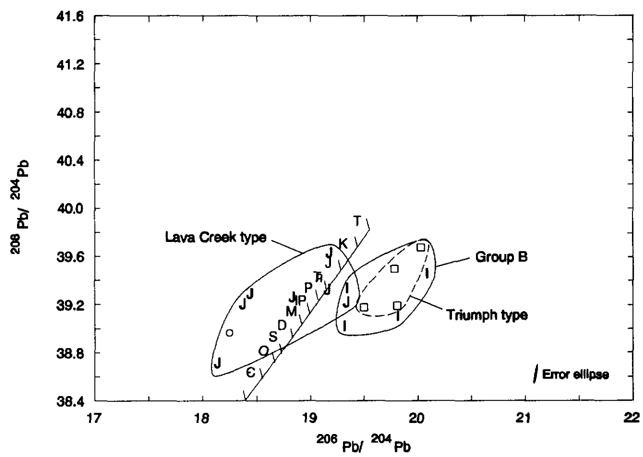
Galena from replacement deposits in the Upper Cambrian or Ordovician Bayhorse Dolostone (C)—that is, from



EXPLANATION

- F Devonian argillite host rock, Milligen Formation, Minnie Moore mine group, sulfide samples
- G Devonian argillite host rock, Milligen Formation, Triumph block, epigenetic sulfide samples
- □ Devonian argillite host rock, Milligen Formation, Triumph block, syngenetic sulfide samples
- ■ Devonian argillite, Milligen Formation, Triumph block, whole-rock samples
- H Devonian argillite host rock, Milligen Formation, epigenetic sulfide samples
- ◇ Devonian argillite host rock, Milligen Formation, syngenetic sulfide samples
- ◆ Devonian argillite, Milligen Formation, whole-rock samples

Figure 4. Lead-isotope ratios for samples of whole rock and galena from Devonian Milligen Formation argillite, central Idaho. Average shale growth curve of Godwin and Sinclair (1982) is also shown. Arrow indicates possible initial ratios for argillite host rocks assuming average shale concentrations of uranium and thorium.



EXPLANATION

- I Devonian argillite host rock, part of Salmon River assemblage, epigenetic sulfide samples
- □ Devonian argillite host rock, part of Salmon River assemblage, syngenetic sulfide samples
- J Mississippian argillite host rock, Alto-Muldoon-Lava Creek area, crosscutting vein sulfide samples
- ○ Mississippian argillite host rock, Phi Kappa mine, bedded sulfide sample

Figure 5. Lead-isotope ratios for samples of vein and syngenetic galena from argillite from the Devonian part of the Paleozoic Salmon River assemblage and from the Mississippian Copper Basin Formation, central Idaho. Average shale growth curve of Godwin and Sinclair (1982) is also shown.

the Pacific and Riverview mining areas (Davis, 1978)—has radiogenic lead, $^{206}Pb/^{204}Pb$=20.5–21.5 and $^{208}Pb/^{204}Pb$=40.8–41.1, that we designate Pacific type (fig. 3). Together with a sample from the Wilbert mine, these samples are from what we call the Bayhorse dolostone group.

DEVONIAN ARGILLITIC ROCKS

Samples of Devonian argillite are from the Milligen Formation (fig. 4) and from the Devonian part of the Paleozoic Salmon River assemblage (Link and others, this volume) (fig. 5). The argillite of the Milligen in the Triumph block is plotted separately, as discussed following. Samples from the Milligen Formation include (1) sulfides from epigenetic veins in the Silver Star Queen mine (Minnie Moore group of mines) (F), Triumph block (G), and other localities in the black shale terrane (H); (2) galena from conformable, possibly syngenetic sulfides of the Triumph (open squares) and Snoose mines (open diamonds); and (3) pyrite-rich argillite whole-rock samples from the Triumph block (filled square) and other parts of the black shale terrane (filled diamonds) (fig. 4). Samples from the Devonian part of the Salmon River assemblage include galena from epigenetic veins (I) and from conformable, possibly syngenetic sulfide ore (open squares) (fig. 5). Some of the ore (the "complex ores") from the Triumph mine is probably syngenetic (Thor Kiilsgaard, oral commun., 1988; Turner and Otto, this volume). Similar deposits, namely the Livingston, Hoodoo, and Snoose deposits, may also contain syngenetic sulfides, but the evidence is less definitive (Hall, 1985; Link and others, this volume). For the purposes of this paper, these four deposits are referred to as syngenetic. A previously reported analysis of whole-rock argillite from the Devonian part of the Salmon River assemblage (Holman Creek sample T–270) is anomalously radiogenic due to uranium enrichment and plots off the diagram (fig. 5) to the right.

Lead isotopes in sulfides and in whole rocks from Milligen and Salmon River argillites form two discrete clusters belonging to group A (Minnie Moore type) and group B (Triumph- and black shale-types) (figs. 4, 5). Minnie Moore-type lead includes the cluster having $^{206}Pb/^{204}Pb$=20.2–20.6 and $^{208}Pb/^{204}Pb$=40.8–41.4, and Triumph- and black shale-type leads include the cluster having $^{206}Pb/^{204}Pb$=19.2–20.2 and $^{208}Pb/^{204}Pb$=39.2–40.0. Lead isotopes for epithermal vein galena (F, fig. 4) in the Minnie

Table 2. Lead-isotope data for central Idaho.

[Letter or other symbol in parentheses following sample description indicates plotting symbol (figs. 3–8). Lead-isotope types ("Type"): BS, black shale, CT, Carrietown; LC, Lava Creek; MM, Minnie Moore; PA, Pacific; TR, Triumph. Host-rock types ("Host"): Dj, Devonian Jefferson Dolostone; Dm, Devonian Milligen Formation; Dmt, Milligen Formation in the Triumph block; Kap, Cretaceous aplite porphyry; Kgd, Cretaceous granodiorite; Kgdh, Cretaceous hornblende granodiorite; Kgp, Cretaceous granite porphyry, Kqd, Cretaceous quartz diorite; Mc, Mississippian Copper Basin Formation; Mm, Mississippian McGowan Creek Formation; Ob, Cambrian or Ordovician Bayhorse Dolostone; Oe, Ordovician Ella Dolostone; Ok, Ordovician Kinnikinic Quartzite; Or, Ordovician Ramshorn Slate; SOsm, Ordovician and Silurian Saturday Mountain Formation; PPd, Pennsylvanian-Permian Dollarhide Formation, undivided; Pdu, Permian upper Dollarhide; Pdm, Permian middle Dollarhide; PPdl, Pennsylvanian-Permian lower Dollarhide; PPw, Pennsylvanian-Permian Wood River Formation; Pzl, Paleozoic carbonate roof pendants in Idaho batholith, undivided; Pzsr, Paleozoic Salmon River assemblage; Tc, Eocene Challis Volcanic Group; Tcr, Eocene Challis Volcanic Group, rhyolite; Tg, Eocene granite; Ys, Middle Proterozoic Swauger Formation. Column labeled "Association" indicates whether sulfide minerals are epigenetic (Epi), possibly syngenetic (Syn), or remobilized syngenetic (Syn-rmbl) based on textural relations or, in the case of whole-rock samples, the ore-deposit type associated with the host rock. Abbreviations for ore-deposit types: combinations of Au-Ag-Pb-Zn-Cu-(Mo), precious- and base-metal veins possibly associated with molybdenum porphyry; (Sn), inferred tin granite based on chemistry and mineralogy of granite; Mo, molybdenum porphyry; U, uranium. Analytical methods: Gel, silica gel; Mth, tetramethyl lead; 3Fil, triple filament. References: (1) Small, 1968; (2) Doe and Rohrbough, 1977; (3) Davis, 1978; (4) Hall and others, 1978; (5) Doe and Sanford, this volume); (6) Sanford and others, 1989, and this work. Data of Small (1968) have been corrected by multiplying the original $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ ratios by 0.99324, 0.99067, and 0.98910, respectively; other data are as published]

Sample no.	Mine or location	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Type	Material	Host	Association	Mining district	County	Method	Refs.
Middle Proterozoic Swauger quartzite host rock, sulfide (A) (fig. 3)												
591	Blue Lead	18.304	15.643	39.125	LC	Galena	Ys	Epi	Junction	Lemhi	Mth	1, 2
Paleozoic carbonate roof pendants in Idaho batholith, sulfide (B) (fig. 3)												
K52-1	Mountain King	17.810	15.593	40.290	CT	Galena	Pzl	Epi	Sheep Mountain-Vienna	Custer	Gel	5
R-154	Mountain King	17.943	15.606	40.299	CT	Galena	Pzl	Epi	Sheep Mountain-Vienna	Custer	Gel	5
Ordovician carbonate host rock, Bayhorse Dolostone group, sulfide (C) (fig. 3)												
D6S3	Pacific	21.452	15.840	41.145	PA	Galena?	OCb	Epi	Bayhorse	Custer	Gel	5
KD68	Pacific Mountain-dump	21.332	15.771	40.948	PA	Galena	OCb	Epi	Bayhorse	Custer	Gel	3
KD69	Pacific Mountain-main ore body	21.431	15.767	40.899	PA	Galena	OCb	Epi	Bayhorse	Custer	Gel	3
KD72-9-124	Pacific Mountain-above main ore body	21.323	15.765	40.796	PA	Galena	OCb	Epi	Bayhorse	Custer	Gel	3
KD73-4-161	Pacific Mountain-top, main ore body	21.347	15.801	40.966	PA	Galena	OCb	Epi	Bayhorse	Custer	Gel	3
KD74-P9-5B	Pacific Mountain	21.375	15.807	41.004	PA	Galena	OCb	Epi	Bayhorse	Custer	Gel	3
KD74-P9-5B	Pacific Mountain	21.355	15.777	40.897	PA	Galena	OCb	Epi	Bayhorse	Custer	Gel	3
SWH 52-67	Rattlesnake prosp.	20.479	15.784	40.874	PA	Galena	OCb	Epi	Bayhorse	Custer	Gel	5
1885 Ag	Riverview	21.073	15.804	40.808	PA	Galena?	OCb	Epi	Bayhorse	Custer	Gel	5
KD-7	Riverview	19.704	15.806	39.281	TR	Galena	OCb	Epi	Bayhorse	Custer	Gel	3
607	Wilbert	20.692	15.957	40.914	MM	Galena	Ok-SOsm	Epi	Dome	Butte	Mth	1, 2
Ordovician carbonate and phyllite host rock, Ella Dolostone group, sulfide (D) (fig. 3)												
79T 23M	Ramshorn	17.776	15.578	39.861	CT	Galena	Or	Epi	Bayhorse	Custer	Gel	5
572	Clayton Silver	18.895	15.736	39.982	CT	Galena	Oe	Epi	Bayhorse	Custer	Mth	1, 2
80WH-1	Clayton Silver	18.944	15.709	39.990	CT	Galena	Oe	Epi	Bayhorse	Custer	Gel	5
SWH 78-65	Clayton Silver	18.965	15.718	39.948	CT	Galena	Oe	Epi	Bayhorse	Custer	Gel	5
SWH120-79	Rob Roy	20.474	15.870	40.143	MM	Galena	Oe	Epi	Bayhorse	Custer	Gel	5
SWH154-76a	Dryden	18.586	15.706	40.608	CT	Galena	SOsm	Epi	Bayhorse	Custer	Gel	5
SWH154-76b	Dryden	18.584	15.690	40.549	CT	Galena	SOsm	Epi	Bayhorse	Custer	Gel	5

Devonian carbonate host rock, Jefferson Dolostone, sulfide (E) (fig. 3)												
590	Gilmore	17.947	15.556	38.746	LC	Galena	Dj	Epi	Gilmore-Texas	Lemhi	Mth	1, 2
80R-1	Hilltop	18.376	15.573	38.955	LC	Galena?	Dj	Epi	Gilmore-Texas	Lemhi	Gel	5
80R-1	Hilltop	18.407	15.606	39.039	LC	Galena?	Dj	Epi	Gilmore-Texas	Lemhi	Gel	5
Devonian argillite host rock, Milligen Formation, Minnie Moore mine group, sulfide (F) (fig. 4)												
578	Silver Star Queen	20.622	16.021	41.320	MM	Galena	Dm	Epi	Mineral Hill	Blaine	Mth	1, 4
579	Silver Star Queen	20.378	15.978	41.275	MM	Galena	Dm	Epi	Mineral Hill	Blaine	Mth	1, 4
580	Silver Star Queen	20.597	16.018	41.387	MM	Galena	Dm	Epi	Mineral Hill	Blaine	Mth	1, 4
581	Silver Star Queen	20.275	15.937	41.199	MM	Galena	Dm	Epi	Mineral Hill	Blaine	Mth	1, 4
581	Silver Star Queen	20.298	15.957	41.229	MM	Galena	Dm	Epi	Mineral Hill	Blaine	Mth	1, 4
581	Silver Star Queen	20.267	15.946	41.215	MM	Galena	Dm	Epi	Mineral Hill	Blaine	Mth	1, 4
WH70-16	Silver Star Queen	20.421	15.962	41.301	MM	Galena	Dm	Epi	Mineral Hill	Blaine	3Fil	2, 4
Devonian argillite host rock, Milligen Formation, Triumph block, epigenetic sulfide (G) (fig. 4)												
8H 91A	Courier	19.910	15.863	39.493	TR	Galena	Dmt	Epi	Warm Spring	Blaine	Gel	6
8H 92C	Independence	19.789	15.820	39.450	TR	Galena	Dmt	Epi	Warm Spring	Blaine	Gel	6
WH74-2B	North Star	19.688	15.906	39.438	TR	Galena	Dmt	Epi	Mineral Hill	Blaine	3Fil	2, 4
575	Triumph	20.515	16.086	40.886	MM	Galena	Dmt	Epi	Warm Spring	Blaine	Mth	1, 2
Devonian argillite host rock, Milligen Formation, Triumph block, syngenetic sulfide (open square) (fig. 4)												
8H112A	Triumph	19.730	15.886	39.729	TR	Galena	Dmt	Syn	Warm Spring	Blaine	Gel	6
8H112B	Triumph	19.454	15.847	39.145	TR	Galena	Dmt	Syn	Warm Spring	Blaine	Gel	6
8H112C1	Triumph	19.493	15.852	39.370	TR	Galena-sphalerite	Dmt	Syn	Warm Spring	Blaine	Gel	6
Devonian argillite, Milligen Formation, Triumph block, whole rock (filled square) (fig. 4)												
8H 98	Ida Harlan	20.118	15.835	39.431	BS	Whole rock, pyrite in argillite	Dmt	Ag-Pb-Zn	Warm Spring	Blaine	Gel	6
Devonian argillite host rock, Milligen Formation, epigenetic sulfide (H) (fig. 4)												
T597M	Crest Lake?-Trail Creek?	19.952	15.797	39.619	BS	Galena	Dm	Epi	Warm Spring	Blaine	Gel	5
8H170	Lost Dump	19.545	15.826	39.408	TR	Galena	Dm	Epi	Warm Spring	Blaine	Gel	6
8H 31	Memorial prospect	19.817	15.829	39.512	TR	Galena	Dm	Epi	Mineral Hill	Blaine	Gel	6
Devonian argillite host rock, Milligen Formation, syngenetic sulfide (open diamond) (fig. 4)												
8H 76C	Snoose	20.117	15.852	39.900	TR	Galena	Dm	Syn?	Mineral Hill	Blaine	Gel	6
8H 76D	Snoose	20.090	15.835	39.828	TR	Galena	Dm	Syn?	Mineral Hill	Blaine	Gel	6
8H 76I	Snoose	20.092	15.836	39.840	TR	Galena	Dm	Syn?	Mineral Hill	Blaine	Gel	6
Devonian argillite, Milligen Formation, whole rock (filled diamond) (fig. 4)												
8H 29	Roadside	19.887	15.753	39.420	BS	Whole rock, pyrite in phyllite	Dm	Ag-Pb-Zn	Mineral Hill	Blaine	Gel	6
8H 28A	Slaughterhouse Creek	19.857	15.789	39.500	BS	Whole rock, quartz and pyrite in phyllite	Dm	Ag-Pb-Zn	Mineral Hill	Blaine	Gel	6
8L625	Slaughterhouse Creek	19.903	15.784	39.363	BS	Whole rock, pyrite in argillite	Dm	Ag-Pb-Zn	Mineral Hill	Blaine	Gel	6
WH2112	"Sawmill Gulch"	19.308	15.673	39.526	BS	Whole rock, argillite	Dm	Ag-Pb-Zn	Warm Springs	Blaine	Gel	2, 4
Devonian argillite host rock, part of Salmon River assemblage, epigenetic sulfide (I) (fig. 5)												
R-366	Empire	19.831	15.673	39.111	BS	Pyrrhotite, sphalerite, and scheelite	Pzsr	Epi	Washington Basin	Custer	Gel	5
T576M	Patti Flynn	19.333	15.809	39.026	TR	Boulangerite and jamesonite	Pzsr	Epi	Fourth of July	Custer	Gel	5
T572M	Timberline	20.099	15.772	39.460	BS	Jamesonite	Pzsr	Epi	Fourth of July	Custer	Gel	5
SWH	Twin Apex	19.357	15.753	39.341	TR	Galena	Pzsr	Epi	Bayhorse	Custer	Gel	5

Table 2. Lead-isotope data for central Idaho—Continued.

Sample no.	Mine or location	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Type	Material	Host	Association	Mining district	County	Method	Refs.
Devonian argillite host rock, part of Salmon River assemblage, syngenetic sulfide (open square) (fig. 5)												
R-13	Hoodo	19.810	15.795	39.192	TR	Barite vein	Pzsr	Syn	Boulder Creek-Slate Creek	Custer	Gel5	
S-16	Hoodo	19.502	15.859	39.177	TR	Galena	Pzsr	Syn-Epi	Boulder Creek-Slate Creek	Custer	Gel	5
80T-1	Livingston	20.030	15.833	39.673	TR	Galena and jamesonite	Pzsr	Syn-rmbl	Boulder	Custer	Gel	5
80T-2	Livingston	19.779	15.786	39.499	TR	Galena	Pzsr	Syn-rmbl	Boulder	Custer	Gel	5
Devonian argillite, part of Salmon River assemblage, whole rock (plots off diagram)												
T-270	Holman Creek-west	24.073	15.954	39.145	BS	Whole rock, argillite	Pzsr	Ag-Pb-Zn	Bayhorse	Custer	Gel	5
Mississippian argillite host rock, Alto-Muldoon-Lava Creek area, crosscutting vein sulfide (J) (fig. 5)												
8H159	Alto	19.175	15.766	39.329	LC	Galena	Mc	Epi	Alto	Custer	Gel	6
8H160	Alto-south	19.191	15.842	39.636	LC	Galena	Mc	Epi	Alto	Custer	Gel	6
7M104	Horseshoe-west	18.153	15.601	38.718	LC	Galena	Mc	Epi	Alder Creek	Custer	Gel	6
7M130	Mackinaw group	18.452	15.702	39.295	LC	Galena	Mc	Epi	Copper Basin	Custer	Gel	6
8H 7A	Muldoon Barite	19.353	15.812	39.222	LC	Galena?	Mc	Epi	Little Wood River	Blaine	Gel	6
7M128	Star Hope	18.384	15.680	39.204	LC	Copper ore	Mc	Epi	Copper Basin	Custer	Gel	6
8H157A	Leadbelt-southeast	18.842	15.788	39.264	LC	Galena	Mm	Epi	Lava Creek	Butte	Gel	6
Mississippian argillite host rock, Phi Kappa mine, bedded sulfide (open circle) (fig. 5)												
574	Phi Kappa	18.248	15.650	38.965	LC	Galena	Mc	Epi	Alto	Blaine	Mth	1, 2
Mississippian argillite host rock, Trail Creek area, whole rock (plots off diagram)												
T-602	Urexco	29.384	16.238	38.881	BS	Whole rock, argillite	Mc	U	Alto	Custer	Gel	5
Pennsylvanian-Permian argillite host rock, Wood River Formation, sulfide (K) (fig. 6)												
T 49M	Bulldozer cut/dump	19.692	15.808	39.383	TR	Galena	PPw	Epi	Boyle Mountain	Blaine	Gel	5
79T565M	Combination?-new vein	19.735	15.800	39.367	TR	Galena	PPw	Epi	Galena	Blaine	Gel	5
79T 5M	Crown Point	19.624	15.800	39.274	TR	Galena	PPw	Epi	Boulder Basin	Blaine	Gel	5
76T 32M	Emperium	19.785	15.788	39.311	BS	Galena	PPw	Epi	Germania Basin	Custer	Gel	5
79T580M	Golden Glow	19.767	15.844	39.677	TR	Galena	PPw	Epi	Boulder Basin	Blaine	Gel	5
8H165A	Golden Glow	19.763	15.831	39.575	TR	Galena	PPw	Epi	Warm Spring	Blaine	Gel	6
8H166	Golden Glow-west	19.746	15.828	39.503	TR	Galena	PPw	Epi	Warm Spring	Blaine	Gel	6
8H 97A	Homestake	19.966	15.910	39.866	TR	Galena	PPw-Dm	Epi	Warm Spring	Blaine	Gel	6
8H 90A	Ohio group	19.895	15.842	39.628	TR	Galena	PPw	Epi	Warm Spring	Blaine	Gel	6
W312A	White Cloud-east	19.404	15.732	39.358	BS	Galena	PPw	Epi	East Fork-Boulder Creek	Custer	Gel	5
Pennsylvanian-Permian argillite host rock, Dollarhide Formation, Carrietown-Westlake area, sulfide (L) (fig. 6)												
76T 23M	Dollarhide	18.081	15.572	39.918	CT	Galena	PPdl	Epi	Carrietown	Camas	Gel	5
T 81M	Horn Silver	18.441	15.652	39.735	CT	Galena	PPdl	Epi	Carrietown	Camas	Gel	5
577	Star	19.158	15.763	39.981	CT	Galena	PPdl	Epi	Mineral Hill	Blaine	Mth	1, 4
T 68M	Carrie Leonard	18.094	15.591	39.975	CT	Galena	PPdl	Epi	Carrietown	Camas	Gel	5
576	Buttercup	18.549	15.675	39.680	CT	Galena	PPd	Epi	Willow Creek	Camas	Mth	1, 2
576	Buttercup	18.808	15.730	39.850	CT	Galena	PPd	Epi	Willow Creek	Camas	Mth	1, 2
8H 5H	Buttercup	18.780	15.683	39.550	CT	Siderite and pyrite	PPd	Epi	Willow Creek	Camas	Gel	6
8H168	Buttercup	18.721	15.680	39.692	CT	Galena	PPd	Epi	Willow Creek	Camas	Gel	6

8H 53A	Hidden Treasure	18.412	15.669	39.690	CT	Galena	PPd	Epi	Little Smokey	Camas	Gel	6
8H 85	Westlake North	19.034	15.764	39.896	CT	Galena	Pdu	Epi	Mineral Hill	Blaine	Gel	6
Pennsylvanian-Permian argillite, Dollarhide Formation, Carrietown-Westlake area, whole rock (filled square) (fig. 6)												
8H167	Buttercup-above and north	19.182	15.697	39.061	BS	Whole rock, pyrite in argillite	PPd	Ag-Pb-Zn	Willow Creek	Camas	Gel	6
8L660	Willow Creek	19.474	15.713	39.208	BS	Whole rock, pyrite in argillite	PPd	Ag-Pb-Zn	Willow Creek	Camas	Gel	6
Pennsylvanian-Permian argillite host rock, Dollarhide Formation, Croy Creek area, crosscutting vein sulfide (M) (fig. 6)												
8H 96	Blumite?	20.363	15.960	40.515	MM	Galena	PPd	Epi	Mineral Hill	Blaine	Gel	6
8H 6A	Jay Gould	20.044	15.972	40.724	MM	Galena?	PPd	Epi	Mineral Hill	Blaine	Gel	6
A176	Liberty	20.013	15.917	40.619	MM	Galena	PPd	Epi	Mineral Hill	Blaine	3Fil	2, 4
8H 87A	Red Cloud	20.131	15.928	40.552	MM	Galena	PPd	Epi	Mineral Hill	Blaine	Gel	6
8H 88A	Red Cloud-west	19.951	15.901	40.507	MM	Galena	Pdm	Epi	Mineral Hill	Blaine	Gel	6
8H 86A	Argent	19.995	15.910	40.615	MM	Galena	Pdu	Epi	Mineral Hill	Blaine	Gel	6
WH70-13B	Eureka	20.220	15.910	40.904	MM	Galena	Pdu	Epi	Mineral Hill	Blaine	3Fil	2, 4
8H 79C	Idahoan	20.168	15.931	40.925	MM	Galena	Pdu	Epi	Mineral Hill	Blaine	Gel	6
8H 79G	Idahoan	20.171	15.921	40.924	MM	Galena	Pdu	Epi	Mineral Hill	Blaine	Gel	6
Pennsylvanian-Permian argillite host rock, Dollarhide Formation, Croy Creek area, bedded sulfide (open circle) (fig. 6)												
8H 79F	Idahoan	20.238	15.933	40.996	MM	Galena	Pdu	Syn?	Mineral Hill	Blaine	Gel	6
Pennsylvanian-Permian argillite, Dollarhide Formation, Croy Creek area, whole rock (filled diamond) (fig. 6)												
8H 79A	Idahoan	20.326	15.816	39.381	BS	Whole rock, pyrite in siltstone	Pdu	Ag-Pb-Zn	Mineral Hill	Blaine	Gel	6
8H 79E	Idahoan	19.787	15.788	39.413	BS	Whole rock, pyrite and sphalerite in siltstone	Pdu	Ag-Pb-Zn	Mineral Hill	Blaine	Gel	6
Pennsylvanian-Permian argillite host rock, Grand Prize Formation, sulfide (N) (fig. 6)												
8H163	Grand Prize #1	19.431	15.812	39.123	TR	Galena	PPg	Epi	East Fork	Custer	Gel	6
Cretaceous intrusive host rock, south-central Idaho batholith, sulfide (O) (fig. 7)												
8H 38A	Franklin	17.832	15.671	40.130	CT	Galena	Kap	Epi	Pine	Elmore	Gel	6
8H 19A	Blackstone	17.504	15.544	38.932	CT	Galena	Kgd	Epi	Volcano	Camas	Gel	6
8H 44A	Marsh Creek	17.076	15.469	39.066	CT	Galena	Kgd	Epi	Featherville	Elmore	Gel	6
Cretaceous intrusive host rock, Vienna district, Idaho batholith, sulfide (P) (fig. 7)												
8H110	Solace	18.353	15.680	39.843	CT	Galena	Kgd	Epi	Vienna	Blaine	Gel	6
8H111A	Solace-below	17.980	15.607	39.794	CT	Galena	Kgd	Epi	Vienna	Blaine	Gel	6
8H108A	Vienna	18.392	15.627	39.583	CT	Galena	Kgd	Epi	Vienna	Blaine	Gel	6
T302M	Webfoot	17.785	15.587	39.895	CT	Galena	Kgd	Epi	Vienna	Blaine	Gel	5
Cretaceous intrusive host rock, southeast Idaho batholith, sulfide (Q) (fig. 7)												
8H 48	"Big Smoky"-north	19.022	15.772	39.590	LC	Galena	Kgd	Epi	Big Smokey	Camas	Gel	6
8H 52A	Cannonball	18.123	15.605	39.939	CT	Galena	Kgd	Epi	Soldier	Camas	Gel	6
8H 47B	El Oro	18.712	15.681	39.411	LC	Galena	Kgd	Epi	Skeleton Creek	Camas	Gel	6
8H 51A	Five Points	18.263	15.634	39.402	CT	Galena	Kgd	Epi	Soldier	Camas	Gel	6
8H 50A	Richard Allen	19.189	15.773	40.038	CT	Galena	Kgd	Epi	Soldier	Camas	Gel	6
8H 4A	Princess Blue Ribbon	18.267	15.656	39.767	CT	Galena	Kgdh	Epi	Camas	Camas	Gel	6
8H 4C	Princess Blue Ribbon	18.171	15.626	39.614	CT	Galena	Kgdh	Epi	Camas	Camas	Gel	6
6H 41A	Texas Star	18.248	15.721	39.749	CT	Galena	Kgdh	Epi	Mineral Hill	Blaine	Gel	6
8H 10A	Utah-Bellevue-west	18.752	15.713	40.027	CT	Galena	Kgdh-Kqd	Epi	Mineral Hill	Blaine	Gel	6
8H162	Croesus	18.665	15.699	39.922	CT	Galena	Kqd	Epi	Mineral Hill	Blaine	Gel	6

Table 2. Lead-isotope data for central Idaho—Continued.

Sample no.	Mine or location	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Type	Material	Host	Association	Mining district	County	Method	Refs.
Cretaceous intrusive rock, southeast Idaho batholith, potassium feldspar (filled square) (fig. 7)												
A230	McCoy mine stock	19.572	15.762	39.309	BS	Potassium feldspar	Kqd	Au-Ag-Pb-Zn	Mineral Hill	Blaine	Gel	2, 4
Cretaceous intrusive host rock, central Idaho batholith, sulfide (R) (fig. 7)												
Buck	Buckskin Drill Road-Thompson Creek	17.577	15.583	40.875	CT	Galena	Kgd	Epi	Bayhorse	Custer	Gel	5
S76-895	Buckskin Drill Road?-Thompson Creek	17.617	15.594	41.014	CT	Galena	Kgd	Epi	Bayhorse	Custer	Gel	5
F112	Carlson Gulch-Grimes Creek	19.144	15.672	39.087	LC	Galena	Kgd	Epi	Grimes Pass	Boise	Gel	5
F107	Comeback	19.097	15.678	39.083	LC	Galena	Kgd	Epi	Grimes Pass	Boise	Gel	5
KR880	Deadwood	19.341	15.754	39.527	TR	Galena	Kgd	Epi	Dixie	Elmore	Gel	5
GHM-1	Greyhound	18.110	15.646	40.565	CT	Galena	Kgd	Epi	Seafoam	Custer	Gel	5
R-151	K.G.	17.990	15.578	39.286	CT	Galena	Kgd	Epi	Seafoam-Deadwood	Valley	Gel	5
F154	Seafoam Lake-west	17.747	15.604	39.765	CT	Galena	Kgd	Epi	Seafoam	Custer	Gel	5
R-155	Silver Bell	17.596	15.574	39.785	CT	Galena	Kgd	Epi	Seafoam	Custer	Gel	5
Cretaceous intrusive rocks, central Idaho batholith, whole rock and potassium feldspar (filled diamond) (fig. 7)												
KR743Kf	Deadwood	19.488	15.716	39.320	BS	Potassium feldspar	Kgd	Zn-Pb-Ag	Dixie	Elmore	Gel	5
LP-1	Lost Packer	19.768	15.742	40.078	PA	Whole rock	Kgd	Cu-Ag-(Mo)	Loon Creek	Custer	Gel	5
W320	Thompson Creek	20.508	15.811	40.329	BS	Potassium feldspar	Kgd	Mo	Bayhorse	Custer	Gel	5
W188	White Cloud-west	19.204	15.694	39.262	LC	Potassium feldspar	Kgp	Mo	East Fork	Custer	Gel	5
Eocene Challis Volcanic Group host rock, Lava Creek district, sulfide (S) (fig. 8)												
638	"Dry Fork Creek"	17.909	15.628	38.696	LC	Galena	Tc	Epi	Lava Creek	Butte	Mth	1, 2
634	Ella	19.166	15.819	39.638	LC	Galena	Tc	Epi	Lava Creek	Butte	Mth	1, 2
640	Hanni 2-Silver Bell?	17.290	15.579	38.508	LC	Galena	Tc	Epi	Lava Creek	Butte	Mth	1, 2
639	Hub	17.868	15.636	38.673	LC	Galena	Tc	Epi	Lava Creek	Butte	Mth	1, 2
8H161A	St. Louis	19.030	15.761	39.334	LC	Galena	Tc	Epi	Lava Creek	Butte	Gel	6
Eocene granite host rock, southeast Idaho batholith, sulfide (T) (fig. 8)												
76T 31	GAF, Big Smokey Creek	18.311	15.622	39.883	CT	Galena	Tg	Epi	Vienna	Blaine	Gel	5
Eocene Challis Volcanic Group, central Idaho batholith, whole rock (filled diamond) (fig. 8)												
LF-1	Little Falls	19.149	15.653	39.186	LC	Whole rock	Tcr	Mo	Bear Creek	Boise	Gel	5
Grey-1	Mt. Greylock	17.536	15.525	39.272	CT	Whole rock	Tcr	Au-Ag-(Mo)	Yankee Fork	Custer	Gel	5
GM-31	Parker Mountain	17.869	15.526	39.634	CT	Whole rock	Tcr	Au-Ag-(Mo)	Parker Mountain	Lemhi	Gel	5
RH-122	Red Mountain Dome	17.493	15.522	39.459	CT	Whole rock	Tcr	Au-Ag-(Mo)	Yankee Fork	Custer	Gel	5
GM-365	Singheiser	18.885	15.645	39.818	CT	Whole rock	Tcr	Ag-Au-(Mo)	Gravel Range	Lemhi	Gel	5
SB215120	Sunbeam	17.803	15.529	39.497	CT	Whole rock	Tcr	Au-Ag-(Mo)	Yankee Fork	Custer	Gel	5

Eocene granite host rock, central Idaho batholith, sulfide (V) (fig. 8)												
584-13	CUMO	19.039	15.656	38.965	LC	Galena	Kgd	Epi	Pioneerville	Boise	Gel	5
SR-1	Sunrise	17.473	15.562	40.348	CT	Galena	Tg	Epi	Yankee Fork	Custer	Gel	5
Eocene granite, central Idaho batholith, whole rock (filled square) (fig. 8)												
KR 56T	Bear River-North	18.842	15.637	39.094	LC	Whole rock	Tg	(Sn)	East Fork?	Custer	Gel	5
KR842T	Big Silver Creek	18.826	15.632	39.108	LC	Whole rock	Tg	(Sn)	Gambrinus	Boise	Gel	5
KR493C	Camas Creek	18.762	15.645	39.755	CT	Whole rock	Tg	(Sn)	Gravel Range	Lemhi	Gel	5
KR804T	Stanley Lake Creek	17.551	15.549	39.330	CT	Whole rock	Tg	(Sn)	Seafoam	Custer	Gel	5
KR803T	Warm Springs Creek	17.580	15.552	39.219	CT	Whole rock	Tg	(Sn)	Stanley?	Boise	Gel	5
KR529T	Wolf Mountain	18.637	15.612	39.055	LC	Whole rock	Tg	(Sn)	Big Creek	Boise	Gel	5

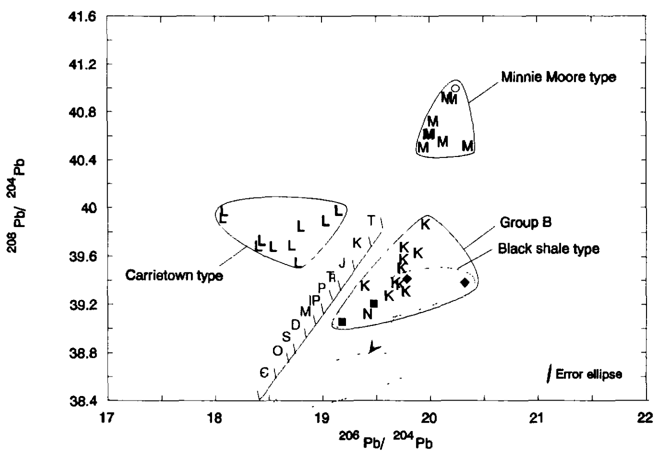
Moore group of mines (mainly the Silver Star Queen mine) and for an epithermal vein galena (G, fig. 4) from the Old Triumph mine (sample number 575) plot in a cluster, which we designate Minnie Moore type, that has high $^{208}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ (fig. 4). The other cluster, group B lead, comprises whole rocks, possibly syngenetic sulfides, and other epithermal veins (figs. 4, 5). Group B lead consists of two types, Triumph (open squares, figs. 4, 5) and black shale (solid squares and diamonds, fig. 4), based on different $^{207}Pb/^{204}Pb$ ratios. Triumph-type lead characterizes the ore in the possibly syngenetic massive sulfide deposits at Triumph, Hoodoo, and Livingston. Lead in the Snoose deposit is Triumph type but is very close to the dividing line between the types.

Uranium and thorium in the whole-rock samples has contributed to the radiogenic lead (^{206}Pb, ^{207}Pb, and ^{208}Pb) since the original sediments were deposited. In order to minimize these contributions, samples that were metal rich and contained visible pyrite, sphalerite, or other sulfides were selected for analysis. Thus the initial lead concentration should be high relative to uranium and thorium concentrations, and the contribution from radiogenic lead produced since sediment deposition should be relatively small. The fact that the lead-isotope ratios for whole-rock argillite (filled squares and diamonds, fig. 4) cluster in a tight group is evidence that this assumption is correct. If the initial lead had been low compared to uranium, the present-day lead-isotope ratios probably would plot as an extended linear array rather than a tight cluster. If there had been significant production of radiogenic lead, the initial $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ values would have been lower than the present-day values and might plot in the field shown by the arrow and dotted outline in figure 4, assuming lead and thorium concentrations for average shale in the Canadian Cordillera (Godwin and Sinclair, 1982). The adjustment of lead-isotope ratios shows that, if significant radiogenic lead has been produced, then the initial whole-rock lead would have been distinctly less radiogenic than lead in galena from veins and syngenetic deposits.

MISSISSIPPIAN ROCKS

Samples from Mississippian rocks include sulfides from veins in the Copper Basin and McGowan Creek Formations (J) and galena from conformable-bedded replacement sulfide ore of the Phi Kappa mine in the Copper Basin Formation (open circle) (fig. 5). One previously reported analysis of whole-rock argillite from the Copper Basin Formation (Urexco sample T–602) is anomalously radiogenic due to uranium enrichment and plots off the diagram (fig. 5) to the right.

The vein and bedded-replacement sulfides hosted by the Copper Basin Formation form a distinct group, the Lava Creek type, that has $^{206}Pb/^{204}Pb$=18.2–19.4 and $^{208}Pb/^{204}Pb$=38.7–39.6 and overlaps only slightly with the group B sulfide samples from the Milligen Formation and Salmon River assemblage.



EXPLANATION

K Pennsylvanian-Permian argillite host rock, Wood River Formation, sulfide samples

L Pennsylvanian-Permian argillite host rock, Dollarhide Formation, Carrietown-Westlake area, sulfide samples

■ Pennsylvanian-Permian argillite, Dollarhide Formation, Carrietown-Westlake area, whole-rock samples

M Pennsylvanian-Permian argillite host rock, Dollarhide Formation, Croy Creek area, crosscutting vein sulfide samples

○ Pennsylvanian-Permian argillite host rock, Dollarhide Formation, Croy Creek area, bedded sulfide sample

◆ Pennsylvanian-Permian argillite, Dollarhide Formation, Croy Creek area, whole-rock samples

N Pennsylvanian-Permian argillite host rock, Grand Prize Formation, sulfide sample

Figure 6. Lead-isotope ratios for samples of vein and conformable epigenetic galena and whole rock from Pennsylvanian and Permian argillite, central Idaho. Average shale growth curve of Godwin and Sinclair (1982) is also shown. Arrow indicates correction for radiogenic lead.

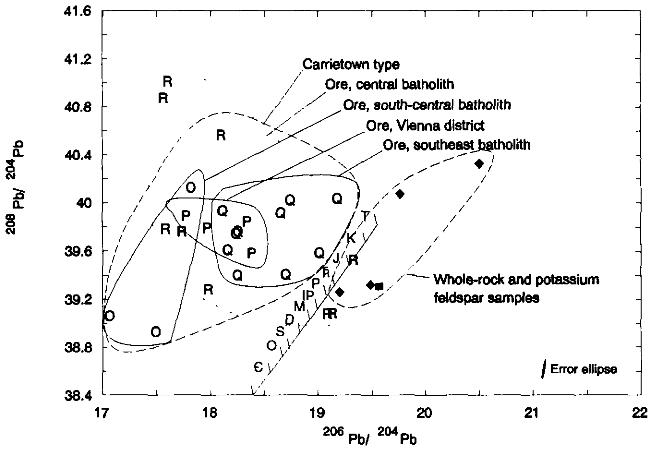
PENNSYLVANIAN AND PERMIAN ROCKS

Samples from Pennsylvanian and Permian rocks include sulfides from epigenetic veins in the Wood River (K), Dollarhide (L, M), and Grand Prize Formations (N) (Sun Valley Group of Mahoney and others, 1991); galena from bedding-conformable concentrations in the Dollarhide Formation (open circle); and pyrite-rich argillite from the Dollarhide Formation (solid diamonds and squares, fig. 6).

Lead-isotope ratios define three types of samples, Carrietown and Minnie Moore (group A) and black shale (group B). Carrietown-type material, which has $^{206}Pb/^{204}Pb$=18.0–19.2 and $^{208}Pb/^{204}Pb$=39.5–40.0, includes some of the vein and bedding-conformable galena hosted by the Dollarhide Formation. Samples in this group are from the Buttercup mine, the Westlake North mine, and all the sampled deposits in the Carrietown mining district. The rest of the vein and bedding-conformable sulfides hosted by the Dollarhide Formation form the radiogenic Minnie Moore type, having $^{206}Pb/^{204}Pb$=19.9–20.4 and $^{208}Pb/^{204}Pb$=40.5–41.1. This type includes samples from the Eureka, Liberty, Jay Gould, and Idahoan mines, for example, all in the Croy Creek area. The isotope ratios in the Croy Creek area are slightly but significantly different from those of the Minnie Moore mine group, indicating a different source, but are probably similar enough to justify grouping them into a single type. Group B, which has $^{206}Pb/^{204}Pb$=19.2–20.4 and $^{208}Pb/^{204}Pb$=39.0–40.0, includes four samples of whole rock from the Dollarhide Formation, one vein galena from the Grand Prize Formation, and all of the vein galena samples from the Wood River Formation. Within group B, the whole-rock samples define the black shale type. The Wood River vein galenas have slightly higher $^{208}Pb/^{204}Pb$ ratios on the average than the other samples in this group. Initial lead-isotope ratios were calculated assuming average shale concentrations of uranium and thorium, as described above for Devonian rocks. The displacement of the isotope ratios to lower values indicates that, if the assumption is valid, initial whole-rock lead was distinctly different from the vein and syngenetic lead hosted by those rocks.

The isotopic ratios of the bedding-conformable concentration of galena in the Dollarhide Formation (bedded sulfide, open circle, fig. 6; sample 8H79E, table 2) show that isotopic characteristics can be used to discriminate between syngenetic and epigenetic sulfides. This sample has textural features suggestive of syngenetic material; however, isotopically it clearly resembles epithermal vein sulfides also hosted by the Dollarhide and is distinct from both the host argillite and all known syngenetic sulfides in the area.

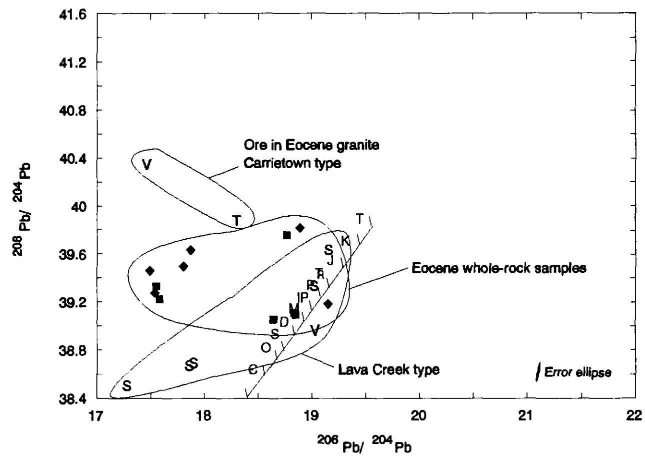
The samples from the Triumph block were plotted separately from others of the Milligen Formation (fig. 4) in order to investigate whether the argillite of the Triumph block correlates with the Middle Pennsylvanian to Lower Permian Dollarhide Formation (Wavra, 1988; Wavra and Hall, 1989) or with the Milligen Formation (Mahoney and others, this volume; Turner and Otto, this volume). Lead-isotope ratios for whole-rock samples cannot resolve this stratigraphic correlation problem; however, lead-isotope ratios from ore deposits support the assignment of the argillite of the Triumph block to the Milligen Formation. Ore deposits in the Triumph block have a lead-isotope signature (Triumph type) almost identical to that of deposits hosted by the Milligen Formation and by the Devonian part of the Salmon River assemblage elsewhere in the black shale terrane. In contrast, the lead-isotope signature of Dollarhide-hosted veins (Carrietown- and Minnie Moore-types) is distinctly different from the lead-isotope signature of deposits in the Triumph block. This evidence supports the correlation of the Triumph argillite with the Milligen Formation.



EXPLANATION

- O Cretaceous intrusive host rock, south-central Idaho batholith, sulfide samples
- P Cretaceous intrusive host rock, Vienna district, east-central Idaho batholith, sulfide samples
- Q Cretaceous intrusive host rock, southeast Idaho batholith, sulfide samples
- ◆ Cretaceous intrusive rock, southeast Idaho batholith, potassium feldspar sample
- R Cretaceous intrusive host rock, central Idaho batholith, sulfide samples
- ◆ Cretaceous intrusive rock, central Idaho batholith, whole-rock and potassium feldspar samples

Figure 7. Lead-isotope ratios for samples of vein galena, whole rock, and potassium feldspar from Cretaceous intrusive rocks, central Idaho. Average shale growth curve of Godwin and Sinclair (1982) is also shown.



EXPLANATION

- S Eocene Challis Volcanic Group host rock, Lava Creek district, sulfide samples
- T Eocene granite host rock, southeast Idaho batholith, sulfide sample
- ◆ Eocene Challis Volcanic Group, central Idaho batholith, whole-rock samples
- V Eocene granite host rock, central Idaho batholith, sulfide samples
- ▪ Eocene granite, central Idaho batholith, whole-rock samples

Figure 8. Lead-isotope ratios for samples of vein galena and whole rock from Eocene intrusive and volcanic rocks, central Idaho. Average shale growth curve of Godwin and Sinclair (1982) is also shown.

CRETACEOUS INTRUSIVE ROCKS

Lead isotopes from four epigenetic vein galena samples (O, P, Q, and R), three feldspar samples (filled diamonds), and one whole-rock sample (filled square) from Cretaceous intrusive rocks include a wide range of $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios (fig. 7). All of the vein galena has $^{206}Pb/^{204}Pb$ less than 19.7. The vein material from the Challis quadrangle (central Idaho batholith, R) has a wider isotopic range than that from the Hailey quadrangle (southern Idaho batholith, O, P, and Q). Vein material from the Hailey quadrangle can be divided into three moderately overlapping groups corresponding to geographic areas within the Idaho batholith. One group (O), having $^{206}Pb/^{204}Pb$=17.0–18.0, includes three samples from the south-central part of the Idaho batholith, encompassing the Featherville, Pine, and Volcano mining districts. A second group (P), having $^{206}Pb/^{204}Pb$=17.8–18.4, comprises all the samples in the east-central part of the Idaho batholith from the Vienna district. A third group (Q), having $^{206}Pb/^{204}Pb$=18.0–19.2, comprises veins in the southeast part of the Idaho batholith, composed of the Big Smoky, Camas, Mineral Hill, Skeleton Creek, and Soldier mining districts.

Lead from a whole-rock sample and four potassium feldspar samples from intrusive rocks of the Cretaceous Idaho batholith is group B, according to our definition. These samples have the same lead-isotope composition as the samples of Devonian and Pennsylvanian-Permian argillite. Lead-isotope ratios of Cretaceous whole-rock and potassium feldspar samples are clearly distinct from those of veins hosted by Cretaceous intrusive rocks. On the basis of our sampling of Cretaceous rocks, it appears that Cretaceous intrusive rocks were not the source for the lead in the Cretaceous-hosted veins.

EOCENE INTRUSIVE AND EXTRUSIVE ROCKS

Samples from Eocene rocks and Eocene-hosted veins include whole rocks from Eocene volcanic and intrusive rocks in the Challis quadrangle and vein galena hosted by intrusive and extrusive rocks from various parts of the study area (fig. 8). Vein galena (S) in the Lava Creek mining district in the Idaho Falls quadrangle is hosted by the Challis Volcanic Group and defines the Lava Creek type, $^{206}Pb/^{204}Pb$=17.3–19.2 and $^{208}Pb/^{204}Pb$=38.5–39.6. These veins

Table 3. Summary of lead-isotope types of sampled rocks and ore deposits in central Idaho.

GROUP A LEAD ISOTOPES	
Carrietown type	
Whole-rock samples	
	Some Eocene granite whole-rock samples, central Idaho batholith.
	Most Eocene volcanic whole-rock samples, central Idaho batholith.
Sulfide samples	
	All replacement deposits in Paleozoic carbonate roof pendants in Idaho batholith.
	Most replacement deposits in Ordovician carbonate and phyllite host rock, Ella Dolostone group.
	All veins in Pennsylvanian-Permian Dollarhide Formation argillite host rock, Carrietown-Westlake area.
	Most veins in Cretaceous intrusive host rock, central Idaho batholith.
	All veins in Cretaceous intrusive host rock, Vienna district, Idaho batholith.
	Most veins in Cretaceous intrusive host rock, southeast Idaho batholith.
	All veins in Cretaceous intrusive host rock, south-central Idaho batholith.
	Vein in Eocene granite host rock, southeast Idaho batholith.
	Vein in Eocene granite host rock, central Idaho batholith.
Lava Creek type	
Whole-rock and potassium feldspar samples	
	Potassium feldspar in Cretaceous intrusive rocks, central Idaho batholith.
	Some Eocene granite whole-rock samples, central Idaho batholith.
Sulfide samples	
	Vein in Middle Proterozoic Swauger Formation host rock.
	All replacement deposits in Devonian carbonate host rock, Jefferson Dolostone group.
	All veins in Mississippian argillite host rock, Alto-Muldoon-Lava Creek area.
	Bedded replacement deposit in Mississippian argillite host rock, Phi Kappa mine.
	Some veins in Cretaceous intrusive host rock, central Idaho batholith.
	All veins in Eocene Challis Volcanic Group host rock, Lava Creek district.
	Vein in Eocene granite host rock, central Idaho batholith.
Minnie Moore type	
Sulfide samples	
	All veins in Devonian Milligen Formation argillite host rock, Minnie Moore mine group.
	All veins in Pennsylvanian-Permian Dollarhide Formation argillite host rock, Croy Creek area.
	Bedded sulfide in Pennsylvanian-Permian Dollarhide Formation argillite host rock, Croy Creek area.
Pacific type	
Whole-rock sample	
	Cretaceous intrusive whole-rock sample, central Idaho batholith.
Sulfide sample	
	Most replacement deposits in Ordovician carbonate host rock, Bayhorse Dolostone.

have a similar range in $^{206}Pb/^{204}Pb$ as veins from the Idaho batholith (Carrietown type); however, the $^{208}Pb/^{204}Pb$ ratios of the Lava Creek type are generally lower for a given $^{206}Pb/^{204}Pb$. Vein galena (T and V) hosted by Eocene granite defines a cluster, having $^{206}Pb/^{204}Pb$=17.5–18.3 and $^{208}Pb/^{204}Pb$=39.9–41.0, that we include in the Carrietown type. Lead in Eocene whole-rock samples forms a broad but well-defined group, having $^{206}Pb/^{204}Pb$=17.5–20.5 and $^{208}Pb/^{204}Pb$=39.1–40.1, that overlaps many of the types defined above, specifically, Carrietown-, Lava Creek-, Pacific-, and black shale-types.

SUMMARY

The characteristic $^{208}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ ratios for sulfide minerals and whole rocks reveal distinct lead isotope signatures or types (see summary, table 3). The first four types are in group A and the fifth is in group B. (1) Carrietown type, which is low in uranogenic lead and high in thorogenic lead, is defined by isotopic ratios in replacement deposits in carbonate roof pendants in the Idaho batholith, replacement deposits in the Middle Ordovician Ella Dolostone (fig. 3), veins hosted by the Middle Pennsylvanian to Lower Permian Dollarhide Formation in the Carrietown and Westlake mining areas (fig. 6), veins in the Cretaceous Idaho batholith of the Hailey quadrangle (fig. 7), and veins hosted by Eocene granite in the Challis quadrangle (fig. 8). (2) Pacific type, which is high in both uranogenic and thorogenic lead, is defined by isotopic ratios in replacement galena in the Upper Cambrian or Lower Ordovician Bayhorse Dolostone (fig. 3). (3) Minnie Moore type, which is high in both uranogenic and thorogenic lead, is defined by isotopic ratios in veins in the Minnie Moore group of mines in the Devonian Milligen

Table 3. Summary of lead-isotope types of sampled rocks and ore deposits in central Idaho—Continued.

GROUP B LEAD ISOTOPES
Triumph type
Syngenetic sulfide samples
All syngenetic sulfides Devonian Milligen Formation argillite host rock including Triumph block.
All syngenetic sulfides in Devonian part of Paleozoic Salmon River assemblage argillite host rock.
Epigenetic sulfide samples
Most epigenetic veins in Devonian argillite host rock, Triumph block.
Some veins in rest of Devonian argillite host rock.
Some veins in Devonian part of Paleozoic Salmon River assemblage argillite host rock.
Most veins in Pennsylvanian-Permian Wood River Formation argillite host rock.
Vein in Pennsylvanian-Permian Grand Prize Formation argillite host rock.
Black shale type
Whole-rock and potassium feldspar samples
Devonian Milligen Formation argillite whole-rock sample, Triumph block.
All other Devonian argillite whole-rock samples.
Mississippian Copper Basin Formation argillite whole-rock sample.
Devonian part of Paleozoic Salmon River assemblage argillite whole-rock sample.
Both Pennsylvanian-Permian Dollarhide Formation argillite whole-rock samples, Carrietown-Westlake area.
All Pennsylvanian-Permian Dollarhide Formation argillite whole-rock samples, Croy Creek area.
Some potassium feldspar in Cretaceous intrusive rocks, central Idaho batholith.
One potassium feldspar in Cretaceous intrusive rock, southeast Idaho batholith.
Sulfide samples
Some veins in Devonian part of Paleozoic Salmon River assemblage argillite host rock.
Some veins in Pennsylvanian-Permian Wood River Formation argillite host rock.

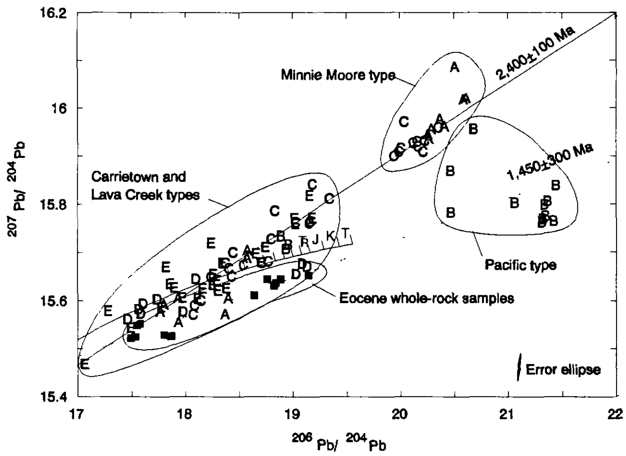
Formation (fig. 4), one sample from the Old Triumph mine in the Triumph block (fig. 4), and veins in the Middle Pennsylvanian to Lower Permian Dollarhide Formation excluding the Carrietown district (fig. 6). (4) Lava Creek type, which is low in both uranogenic and thorogenic lead, is defined by isotopic ratios in a vein in Precambrian quartzite (fig. 3), replacement deposits in the Upper Devonian Jefferson Dolostone (fig. 3), veins and replacements in Mississippian argillite of the Copper Basin Formation (fig. 5), and Eocene Challis Volcanic Group volcanic-hosted veins in the Idaho Falls quadrangle (fig. 8). (5) Group B (consisting of black shale- and Triumph-types, which are not distinguishable using $^{208}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$), which is high in uranogenic lead and low in thorogenic lead, is defined by isotopic ratios in all the whole-rock argillite (figs. 4, 6), all syngenetic deposits (figs. 4, 5), most veins in the Devonian Milligen Formation including argillite of the Triumph block (fig. 4), some veins in the Devonian part of the Salmon River assemblage (fig. 5), and most veins in the Middle Pennsylvanian and Lower Permian Wood River and Grand Prize Formations (fig. 6).

The Carrietown-, Minnie Moore-, and Pacific-types together correspond to group A of Doe (Doe and Delevaux, 1985; Doe and Sanford, this volume) and to area Ia of Zartman (Zartman, 1974). The Triumph- and black shale-types correspond to group B of Doe and to area II of Zartman. The Lava Creek type, which has affinities to both groups A and B, is here included in group A because it is more similar to area Ia of Zartman than to area II, as discussed below. Thus, some of the samples classified as group B in Doe and Sanford (this volume) are here classified as group A.

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

The $^{207}Pb/^{204}Pb$ data are presented separately for groups A and B (figs. 9, 10). Carrietown-, Lava Creek-, and Minnie Moore-types display a fairly linear trend that yields a calculated model age of 2,400±100 Ma using the model 2 least-squares regression of York (1969). Eocene igneous whole rocks of group A also exhibit a linear trend yielding a calculated model age of 1,450±300 Ma. These model ages agree well with previously calculated model ages (Small, 1968; Doe and Delevaux, 1985) and probably indicate times of major orogenic reworking of crustal source rocks. Pacific-type lead does not show a trend by itself but is consistent with a model age of ~1,450 Ma for the source rock.

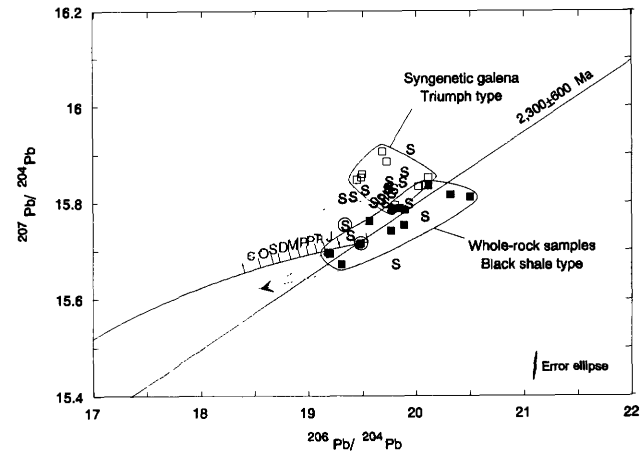
$^{207}Pb/^{204}Pb$ data for group B samples suggest two slightly overlapping groups (fig. 10). Triumph-type lead in syngenetic galena from the Triumph, Hoodoo, and Livingston mines in general has higher $^{207}Pb/^{204}Pb$ than black shale-type lead from whole rocks of the Milligen and Dollarhide Formations. Triumph-type lead is defined by isotopic ratios of syngenetic sulfides in Devonian argillic host rock (figs. 4, 5, 10). Black shale-type lead is defined by isotopic ratios of Devonian, Pennsylvanian, and Permian argillite



EXPLANATION

- A Undifferentiated Paleozoic carbonate, Devonian argillite and carbonate host rock, sulfide samples
- B Ordovician carbonate host rock, sulfide samples
- C Mississippian, Pennsylvanian, and Permian host rocks, sulfide samples
- D Cretaceous and Eocene igneous host rock, central batholith, sulfide samples
- E Cretaceous and Eocene host rock, southern batholith, sulfide samples
- ■ Eocene rhyolite and granite, whole-rock samples

Figure 9. Lead-isotope ratios for samples in group A, central Idaho. Symbol A includes Milligen Formation (including Triumph block argillite), Ramshorn slate, Saturday Mountain Formation, Jefferson Dolostone, and Paleozoic carbonate roof pendants within the Idaho batholith. Average shale growth curve of Godwin and Sinclair (1982) is also shown. Model age is calculated from vein lead samples excluding those from Bayhorse and Ella carbonate rocks.



EXPLANATION

- ■ Paleozoic argillite and Cretaceous intrusive rocks, whole-rock and potassium feldspar samples
- □ Paleozoic sedimentary host rock, syngenetic sulfide samples
- S Paleozoic sedimentary host rock, epigenetic sulfide samples
- ○ Igneous host rock, sulfide sample

Figure 10. Lead-isotope ratios for group B, central Idaho. Dashed curve shows isotopic ratios of black shale type assuming decay since the Devonian using data for average shale from Godwin and Sinclair (1982). Model age is calculated from whole-rock argillite samples only. Arrow indicates correction for radiogenic lead.

whole rocks (figs. 4, 6, 10) and Cretaceous igneous whole rocks and potassium feldspar (figs. 7, 10). Even if age corrections assuming decay of uranium and thorium are applied, the Triumph- and black shale-types are distinct. Least-squares regression of $^{207}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ data for argillite whole-rock samples yields a model age of 2,300±600 Ma. Although the error is large, the model age agrees with that from group A lead (Carrietown-, Lava Creek-, and Minnie Moore-types) (fig. 9), which suggests a similar age for the source rock. Galena samples from sediment-hosted veins have $^{207}Pb/^{204}Pb$ values that span the range of Triumph and black shale lead-isotope ratios, but most correlate isotopically with the Triumph-type syngenetic galena rather than with the black shale-type whole-rock argillite. The distinction between Triumph- and black shale-type lead is not apparent on the $^{208}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ diagram, unless the whole-rock lead is corrected for radiogenic growth since sediment deposition. Our classification scheme for central Idaho lead isotopes is summarized in figure 11.

For classifying lead-isotope ratios from other sources, straight lines are shown that closely correspond to boundaries between the defined types (heavy lines in fig. 11). Only relatively few samples having lead isotope ratios in the areas of overlap might be misclassified using these fields. Equations for the field boundaries are as follows.

Boundary between Carrietown–Minnie Moore and Lava Creek–Triumph–black shale types:

$$^{208}Pb/^{204}Pb=0.516\times{}^{206}Pb/^{204}Pb+29.778$$

Boundary between Carrietown- and Minnie Moore-types:

$$^{206}Pb/^{204}Pb=19.6$$

Boundary between Lava Creek and Triumph–black shale types:

$$^{206}Pb/^{204}Pb=19.27$$

Boundary between Triumph- and black shale-types and between Minnie Moore- and Pacific-types:

$$^{207}Pb/^{204}Pb=0.149\times{}^{206}Pb/^{204}Pb+12.855$$

DISCUSSION

The isotopic evidence reveals three major crustal zones (table 1, also compare figs. 12 and 1): a broad western zone that corresponds to the Cretaceous intrusive rock terrane and small parts of the black shale terrane and is characterized isotopically by low $^{206}Pb/^{204}Pb$ and high $^{208}Pb/^{204}Pb$ (Carrietown type); a narrow arcuate central zone corresponding to black shale terrane and characterized by high $^{206}Pb/^{204}Pb$ (Minnie Moore-, Pacific-, Triumph-, and black shale-types); and a broad eastern zone that corresponds to flysch, shelf carbonate and Precambrian terranes and is characterized by low $^{206}Pb/^{204}Pb$ and low $^{208}Pb/^{204}Pb$ (Lava Creek type).

We determined the sources of lead in the ore deposits for each crustal zone and terrane by a process of elimination. If the isotopic ratios of lead in a potential source rock do not match those of the veins, then that source rock is eliminated as a possibility. Other factors, such as the relative ages of rocks and veins, can further eliminate source-rock candidates; for example, the source rock had to exist at the time the veins formed. The effect of radiogenic lead-isotope production in the source rock with time is considered when testing for any match. Summaries of the sulfide and rock samples in each lead-isotope type are given by crustal zone and terrane in table 1 and by isotope type in table 3.

WESTERN CRUSTAL ZONE—CARRIETOWN TYPE

Most veins in the western crustal zone (Cretaceous intrusive rock terrane) probably derived their lead (Carrietown type) from a Precambrian source beneath the exposed Idaho batholith. The three samples from widely separated Cretaceous intrusions have black shale-type lead (fig. 7). Unless further sampling of Cretaceous rocks discloses a larger range of isotopic ratios—specifically, lower $^{206}Pb/^{204}Pb$ and higher $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$—the Cretaceous igneous rocks cannot be the source for the lead in Cretaceous-hosted veins. Similarly, the vein lead could not have come from Paleozoic argillite. The limited areal distribution of argillite probably also rules argillite out as a direct source of lead in the Idaho batholith-hosted veins. The only known rocks having the appropriate isotopic ratios are some of the Eocene intrusive and volcanic rocks (figs. 8, 11); however, most of the veins are Cretaceous in age (Snee and Kunk, 1989; Darling and others, this volume; Link and others, this volume; Whitman, this volume; Park, in press), which makes an Eocene source rock impossible. Further, Eocene rocks are mostly too low in $^{208}Pb/^{204}Pb$ to account for the entire range in Cretaceous-hosted veins. This virtually eliminates every conceivable source except the unexposed Precambrian basement.

The high $^{207}Pb/^{204}Pb$ of Carrietown-type lead (fig. 9) suggests that the Precambrian source for the intrusive-rock-hosted veins contained a significant component of Archean as well as Proterozoic material. The absence of this Archean signature in Eocene intrusive rocks suggests that the veins and the intrusive rocks had different sources.

Carrietown-type lead in sedimentary rocks of the Middle Pennsylvanian to Lower Permian Dollarhide in the Carrietown and Westlake areas, in Paleozoic carbonate roof pendants within the Idaho batholith, and in Eocene granite host rock (table 3) strongly suggests that veins and replacements in these sedimentary and intrusive rocks had the same type of source as veins in the Cretaceous intrusive rocks themselves. The proximity of all of these rocks to the Cretaceous intrusive rock terrane is consistent with a common "plumbing system" in the batholith and adjacent sedimentary rocks. This interpretation is consistent with the conclusions of other workers that Cretaceous intrusive activity, adjacent metamorphism of the Dollarhide Formation, and vein mineralization are all closely associated (Darling and others, this volume; Whitman, this volume; Park, in press). Although the Cretaceous intrusions probably provided the heat for hydrothermal circulation, the metals apparently came from the underlying basement and not from the intrusions themselves. For the Eocene intrusion-hosted veins, the plumbing system would have had to develop later in the Eocene rocks but nevertheless appears to have tapped a similar source, one that may also have been enriched in Archean material. What is certain is that the vein lead in these sedimentary and igneous rocks did not come from the host rocks themselves.

Significantly, Dollarhide-hosted, bedding-conformable galena having Carrietown-type lead must also have had a source external to the host rock. Thus, the conformable sulfide in the Dollarhide Formation is simply vein filling that happened to be deposited parallel with bedding, and textural evidence is not a completely reliable guide to syngenetic material.

The recognition of Lava Creek-type lead suggests the reclassification of some samples that were classified as group B in earlier studies (Sanford and others, 1989; Doe and Sanford, this volume). The most important change is that the samples of galena and whole rock from the area northeast of Idaho City (left side of fig. 12 near lat 44°) are group A rather than group B. The occurrence of group B lead in the midst of the Idaho batholith had been puzzling; however, classifying this lead as group A is now consistent with the interpretation that Lava Creek-type lead is a low-$^{208}Pb/^{204}Pb$ variety of group A lead that has a Precambrian source and is associated with Tertiary intrusive rocks (see discussion of eastern crustal zone below).

CENTRAL CRUSTAL ZONE—TRIUMPH- AND BLACK SHALE-TYPES

The narrow, arcuate black shale terrane contains four of the six lead-isotope types, all of which have relatively high $^{206}Pb/^{204}Pb$. The most widespread types in this terrane are Triumph- and black shale-types, which together define group B lead. The association of group B lead and argillitic sedimentary rocks is striking (table 1 and compare figs. 1 and 12) and would appear to support the lateral-secretion theory whereby lead is leached from the black shale and then deposited nearby as ore (Hall and others, 1978). However, the difference in $^{207}Pb/^{204}Pb$ between Triumph- and black

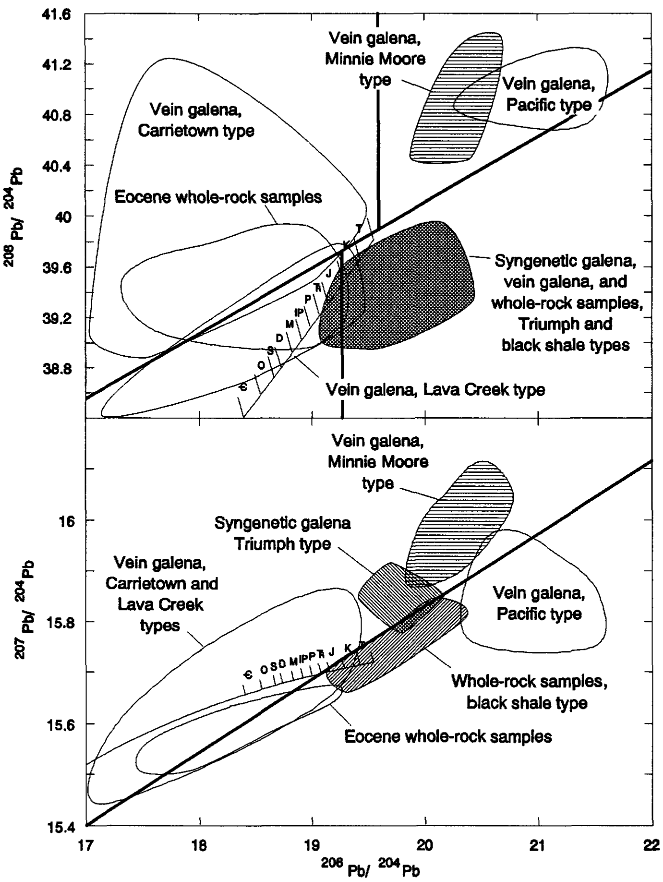


Figure 11. Summary of lead-isotope types, central Idaho. Outlined areas indicate isotopic range of lead-isotope types described in text. Heavy lines are the straight boundaries that best approximate the observed boundaries and are to be used for samples whose association is unknown. Equations for these lines are given in text.

shale-types indicates that the ore was not derived exclusively from the sedimentary rocks but that some or all of it was derived from a different source, probably the underlying Precambrian basement.

Black shale-type lead probably represents the average isotopic composition of upper crustal rocks east and northeast of the black shale terrane. This type of lead characterizes all of the argillite of Devonian and Pennsylvanian-Permian age in the black shale terrane. Sedimentologic and paleogeographic studies suggest that this detrital material was eroded from highlands to the east and northeast (Mahoney and others, this volume; Turner and Otto, this volume). The low $^{207}Pb/^{204}Pb$ values indicate that the eroded crust consisted of Proterozoic rocks having no detectable component of Archean material.

Triumph-type lead in the syngenetic deposits probably consists mostly or entirely of hydrothermally derived lead having an Archean component. A second component of sediment-derived black shale-type lead is possible but not required by the data. The Triumph-type lead in these deposits has significantly higher $^{207}Pb/^{204}Pb$ than the black shale-type lead in the argillite host rock (fig. 10). Although some of the lead in the syngenetic deposits having Triumph-type lead may have come from the enclosing sediments, there is clearly a high $^{207}Pb/^{204}Pb$ component that suggests a contribution of Archean material in the source. On the basis of similarity with the Triumph deposit (Turner and Otto, this volume), these possibly syngenetic deposits probably formed from "black smokers" in a submarine rift basin. Our data are consistent with such a model. The ore-forming hydrothermal fluid appears to have tapped a deep source of continental crustal lead having an Archean component while, contemporaneously, lead contained in detrital material was being eroded from a Proterozoic craton having no recognizable Archean component. The Triumph-type lead-isotope ratios thus clearly indicate a component of hydrothermally derived lead from depth. An additional component from the black shale is consistent with the data but not required. A modern example of such a mixing process has been documented for active mid-ocean ridge hydrothermal vents (LeHuray and others, 1988).

Epigenetic veins hosted by the Devonian Milligen and Salmon River argillite were probably either part of the same hydrothermal system that produced the syngenetic Triumph-type deposits, and thus obtained their lead directly from the same underlying basement, or they formed later by remobilization of syngenetic sulfides from a Triumph-like source. As with the syngenetic deposits, the epigenetic veins could not have obtained their lead exclusively from the detrital argillite host rock. Comparison of all central Idaho veins having group B lead with syngenetic deposits and whole-rock argillite shows that lead in almost all these veins isotopically resembles lead in syngenetic deposits and does not match lead in argillitic whole rocks (fig. 10), despite the close spatial association of veins and argillite.

If epigenetic veins in Devonian argillite constituted the feeder system for Triumph-like syngenetic deposits, this would explain the similar isotopic signatures. Examples of such epigenetic feeder veins and contemporaneous syngenetic exhalative deposits are in Alaska, Australia, Canada, and Ireland (Church and others, 1987). The contemporaneous formation of both bedded and crosscutting ore helps explain the commonly contradictory textural evidence that has prevented geologists from concluding definitively whether the deposits are syngenetic or epigenetic.

On the other hand, the presence of Tertiary-mineralized rock in the study area (Burton and Link, this volume) suggests that some or all veins in the Devonian argillite postdate the syngenetic deposits and that vein sulfide deposits represent remobilized syngenetic sulfides.

Vein deposits in the Wood River Formation probably reflect a remobilized Triumph-like sulfide component. Because they are in Pennsylvanian-Permian host rocks, which are not host to any syngenetic massive sulfide

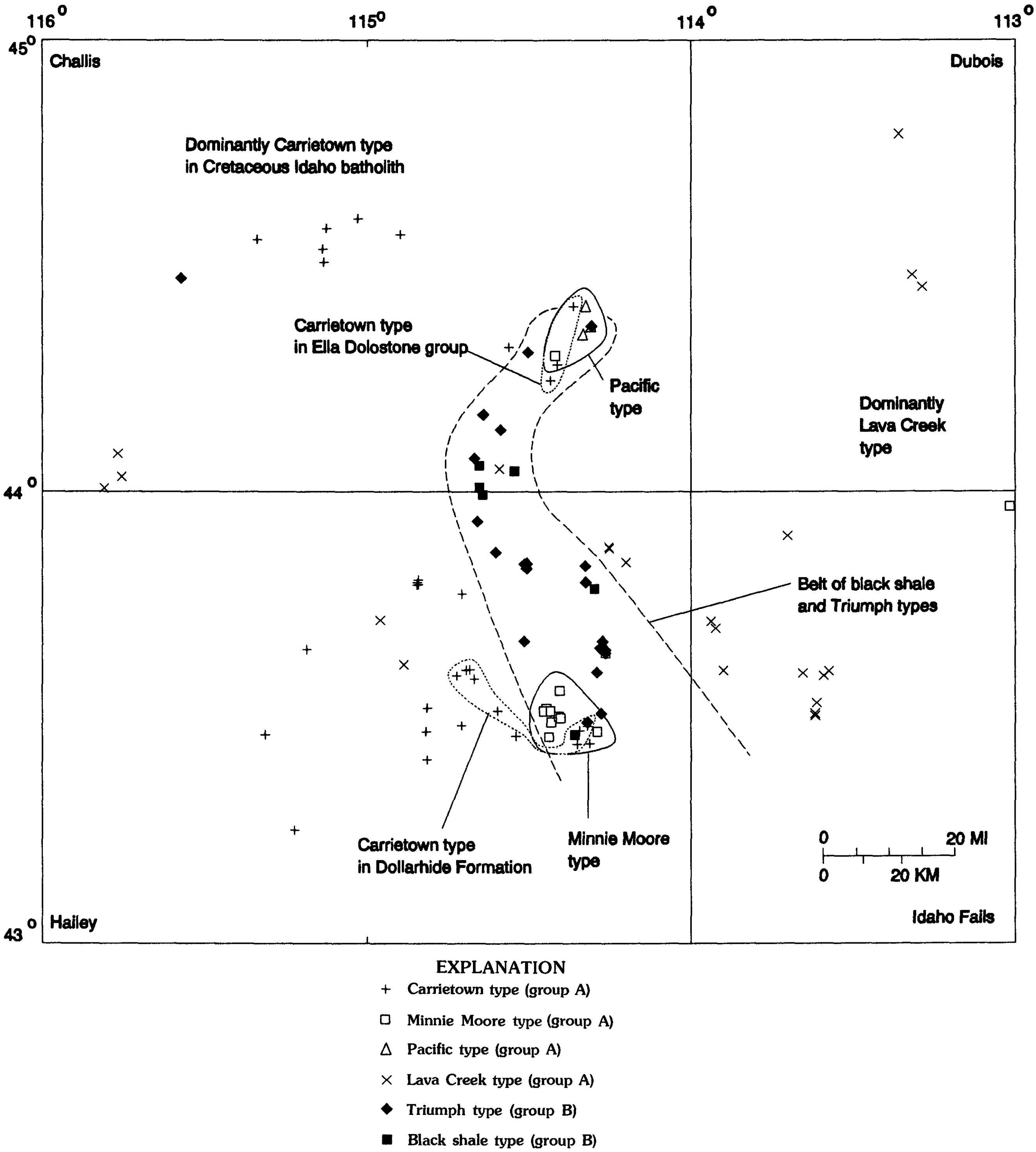


Figure 12. Map showing distribution of lead-isotope types for galena and other sulfide minerals, central Idaho.

deposits, they probably were not part of a feeder system for syngenetic deposits. As above, the Triumph-type lead in these veins indicates that the source was not exclusively argillite such as the Milligen Formation or Salmon River assemblage. Because of the isotopic similarity to lead in syngenetic deposits such as the Triumph, the source for vein lead in the Wood River Formation may have been Triumph-like deposits, isotopically similar disseminated syngenetic sulfides within the argillite, or an isotopically similar source at depth. Although more work is needed, the most likely interpretation at present is that veins in Devonian argillite that have Triumph-type lead were part of the feeder system for contemporaneous syngenetic deposits, whereas those in Pennsylvanian-Permian argillite consist of remobilized Devonian syngenetic sulfides. Because of the independent evidence for remobilization of older sulfides (Burton and Link, this volume), this is the most likely explanation.

The following model is consistent with the data and is preferred to the lateral-secretion model. Lead in the syngenetic sulfides came from Precambrian rocks at depth, was transported upward by hydrothermal fluids, and was deposited at submarine vents. During transport, this lead may have mixed with a second component of lead from detrital material in the host argillite, but this mixing is not required. Some argillite-hosted vein deposits probably formed contemporaneously with syngenetic exhalative deposits, whereas others probably formed by later remobilization of this syngenetic lead. Crosscutting veins in Devonian argillite were probably contemporaneous with syngenetic deposits, whereas those in Pennsylvanian-Permian argillite probably formed by remobilization.

Lead-isotope data can help resolve stratigraphic correlation problems. All whole-rock argillite samples in this study have black shale-type lead (except for two anomalously radiogenic uranium enriched samples). We therefore cannot use the lead isotopes in whole rocks to resolve stratigraphic problems. Lead-isotope signatures of ore deposits can help, however, in correlating units and, in the present case, confirm the assignment of the Triumph argillite to the Milligen Formation.

The anomalously radiogenic samples suggest a Permian-Triassic age for uranium enrichment. A least-squares regression line for $^{207}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ data from whole-rock argillite is defined by a cluster of values having intermediate lead-isotope ratios and by the two most radiogenic samples (Urexco and Holman Creek), which plot off the diagrams to the right (table 2). Based on the model 2 regression of York (1969), uranium enrichment in argillite probably occurred about 250–210 Ma; that is, Permian to Triassic. This age is only an estimate and should be used with caution. Permian and Triassic ages for uranium enrichment correspond to uranium enrichment events in the Colorado Plateau (Granger and Finch, 1988).

CENTRAL CRUSTAL ZONE—MINNIE MOORE- AND PACIFIC-TYPES

Clusters of deposits having Minnie Moore-type lead and Pacific-type lead are at the south and north ends of the black shale terrane, respectively. The isotopic characteristics of these lead types rule out as possible sources all rocks analyzed so far within the study area because none are sufficiently radiogenic. The problem of source for this highly radiogenic vein lead has been discussed previously (for example, Doe and others, 1979).

The most likely source of the radiogenic lead is recycled upper crustal Precambrian material having an Archean component that had high initial $^{238}U/^{204}Pb$ (μ) and Th/U. Archean model ages in the range 2,700–2,300 Ma suggest that this time was a period of widespread crustal disturbance that created variations in μ. Zartman (1992) argued that variation in μ in Archean source rocks has produced some high values of $^{207}Pb/^{204}Pb$. By extension, an Archean crustal component having even higher values of μ, as well as higher initial Th/U, would explain the radiogenic Minnie Moore- and Pacific- types of lead. Recycling in Proterozoic time would further accentuate these variations. Later selective leaching of the radiogenic lead component might also have contributed to the radiogenic lead in the Pacific- and Minnie Moore-types. Further, the Pacific-type lead discussed here has some affinities with lead in some of the classic Mississippi Valley-type deposits, which these deposits resemble. An explanation for the source of lead in Mississippi Valley-type deposits might help explain the source of the central Idaho carbonate-hosted lead.

EASTERN CRUSTAL ZONE—LAVA CREEK TYPE

The eastern crustal zone is dominated by Lava Creek-type lead in veins and replacements in flysch, shelf-carbonate, and Precambrian terranes. The most likely source for Lava Creek-type lead is Precambrian basement. The lead in Paleozoic argillite and Cretaceous intrusive rocks is too radiogenic for these rocks to be sources. Some Eocene whole rocks have lead-isotope ratios similar to those in the veins, and Eocene rocks could have been the source. Close spatial proximity of ore deposits and known Eocene intrusive rocks supports this supposition. Precambrian basement is the probable source for most of the lead in the western and central crustal zones, as discussed above, and therefore it could also be a source in the eastern crustal zone. Compared to the other crustal zones, the eastern crustal zone has the thinnest sedimentary cover over the basement, which makes a Precambrian source more likely.

Lava Creek-type lead seems to characterize Eocene-age veins. All of the veins in the eastern crustal zone are Eocene, as are many veins in the Trans-Challis fault zone northeast of Idaho City (Ronald G. Worl, oral commun., 1992). All of these veins yield Lava Creek-type lead. Thus, the Lava Creek-type lead appears to be unique to and characteristic of Eocene veins.

The fact that Carrietown- and Lava Creek-types completely overlap when plotted on a $^{207}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ diagram (fig. 9) indicates that there is no detectable difference in the age of the underlying crust in the western and eastern parts of central Idaho. Both sets of data are consistent with ~2,400-Ma crust that has a similar history; however, the consistently higher proportion of thorogenic lead in the Carrietown type suggests a major, long-lasting, compositional difference in the crust between these two zones.

In summary, lead isotopes reveal whether the galena in central Idaho ore deposits is syngenetic or epigenetic and indicate the type of source rock, whether Paleozoic argillaceous sedimentary rocks, remobilized syngenetic sulfides, Cretaceous-Eocene intrusive rocks, or a fourth, as yet unidentified, upper crustal rock type. The variety of lead-isotope types that originated in Precambrian basement suggests that the Precambrian basement is quite heterogeneous. Incorporation of data from other isotopes such as sulfur should help to identify better the source rocks.

REFERENCES CITED

Church, S.E., Gray, J.E., Delevaux, M.H., and LeHuray, A.P., 1987, Pb-isotope signatures of Devonian-Mississippian massive sulfide deposits in Alaska and their significance to mineral exploration, *in* Elliott, I.L., and Smee, B.W., eds., GEOEXPO/86, Exploration in the North American Cordillera: Association of Exploration Geochemists, Rexdale, p. 132–141.

Criss, R.E., Champion, D.E., and McIntyre, D.H., 1985, Oxygen isotope, aeromagnetic, and gravity anomalies associated with hydrothermally altered zones in the Yankee Fork mining district, Custer County, Idaho: Economic Geology, v. 80, p. 1277–1296.

Criss, R.E., Ekren, E.B., and Hardyman, R.F., 1984, Casto ring zone—A 4,500-km^2 fossil hydrothermal system in the Challis volcanic field, central Idaho: Geology, v. 12, p. 331–334.

Davis, K.E., 1978, Lead isotope ratios of galena from the Bayhorse mining district, Custer County, Idaho: Isochron/West, no. 22, p. 15–16.

Doe, B.R., 1978, The application of lead isotopes to mineral prospect evaluation of Cretaceous-Tertiary magmatothermal ore deposits in the western United States, *in* Watterson, J.R., and Theobald, P.K., eds., Geochemical exploration 1978: International Geochemical Exploration Symposium, 7th, Proceedings; Association of Exploration Geochemists, p. 227–232.

Doe, B.R., and Delevaux, M.H., 1985, Lead-isotope characteristics of ore systems in central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–O, p. 181.

Doe, B.R., and Rohrbough, Randall, 1977, Lead isotope data bank—3,458 samples and analyses cited: U.S. Geological Survey Open-File Report 77–418, 137 p.

Doe, B.R., Steven, T.A., Delevaux, M.H., Stacey, J.S., Lipman, P.W., and Fisher, F.S., 1979, Genesis of ore deposits in the San Juan volcanic field, southwestern Colorado—Lead isotope evidence: Economic Geology, v. 74, p. 1–26.

Fisher, F.S., and Johnson, K.M., eds., 1987, Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadrangle, Idaho": U.S. Geological Survey Open-File Report 87–480, 251 p.

Godwin, C.I., and Sinclair, A.J, 1982, Average lead isotope growth curves for shale-hosted zinc-lead deposits, Canadian Cordillera: Economic Geology, v. 77, p. 675–690.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

Hall, W.E., Rye, R.O., and Doe, B.R., 1978, Wood River mining district, Idaho—Intrusion-related lead-silver deposits derived from country rock source: U.S. Geological Survey Journal of Research, v. 6, p. 579–592.

Holmes, A., 1946, An estimate of the age of the earth: Nature, v. 159, p. 127.

Houtermans, F.G., 1946, The isotopic ratios in natural lead and the age of uranium: Naturwissenschaften, v. 33, p. 185–186.

Howe, S.S., and Hall, W.E., 1985, Light-stable-isotope characteristics of ore systems in central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–P, p. 183–192.

LeHuray, A.P., Church, S.E., Koski, R.A., and Bouse, R.M., 1988, Pb isotopes in sulfides from mid-ocean ridge hydrothermal sites: Geology, v. 16, p. 362–365.

Ludwig, K.R., 1990, ISOPLOT—A plotting and regression program for radiogenic-isotope data, for IBM-PC compatible computers, version 2.11: U.S. Geological Survey Open-File Report 88–557, 32 p.

Mahoney, J.B., Link, P.K., Burton, B.R., Geslin, J.K., and O'Brien, 1991, Pennsylvanian and Permian Sun Valley Group, Wood River basin, south-central Idaho, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 2, p. 551–580.

Park, W.A., in press, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–CC.

Rember, W.C., and Bennett, E.H., 1979, Geologic map of the Idaho Falls quadrangle, Idaho: Idaho Bureau of Mines and Geology, Geologic Map Series.

Russell, R.D., and Farquhar, R.M., 1960, Lead isotopes in geology: New York, Interscience, 141 p.

Sanford, R.F., Wooden, J.L., Doe, B.R., and Worl, R.G., 1989, Origin of ore deposits in the central Idaho black shale belt—Implications from lead isotopes, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 39–52.

Small, W.D., 1968, Cordilleran geochronology deduced from hydrothermal leads: Vancouver, University of British Columbia, Ph.D. thesis, 96 p.

Snee, L.W., and Kunk, M.J., 1989, $^{40}Ar/^{39}Ar$ thermochronology of mineral deposits in the southern part of the Idaho batholith, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 37–38.

Wavra, C.S., 1988, Shale-hosted massive sulfide potential of the upper member/Permian Dollarhide Formation, south-central Idaho: Geological Society of America Abstracts with Programs, v. 20, p. 473–474.

Wavra C.S., and Hall, W.E., 1989, Potential for shale-hosted massive sulfides in the Permian Dollarhide Formation, south-central Idaho: U.S. Geological Survey Open-File Report 88–376, 9 p.

Worl, R.G., and Johnson, K.M., 1989, Geologic terranes and mineral deposit types in the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 3–16.

York, D., 1969, Least-squares fitting of a straight line with correlated errors: Earth and Planetary Science Letters, v. 5, p. 320–324.

Zartman, R.E., 1974, Lead isotope provinces in the Cordillera of the western United States and their geologic significance: Economic Geology, v. 69, p. 792–805.

———1992, Archean crustal lead in the Helena embayment of the Belt basin, Montana, *in* Bartholomew, M.J., Hyndman, D.W., Mogk, D.W., and Mason, R., eds., Basement tectonics 8—Characterization and comparison of Ancient and Mesozoic continental margins: International Conference on Basement Tectonics, 8th, Butte, Montana, Proceedings; Dordrecht, The Netherlands, Kluwer Academic Publishers, p. 699–709.

Geochemical Evidence for Epithermal Metallization in Mississippian Turbidites of the McGowan Creek Formation, Lava Creek Mining District, South-Central Idaho

By James A. Erdman, Falma J. Moye, Betty Skipp, *and* Paul K. Theobald

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–O

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Geochemical Evidence for Epithermal Metallization in Mississippian Turbidites of the McGowan Creek Formation, Lava Creek Mining District, South-Central Idaho

By James A. Erdman,[1] Falma J. Moye,[2] Betty Skipp,[1] *and* Paul K. Theobald[1]

ABSTRACT

Geochemical and biogeochemical exploration techniques were used in the Lava Creek mining district, Butte County, Idaho, north of Craters of the Moon National Monument, to assess the potential for new metallic mineral occurrences. Heavy-mineral concentrates, stream sediments and soils, and big sagebrush (*Artemisia tridentata*) were collected in two drainage-basin surveys: an initial reconnaissance-scale survey in 1987 and a followup survey in 1988. Samples of sagebrush were analyzed to investigate this species' ability to selectively remove hydromorphically bound, ore-related elements from the weathering environment. The plant results were compared with those from the more conventional sample media.

The Lava Creek district was selected because of its historical importance as a producer of silver and byproduct gold and because it is the only area of major mineralized rock in the southeastern Challis volcanic field. Much of this district is underlain by rocks of the Eocene Challis Volcanic Group that have been so hydrothermally altered that outcrop is limited. Thus, metal-bearing rocks are poorly exposed. Biogeochemical methods are especially effective under such conditions.

Anomalous concentrations of the element suite antimony, gold, and bromine in sagebrush collected in the reconnaissance-scale survey are characteristic of epithermal precious-metal deposits. These anomalies include two areas in the Lava Creek district where a possible host is turbidite of the Lower Mississippian McGowan Creek Formation. Exposures of this formation are peripheral to the formerly active mines and the new Champagne (gold) mine. The anomaly threshold for gold in sagebrush is 0.4 ppb (dry-weight basis); the highest concentration, 1.0 ppb, was in a sample from the Champagne mine.

Followup sampling of sagebrush, stream sediments, and soils from these two areas led to the discovery in one of the areas of a north-northeast-trending, silicified fault breccia near a poorly exposed dacite porphyry intrusive body. This fault may be the focal point of a hydrothermal system that contains anomalous concentrations of silver, mercury, and arsenic. Gold, antimony, bromine, and tungsten anomalies are less widespread. Cinnabar was identified in several samples of heavy-mineral concentrates from the fault area.

Based on the results of the followup geochemical survey, we conclude that this fault breccia is a prospective target. Detailed studies should be conducted to identify the mineral deposit type(s) that might be present.

INTRODUCTION

The Lava Creek mining district (fig. 1), a major silver producer in Idaho in the 1880's, is in Butte County in the almost treeless foothills of the Pioneer Mountains where the dominant vegetation is big sagebrush (*Artemisia tridentata* Nutt.). A rock geochemical survey (Hillier and others, 1983) in rocks of the Eocene Challis Volcanic Group, host for the former and present mining, shows extremely high values for not only Ag, but also As, Sb, Hg, Mn, Bi, Pb, Cu, Zn, Cd, and some Au. Species of the genus *Artemisia* have been used in gold prospecting in arid regions of the former Soviet Union since the 1960's, and big sagebrush currently is being sampled extensively, especially in Nevada, in the search for disseminated, bulk-minable gold deposits.

The arid environment supports a vegetation type characterized by sagebrush but commonly containing less conspicuous shrubs such as antelope bitterbrush (*Purshia tridentata* [Pursh] DC.) and rabbitbrush (*Chrysothamnus* sp.). Small stands of mainly Douglas-fir (*Pseudotsuga menziesii* [Mirb.] Franco) commonly dominate slopes at higher elevations, primarily north exposures. Some of the smaller

[1]U.S. Geological Survey, Denver Federal Center, Denver, Colorado 80225.

[2]Department of Geology, Idaho State University, Pocatello, Idaho 83209.

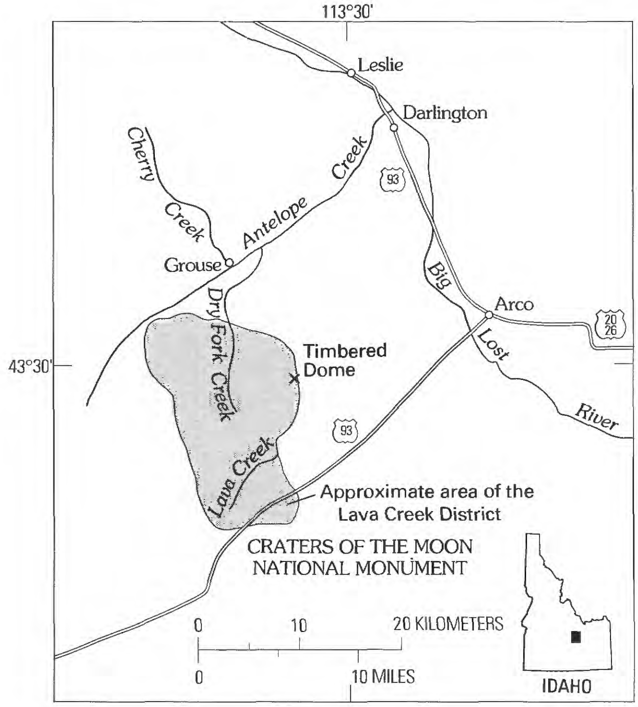


Figure 1. Index map showing area of the Lava Creek mining district, central Idaho. The center of the Lava Creek mining district is about 14 mi (22 km) southwest of Arco, Idaho. The elevation ranges from about 5,600 ft (1,700 m) at the mouth of Lava Creek in the southern part of the district to 8,356 ft (2,550 m) at Timbered Dome, the predominant geographic feature in the study area.

EXPLANATION

Qal	**Alluvium, gravel, and other unconsolidated sediments (Quaternary)**
Qb	**Basalt (Quaternary)**
Tj	**Jasperoid (Tertiary)**
Th	**Volcanics at Heise (Tertiary)**
Te	**Extrusive rocks (Tertiary)**—Including the Challis Volcanic Group (Eocene)
Ti	**Intrusive rocks (Tertiary)**—Including subvolcanic rocks of the Challis Volcanic Group (Eocene) and prophyritic rocks not know to be Challis
PMcb	**Primarily carbonate rocks**—Including the Snaky Canyon Formation (Lower Permian to Upper Mississippian), Bluebird Mountain Formation (Upper Mississippian), Lower Mississippian rocks in the Beaverhead Mountains, Lemhi County, and Arco Hills Formation (Upper Mississippian)
Mcb	**Carbonate units**—Including the Surrett Canyon (Upper Mississippian), South Creek (Upper Mississippian), Scott Peak (Upper Mississippian), and Middle Canyon (Upper and Lower Mississippian) Formations
Mcu	**Copper Basin Formation (Upper and Lower Mississippian)**
Mmg	**McGowan Creek Formation (Lower Mississippian)**
DOu	**Undifferentiated rocks**—Including the Carey Dolomite (Middle and Lower Devonian:), Fish Haven Dolomite (?)(Upper Ordovician), Kinnikinic Quartzite (?)(Middle Ordovician), and Summerhouse Formation (Lower Ordovician). The latter three units may all be Summerhouse Formation

Contact—Dashed where uncertain

Fault—Dashed where uncertain

valleys harbor stands of alder (*Alnus* sp.), aspen (*Populus tremuloides* Michx.), and poplar (*Populus* sp.) along the banks of perennial streams. The abundance of perennial grasses makes the district good pasturage for cattle and sheep (Anderson, 1929).

The dominance of sagebrush in the district and the poor outcrop prompted us to use biogeochemistry as a tool, along with conventional heavy-mineral concentrates, to assess the precious-metal potential of the area. The biogeochemical method uses the element concentrations of plants to outline secondary dispersion halos in soils and overburden (Boyle, 1984, p. 10). Sampling of stream sediments and, to a limited extent, soils was included in the followup survey.

This paper describes the geochemical results from a reconnaissance-scale drainage-basin survey of the district conducted in 1987 and a detailed followup survey in 1988 of two selected areas underlain by Mississippian sedimentary rocks. To our knowledge, this is the first study in the Western United States to describe the use of sagebrush in a drainage-basin type of geochemical survey. Erdman designed the study and conducted the geochemical surveys. Skipp mapped the geology, and Moye assessed the mineral potential using the geological and structural information. Theobald identified the heavy minerals.

RATIONALE FOR THE STUDY

The Lava Creek mining district is considered one of the most promising areas in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle for the occurrence of disseminated gold, based in large measure on the rock geochemical study by Hillier and others (1983). Other lines of evidence support this conclusion. Jasperoid (fig. 2), "a distinctive alteration type formed by intense silicification of marine sediments" (Nelson, 1990), is present as extensive outcrops in parts of the district. These outcrops were studied recently by Wilson and others (1988) and Soulliere and others (1988). The marine sedimentary rocks, turbidites of the Lower Mississippian McGowan Creek

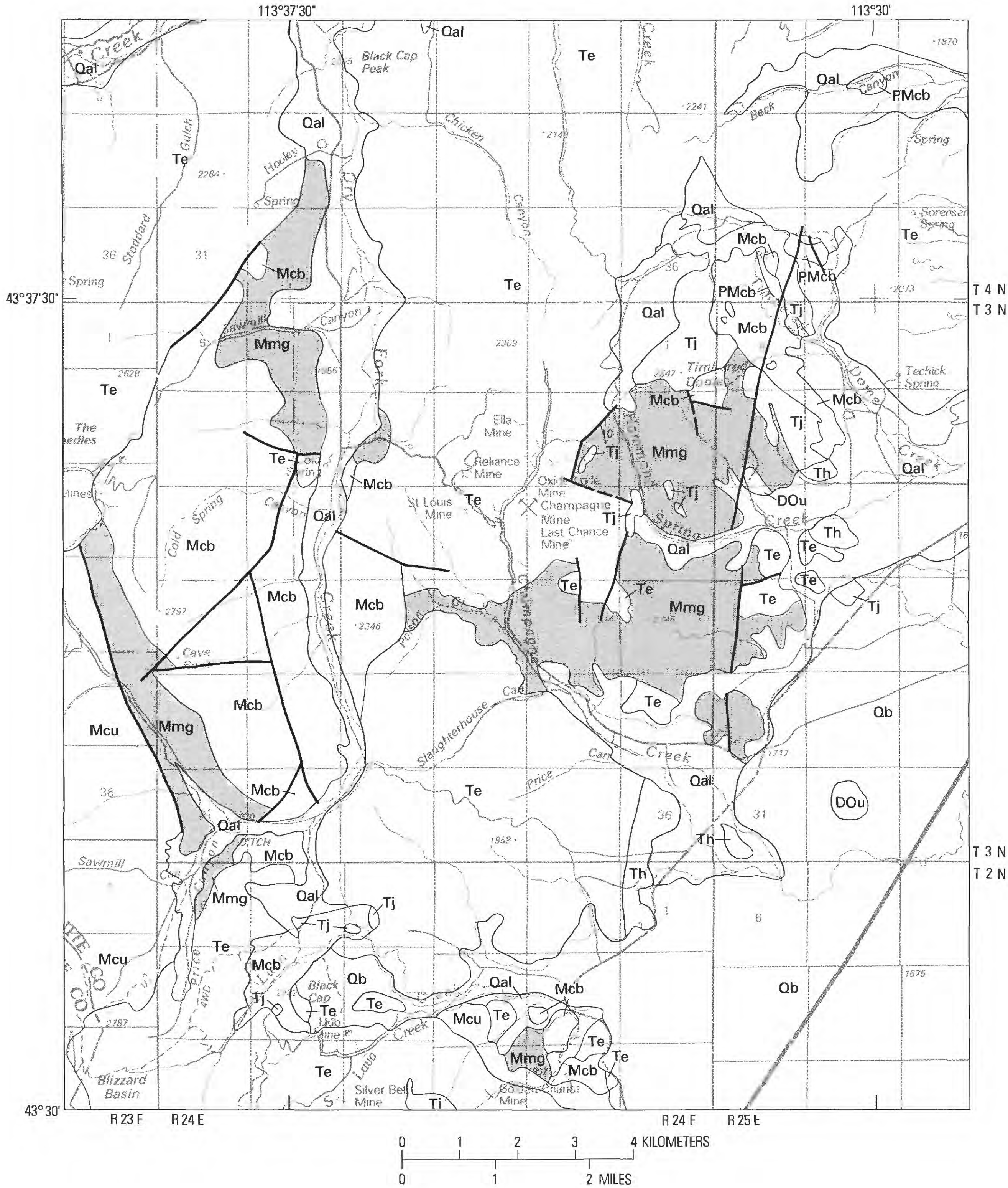


Figure 2 (above and facing column). Map showing generalized geology of the Lava Creek district, central Idaho. Mines in rocks of the Eocene Challis Volcanic Group are shown. Modified from Soulliere and others (1988). Base from U.S. Geological Survey Arco (scale 1:100,000) quadrangle (1988).

Formation, are one of the major sedimentary units exposed in the area and underlie much of the Eocene Challis Volcanic Group, which dominates the district. Carbonate facies of the McGowan Creek might have been physically reactive with hydrothermal fluids and, therefore, are potentially ideal sites for precipitation of metals. Recent papers on gold deposits in turbidite or black shale sequences (Boyle, 1986; Korobeynikov, 1986) are especially relevant to this study.

Acknowledgments.—This study was supported in part by a grant to Falma Moye from the Idaho State Board of Education. The plant samples were sent to the laboratory of Minerals Exploration and Environmental Geochemistry, Reno, Nevada, for preparation, and were analyzed by US Mineral Laboratories, Inc. (formerly Geochemical Services, Inc.), North Highlands, California, and Nuclear Activation Services, Ltd., Ann Arbor, Michigan. The heavy-mineral concentrates, stream sediments, and soils were analyzed in the laboratories of U.S. Geological Survey, Denver, Colorado; analysts: John Bullock (emission spectrography of sediments and soils), Betty Adrian and Mollie Malcolm (emission spectrography of heavy-mineral concentrates), Karen Slaughter (inductively coupled plasma analysis of sediments and soils), Phil Hageman (mercury analysis of stream sediments and soils by cold-vapor atomic absorption spectroscopy), and Rich O'Leary (gold analysis of sediments and soils by flameless atomic absorption spectroscopy). We thank Idaho Gold Corp. (Arco, Idaho), a subsidiary of Bema Gold, Inc. (Vancouver, Canada), for permission to sample at the Champagne mine.

GEOLOGIC SETTING

The geology of the Lava Creek mining district (Skipp and others, 1990) consists of Paleozoic sedimentary rocks overlain by rocks of the Eocene Challis Volcanic Group and intruded by cogenetic Eocene dikes, plutons and domes. Locally, these units are capped by Miocene basalt and rhyolite that are related to volcanism of the Snake River Plain.

Paleozoic strata in the area that range in age from Ordovician through Pennsylvanian are allochthonous and make up part of the Grouse thrust plate (Skipp, 1987; Link and others, 1988). Mississippian rocks include the McGowan Creek Formation at the base overlain by the Middle Canyon, Scott Peak, South Creek, Surrett Canyon, and Bluebird Mountain Formations. The Lower Mississippian McGowan Creek Formation is part of a westward-thickening wedge of flysch sediments deposited on the east side of the Antler foreland basin. It is about 2,000 ft (610 m) thick and is divided into three parts on the eastern flank of Timbered Dome (Skipp and others, 1990). The lower 350 ft (107 m) consists of interbedded siltstone, sandstone, and granite- to pebble-conglomerate; the middle 1,300 ft (396 m) consists of mudstone, siltstone, very fine grained sandstone, and minor limestone; and the uppermost 370 ft (113 m) consists mostly of gray laminated mudstone. West of the Champagne Creek thrust fault, however, the middle and upper parts of the McGowan Creek Formation contain minor gravel- to pebble-conglomerate and siltstone. The overlying Middle Canyon Formation, 508 ft (155 m) thick, comprises thin- to medium-bedded silty spiculitic limestone deposited in a foredeep in front of a prograding Late Mississippian carbonate bank. The remainder of the Mississippian formations are chiefly thick bedded, variably cherty, pure fossiliferous limestone of the Late Mississippian carbonate bank.

The Eocene Challis Volcanic Group is the main rock unit in the area and rests unconformably on the Paleozoic strata. The Challis Volcanic Group in the Lava Creek area includes a great variety of rock compositions. Radiometric ages (K/Ar and $^{40}Ar/^{39}Ar$) define a time span for the unit from 49 to 44 Ma (Moye, unpublished data); however, field relations suggest that the most voluminous activity was between 49 and 47 Ma. The basal Challis unit comprises a pre-volcanic conglomerate that grades from an orthoconglomerate lacking volcanic debris to a tuffaceous paraconglomerate that signifies the onset of volcanism. The earliest volcanism recorded in the area is dominantly andesite lava and tuff breccia. These rocks are 0–1,148 ft (350 m) thick and fill local paleotopography. They comprise a great number of individual flows that erupted from scattered vents throughout the region. Flows are generally porphyritic and contain plagioclase, pyroxene, and hornblende phenocrysts. This dominantly andesite volcanism was followed by a period of voluminous intermediate volcanism, including dacite to quartz latite lava flow rocks and ash-flow tuff and interbedded tuffaceous sedimentary rocks. The intermediate volcanic rocks are as much as 2,625 ft (800 m) thick, and a number of individual flow units were erupted from different sources. Dacite lava flow rocks are strongly porphyritic and contain as much as 40 percent phenocrysts of plagioclase, biotite, hornblende, and pyroxene in varied proportions. The flows generally are of small aerial extent and most likely were ponded in paleotopographic depressions. Dacite ash-flow tuff is poorly to densely welded, crystal-lithic tuff, and lithic content and size vary with distance from source. Sources for older dacite ash-flow tuff have been recognized to the north of the Lava Creek district near Lehman Basin. The final pulse of magmatic activity was small-volume explosive silicic activity and emplacement of rhyolite and dacite domes, dikes, and shallow granitic intrusions.

Evidence for both Mesozoic and Cenozoic tectonism is present in the structurally complex Lava Creek district. Paleozoic strata were thrust faulted and folded during middle to late Mesozoic compressional tectonism. Early Tertiary structures characteristically are northeast- and north-northwest-trending high-angle faults that show evidence of pre- and post-volcanic displacement. North-northeast-trending faults parallel with the margins of the Snake River Plain have only post-Eocene displacement.

MINERAL DEPOSITS

Historically, the Lava Creek district is primarily known for its silver deposits because during the main period of activity in the late 1800's most of the ore recovered was from oxidized bonanza silver ore. Recently, however, the new Champagne mine is an oxidized low-grade gold mine (The Mining Record, 1989).

Anderson (1929, 1947) provided the most detailed descriptions of ore deposits in the Lava Creek district. Ore deposits in the district are associated with the youngest silicic intrusive phase of the Challis Volcanic Group and are typically breccia or fissure deposits. The breccia deposits are generally considered to be hydrothermal explosion breccias that are localized along north-northwest-trending structures or at the intersection of northeast- and northwest-trending structures. Breccia zones show evidence of multiple periods of brecciation and silicification. Open-space textures are present but not common. Deposits are primarily in rocks of the Challis Volcanic Group or in cogenetic intrusive rocks, although there are some fissure deposits in limestone. In highly silicified hydrothermal breccia zones, original rock textures are partly to completely replaced by silica. Locally, tuffaceous sedimentary beds in the Challis Volcanic Group are pyritized and silicified but are not considered ore grade.

Ore of the Lava Creek mining district shows much mineralogical variation but can generally be classified as epithermal precious-metal deposits. These deposits also contain lead and zinc, as well as bismuth, tungsten, and antimony (Anderson, 1929). Silver was recovered from argentiferous galena, aikinite (a comparatively rare copper-lead-bismuth sulfide mineral that locally has high silver values), or surficial oxidized ore.

One of the most striking features about the Lava Creek district is the widespread hydrothermal alteration (Skipp, 1988; Skipp and others, 1990). Sericitization is the predominant alteration type; near vein margins rock textures are completely obliterated by sericite and quartz, and farther from veins more selective replacement of feldspar has occurred. In the Champagne Creek area, mineralized zones are characterized by complete silicification to chalcedonic quartz and by intense clay alteration to halloysite forming a dense white chertlike rock (Anderson, 1929).

Anderson (1929) suggested that deposits in the southwestern part of the district, in the Lava Creek area, formed at higher temperature than those in the Champagne Creek area. This is consistent with his observations, and our own, that the deposits to the southwest are associated with deeper granitic intrusive rocks and those in the Champagne Creek area are associated with shallow epizonal intrusions where hydrothermal systems may have vented to the surface as hot springs.

FIELD METHODS

RECONNAISSANCE-SCALE AND FOLLOWUP SURVEYS

A reconnaissance-scale survey of the district was conducted July 8–17, 1987, using the U.S. Geological Survey Grouse 1:62,500-scale topographic quadrangle as a base map. Two types of samples—heavy-mineral concentrates from stream sediments and current growth of big sagebrush—were collected at the mouths of 30 drainage basins of mostly first- and second-order streams (fig. 3). These sample localities were chosen to reflect areas of mineralized rock in the Lava Creek and Champagne Creek watersheds and to assess the possibility of mineralized rock in adjoining areas. The Timbered Dome area was included because of extensive outcrops of jasperoid. The sample localities include most of the critical parts of the district, except for the far western edge. Inclement weather forced an end to the field effort, and access to private holdings in the southwest part of the district was denied by the owner.

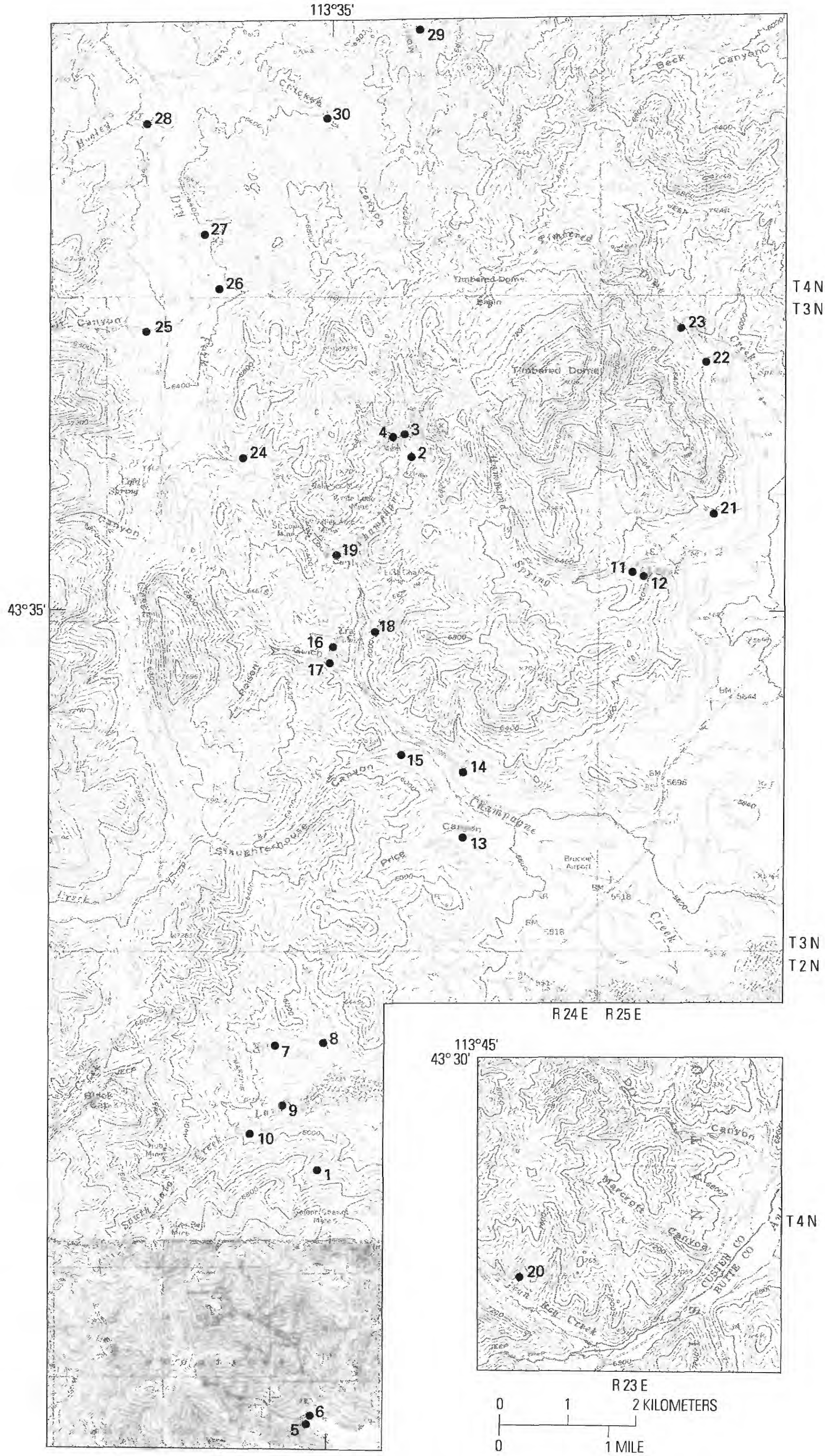
Based on the interpretation of analytical results from the reconnaissance survey, two areas were selected for a more detailed study May 19–25, 1988: Hammond Spring Creek on the east and Sawmill Canyon to the west (fig. 4). Samples of stream sediments (or, much less commonly, soils) and sagebrush were collected from 62 localities in these two areas, using advance sheets of the U.S. Geological Survey Mackay 4 Southeast and Southwest 1:24,000-scale topographic maps. Soil and sagebrush samples were also collected near four drillholes in exposed or thinly covered oxidized ore that is hosted by Challis Volcanic Group at the new Champagne mine (The Mining Record, 1989). Heavy-mineral concentrates were collected from three localities.

Analytical results for the area around Hammond Spring Creek were clearly more promising than those for Sawmill Canyon, and therefore we report only the former here.

SAMPLE MEDIA AND COLLECTION

HEAVY-MINERAL CONCENTRATES

The nonmagnetic fraction of a heavy-mineral concentrate sample is useful in detecting mineralized areas because primary and secondary ore minerals are commonly in this fraction. This is particularly true for uncommon minerals such as gold and cinnabar (mercury sulfide). The concentration of ore and ore-related minerals in these samples facilitates determination of elements that are not easily detected in bulk stream-sediment samples. The contrast in metal content between geochemical anomalies and normal background is greatly expanded such that anomalies which are fairly subtle in stream sediments may be strong and easily recognized.



113°35'
43°35'
T4N
T3N
T3N
T2N
R 24 E
R 25 E
113°45'
43° 30'
T4N
R 23 E
0
1
2 KILOMETERS
0
1 MILE
Timbered Dome
Champagne
Canyon
Slaughterhouse
Price
Creek
Spring
Dry
Canyon
Marcroft
Canyon
CUSTER CO
BUTTE CO
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30

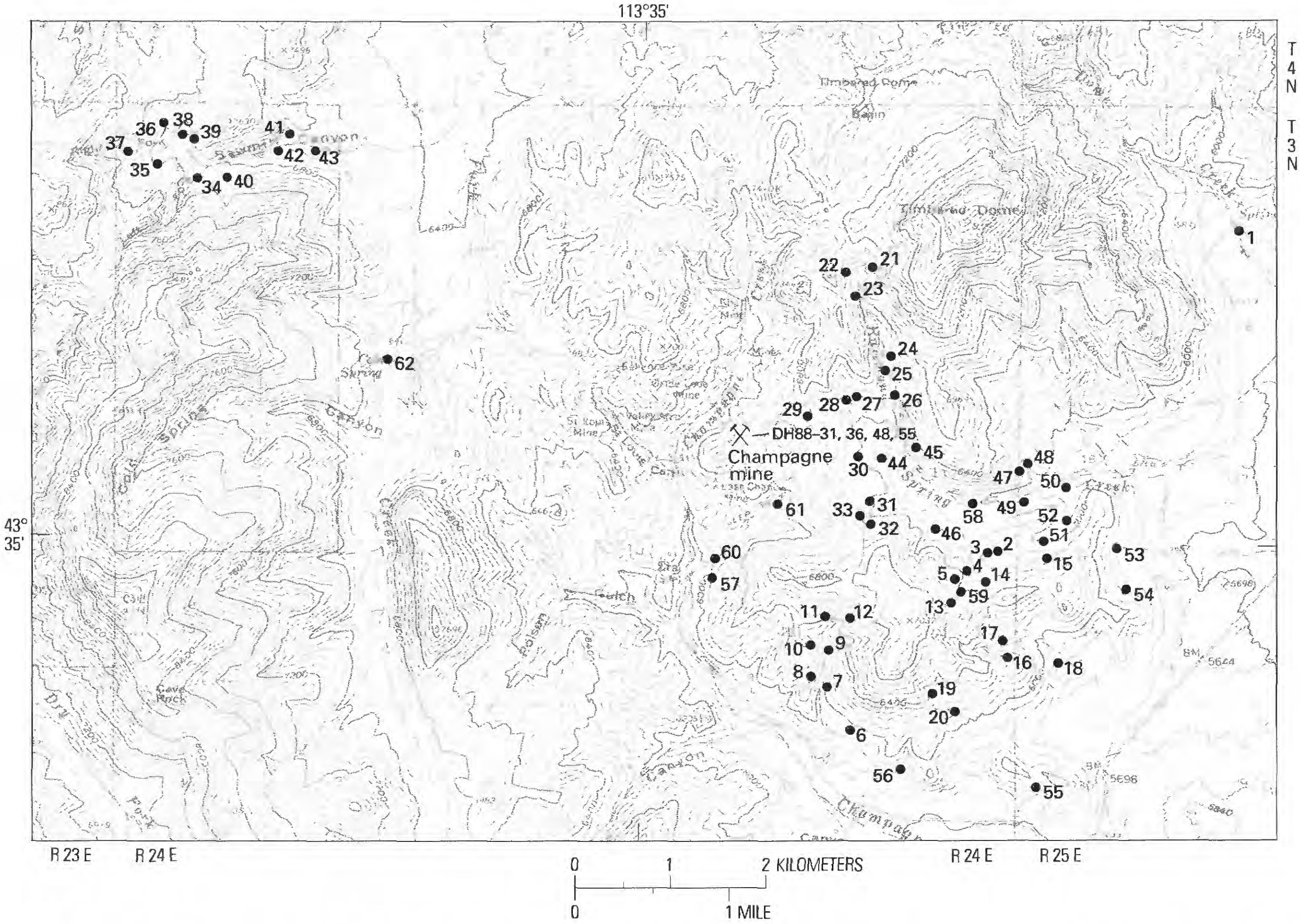


Figure 4. Map showing sample localities, followup geochemical survey, Lava Creek mining district, central Idaho. Base from U.S. Geological Survey Grouse (scale 1:62,500) quadrangle.

Heavy-mineral concentrates were separated from stream-sediment samples collected from sites favorable for the accumulation of heavy minerals in the active channels. The concentrate samples represent weathered surface-rock material mechanically transported to the sample locality. Each sample consisted of two 6- by 10-in. cloth bags filled with the sediments that passed a 10-mesh screen (2-mm opening). The <10-mesh fraction of the bulk sediment samples was panned in the field until most of the quartz, feldspar, clay, and organic matter was removed. This panned concentrate was then placed in paper bags for further processing in the laboratory.

Figure 3 (facing page). Map showing sample localities, reconnaissance-scale geochemical survey, Lava Creek mining district, central Idaho. Base from U.S. Geological Survey Grouse (scale 1:62,500), Inferno Cone (scale 1:24,000), and Blizzard Mountain South (scale 1:24,000) quadrangles.

STREAM SEDIMENTS AND SOILS

Stream-sediment samples represent the rock material eroded from the drainage basin upstream from each locality. Such information is useful in identifying those basins that contain concentrations of elements that may be related to mineral deposits. The stream-sediment samples consist of active alluvium collected, with few exceptions, from the bases of small ephemeral drainages.

In the followup survey, 54 stream-sediment samples and 2 soil samples were collected from 56 of the 62 localities shown on figure 4. Four additional soil samples were collected from the Champagne mine. Of the six localities where stream sediments were not collected, five (localities 1, 22, 58, 61, and 62) are springs and the sixth (locality 27) is the dump at a prospect pit. The sediment and soil samples were sieved to <10 mesh and placed in small HUBCO bags for further processing and analysis.

SAGEBRUSH

Plant samples provide information on soluble metals that are transported, under the conditions of this study, down the drainage basin and that are available at depth, although the roots of sagebrush are not known to extend downward more than several meters (Weaver and Clements, 1938, p. 320). Sturges (1977) reported that big sagebrush has both a fibrous root system that can feed near the surface and a tap-root that can draw moisture and nutrients from deep in the soil profile. Quantitative data on the depth and lateral spread of the root systems of big sagebrush are given in Tabler (1964), Sturges (1977), and Sturges and Trlica (1978).

Despite the relatively shallow rooting depths of sage-brush, Kovalevskii (1979) reported that *Artemisia* responds to geochemical features at depths of 40–60 m. He stated that plants may respond to orebodies 100 m or more below the surface where the roots tap hydrogeochemical haloes.

In the reconnaissance survey, each of the 30 big sage-brush samples was a composite of new growth (stems and leaves combined) from several shrubs that grew at the edge of the active channel of small streams or in the dry washes; sagebrush is usually not present in the channel proper where the flow is fairly constant. The samples were com-posites of clippings from several shrubs within about 5 m of one another; each sample of about 50 g was placed in a small cloth bag.

For the followup survey, leaves were stripped from the new growth and, to some extent, the previous year's growth. Sagebrush samples were collected from 58 of the 62 locali-ties shown in figure 4, and four additional samples were col-lected from localities at the Champagne mine. No sagebrush grew at four of the sample localities (34, 35, 39, and 40), in the upper reaches of Sawmill Canyon.

SAMPLE PREPARATION AND ANALYSIS

HEAVY-MINERAL CONCENTRATES

The panned concentrates were air dried and sieved to less than 0.425 mm (<35 mesh); the light minerals remaining in the <35-mesh fraction were removed by heavy-liquid flo-tation (bromoform, specific gravity 2.85). The heavy-min-eral concentrate samples were then separated into magnetic, weakly (para-) magnetic, and nonmagnetic fractions by plac-ing the sample in contact with the face of a large electromag-net (in this case a modified Frantz Isodynamic Separator). The most magnetic material (removed at a setting of 0.25 ampere), primarily magnetite, was not analyzed. The second fraction (removed at a setting of 1.75 ampere), mostly ferro-magnesian silicate minerals and iron and manganese oxide minerals, was archived; this paramagnetic fraction may contain limonite and manganese oxide minerals, which may contain high trace-metal values related to mineral deposits. The third fraction (the nonmagnetic material, which may include the nonmagnetic ore minerals zircon, sphene, and so on) was split using a Jones splitter.

One split of the nonmagnetic fraction was ground and chemically analyzed for 35 elements by direct-current arc emission spectrography (Grimes and Marranzino, 1968). The second split was used for mineralogical studies of indi-vidual grains with a conventional binocular microscope at low magnification (10 ×–30 ×).

STREAM SEDIMENTS AND SOILS

The stream-sediment and soil samples were air dried, then sieved using a 35-mesh stainless-steel sieve. The part of sample passing through the sieve was saved for analysis. Gold was determined by flameless atomic absorption spec-troscopy using a graphite furnace (O'Leary and Meier, 1986). Antimony, arsenic, bismuth, cadmium, and zinc were determined by inductively coupled argon plasma-atomic emission spectroscopy after a hydrochloric acid-hydrogen peroxide digestion (Crock and others, 1987). Mercury was determined by continuous-flow, cold-vapor atomic absorp-tion spectroscopy (Crock and others, 1987). A 35-element suite was determined by direct-current arc emission spec-trography (Grimes and Marranzino, 1968).

SAGEBRUSH

The sagebrush samples were sent to Minerals Explora-tion and Environmental Geochemistry, Reno, Nevada, for preparation before analysis. After washing, they were air dried for a day with a finish drying by microwave for 15 min-utes. The dried samples were then macerated and homoge-nized through a 2-mm sieve in a Wiley mill.

Splits of the 30 reconnaissance samples were further processed as follows. A 30-gram split of dry plant material was ashed at 480°C for 36 hours, and the remaining ash was sent to Geochemical Services Inc. (GSi), Rocklin, Califor-nia, for analysis. At the GSi labs, the ash was digested in a strong oxidizing acid mixture, heated to 80°C, and cooled. Metals were extracted from the ash into an organic solution and the solution analyzed by inductively coupled plasma emission spectroscopy for 13 trace elements and by graphite furnace atomic absorption for ultra-trace-amounts of gold. The suite of elements used from the GSi results include Ag, Cd, Cu, Ga, Mo, and Pb. An 8-gram split of dry material was pelletized to a 40-mm-diameter wafer for analysis of 17 ele-ments, including mercury, by direct instrumental neutron activation analysis by Nuclear Activation Services, Ann Arbor, Michigan. The suite of elements from the neutron activation analysis includes As, Au, Ba, Br, Co, Cr, Fe, Hg, Sb, U, W, and Zn.

The 62 sagebrush samples from the followup survey were prepared by Minerals Exploration and Environmental Geochemistry and the pelletized wafers sent to Nuclear Activation Services for analysis by direct instrumental neutron activation analysis. Analytical splits of six of the samples were submitted as a check on laboratory analytical precision, which was satisfactory except for antimony. The accuracy of the activation analysis method was checked by several internal plant standards that Minerals Exploration Geochemistry included in the sample suite; correspondence of the concentrations determined in these standards with established values was excellent.

RESULTS AND DISCUSSION

Geochemical anomalies were delineated using several criteria: published norms, frequency distributions in histograms, cumulative probability plots, and multivariate R-mode factor analysis (sagebrush only).

All of the analytical results are given in Adrian and others (1990).

RECONNAISSANCE-SCALE SURVEY

HEAVY-MINERAL CONCENTRATES

The heavy-mineral concentrates include a variety of accessory minerals from the country rocks and a few ore minerals. Three minerals—barite, apatite, and zircon—constitute the greater part of the concentrates, either singly or in combination. Most of the barite-rich samples are from drainage basins underlain by the Challis Volcanic Group and presumably reflect barite-bearing veins that cut these rocks. Apatite, with or without barite or zircon, is the most abundant mineral throughout the area. Zircon, although invariably present in the concentrates, is most abundant in the northern part of the area, around and to the north of Champagne Creek. Corundum is fairly common in concentrates from the northern part of the district. It has a mottled or zoned blue color and is in flat hexagonal plates in most of the samples. In samples from some of the northern localities it is more blocky, favoring a dipyramidal form, and has a purple tint that resembles fluorite, which was not identified in any of the samples.

Pyrite is present in most concentrate samples, even those from drainages underlain by Paleozoic sedimentary rocks in the Timbered Dome area. No pyrite was identified in the sample from locality 25 at Sawmill Canyon or in samples from localities 11 and 12 in the Hammond Spring Creek drainage where the sagebrush samples were anomalous (fig. 5).

Ore minerals in the concentrate samples include cinnabar, sphalerite, cuprite, arsenopyrite, cerussite, lead, and native gold. Cinnabar, gold, cuprite, and, in some places, arsenopyrite are present in samples from a tight cluster of localities around Lava Creek (fig. 5). Sphalerite, cerussite, and galena in concentrates close to Champagne Creek

Table 1. Summary of selected data for 30 heavy-mineral concentrate samples from the reconnaissance-scale geochemical survey, Lava Creek mining district, Idaho.

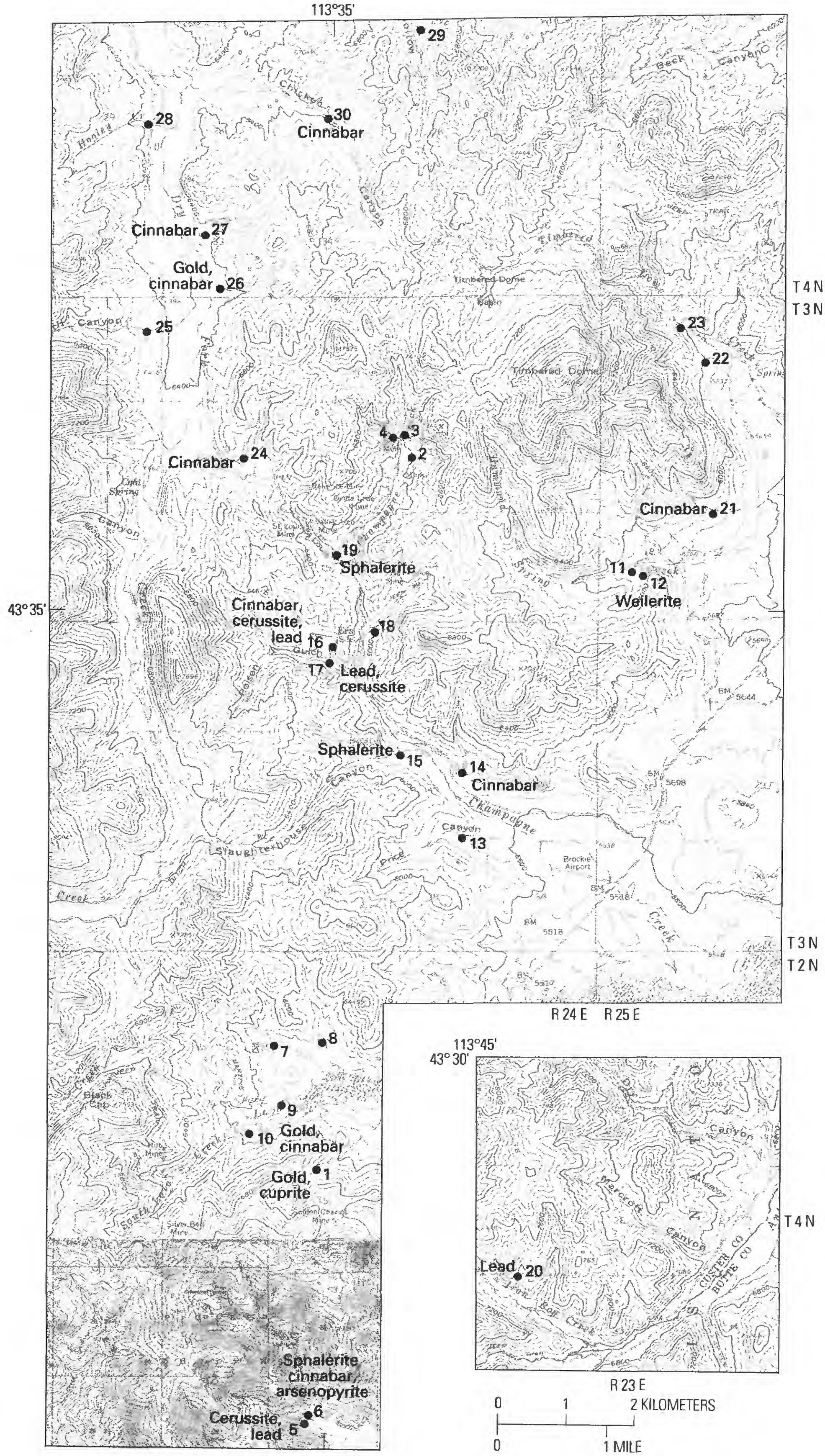
[In parts per million except where noted. All elements by emission spectrography; see text for details. Ratio is the number of samples in which the element was detected to number of samples analyzed]

Element	Minimum	Maximum	Median	Ratio
Fe (percent)	<0.1	7	0.5	29:30
Ag	<1	500	<1	6:30
As	<500	>20,000	<500	1:30
Au	<20	150	<20	2:30
B	<20	200	20	24:30
Ba	1,500	>10,000	>10,000	30:30
Bi	<20	150	<20	2:30
Cu	<10	2,000	<10	14:30
Pb	<20	10,000	30	27:30
Sb	<200	300	<200	2:30
Sn	<20	>1,000	<20	10:30
W	<50	100	<50	1:30
Zn	<500	>20,000	<500	2:30

Table 2. Summary of data for 30 sagebrush samples from the reconnaissance-scale geochemical survey, Lava Creek mining district, Idaho

[In parts per million unless otherwise noted. Concentrations on a dry-weight basis; see text for details. Ratio is the number of samples in which the element was detected to number of samples analyzed]

Element	Minimum	Maximum	Median	Ratio
Neutron activation analysis				
As	<0.01	0.13	0.04	29:30
Au (ppb)	<0.2	1.0	0.3	20:30
Ba	<20	50	20	14:30
Br	1.2	17	3.7	30:30
Co	<0.3	0.6	<0.3	12:30
Cr	<0.3	1.0	<0.3	15:30
Fe (percent)	0.006	0.023	0.011	30:30
Hg (ppb)	<50	60	<50	2:30
Sb	0.04	0.36	0.06	30:30
Se	<0.5	1.8	<0.5	3:30
U	<0.02	0.05	<0.02	10:30
W	<~0.04	0.10	<~0.04	6:30
Zn	23	65	30	30:30
Inductively coupled plasma emission spectrometry				
Ag (ppb)	1.1	31	4.5	30:30
Cd (ppb)	19	300	45	30:30
Cu	3.3	21	11	30:30
Ga (ppb)	<28	48	<28	5:30
Mo (ppb)	64	660	240	30:30
Pb (ppb)	140	2,400	210	30:30


113°35'
29
30
Cinnabar
28
Cinnabar 27
Gold,
cinnabar 26
25
T4N
T3N
23
22
4
3
2
Cinnabar 24
Cinnabar 21
19
Sphalerite
11
12
Weilerite
43°35'
Cinnabar,
cerussite,
lead 16
18
17 Lead,
cerussite
Sphalerite 15
14
Cinnabar
13
T3N
T2N
R 24 E R 25 E
8
7
9
10 Gold,
cinnabar
Gold, 1
cuprite
Sphalerite
cinnabar,
arsenopyrite
Cerussite, 6
lead 5
113°45'
43° 30'
T4N
Lead 20
R 23 E
0 1 2 KILOMETERS
0 1 MILE

presumably reflect contamination from the mining activity in that area. Cinnabar and gold are peripheral to the area of mining at Champagne Creek and extend northward along the east side of Dry Fork Creek.

A single grain of weilerite was identified in the concentrate from locality 12 (fig. 5). Weilerite is a barium aluminum arsenate that can form in a hydrothermal environment (Botinelly, 1976). Although its singular presence is scant evidence for such an environment, it supports other more convincing data described below. Interestingly, no barite was identified in the concentrate sample containing weilerite.

The minerals incorporated in the heavy-mineral concentrates must be derived from surface material. Thus, the ore minerals in the concentrates neatly define the areas of exposed mineralization, mainly in rocks of the Challis Volcanic Group. Except for cinnabar at localities 14 and 21 in the Mississippian sedimentary rocks (fig. 5), these concentrates correlate with known mines and prospects.

The distribution of concentrates that have anomalous element concentrations generally mirrors the distribution of the ore-related minerals described above. Further discussion of the results therefore is unwarranted; however, data for selected elements are summarized in table 1.

SAGEBRUSH

Analytical data for the 30 sagebrush samples are summarized in table 2. Because of excessive censoring (>50 percent), gallium, mercury, selenium, uranium, and tungsten are not included in the multivariate interpretation described following. Cobalt and chromium are not included, both because of censoring and because of their limited importance.

The gold concentrations, dry-weight basis, of sagebrush range from less than 0.2 to 1.0 ppb, and the median is 0.3 ppb. We tested the possibility that the skewed distribution represents two populations by using a statistic proposed by Miesch (1981). The statistic is a standardized gap, which, if significant, can be taken as the separation between two geochemical populations. The results indicate a significant gap at 0.35 ppb (confidence level, ~90 percent); 10 of the 30 samples contain gold at anomalous levels of 0.4 ppb or greater.

A varimax solution of R-mode factor analysis, widely used in exploration geochemistry, reduced a log-transformed data base of 11 elements to help define interrelated element assemblages. (The censored data were first replaced with values 0.7 of the lower limits of determination because the procedure will not accept qualified data. This modified data set was then transformed to a log base 10 because of the tendency of geochemical data to be positively skewed.) Molybdenum was eliminated from the data set because it constituted a single-element factor when a five-factor solution was selected. A four-factor model proved to be the most interpretable. Correlations between the factor scores and the concentrations of each dominant element (correlation coefficients >0.3) for each factor are given in figure 6. Approximately 70 percent of the standardized data are accounted for in this four-factor model.

We consider factor three, in the third column, most important because the key elements antimony, gold, and, to a lesser extent, bromine reflect an epithermal precious-metal suite. Colin Dunn (Geological Survey of Canada, oral commun., 1990) has also found a weak positive correlation between bromine and gold in his biogeochemical studies in the boreal forests of Canada. In his work in northern Saskatchewan, for example, he noted "that Br levels in ashed alder twigs are commonly elevated near gold mineralization" (Dunn and Hoffman, 1986, p. 378). This association of bromine and gold in plant tissue is probably widespread. Leslie Thompson (Gold Fields Mining Corp., Denver, oral commun., April 1990) found that bromine is associated with gold, arsenic, and antimony in samples of mesquite (*Prosopis* sp.) and creosote-bush (*Larrea tridentata* [DC.] Coville) from the deserts of southern California.

Factor scores represent a measure of the effect of a factor on each sample. The locations of those sagebrush samples having higher scores for the antimony-gold-bromine factor (fig. 7) point not only to the Lava Creek and Champagne Creek subdistricts, especially where cinnabar is present in the concentrates, but also to Sawmill Canyon and Hammond Spring Creek in terrain characterized by turbidite of the McGowan Creek Formation (Erdman and others, 1988). Sagebrush samples from these latter two areas contained the only detectable mercury and selenium, but cinnabar was identified in the associated concentrates only at locality 14.

Although chemical anomalies were identified in concentrates from drainage basins in and adjacent to mines in rocks of the Challis Volcanic Group of the Lava Creek district, none were associated with anomalies in sagebrush that grew in sediments derived from the McGowan Creek Formation. The source of the Sawmill Canyon and Hammond Spring Creek biogeochemical anomalies appeared to be hidden.

FOLLOWUP SURVEY

Results from the 10 localities sampled in Sawmill Canyon (fig. 4) indicate no prospective target, whereas results from the Hammond Spring Creek area, which is almost entirely underlain by the McGowan Creek Formation, led us

Figure 5 (facing page). Map showing distribution of selected ore minerals in heavy-mineral concentrates, reconnaissance-scale geochemical survey, Lava Creek mining district, central Idaho.

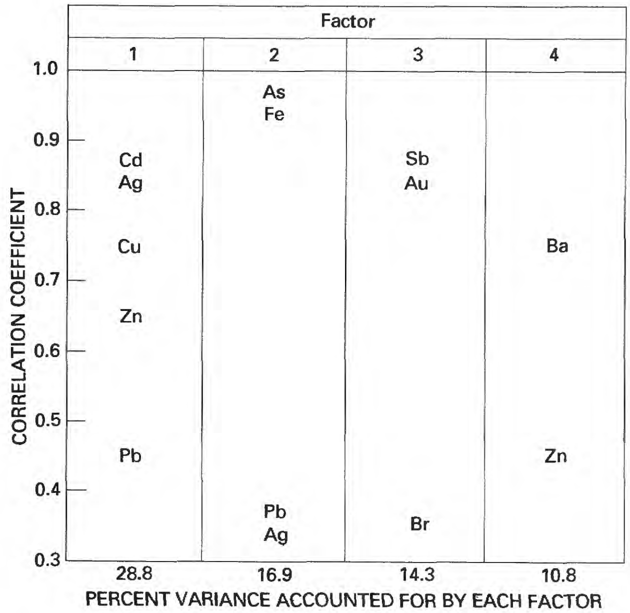


Figure 6. Factor compositions for sagebrush from reconnaissance-scale survey, Lava Creek mining district, central Idaho. Elements are arranged in descending order of correlation coefficient, which indicates importance of element.

Table 3. Summary of selected data for 60 stream-sediment and soil samples from the followup geochemical survey, Lava Creek mining district, Idaho

[All values in parts per million except where noted; see text for details. Ratio is the number of samples in which the element was detected to number of samples analyzed]

Element	Minimum	Maximum	Median	Ratio
Graphite-furnace atomic absorption spectrometry				
Au (ppb)	<2	70	<2	8:60
Emission spectrography				
Ag	<0.5	5	<0.5	23:60
B	<10	300	50	59:60
Pb	<10	150	10	51:60
V	70	700	150	60:60
Inductively coupled plasma spectrometry				
As	<5	48	20	57:60
Bi	<2	6	<2	2:60
Cd	0.1	3.1	0.7	60:60
Sb	<2	9	2	32:60
Zn	21	490	94	60:60
Cold-vapor atomic absorption spectrometry				
Hg	<0.02	1.2	0.08	56:60

to a prospective target at the head of a small unnamed tributary (Erdman and others, 1989). For this reason, we report only the latter results here.

From casual observation, this unnamed tributary (fig. 8) lacks any clear evidence of mineralized rocks; however, a well-exposed fault breccia is present at the very head of the tributary valley. Slickensides were seen on float at several of the sample localities near the head of the tributary. A small, propylitically altered dacite porphyry intrusive body is near the north-northeast-trending fault breccia. The altered dacite porphyry and explosive hydrothermal breccia may constitute the upper part of a breccia pipe. The geochemical results for both sediments and sagebrush, and the occurrence of cinnabar in concentrates, may indicate leakage of hydrothermal solutions from depth.

STREAM SEDIMENTS AND SOILS

Data for selected elements from the 60 sediment and soil samples are summarized in table 3. Few of the samples contain detectable gold; of the eight samples in which gold was detected, the three highest concentrations, 10, 40, and 70 ppb, are in samples from the Champagne mine. The five remaining samples contain only 2 ppb gold and are from widely scattered localities. Bismuth concentrations also are above the detection limit only in soils from the Champagne mine; the two samples in which it was detected (4 and 6 ppm) are from the mine. Lead concentrations are also anomalous only in soils from the mine; three of the four samples from the mine contain from 70 to 150 ppm Pb, as compared to samples elsewhere that contain from less than 10 to 30 ppm Pb.

Soils associated with oxidized gold ore at the Champagne mine are characterized by a unique gold-bismuth-lead suite; sediments derived from the turbidite are characterized by a unique suite of boron, cadmium, vanadium, and zinc. Boron concentrations of 100–300 ppm are associated with mineralized rock in the McGowan Creek Formation, as compared to background amounts of from less than 10 to 70 ppm. Threshold values (and background ranges) of the other remaining three elements are cadmium, >1 ppm (0.1–0.8); vanadium, ≥500 ppm (70–200); and zinc, ≥150 ppm (21–100).

In general, soils from the Champagne mine and sediments from the fault-breccia area yielded above-median concentrations of silver, arsenic, and mercury. Because of its volatility, mercury is used as a pathfinder for buried metallic orebodies (Maciolek and Jones, 1987). As shown in figure 9,

Figure 7 (facing page). Map showing location of sagebrush samples having high scores for the antimony-gold/bromine factor, reconaissance-scale geochemical survey, Lava Creek mining district, central Idaho. Large asterisk indicates greater than 90th percentile; small asterisk indicates 80th–90th percentile.

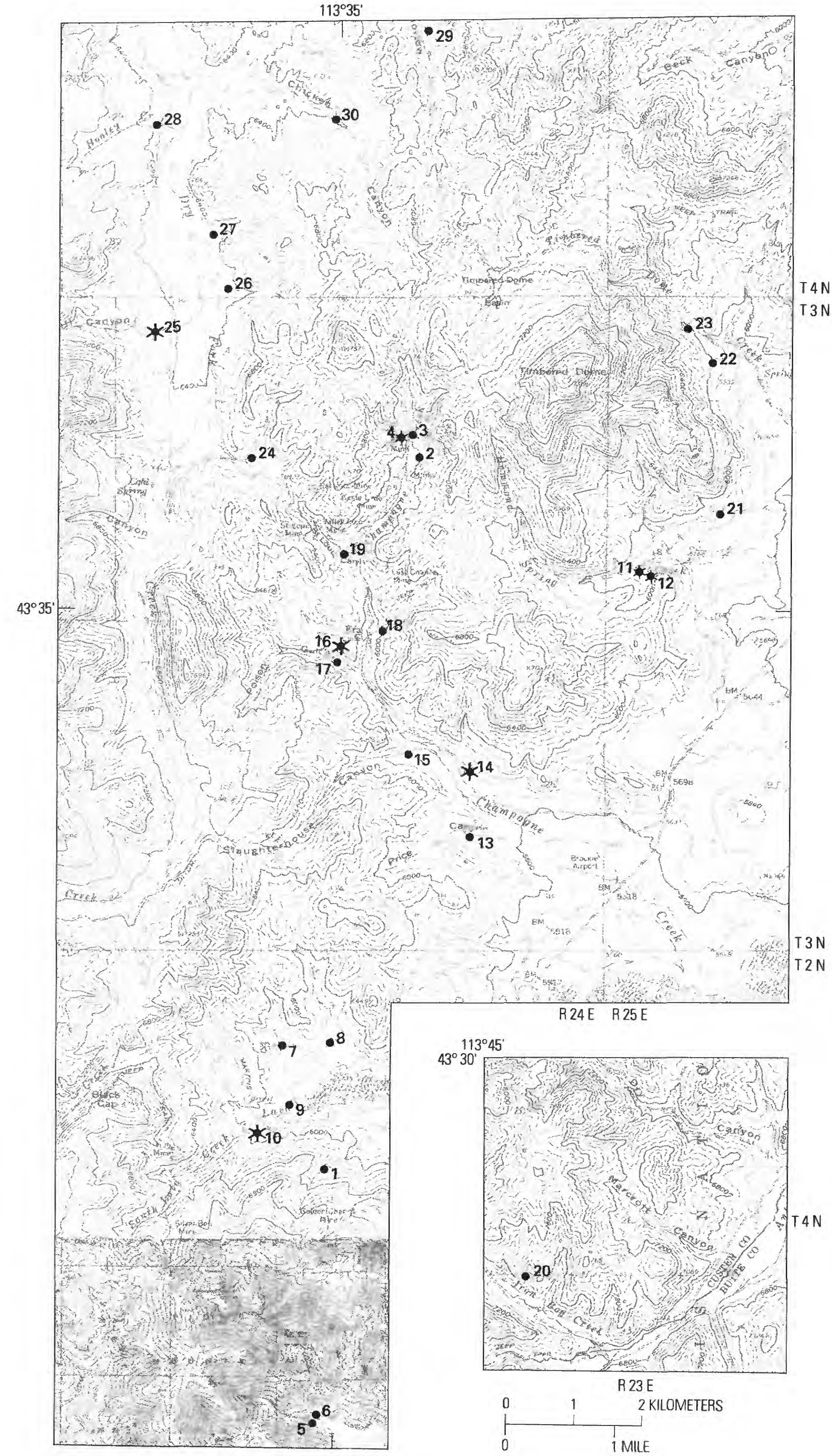

113°35'
43°35'
T4N
T3N
T3N
T2N
R 24 E R 25 E
113°45'
43° 30'
T4N
R 23 E
0 1 2 KILOMETERS
0 1 MILE
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
Champagne
Creek
Timbered Dome
Slaughterhouse
Canyon
Price
Beck Canyon
Marcott Canyon
Iron Bog Creek
CUSTER CO
BUTTE CO

Figure 8. Mineralized target area, a silicified and brecciated, north-northeast-trending fault in the McGowan Creek Formation, Lava Creek mining district, central Idaho.

the mercury content delineates a broad halo over the fault-breccia area. Anomalous mercury concentrations in sediments from a tributary of Champagne Creek west of the divide from the Hammond Spring Creek drainage system suggest that the mineralized area associated with the fault might be more extensive.

In contrast to the distribution pattern of anomalous mercury in sediments, anomalous concentrations of silver delineate a more restricted target that coincides with the fault (fig. 10). Samples from localities peripheral to this target contain silver at or below the 0.5-ppm limit of determination.

In the course of sampling, Erdman observed that *Eriogonum ovalifolium* Nutt., one of many species of wild buckwheat, dominated parts of a slope that lay on strike with the mineralized fault (fig. 11). A century ago Lidgey (1897) reported, "In Montana experienced miners look for silver wherever the '*Eriogonum ovalifolium*' flourishes. This plant grows in low dense bunches; its small leaves coated with thick white down." A soil sample from the locality where this plant is abundant contains 3 ppm Ag, second only to a

Table 4. Summary of data for 62 sagebrush samples from the followup geochemical survey, Lava Creek mining district, Idaho
[Asterisk (*) indicates element was used in factor analysis. Concentrations on a dry-weight basis. In parts per million unless otherwise noted. All elements by neutron activation analysis; see text for details. Ratio is the number of samples in which the element was detected to number of samples analyzed]

Element	Minimum	Maximum	Median	Ratio
*As	<0.01	0.12	0.03	59:62
*Au (ppb)	<0.1	1.0	0.14	29:62
*Ba	<20	130	30	59:62
*Br	0.69	14	1.6	62:62
Co	<0.3	0.5	<0.3	6:62
Cr	<0.3	0.9	0.5	50:62
*Fe (percent)	<0.005	0.017	0.008	58:62
Hg	<50	50	<50	4:62
*Mo	0.18	1.4	0.50	62:62
*Sb	0.03	0.15	0.045	62:62
*Se	<0.5	6.1	<0.5	27:62
U	<0.2	0.07	<0.2	7:62
*W	0.09	2.0	0.46	62:62
*Zn	31	100	50	62:62

soil sample from the Champagne mine that contains 5 ppm Ag. Species of the genus *Eriogonum* are commonly associated with copper and molybdenum occurrences in the Western United States (U.S. Geological Survey, unpublished data), but this is the first case where *Eriogonum* seems to be so strongly controlled by a silver-rich substrate.

SAGEBRUSH

Data for selected elements from the 62 sagebrush samples are summarized in table 4. R-mode factor analysis was used with 10 of the elements listed in the table. The optimum model derived from this analysis was again four factors, as shown in figure 12. This data set represents sagebrush growing on country rock dominated by black marine shale of the McGowan Creek Formation. Factor one is characterized by the element suite, in descending order of importance, arsenic, antimony, iron, bromine, and gold.

The geochemical signature of sagebrush associated with both the Champagne mine and the fault consists of the element assemblage gold, antimony, arsenic, bromine, and barium. The highest gold values, 1.00 and 0.80 ppb, are in two samples of sagebrush collected above the oxidized orebody at the Champagne mine; gold concentrations in the other two samples of sagebrush collected at the Champagne mine are only slightly anomalous at 0.4 ppb. Overall, the gold results for sagebrush are more useful than those for the sediments, but we do not recommend reliance on gold biogeochemical data alone.

Concentrations of selenium and zinc are anomalous in samples of sagebrush from suspected mineralized localities in the McGowan Creek Formation but not from the Champagne mine. Evidence from our data, as well as from data from similar marine sedimentary rocks north of the Lava Creek district (Sandra Soulliere, U.S. Geological Survey, unpublished data), suggests that marine black shale (turbidite) of the McGowan Creek Formation is enriched in selenium and zinc and that this enrichment is reflected in the sagebrush growing in soils derived from this shale. Where these elements have been remobilized through hydrothermal processes, the selenium content of sagebrush from altered areas is well above the normal maximum concentration of 1.1 ppm in sagebrush in the Western United States (Gough and Erdman, 1983).

Because arsenic dominates the element suite that characterizes factor one, the distribution of anomalous arsenic in sagebrush mirrors the distribution of high scores for this factor. Arsenic is a well-known pathfinder element in gold exploration and is more easily interpreted than multielement factors. The pattern of high arsenic in sagebrush (fig. 13) is remarkably similar to the pattern of high mercury in sediments (fig. 9) and reflects the mobility of arsenic as an anion in arid environments.

The distribution of anomalous tungsten in sagebrush (fig. 14) centers on the single drainage just to the southwest of the zone of silver-rich sediments and the fault breccia. By itself tungsten might not warrant further notice, but the associated mercury anomalies in the sediments and the presence of cinnabar in the concentrate from the reconnaissance survey raise the possibility of an epithermal precious-metal occurrence. In his biogeochemical studies of gold occurrences in turbidite of Nova Scotia, Dunn (1989) found zones of weak tungsten enrichment near some gold deposits.

HEAVY-MINERAL CONCENTRATES

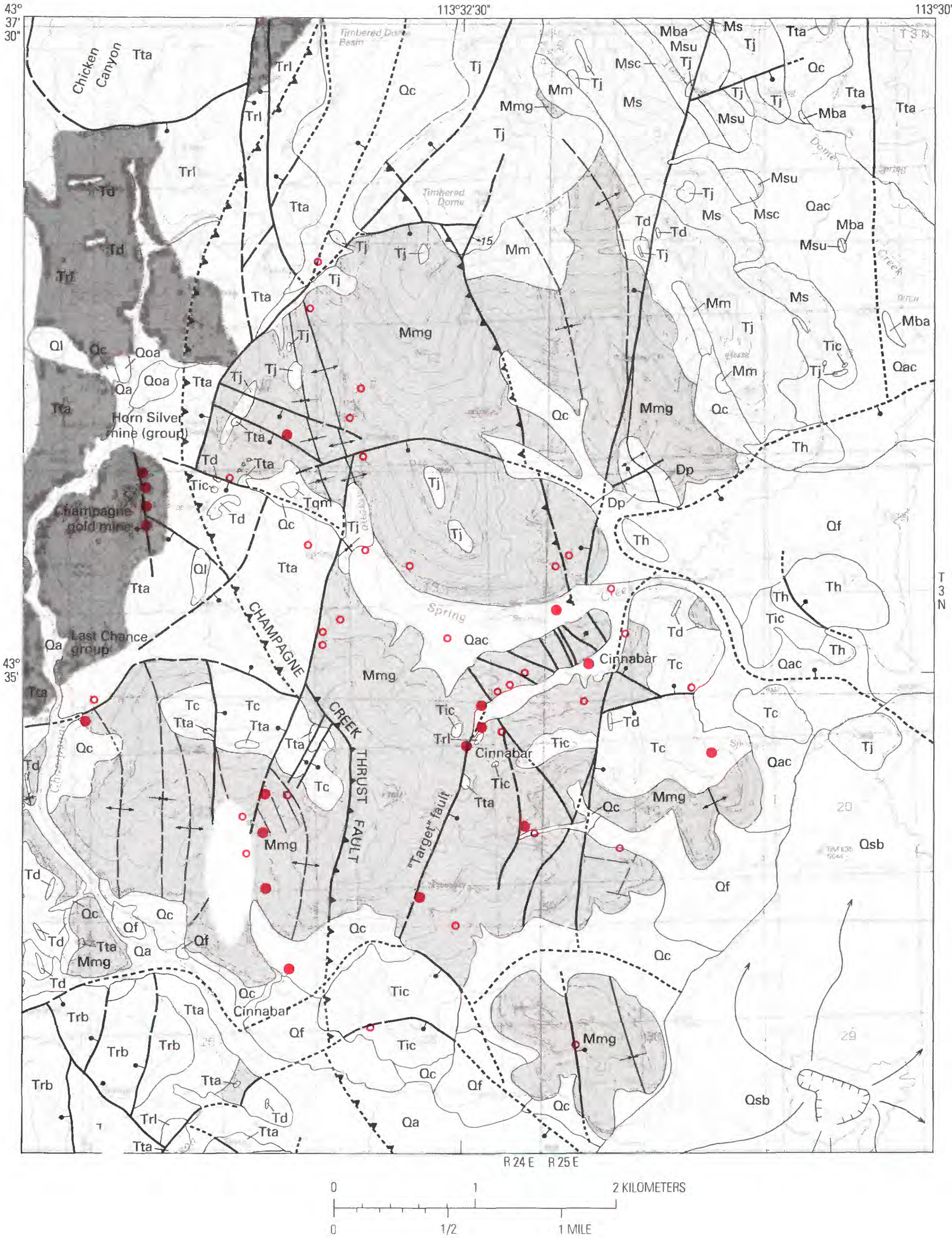
A microscopic inspection of a heavy-mineral concentrate sample taken immediately downstream from the breccia zone (fig. 8) shows that barite is dominant in the mineral assemblage and that sparse, small fragments of probably sedimentary pyrite and, most importantly, a single grain of cinnabar are also present. The cinnabar provides evidence of low-temperature mineralization associated with the breccia exposure upstream and supports the strong mercury anomalies in the sediments.

Analysis of the powdered split of this concentrate produced a 5-ppm Ag anomaly and an unexplained 2,000-ppm La anomaly but no base-metal anomalies. The silver anomaly is consistent with the 3-ppm Ag anomaly in the soil sample on a nearby slope that supported the indicator plant, *Eriogonum ovalifolium.*

CONCLUSIONS

We now have evidence, initially obtained using plant chemistry, of epithermal precious-metal mineralization in the Hammond Spring Creek drainage of the Lava Creek mining district, Idaho, east of the outcrop area of the Eocene Challis Volcanic Group, in an area where the country rock is predominantly turbidite of the Lower Mississippian McGowan Creek Formation. The key elements that drew us to this conclusion are gold, antimony, and bromine and, to a lesser extent, arsenic, mercury, and selenium. On the basis of analytical results in samples of big sagebrush and heavy-mineral concentrates from a reconnaissance-scale survey, the source was considered blind.

A followup survey of two selected areas in the turbidite terrain, during which sagebrush and stream sediments were sampled, reveals that the most prospective area for metallization is on the east edge of the Lava Creek district: an explosive hydrothermal breccia and weakly exposed, propylitically altered dacite porphyry intrusive body subjacent to a clearly defined north-northeast-trending fault. The geochemical results and the presence of cinnabar in concentrates from the anomalous area may indicate


43° 37' 30"
113°32'30"
113°30'
43° 35'
T 3 N
R 24 E R 25 E
Chicken Canyon
Horn Silver mine (group)
Champagne gold mine
Last Chance group
CHAMPAGNE CREEK THRUST FAULT
"Target" fault
Spring Creek
Cinnabar
Tta
Trl
Td
Tj
Qc
Qoa
Qa
Ql
Mmg
Mm
Ms
Msu
Msc
Mba
Tic
Tc
Th
Dp
Qf
Qac
Qsb
Trb
Tqm
0 1 2 KILOMETERS
0 1/2 1 MILE

EXPLANATION

Qa	Alluvium (Holocene)
Qc	Colluvium (Holocene and Pleistocene)
Qac	Alluvium and colluvium, undifferentiated (Holocene and Pleistocene)
Qf	Alluvial fan deposits (Holocene and Pleistocene)
Ql	Landslide deposits (Holocene and Pleistocene)
Qoa	Older alluvial deposits (Holocene and Pleistocene)
	LAVA FLOWS AND RELATED DEPOSITS OF THE SNAKE RIVER GROUP
Qsb	Pahoehoe basalt flows and shield volcano (Pleistocene)
	ROCKS OF THE HEISE VOLCANIC FIELD
Th	Tuffs of Blue Creek and Blacktail, undivided (Miocene)
	IDAVADA VOLCANICS
Tic	Tuffs of Little Chokecherry Canyon (Miocene)
	SILICIFIED ROCK
Tj	Jasperoid (Eocene or younger)
	CHALLIS VOLCANIC GROUP
Td	Dikes
Trb	Rhyodacite breccia
Trl	Rhyodacite lava flows
Tqm	Quartz monzonite porphyry
Tta	Tuffs of Antelope Creek and granite porphyry of Anderson (1947)
Tc	Basal conglomerate, breccia, and sandstone
	SEDIMENTARY ROCKS
Mba	Bluebird Mountain and Arco Hills formation, undivided (Upper Mississippian)
Msu	Surrett Canyon Formation (Upper Mississippian)
Msc	South Creek Formation (Upper Mississippian)
Ms	Scott Peak Formation (Upper Mississippian)
Mm	Middle Canyon Formation (Upper and Lower Mississippian)
Mmg	McGowan Creek Formation (Lower Mississippian)
Dp	Picabo(?) Formation (Upper Devonian)

Contact—Dashed where approximately located

15

Fault—Dashed where approximately located; dotted where concealed. Bar and ball on downthrown side. Single arrow indicates apparent relative lateral movement. Arrow and number (in degrees) indicate dip of fault plane

Thrust fault—Sawteeth on upper plate. Dashed where approximately located; dotted where concealed

Photo lineament—Probable fault or fracture having undetermined offset

Anticline—Showing trace of axial plane. Dashed where approximately located; dotted where concealed

Syncline—Showing trace of axial plane. Dashed where approximately located; dotted where concealed

Argillic alteration

Tectonic breccia

Flow direction in basalt

Outline of basalt vent crater

leakage of hydrothermal solutions from depth. The source of the anomaly in the area on the west side of the district was not identified.

The unique response of sagebrush to fairly subtle metallization in the Lava Creek district led to the discovery of a silver- and mercury-rich brecciated zone in the Mississippian turbidite and underscores the value of biogeochemical methods in an initial mineral assessment. The followup survey of selected areas to locate the sources of the geochemical anomalies shows, however, that either sediments or sagebrush would have provided equally satisfactory results.

REFERENCES CITED

Adrian, Betty, Erdman, J.A., Bullock, J.H., and Hageman, P.L., 1990, Analytical results and locality map of heavy-mineral-concentrate, stream-sediment, soil, and big sagebrush samples from the Lava Creek mining district, within the Idaho Falls 1°×2° quadrangle, Butte, County, Idaho: U.S. Geological Survey Open-File Report 90–545.

Anderson, A.L., 1929, Geology and ore deposits of the Lava Creek district, Idaho: Idaho Bureau of Mines and Geology Pamphlet 32, 70 p.

———1947, Epithermal mineralization at the Last Chance and Hornsilver Mines, Lava Creek district, Butte County, Idaho: Geological Society of America Bulletin, v. 58, p. 451–482.

Botinelly, T., 1976, A review of the minerals of the alunite-jarosite, beudantite, and plumbogummite groups: U.S. Geological Survey Journal of Research, v. 4, p. 213–216.

Boyle, R.W., 1984, The prospect for geochemical exploration—Predictable advances and new approaches: Journal of Geochemical Exploration, v. 21, p. 1–18.

———1986, Gold deposits in turbidite sequences—Their geology, geochemistry and history of the theories of their origin, *in* Keppie, J.D., Boyle, R.W., and Haynes, S.J., eds., Turbidite-hosted gold deposits: Geological Association of Canada Special Paper 32, p. 1–13.

Crock, J.G., Briggs, P.H., Jackson, L.L., and Lichte, F.E., 1987, Analytical methods for the analysis of stream sediments and rocks from Wilderness Study Areas: U.S. Geological Survey Open-File Report 87–84, 35 p.

Figure 9 (left and facing page). Map showing location of stream-sediment and soil samples that contain anomalous amounts of mercury (≥0.10 ppm)(solid red circles), followup survey, Lava Creek mining district, central Idaho. Open red circles indicate samples containing less than 0.10 ppm Hg; observed range of values <0.02–1.2 ppm Hg. Geology modified from Skipp and others (1990).

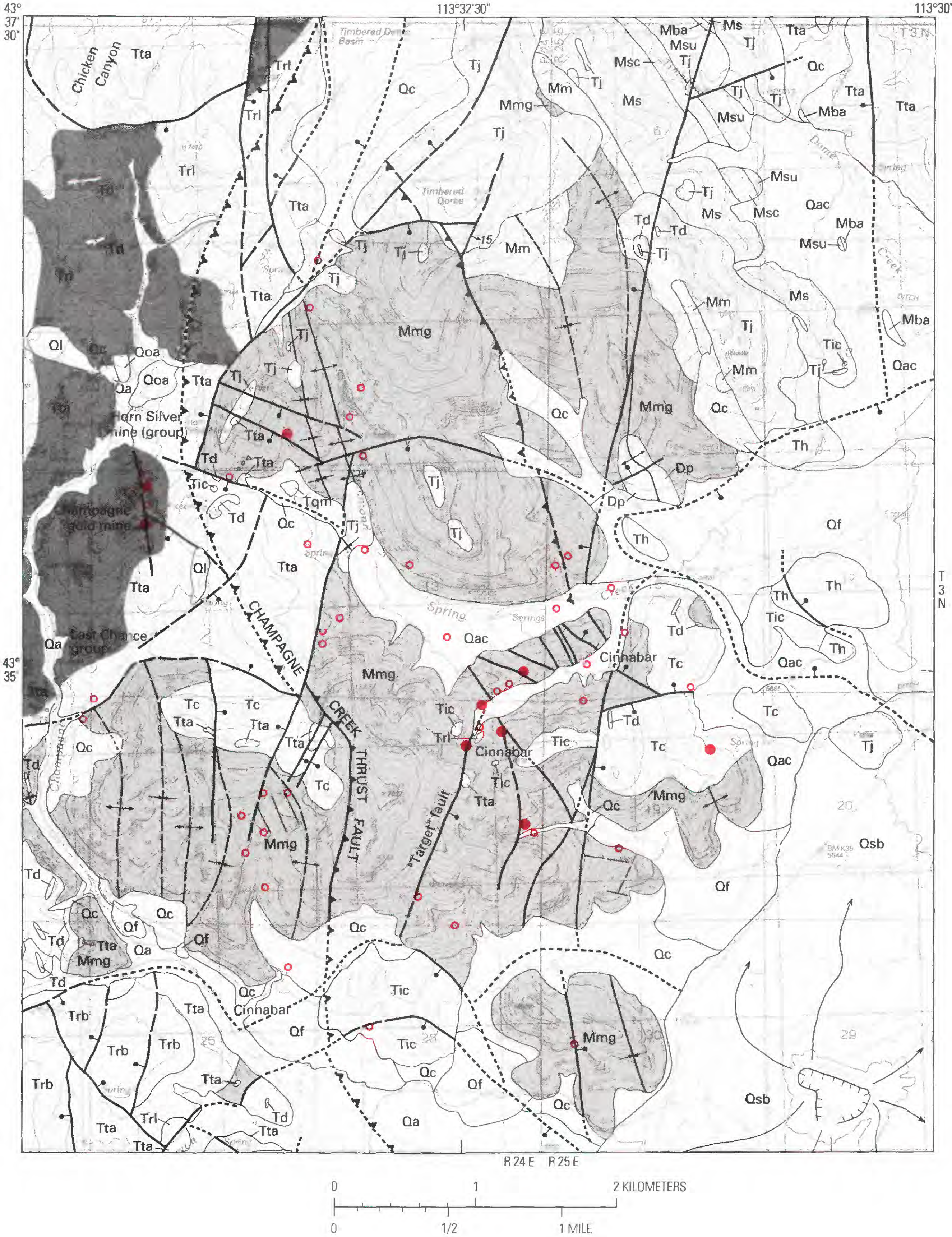

43°
37'
30"
113°32'30"
113°30'
43°
35'
T
3
N
Chicken Canyon
Horn Silver mine (group)
Champagne gold mine
Last Chance group
CHAMPAGNE CREEK THRUST FAULT
"Target" fault
Cinnabar
Timbered Dome
Spring
Creek
Tta
Trl
Td
Tj
Qc
Mmg
Mm
Ms
Msu
Msc
Mba
Qac
Tic
Th
Dp
Qf
Qoa
Qa
Ql
Tc
Tqm
Qsb
Trb
R 24 E
R 25 E
0
1
2 KILOMETERS
0
1/2
1 MILE

Figure 11. Indicator plant *Eriogonum ovalifolium* (foreground) in argentiferous soil near mineralized fault, Lava Creek mining district, central Idaho.

Figure 10 (facing page). Map showing location of stream-sediment and soil samples that contain anomalous amounts of silver (≥1.0 ppm)(solid red circles), followup survey, Lava Creek mining district, central Idaho. Open red circles indicate samples containing less than 1.0 ppm Ag; observed range of values <0.5–5.0 ppm Ag. Geology modified from Skipp and others (1990); see figure 9 for explanation.

Dunn, C.E., 1989, A multi-element and multi-media approach to geochemical exploration for gold in eastern Nova Scotia, Canada: International Geochemical Exploration Symposium, 13th, Rio de Janeiro, Brazil, October 1989, Abstracts, p. 5–6.

Dunn, C.E., and Hoffman, Eric, 1986, Multi-element study of vegetation from a zone of rare-earth rich allanite and apatite in northern Saskatchewan, Canada: Applied Geochemistry, v. 1, p. 375–381.

Erdman, J.A., Moye, Falma, and Theobald, P.K., 1989, Geochemical expression of poorly exposed metallization in the southeastern Challis volcanic field, Idaho—Emphasis on biogeochemical methods, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 91.

Erdman, J.A., Skipp, Betty, Theobald, P.K., and Moye, Falma, 1988, Drainage-basin geochemical survey of Lava Creek district in central Idaho using sagebrush and heavy-mineral concentrates—Preliminary evidence for sediment-hosted gold in McGowan Creek Formation: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 414.

Gough, L.P., and Erdman, J.A., 1983, Baseline elemental concentrations for big sagebrush from Western U.S.A.: Journal of Range Management, v. 36, p. 718–722.

Grimes, D.J., and Marranzino, A.P., 1968, Direct-current arc and alternating-current spark emission spectrographic field methods for the semiquantitative analysis of geologic materials: U.S. Geological Survey Circular 591, 6 p.

Hillier, B.M., Grimes, D.J., Vaughn, R., Arbogast, B., and McDougal, C., 1983, Analyses of samples from the Lava Creek mining district, Blaine and Butte Counties, Idaho: U.S. Geological Survey Open-File Report 83–705, 11 p.

Korobeynikov, A.F., 1986, Gold distributions in black shale associations: Geochemistry International, v. 23, p. 114–124.

Kovalevskii, A.L., 1979, Biogeochemical exploration for mineral deposits: Faridabad, India, Oxonian Press Pvt. Ltd., 136 p. Available from U.S. Department of Commerce, National Technical Information Service, 5285 Port Royal Road, Springfield, Va. 22161.

Lidgey, E., 1897, Some indications of ore deposits: Transactions, Australasian Institute of Mining Engineers, v. 4, p. 110–122.

Link, P.K., Skipp, Betty, Hait, M.H., Jr., Janecke, Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knobe, Pioneer, Boulder, and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 5–42.

Maciolek, J.B., and Jones, V.T., 1987, Mobile mercury application in the search for gold: Explore, no. 61, p. 6–7.

Miesch, A.T., 1981, Estimation of the geochemical threshold and its statistical significance: Journal of Geochemical Exploration, v. 16, p. 49–76.

Nelson, C.E., 1990, Comparative geochemistry of jasperoids from Carlin-type gold deposits of the western United States: Journal of Geochemical Exploration, v. 36, p. 171–195.

O'Leary, R.M., and Meier, A.L., 1986, Analytical methods used in geochemical exploration: U.S. Geological Survey Circular 948, 48 p.

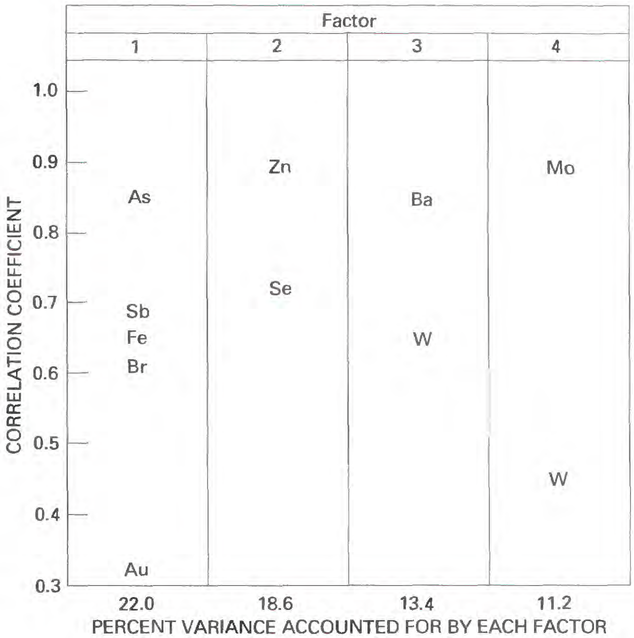


Figure 12. Factor compositions for sagebrush from followup survey, Lava Creek mining district, central Idaho. Elements are arranged in descending order of correlation coefficient, which indicates importance of the element.

Skipp, Betty, 1987, Basement thrust sheets in the Clearwater orogenic zone, central Idaho and western Montana: Geology, v. 15, p. 220–224.

———1988, Geologic map of Mackay 4 (Grouse) SE Quadrangle, Butte County, Idaho: U.S. Geological Open-File Report 88–423, scale 1:24,000.

Skipp, Betty, Kuntz, M.A., and Morgan, L.A., 1990, Geologic map of Mackay 4 (Grouse) SE Quadrangle, Butte County, Idaho: U.S. Geological Open-File Report 89–431, scale 1:24,000.

Soulliere, S.J., Wilson, A.B., Moye, F.J., and Rhea, K.P., 1988, Analytical results and sample locality map of jasperoid samples from Mackay, Idaho: U.S. Geological Survey Open-File Report 88–589, 28 p.

Sturges, D.L., 1977, Soil water withdrawal and rooting characteristics of individual big sagebrush plants: American Midland Naturalist, v. 98, p. 257–274.

Sturges, D.L., and Trlica, M.J., 1978, Root weights and carbohydrate reserves of big sagebrush: Ecology, v. 59, p. 1282–1285.

Tabler, R.D., 1964, The root system of *Artemisia tridentata* at 9,500 feet in Wyoming: Ecology, v. 45, p. 633–636.

The Mining Record, 1989, Bema Gold's Champagne Mine in production: February 8, v. 100, p. 25.

Weaver, J.E., and Clements, F.E., 1938, Plant ecology: McGraw-Hill, New York, 601 p.

Wilson, A.B., Soulliere, S.J., Worl, R.G., Skipp, Betty, and Rhea, K.P., 1988, Geology and geochemistry of jasperoid near Mackay, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 183–192.

Figure 13 (facing page). Map showing location of big sagebrush samples that contain anomalous amounts of arsenic (≥0.05 ppm)(solid red circles), followup survey, Lava Creek mining district, central Idaho. Open red circles indicate samples containing less than 0.05 ppm As; observed range of values <0.01–0.12 ppm As. Geology modified from Skipp and others (1990); see figure 9 for explanation.

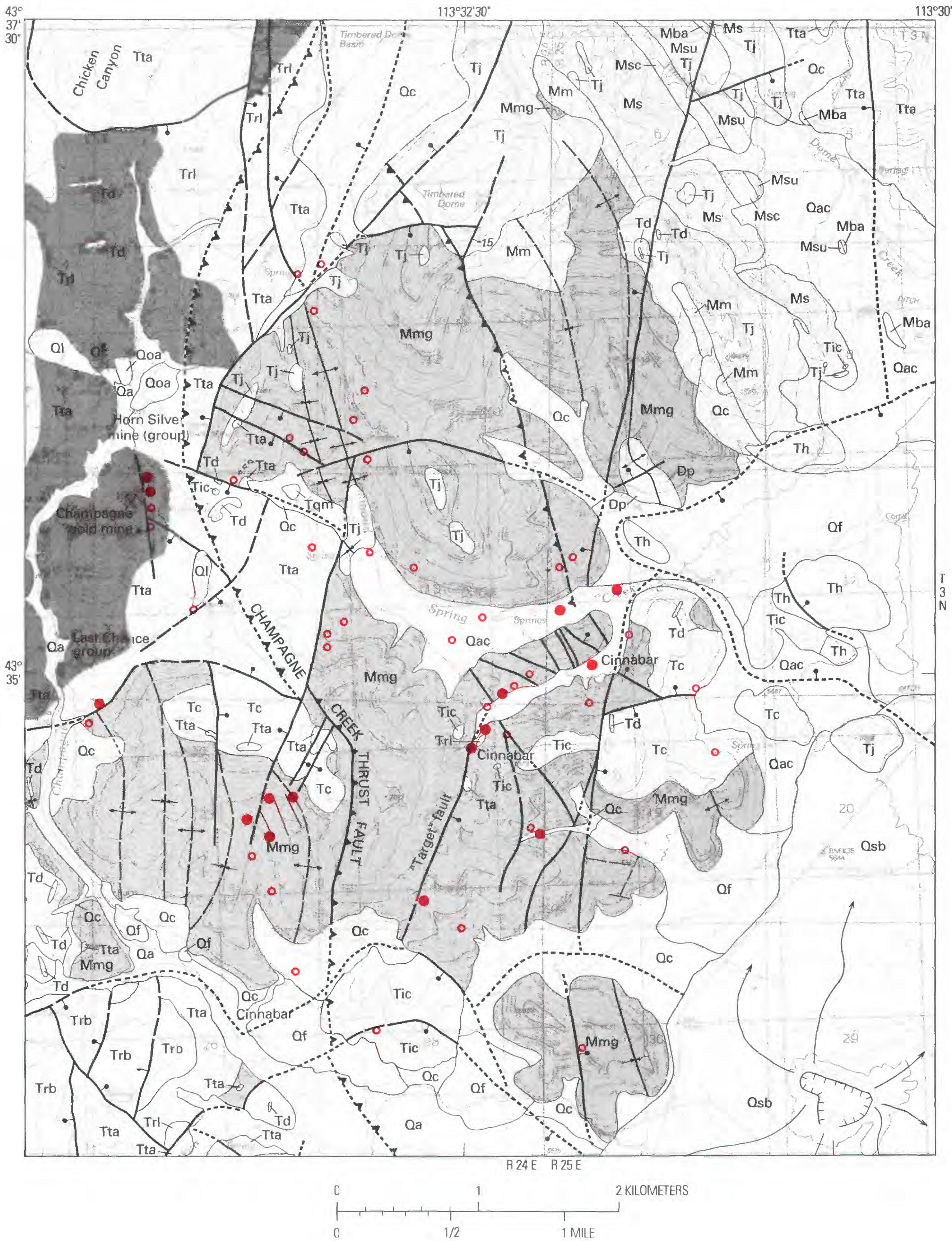
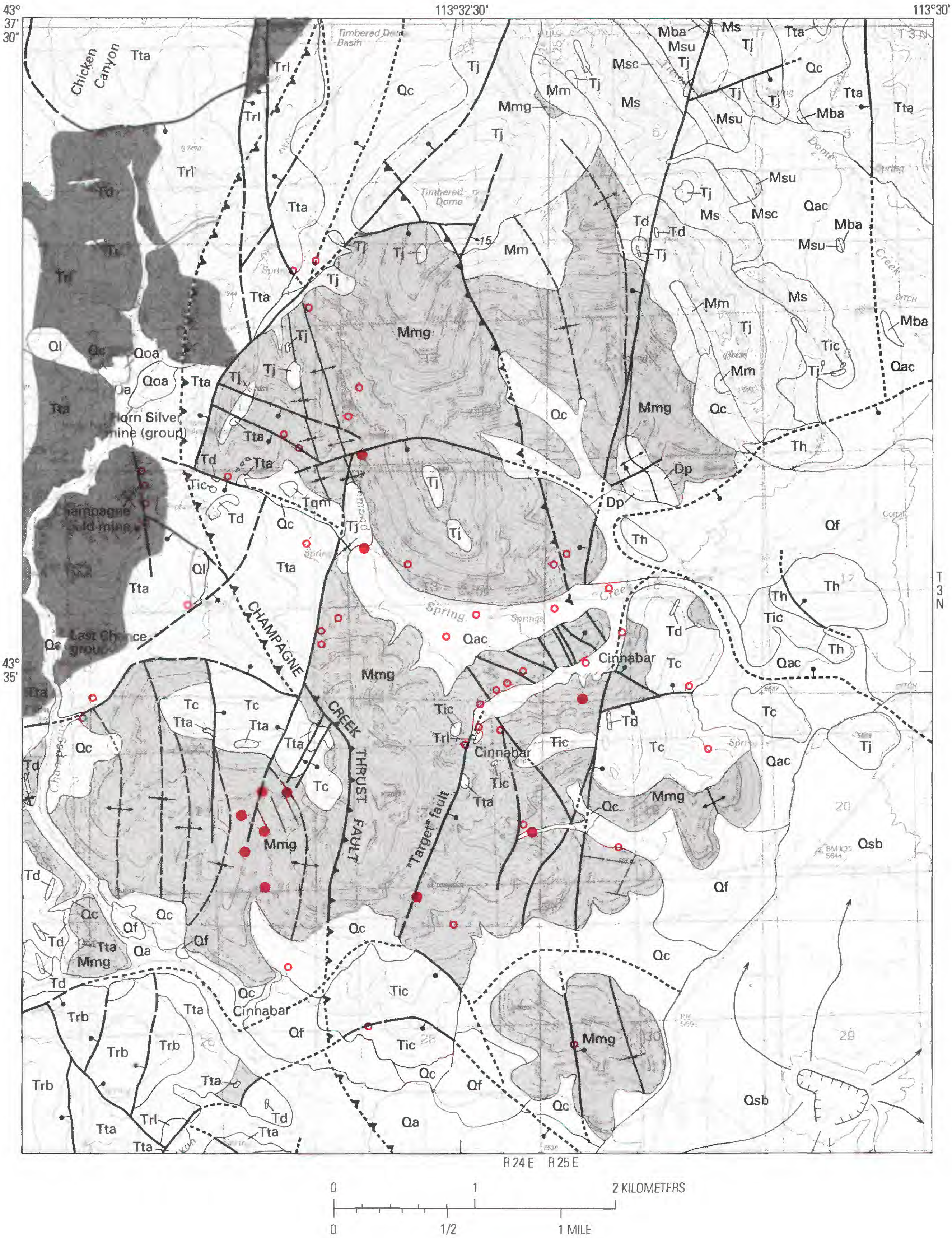

43°
37'
30"
113°32'30"
113°30'
43°
35'
T
3
N
R 24 E R 25 E
0 1 2 KILOMETERS
0 1/2 1 MILE
Chicken Canyon
Horn Silver mine (group)
Champagne gold mine
Last Chance group
CHAMPAGNE CREEK THRUST FAULT
"Target" fault
Cinnabar
Spring Creek
Timbered Dome
Timbered Dome Basin
Dome Creek
Tta
Trl
Td
Tj
Qc
Mmg
Mm
Ms
Msu
Msc
Mba
Qac
Tic
Tc
Th
Dp
Qf
Ql
Qoa
Qa
Tqm
Trb
Qsb
15
20
29
30

Figure 14 (facing page). Map showing location of big sagebrush samples that contain anomalous amounts of tungsten (≥0.77 ppm)(solid red circles), followup survey, Lava Creek mining district, central Idaho. Open red circles indicate samples containing less than 0.77 ppm W; observed range of values <0.09–2.0 ppm W. Geology modified from Skipp and others (1990); see figure 9 for explanation.


43°
37'
30"
113°32'30"
113°30'
Chicken Canyon
Timbered Dome Basin
Timbered Dome
Horn Silver mine (group)
Champagne gold mine
Last Chance group
43°
35'
CHAMPAGNE CREEK THRUST FAULT
"Target" fault
Cinnabar
Tta
Trl
Td
Tj
Qc
Mmg
Mm
Ms
Msc
Msu
Mba
Qac
Tic
Th
Dp
Qf
Qoa
Ql
Tqm
Tc
Trb
Qsb
Qa
T 3 N
R 24 E
R 25 E
0
1
2 KILOMETERS
0
1/2
1 MILE

Petrogenesis of Silver-Lead-Zinc Veins on the Eastern Margin of the Idaho Batholith in the Carrietown Mineralized Area, Blaine and Camas Counties, South-Central Idaho

By Robert S. Darling, Harry W. Campbell, *and* Paul Karl Link

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–P

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLES

Petrogenesis of Silver-Lead-Zinc Veins on the Eastern Margin of the Idaho Batholith in the Carrietown Mineralized Area, Blaine and Camas Counties, South-Central Idaho

By Robert S. Darling,[1] Harry W. Campbell,[2] *and* Paul Karl Link[3]

ABSTRACT

The Carrietown mineralized area, approximately 30 km southwest of Ketchum and 25 km northwest of Fairfield, Idaho, includes the Carrietown mining district, the eastern part of the Little Smoky Creek mining district, and the northern part of the Willow Creek mining district in the Dollarhide Mountain and Buttercup Mountain 7.5-minute quadrangles. The mineralized area covers approximately 75 km^2 in the southwestern part of the central Idaho black shale mineral belt.

Silver-lead-zinc deposits in the Carrietown area have similarities in ore mineralogy, deposit type, geologic setting, and petrogenesis. All of the deposits are epigenetic vein-type bodies containing galena, sphalerite, tetrahedrite, pyrite, arsenopyrite, pyrrhotite, and chalcopyrite. Galena and tetrahedrite are the principal silver-bearing phases. Distribution of deposits is controlled by both structure and lithology; silver-lead-zinc deposits are best developed in northeast-trending fault zones in rocks of the metamorphosed Middle Pennsylvanian to Lower Permian Dollarhide Formation and the Cretaceous Idaho batholith. These mineralized fault zones are (1) near contacts between carbonaceous marble and banded quartzite of the lower member of the Dollarhide Formation (previously known as the Carrietown sequence); (2) near contacts between banded quartzite of the Dollarhide Formation and Cretaceous intrusive rocks; and (3) in shear zones within the intrusive rocks. Lead-isotopic data suggest that lead was derived from Precambrian basement, whereas sulfur-isotopic data suggest a mixed igneous and sedimentary sulfur source. Stability of iron-bearing phases in ore assemblages indicates that conditions were reducing (low fO_2) in the Dollarhide Formation and that sulfur fugacities were lower in banded quartzite than in carbonaceous marble. Temperatures of ore formation estimated from several independent geothermometers suggest mesothermal to hypothermal (250°C–450°C) mineralization. Field, thermal, mineralogical, and available radiometric data indicate that mineralization was related to Cretaceous igneous activity.

INTRODUCTION

Silver-lead-zinc deposits in the Carrietown mineralized area of Idaho are in the northern half of the Dollarhide Mountain 7.5-minute topographic quadrangle and western half of the Buttercup Mountain 7.5-minute topographic quadrangle. The quadrangles are in the central Smoky Mountains, a north-trending range that extends from Galena Summit on the north to the northern margin of the Snake River Plain near the town of Fairfield. The principal divide of the range forms the boundary between Blaine and Camas Counties (fig. 1).

The old village of Carrietown is in the northern part of the Dollarhide Mountain quadrangle and was the principal settlement when mining operations flourished in all three districts in the late 1800's and early 1900's. Today, however, the village is abandoned and only a few buildings remain. Carrietown can be reached by light-duty roads from the towns of Ketchum, 30 km to the east, and Fairfield, 25 km to the south (fig. 1).

Umpleby (1915) called the entire area the Rosetta district. He believed that the Carrietown ores were epigenetic and related to Cretaceous igneous activity, a conclusion with which we substantially agree. Ross (1930) distinguished the Little Smoky Creek and Willow Creek districts. Gehlen (1983) examined mines in the eastern part of the Little Smoky Creek district. Hall (1985) defined only the Carrietown district, as did Darling (1987, 1988). Federspiel and others (1987, 1992) described parts of the Little Smoky and Rosetta districts. We use the term "Carrietown mineralized area" to include the Carrietown mining district, the eastern part of the Little Smoky Creek mining

[1]Department of Geology, State University of New York College at Cortland, P.O. Box 2000, Cortland, New York 13045.

[2]Western Field Operations Center, U.S. Bureau of Mines, E. 360 3rd Ave., Spokane, Washington 99202.

[3]Department of Geology, Idaho State University, Pocatello, Idaho 83209.

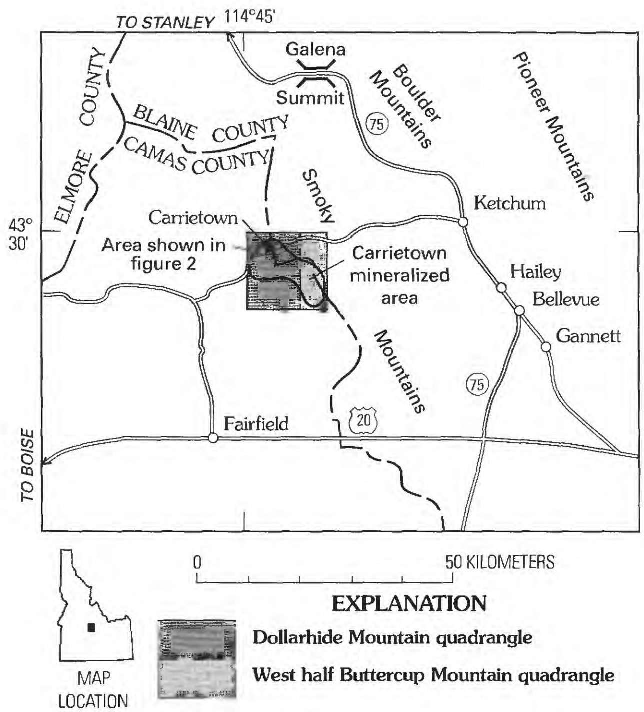


Figure 1. Map showing location of the Dollarhide Mountain 7.5-minute topographic quadrangle and the western half of the Buttercup Mountain 7.5-minute topographic quadrangle, Blaine and Camas Counties, Idaho.

district, and the northern part of the Willow Creek mining district. Silver-lead-zinc deposits in these areas share a genetic mineral deposit type, that of shear-zone-hosted veins, and have similar ore mineral assemblages and pressure-temperature-chemical conditions of ore formation. These mineral deposits are all near the contact between the Idaho batholith and the Middle Pennsylvanian to Lower Permian Dollarhide Formation.

Recognition in the 1800's of syngenetic lead-zinc mineral deposits in Paleozoic rocks of central Idaho (Hall, 1985; Turner and Otto, this volume) has led to renewed interest in the sediment-hosted deposits along the eastern margin of the Idaho batholith. The principal objectives of our research were to characterize the geology of silver-lead-zinc deposits in the Carrietown area and to determine their petrogenesis.

Acknowledgments.—This research was supported by an Idaho State University graduate student research grant and by the U.S. Geological Survey. The paper presented here is an extension of the senior author's M.S. thesis research completed at Idaho State University. Research by Campbell was conducted under the Idaho Land Assessment Program of the U.S. Bureau of Mines. Earlier versions of the manuscript benefited greatly from reviews by M.L. Sorensen and D. Frishman.

GEOLOGIC SETTING

The Carrietown area is on the southwestern margin of the central Idaho black shale mineral belt of Hall (1985), at the junction between folded Paleozoic strata and Cretaceous intrusive rocks of the Atlanta lobe of the Idaho batholith (Link and others, this volume; Lewis, in press). In the Dollarhide Mountain and Buttercup Mountain quadrangles, folded rocks of the Pennsylvanian and Permian Dollarhide Formation are intruded by three phases of the Late Cretaceous Idaho batholith. The Dollarhide Formation is preserved as a large, almost completely isolated roof pendant between the Atlanta lobe to the west and the Deer Creek stock to the east (fig. 2). In both quadrangles, the Cretaceous intrusive rocks and the Dollarhide Formation are unconformably overlain by rocks of the Eocene Challis Volcanic Group and intruded by Eocene dacite porphyry dikes and small stocks (fig. 2). Figures 3–6 are geologic-topographic maps of the mines investigated in this report.

DOLLARHIDE FORMATION

The Middle Pennsylvanian to Lower Permian Dollarhide Formation is named for exposures on Dollarhide Summit in the northern part of the Dollarhide Mountain 7.5-minute quadrangle. The unit was first defined by Hall (1985) and has been described by various workers including Wavra (1985), Wavra and others (1986), Geslin (1986), Link and others, (1988), Whitman (1990, this volume), O'Brien (1991), Mahoney and others (1991), and Link and others (this volume). The unit is divided into three informal members: (1) a lower member (800 m) consisting of rhythmically interbedded, very fine grained, light-gray to light-brown micritic sandstone and dark-gray carbonaceous silty micrite and subordinate medium- to thick-bedded light-brown micritic sandstone and light-gray lenticular conglomerate containing both extra- and intra-basinal clasts; (2) a middle member (300 m) consisting of fine-grained light-brown micritic sandstone and light-gray sandy micrite and subordinate dark-gray to black carbonaceous siltstone and lenticular conglomerate; (3) an upper member (900 m) consisting of thin-bedded to laminated dark-gray carbonaceous siltstone and light-gray silty micrite and minor light-brown micritic sandstone and conglomerate (Mahoney and others, 1991).

The lower and middle members of the Dollarhide Formation are widely exposed in the Dollarhide Mountain quadrangle, whereas the upper member is more abundant in the Buttercup Mountain quadrangle (fig. 2). The lowest exposed part of the lower member contains dark-brown banded quartzite, biotite phyllite, and minor calc-silicate hornfels. These metasedimentary rocks were informally named the "Carrietown sequence" by Skipp and Hall (1980) and mapped and described in detail by Geslin (1986) and Darling (1987, 1988, 1989). Following the interpretation of W.E.

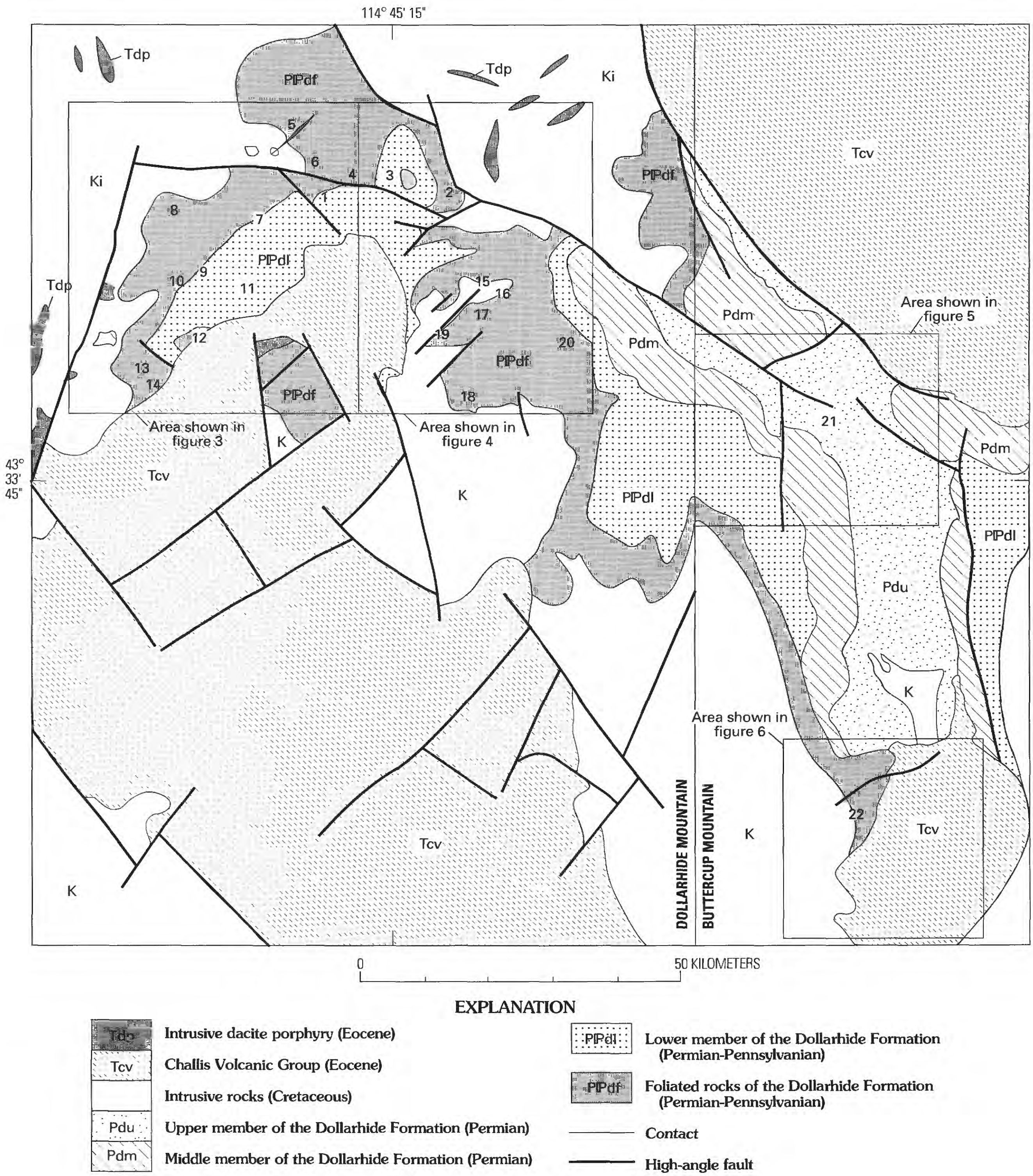
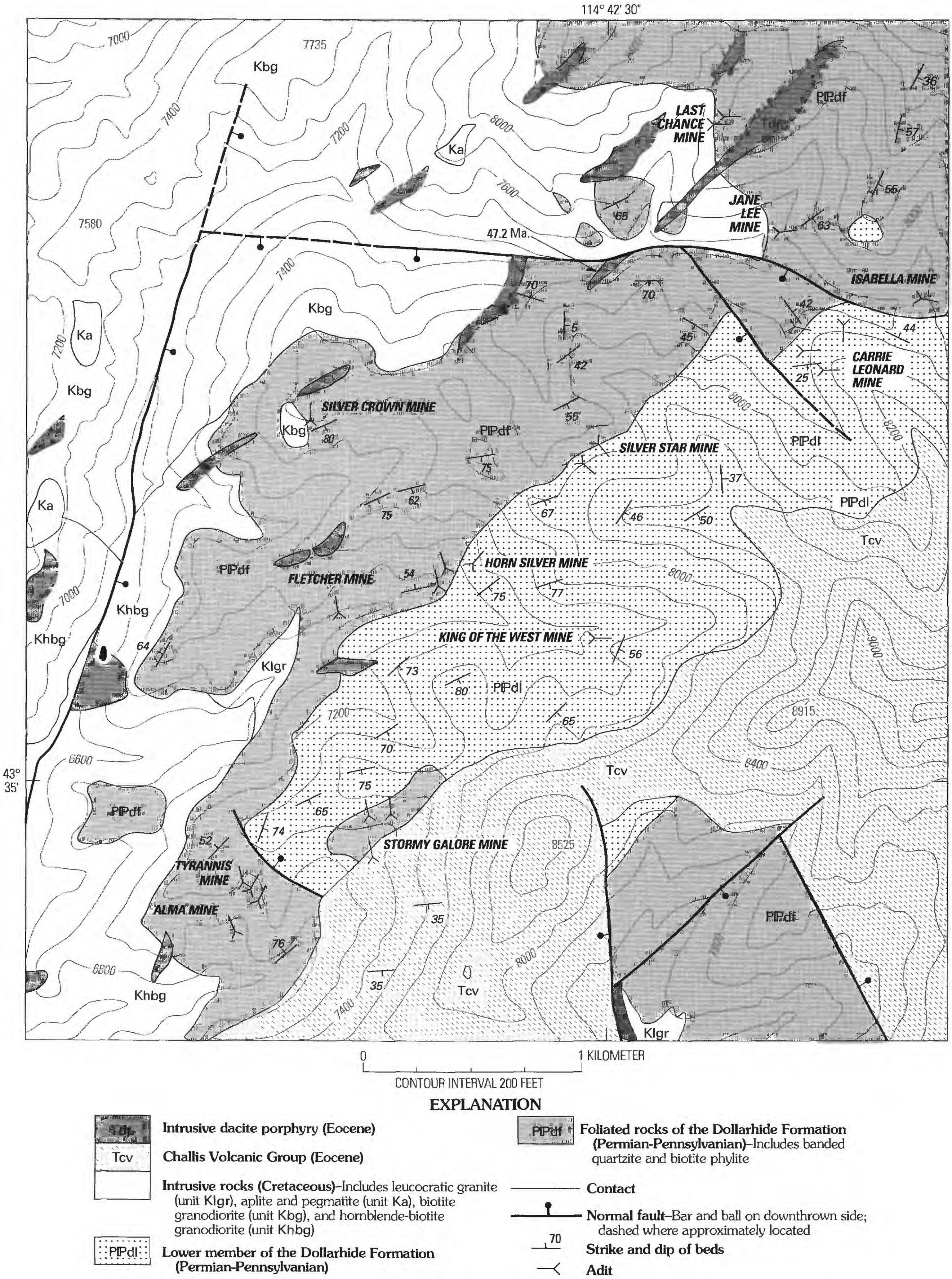


Figure 2. Map showing simplified geology of the Dollarhide Mountain 7.5-minute topographic quadrangle and the western half of the Buttercup Mountain 7.5-minute topographic quadrangle, Blaine and Camas Counties, Idaho. Detailed geology of selected areas is shown in figures 3–6. Mines and prospects in the Carrietown mineralized area (location shown in figure 1) are shown by number: 1, Carrie Leonard mine; 2, Margaret mine; 3, Dollarhide mine; 4, Isabella mine; 5, Last Chance mine; 6, Jane Lee mine; 7, Silver Star mine; 8, Silver Crown mine; 9, Horn Silver mine; 10, Fletcher mine; 11, King of the West mine; 12, Stormy Galore mine; 13, Tyrannis mine; 14, Alma mine; 15; Red Star prospect; 16, Taft prospect; 17, Hidden Treasure mine; 18, Smoky Bullion prospect; 19, Main Moly prospect; 20, Pine Creek mine; 21, unnamed mine; 22, Buttercup mine. A summary of geological information on these mines and prospects is given in table 2. Compiled from Gehlen (1983), Geslin (1986), Darling (1987, 1989, unpub. mapping), Whitman (1990), and O'Brien (1991).


114° 42' 30"
43° 35'
Kbg
Ka
Kbg
Khbg
Klgr
PIPdf
PIPdl
Tcv
47.2 Ma.
LAST CHANCE MINE
JANE LEE MINE
ISABELLA MINE
CARRIE LEONARD MINE
SILVER CROWN MINE
SILVER STAR MINE
HORN SILVER MINE
FLETCHER MINE
KING OF THE WEST MINE
STORMY GALORE MINE
TYRANNIS MINE
ALMA MINE
7000
7735
7400
7200
7580
8000
7600
6600
6800
8200
8400
8525
8915
9000
0
1 KILOMETER
CONTOUR INTERVAL 200 FEET
EXPLANATION
Tdp
Intrusive dacite porphyry (Eocene)
Tcv
Challis Volcanic Group (Eocene)
Intrusive rocks (Cretaceous)–Includes leucocratic granite (unit Klgr), aplite and pegmatite (unit Ka), biotite granodiorite (unit Kbg), and hornblende-biotite granodiorite (unit Khbg)
PIPdl
Lower member of the Dollarhide Formation (Permian-Pennsylvanian)
PIPdf
Foliated rocks of the Dollarhide Formation (Permian-Pennsylvanian)–Includes banded quartzite and biotite phylite
Contact
Normal fault–Bar and ball on downthrown side; dashed where approximately located
70
Strike and dip of beds
Adit

Hall (unpublished mapping, 1980–1984), these workers also inferred that the contact between the Dollarhide Formation and the Carrietown sequence is a thrust fault for reasons discussed in Darling (1987, p. 59). Subsequent studies by Whitman (1990, this volume) revealed, however, that the boundary between the two units is lithologic and that no evidence exists to support a structural contact. Following Whitman (1990, this volume), the term Carrietown sequence is abandoned, and the rocks are now interpreted as foliated and light-colored parts of the lower and middle members of the Dollarhide Formation.

CRETACEOUS INTRUSIVE ROCKS

The Dollarhide Formation is intruded by three separate phases of the Idaho batholith, from oldest to youngest, (1) hornblende-biotite granodiorite, (2) biotite granodiorite, and (3) leucocratic granite (figs. 3–6). The hornblende-biotite granodiorite forms a small stock in the western part of the Dollarhide Mountain quadrangle (fig. 3) and the northern part of the Willow Creek area (fig. 6). This pluton has been intruded by biotite granodiorite that is volumetrically the most abundant Cretaceous intrusive phase in both quadrangles. Both granodiorite bodies are host to numerous aplite and simple pegmatite dikes and pods. Leucocratic granite is the youngest phase and forms small stocks or dikes in both granodiorite bodies but is especially common near contacts between granodiorite and surrounding country rocks (figs. 3–5).

The modal mineralogy of Cretaceous intrusive rocks in the Carrietown district is illustrated in figure 7. This plot is based on results of point counting 20 stained slabs of rocks (1,000 points) using a transparent grid overlay (Darling, 1987). Similar plots for Cretaceous intrusive rocks in the Little Smoky Creek and Willow Creek districts are in Gehlen (1983) and Geslin (1986), respectively. Hornblende-biotite granodiorite is dominated by medium- to coarse-grained normally zoned plagioclase, medium- to coarse-grained quartz, and fine- to medium-grained, interstitial, poikilitic, perthitic microcline. Biotite (~9 percent) forms medium- to coarse-grained anhedral books, and hornblende (~5–6 percent) forms medium- to coarse-grained, dark-brown to black, locally twinned prisms. Some rocks are so alkali feldspar deficient that they are more properly named tonalite (fig. 7). Biotite granodiorite contains medium- to coarse-grained normally zoned plagioclase, medium- to coarse-grained quartz, and fine- to medium-grained, interstitial and separate crystals of perthitic microcline. Biotite is less abundant than in the older phase (~4–5 percent) and is present as fine-grained disseminated flakes and books. Both types of granodiorite have abundant myrmekitic intergrowths and apatite, magnetite, titanite, and zircon accessory phases. Leucocratic granite is dominated by subequal amounts of fine- to medium-grained plagioclase, quartz, and microcline. The granite has abundant micrographic and myrmekitic intergrowths and garnet and allanite accessory phases.

The Cretaceous intrusive rocks in the study area have not been radiometrically dated; however, biotite granodiorite of the Rooks Creek stock, 18 km to the northeast, has been dated at 92.2±2.2 Ma by K-Ar methods on biotite (Marvin and Dobson, 1979). A minimum age of 77.4±2.6 Ma is indicated for leucocratic granite in the Dollarhide Mountain quadrangle based on a K-Ar age for hydrothermal muscovite (M. Tschanz, 1982, unpublished report; cited in Gehlen, 1983) (see discussion on age of mineralization). The above field and radiometric dates are consistent with data collected on Cretaceous intrusive rocks in the Challis 1°×2° quadrangle (Kiilsgaard and Lewis, 1985) and elsewhere in the Hailey 1°×2° quadrangle (Lewis, in press).

METAMORPHISM

In the Dollarhide Mountain and Buttercup Mountain quadrangles, the lower member of the Dollarhide Formation has been metamorphosed proximal to intrusive rocks of the Idaho batholith. Metamorphosed Dollarhide Formation contains sillimanite, andalusite, staurolite, tremolite, and wollastonite, an assemblage that indicates metamorphic temperatures of 500°C–600°C and pressures of 2.5–3.5 kb (Whitman, 1990, this volume). The metamorphism is dated at 83.9±3.4 Ma by whole-rock K-Ar methods (R.L. Armstrong, written commun., 1990), and this age, combined with the estimated contact metamorphic pressure, indicates an emplacement depth of 8–11 km for the Idaho batholith (Whitman, 1990, this volume).

Whitman (1990, this volume) divided the lower member of the Dollarhide Formation into three informal textural units. Unit 1 consists of purplish-brown phyllite, local fine- to medium-grained black to brown schist, and purplish-brown fine-grained phyllitic quartzite; unit 2 consists of purple-brown phyllite and phyllitic quartzite, centimeter-scale banded quartzite containing locally stratiform sulfide minerals (pyrrhotite and chalcopyrite), and gray to white calcareous to siliceous, massive to finely laminated marble and calc-silicate hornfels; and unit 3 consists of black carbonaceous marble, gray to white calc-silicate hornfels, and very fine grained argillite. Units 1 and 2 have strong metamorphic fabric, and marble of unit 3 displays less fabric. Units 1 and 2 are shown as foliated Dollarhide Formation (unit PIPdf) in figures 2–6.

Figure 3 (facing page). Map showing geology and topography of the Carrietown area, Idaho. Location of map area is shown in figure 2. Geology modified from Darling (1987, 1989). Base from U.S. Geological Survey Dollarhide Mountain 7.5-minute quadrangle.

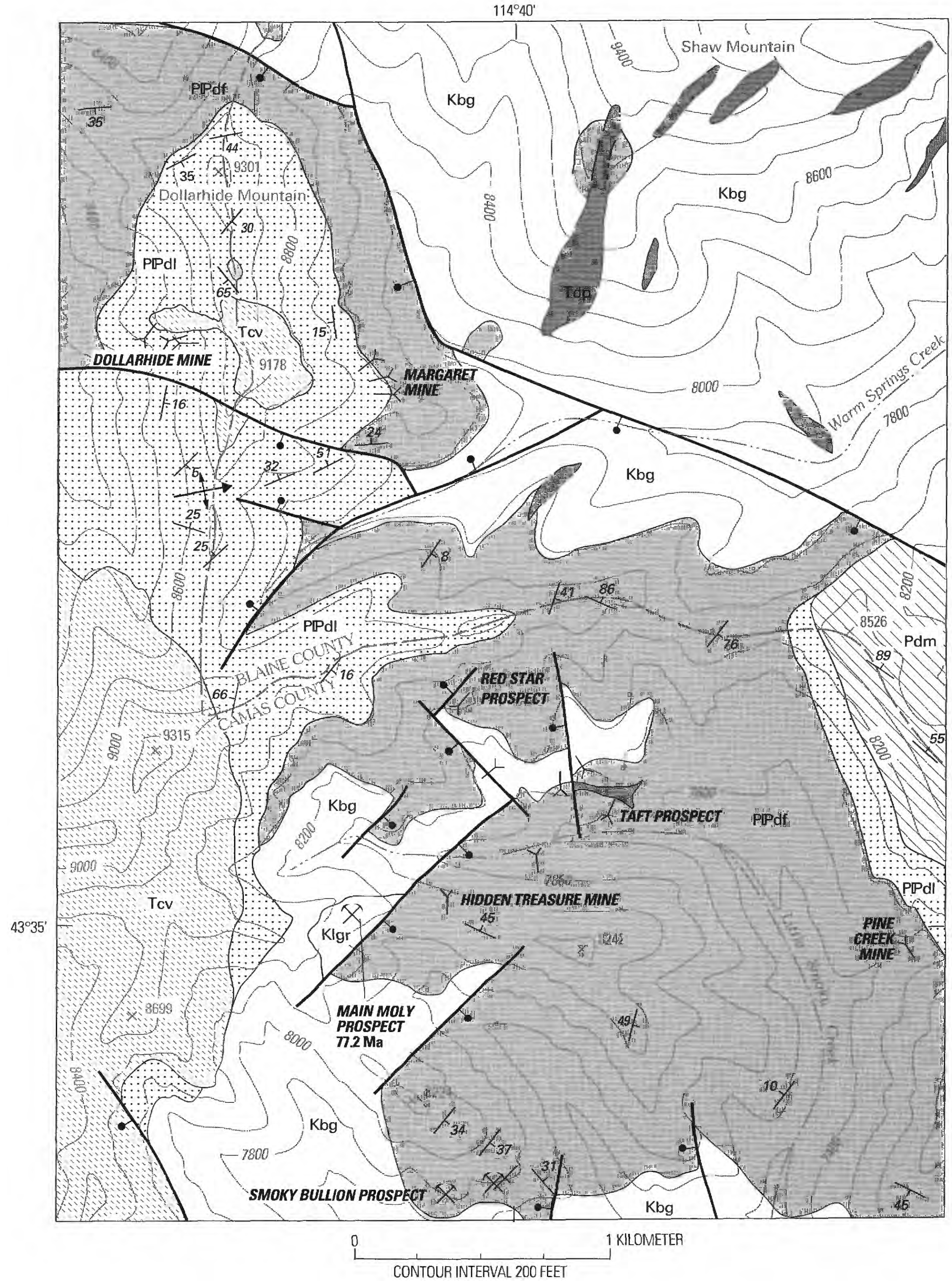


Figure 4 (above and facing column). Map showing geology and topography of the Warm Springs (top part of map area) and Little Smoky Creek areas (bottom part of map area), Idaho. Location of map area is shown in figure 2. Geology modified from Darling (1987, 1989) and Gehlen (1983). Base from U.S. Geological Survey Dollarhide Mountain 7.5-minute quadrangle.

EXPLANATION

- To — **Intrusive dacite porphyry (Eocene)**
- Tcv — **Challis Volcanic Group (Eocene)**
- **Intrusive rocks (Cretaceous)**—Includes leucocratic granite (unit Klgr) and biotite granodiorite (unit Kbg)
- Pdm — **Middle member of the Dollarhide Formation (Permian)**
- PIPdl — **Lower member of the Dollarhide Formation (Permian-Pennsylvanian)**
- PIPdf — **Foliated rocks of the Dollarhide Formation (Permian-Pennsylvanian)**—Includes banded quartzite and biotite phyllite

- **Contact**
- **Normal fault**–Bar and ball on downthrown side; dashed where approximately located
- 70 **Strike and dip of beds**
- **Anticline**–Showing direction of plunge
- **Adit**
- **Prospect**

EOCENE EXTRUSIVE AND INTRUSIVE ROCKS

In the Dollarhide Mountain and Buttercup Mountain quadrangles, both the Dollarhide Formation and the Cretaceous intrusive rocks are unconformably overlain by rocks of the Eocene Challis Volcanic Group, which formed during widespread magmatism between 51 and 40 Ma (McIntyre and others, 1982; Moye and others, 1988; Moye and others, in press). Volcanic rocks in the Carrietown area consist of andesite, dacite, and rhyolite porphyry tuff, lava flow rocks, tuff breccia, and volcaniclastic sedimentary rocks. Detailed descriptions and petrologic interpretations of these rocks are given in Gehlen (1983). The volcanic rocks are intruded by numerous Eocene hypabyssal dacite porphyry dikes and small stocks. These dikes are present throughout the study area but are especially abundant in the northern Dollarhide Mountain quadrangle near Shaw Mountain (fig. 4). The intrusive rocks are coeval, subvolcanic equivalents of the Challis Volcanic Group (Stewart and others, in press). One dike 0.7 km west of Carrietown has been dated at 47.2±0.9 Ma by K-Ar methods on biotite (W.E. Hall, unpublished report, 1978) (fig. 3).

HISTORY AND PRODUCTION

Mining in the Carrietown area began in the early 1880's, but production from many mines decreased substantially by 1890, and almost all operations ceased by the turn of the century (Umpleby, 1915; Ross, 1930). Since then, the districts have been inactive except for small-scale mining and annual assessment work. In the early 1980's, however, when silver prices soared, both local and major mining companies explored the Dollarhide Formation for undiscovered silver-lead-zinc deposits. This exploration abated with depressed silver prices in the late 1980's.

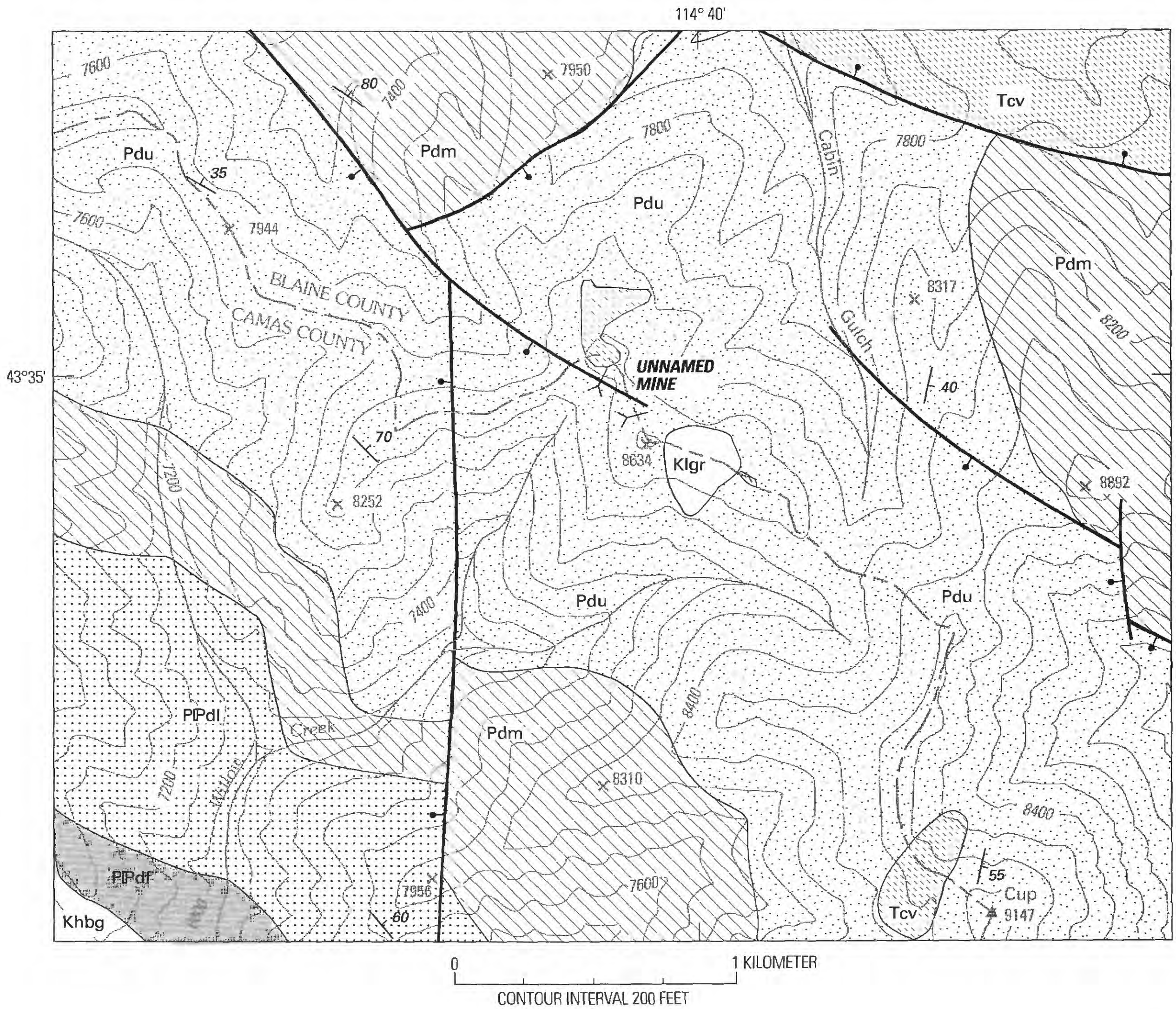
Recorded metal production from the Carrietown area is listed in table 1. The districts in the area produced $6.6 million in silver, $915,770 in lead, and $273,000 in zinc (expressed in 1990 dollars). Small quantities of gold and copper were locally recovered as well (table 1). Refer to Ross (1930), Federspiel and others (1992), and unpublished U.S. Bureau of Mines files (Spokane, Washington) for specific dates of mine operation. The average ore grade was 15–400 ounces of silver per ton (Umpleby, 1915), but local miners working in the area reported grades as high as 1,000 ounces of silver per ton.

SILVER-LEAD-ZINC DEPOSITS

Silver-lead-zinc deposits in the Carrietown area are tabular, vein-type bodies that generally occupy northeast-trending fault zones in the Dollarhide Formation and its metamorphosed equivalents. The mineralized shear zones dip moderately to steeply southeast and northwest (fig. 8) and locally are parallel with bedding in the host rocks. Significant geologic information for mines in all three districts is summarized in table 2. Veins are generally (1) at or near the contact between banded quartzite (unit 2, PIPdf) and carbonaceous marble (unit 3, PIPdl) of the Dollarhide Formation, (2) near contacts between the banded quartzite and Cretaceous granodiorite, and (3) in minor shear zones in the Cretaceous intrusive rocks (figs. 3–6). Veins have an average width of 1–2 m and extend into the country rock for several tens of meters. Within the veins, ore minerals are present in a series of irregular, lenslike pods that are parallel with the vein walls.

ORE PETROGRAPHY

Ore minerals from mines in the Carrietown area include galena, sphalerite, tetrahedrite, pyrite, arsenopyrite, pyrrhotite, chalcopyrite, and cubanite. Tennantite, boulangerite, and molybdenite are present in minor amounts. Quartz and siderite are the principal gangue minerals where the ore is hosted by the banded quartzite (unit 2) of the Dollarhide Formation, but ore hosted by carbonaceous marble or limestone has a quartz-calcite gangue (see table 2). A



EXPLANATION

- Tcv — **Challis Volcanic Group (Eocene)**
- **Intrusive rocks (Cretaceous)**—Includes leucocratic granite (unit Klgr), and hornblende-biotite granodiorite (unit Khbg)
- Pdu — **Upper member of the Dollarhide Formation (Permian)**
- Pdm — **Middle member of the Dollarhide Formation (Permian)**
- PIPdl — **Lower member of the Dollarhide Formation (Permian-Pennsylvanian)**
- PIPdf — **Foliated rocks of the Dollarhide Formation (Permian—Pennsylvanian)**—Includes banded quartzite and biotite phyllite
- **Contact**
- **Normal fault**—Bar and ball on downthrown side
- 70 — **Strike and dip of beds**
- **Adit**

Figure 5. Map showing geology and topography in the Cabin Gulch–Willow Creek area, Idaho. Location of map area is shown in figure 2. Geology modified from O'Brien (1991, plate 1). Base from U.S. Geological Survey Buttercup Mountain 7.5-minute quadrangle.

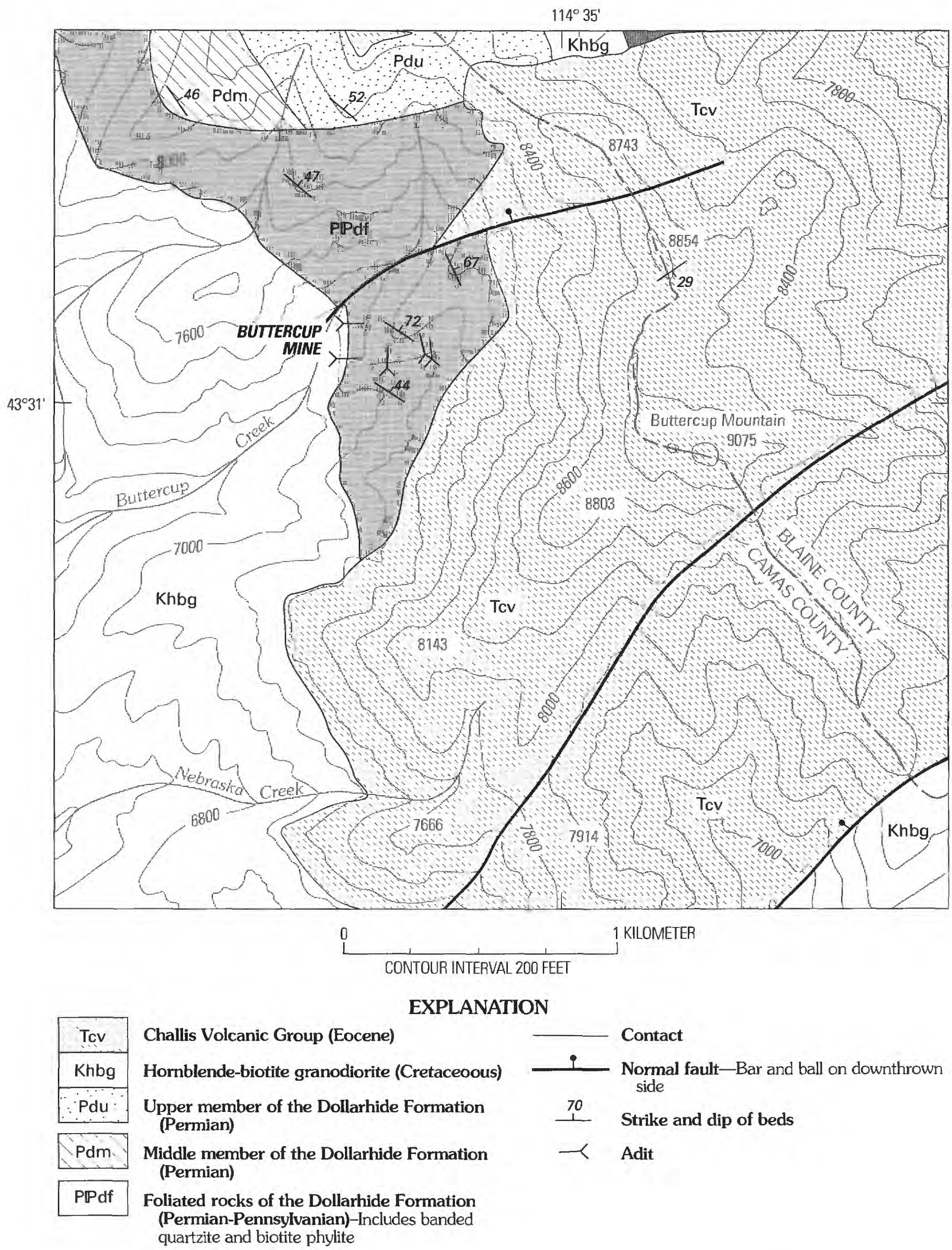


Figure 6. Map showing geology and topography of the Buttercup Mountain area, Idaho. Location of map area is shown in figure 2. Geology modified from Geslin (1986), Whitman (1990), and O'Brien (1991). Base from U.S. Geological Survey Buttercup Mountain 7.5-minute quadrangle.

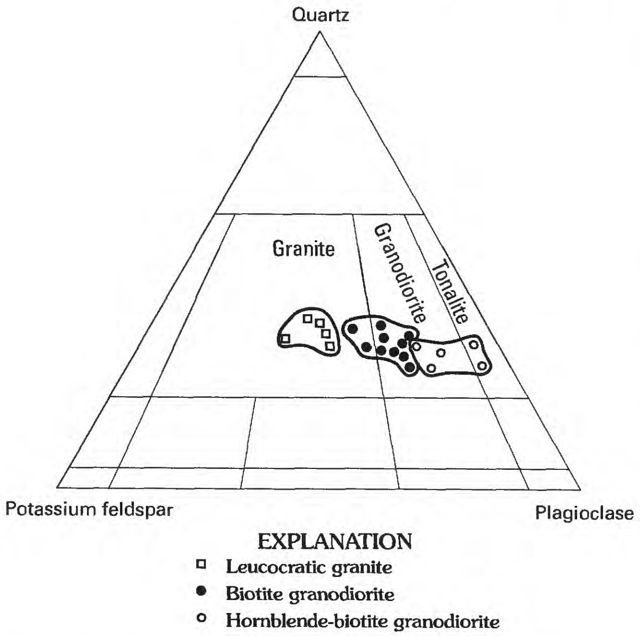


Figure 7. Modal mineralogy of Cretaceous intrusive rocks in the study area. Specified fields according to Streckeisen (1976).

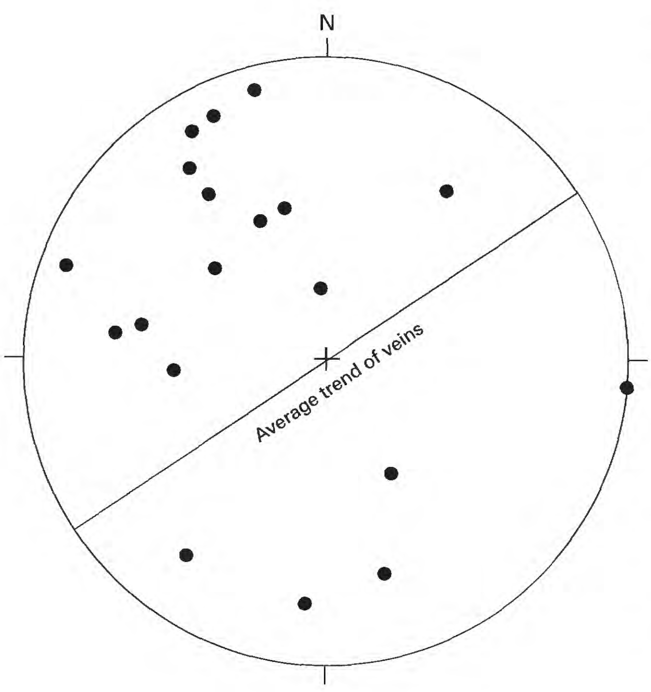


Figure 8. Schmidt equal-area lower hemisphere plot of poles to veins listed in table 2. Note average northeast trend of veins.

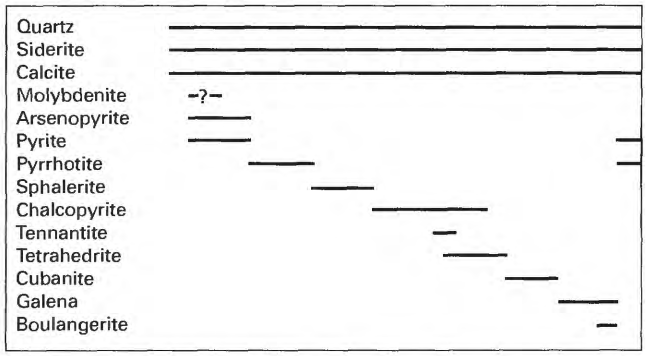


Figure 9. General paragenetic sequence for hypogene minerals in the Carrietown mineralized area, Idaho. Lines represent periods of deposition; time progressing to the right.

general paragenetic sequence for mineralization is illustrated in figure 9. This sequence is based on field observations and petrographic studies of 74 polished sections from deposits in the Carrietown area (Darling, 1987). Figure 9 shows an early phase of iron and arsenic±molybdenum(?) metallization (pyrite, pyrrhotite, arsenopyrite, and molybdenite), followed by a middle phase of zinc and copper metallization (sphalerite and chalcopyrite) that overlapped with, and was superseded by, a late stage of copper, antimony, lead, and silver metallization (tetrahedrite, tennantite, galena, and boulangerite).

Many of the minerals listed in figure 9 are very coarse grained (1–2 cm) and are easily recognized in hand sample; however, cubanite, boulangerite, and tennantite are microscopic. Mineralogical and textural identifications of sulfide minerals in polished section were made according to Craig and Vaughan (1981), Picot and Johan (1982), and Ramdohr (1980). Common ore microtextures include (1) chalcopyrite blebs in sphalerite, (2) cubanite exsolution lamellae in chalcopyrite, (3) pyrite and arsenopyrite intergrowths, (4) comb-structured quartz and siderite, (5) crustiform siderite, (6) deformed galena, (7) tetrahedrite inclusions in galena, and (8) spindle-shaped deformation twins in chalcopyrite.

Umpleby (1915) reported that galena and tetrahedrite are the principal silver-bearing phases in the Carrietown district. Tennantite and boulangerite, although volumetrically insignificant, could also be silver-bearing phases. The manner in which silver is present in galena is not known, but microscopic tetrahedrite inclusions are a probable source. Regardless of the source, the silver-bearing phases were deposited relatively late in the mineralization sequence (fig. 9).

Table 1. Recorded production from the Carrietown, northern Willow Creek, and eastern Little Smoky Creek mining districts, Idaho.
[Compiled from Ross (1930), Federspiel and others (1992), and unpublished files of the U.S. Bureau of Mines, Spokane, Washington. Ounces are troy ounces; NR indicates no record. Commodity prices used in calculations (Mineral Commodity Summaries, U.S. Bureau of Mines, 1990): gold, $385 per ounce; silver, $5.55 per ounce; copper, $1.32 per pound; lead, $0.40 per pound; zinc, $0.83 per pound]

	Gold (ounces)	Silver (ounces)	Copper (ounces)	Lead (pounds)	Zinc (pounds)
Carrietown mining district					
Carrie Leonard	23	869,000	645	989,000	171,000
Margaret	NR	NR	NR	NR	NR
Dollarhide	0.28	3,420	0	18,928	0
Isabella	10.4	19,000	600	61,500	31,000
Last Chance	NR	NR	NR	NR	NR
Jane Lee	NR	NR	NR	NR	NR
Silver Star	118.92	42,511	600	287,850	12,587
Silver Crown	0.62	767.3	0	8,430	0
Horn Silver	21.92	7,187	984	11,601	5,721
Fletcher	NR	NR	NR	NR	NR
King of the West	95.02	108,263	1,897	364,048	99,637
Stormy Galore	16.87	55,450	234	51,995	3,277
Tyrannis	1.96	32,486.1	0	472,456	6,191
Alma	NR	NR	NR	NR	NR
Eastern Little Smoky Creek mining district					
Red Star	NR	NR	NR	NR	NR
Taft	0	20	0	341	0
Hidden Treasure	1.2	30	7	1,038	0
Smoky Bullion	8.15	828	148	22,238	0
Main Moly	NR	NR	NR	NR	NR
Pine Creek	NR	NR	NR	NR	NR
Northern Willow Creek mining district					
Buttercup	14.11	57,731.1	500	1,000,674	73,613
Unnamed mine	NR	NR	NR	NR	NR
Total production	312.45	1,196,694	5,615	2,289,425	329,413
Dollar value	$120,293	$6,641,652	$7,412	$915,770	$273,413
Total dollar value all commodities		$7,958,540			

PETROGENESIS OF SILVER-LEAD-ZINC ORES

In recent years, much attention has been focused on stratiform, syngenetic ore deposits in Paleozoic rocks of the black shale belt of Idaho. Ore deposits of inferred syngenetic origin have been described in the Triumph mine (Triumph-Parker mineral belt, 40 km to the east) and in the Hoodoo and Livingston mines (Slate Creek district, 70 km to the north) (Hall, 1985; Turner and Otto, this volume). In the Carrietown mineralized area, however, the orebodies are interpreted to have an epigenetic origin. This interpretation is supported by the strong structural control of the deposits and the ubiquitous open-space-filling textures (crustiform and comb-structured quartz and siderite). Minor stratiform pyrrhotite and chalcopyrite in banded quartzite (unit 2) of the lower member of the Dollarhide Formation is the only syngenetic sulfide mineralization inferred.

LITHOLOGIC CONTROLS

The abundance of silver-lead-zinc sulfide minerals in veins is strongly controlled by lithology because the most productive veins are hosted by Paleozoic metasedimentary rocks rather than Cretaceous intrusive rocks. This lithologic control is best illustrated by quartz veins that can be traced from granodiorite into banded quartzite, where, after they enter the quartzite, the abundance of vein sulfide minerals increases (for example, Silver Crown and Last Chance mines). It is inferred that rocks of the Dollarhide Formation, which contain syngenetic pyrrhotite and carbonaceous material, provided a reducing environment for the deposition of metal-sulfide minerals, an environment not normally expected in the more oxidized (magnetite-bearing) Cretaceous intrusive rocks.

Gangue mineralogy is also strongly controlled by host-rock lithology. Deposits hosted by quartzite in the

Table 2. Summary of geological information on mines and prospects in the Carrietown mineralized area, Idaho.
[Mines are shown by number in fig. 2. Minerals are listed in order of decreasing abundance]

Mine or prospect	Host rock	Vein attitudes	Ore minerals	Gangue minerals	Reference
1. Carrie Leonard	Lower member, Dollarhide Formation	N. 40° E., 40° S.; N. 55° E., 58° S.	Sphalerite, galena, pyrite	Calcite, quartz	Darling (1987), Umpleby (1915).
2. Margaret	Lower member, Dollarhide Formation	Not available	Pyrite, galena, sphalerite,	Quartz, calcite	Darling (1987), Umpleby (1915).
3. Dollarhide	Lower member, Dollarhide Formation	N. 55° W., 60° S.; N. 55° W., 70° N.	Sphalerite, galena, pyrite	Calcite, quartz	Darling (1987), Umpleby (1915).
4. Isabella	Dollarhide Formation (foliated, banded quartzite±phyllite)	Not available	Pyrite, sphalerite, pyrrhotite	Quartz, siderite	Darling (1987), Umpleby (1915).
5. Last Chance	Dollarhide Formation (foliated, banded quartzite±phyllite)	N. 85° E., 20° S.; N. 75° E., 64° N.	Pyrite, chalcopyrite, arsenopyrite, pyrrhotite, sphalerite, galena	Siderite, quartz	Darling (1987), Umpleby (1915).
6. Jane Lee	Dollarhide Formation (foliated, banded quartzite±phyllite)	Not available	Pyrite, chalcopyrite, arsenopyrite, pyrrhotite, sphalerite, galena	Siderite, quartz	Campbell (unpub. data).
7. Silver Star	Lower member, Dollarhide Formation; Dollarhide Formation (foliated, banded quartzite±phyllite)	N. 65° E., 43° S.; N. 65° E., 80° S.	Pyrite, sphalerite, galena, pyrrhotite, chalcopyrite, arsenopyrite	Quartz, calcite, siderite	Darling (1987), Umpleby (1915)
8. Silver Crown	Dollarhide Formation (foliated, banded quartzite±phyllite); Cretaceous biotite granodiorite	N. 20° E., 81° S.	Pyrite, sphalerite, galena, arsenopyrite, molybdenite	Quartz	Darling (1987), Umpleby (1915).
9. Horn Silver	Dollarhide Formation (foliated, banded quartzite±phyllite); lower member, Dollarhide Formation	Not available	Galena, sphalerite, pyrite, pyrrhotite, arsenopyrite, chalcopyrite	Quartz, calcite, siderite	Darling (1987), Umpleby (1915).
10. Fletcher	Dollarhide Formation (foliated, banded quartzite±phyllite)	N. 5° W., 42° E.; N. 10° E., 53° S.; N. 60° E., 79° S.; N. 55° E., 69° S.	Galena, sphalerite, pyrite, chalcopyrite, tetrahedrite, pyrrhotite	Quartz, siderite	Darling (1987).
11. King of the West	Lower member, Dollarhide Formation	N. 75° E., 84° S.; N. 85° W., 72° N.	Sphalerite, galena, pyrite, tetrahedrite	Quartz, calcite	Darling (1987), Umpleby (1915).
12. Stormy Galore	Dollarhide Formation (foliated, banded quartzite±phyllite)	N. 7° E., 60° S.	Tetrahedrite, galena, pyrite	Quartz, siderite	Darling (1987), Umpleby (1915).
13. Tyrannis	Dollarhide Formation (foliated, banded quartzite±phyllite)	N. 60° E., 37° N.	Galena, pyrite, chalcopyrite, pyrrhotite, arsenopyrite	Quartz, siderite	Darling (1987), Umpleby (1915).
14. Alma	Dollarhide Formation (foliated, banded quartzite±phyllite)	Not available	Galena, sphalerite, pyrite, pyrrhotite, tetrahedrite, arsenopyrite	Siderite, quartz	Darling (1987).
15. Red Star	Dollarhide Formation (foliated, banded quartzite+phyllite)	Not available	Galena, sphalerite, pyrite, chalcopyrite, pyrrhotite	Quartz	Gehlen (1983), Ross (1930).
16. Taft	Dollarhide Formation (foliated, banded quartzite±phyllite)	N. 75° E., 45° S.	Galena, chalcopyrite, pyrite, chalcopyrite, pyrrhotite	Quartz	Gehlen (1983), Ross (1930).

17.	Hidden Treasure	Dollarhide Formation (foliated, banded quartzite+phyllite)	Not available	Galena, sphalerite, pyrite, chalcopyrite, pyrrhotite	Quartz, siderite	Gehlen (1983), Ross (1930).
18.	Smoky Bullion prospect	Dollarhide Formation (foliated, banded quartzite±phyllite)	Not available	Pyrite, galena, sphalerite, chalcopyrite, molybdenite	Quartz, siderite	Gehlen (1983).
19.	Main Moly prospect	Cretaceous leucocratic granite	N. 5° E., 90°	Molybdenite, galena, pyrite, calcite, chlorite	Quartz, muscovite	Gehlen (1983).
20.	Pine Creek	Dollarhide Formation (foliated, banded quartzite±phyllite)	Not available	Pyrite, sphalerite, galena	Quartz, siderite	Link (unpub. data).
21.	Unnamed mine	Upper member, Dollarhide Formation	Not available	Pyrite, sphalerite, galena	Quartz, calcite	Link (unpub. data).
22.	Buttercup	Dollarhide Formation (foliated, banded quartzite±phyllite)	Not available	Pyrite, sphalerite, galena, tetrahedrite	Quartz, calcite, siderite	Geslin (1986), Ross (1930).

Dollarhide Formation have a quartz-siderite gangue, whereas those hosted by carbonaceous marble have a quartz-calcite assemblage. The factor controlling gangue mineralogy is interpreted to be the Fe^{2+}/Ca^{2+} ratio of ore-forming fluids in equilibrium with their host rock. This control is explained by the reaction

$$FeCO_{3\,(s)} + Ca^{2+}_{(aq)} \rightarrow CaCO_{3\,(s)} + Fe^{2+}_{(aq)} \qquad (1)$$

where high Fe^{2+}/Ca^{2+} favors formation of siderite and low Fe^{2+}/Ca^{2+} favors formation of calcite. Because siderite is the principal carbonate phase in ores hosted by banded quartzite, ore-forming fluids had relatively high Fe^{2+}/Ca^{2+} ratios, whereas fluids precipitating calcite-rich veins in the carbonaceous marbles had relatively low Fe^{2+}/Ca^{2+} ratios. Ferrous-iron-bearing phases (pyrite, sphalerite) are common in veins from both types of host rocks; however, calcium-bearing phases are absent in veins hosted by banded quartzite. Therefore, the Fe^{2+} content of ore-forming fluids could have remained constant while the Ca^{2+} content varied as fluids interacted with host rocks containing different amounts of calcium (high Ca^{2+} marble versus low Ca^{2+} quartzite).

STRUCTURAL CONTROLS

The average northeast trend of the veins (fig. 8) suggests that the veins formed in regional northeast-striking shear zones in central Idaho. The age of these structures has been a matter of debate. Although many Eocene shear zones related to the Trans-Challis fault system strike northeast (Bennett, 1986; Kiilsgaard and others, 1986), northeast-striking Cretaceous shear zones have been documented regionally. Lund and others (1986, p. 995) indicated that northeast-striking fractures in the Buffalo Hump district, 50 km east of Riggins, Idaho, host Cretaceous lode deposits in the Idaho batholith and were formed during a regional extension event about 71 Ma. Gammons and others (1985) documented similar Cretaceous structures in the Big Creek district, 80 km southeast of Riggins, Idaho. In both the Big Creek and Buffalo Hump districts, as in the Carrietown area, mineral deposits are concentrated in roof pendants in the batholith.

A regional network of Cretaceous northeast-striking faults and shear zones are in the Smoky Mountains (Park, in press), some of which are manifest in bedding-parallel faults in the Dollarhide Formation. At least some silver-lead-zinc deposits in the Carrietown area (for example, Silver Star and Fletcher mines) are interpreted to have formed in bedding-parallel faults because the veins are parallel with the bedding of their host rocks. In the Last Chance mine (fig. 3), the orebody formed where the fault plane was locally flattened from an otherwise uniform steep dip. The Carrie Leonard orebody (fig. 3) formed at the intersection of two northeast-striking faults.

In previous work, Darling (1987, 1988) suggested that the abundance of mineral deposits along the contact between banded quartzite and carbonaceous marble was caused by ore fluids migrating along the thrust fault that separated the two units. Although the concept of a regional thrust fault has been disproven, differences in rheological behavior between quartzite and marble during Mesozoic folding may have caused development of open fissures for the movement of ore fluids, as suggested by Whitman (1990, this volume).

THERMAL CONDITIONS OF MINERALIZATION

SULFUR-ISOTOPE GEOTHERMOMETRY

Howe and Hall (1985), in an isotopic study of the black-shale belt, performed nine sulfur-isotope analyses in the Carrietown area, four of which were used to calculate temperatures of ore formation by sulfur-isotope fractionation methods (Ohmoto and Rye, 1979). The analyses yielded temperatures of 269±45°C on a sphalerite-galena pair from the Carrie Leonard mine and 375±45°C on a pyrite-galena pair from an unnamed dump (Howe and Hall, 1985, p. 192). In polished section, both sulfide pairs commonly exhibit mutual grain boundaries, suggesting that isotopic equilibrium was obtained and that the calculated temperatures are reliable; however, because galena is paragenetically later than pyrite and sphalerite (fig. 9), the temperatures may only reflect isotopic equilibrium during deposition of galena.

CUBANITE EXSOLUTION GEOTHERMOMETRY

A minimum temperature of ore formation of 250°C–300°C is indicated by cubanite exsolution lamellae in chalcopyrite (Ramdohr, 1980, p. 639). Cubanite exsolution lamellae are almost ubiquitous in chalcopyrite from mines near Carrietown, but, because they can only be observed under the microscope, their presence was not established in chalcopyrite from mines along Little Smoky Creek and from the Buttercup mine.

ARSENOPYRITE GEOTHERMOMETRY

Clark (1960) showed that the equilibrium assemblage of pyrite and arsenopyrite melts to form pyrrhotite and liquid from 491°C±12°C (at 1 bar) to as much as 528°C±10°C (at 2,070 bars). Thus, the pyrite-arsenopyrite intergrowths at many of the mines establish a thermal maximum at 491°C–528°C (if $P_T < 2,070$ bars). Furthermore, two arsenopyrite separates from the Alma and Tyrannis Mines were X-rayed to determine their atomic percent arsenic content by measurement of the *d*-spacing of the (131) plane. The samples were prepared and analyzed according to methods outlined in Kretschmer and Scott (1976). Arsenopyrite from the Alma and Tyrannis mines contain 31.55±0.45 and 32.68±0.45 mole percent arsenic, amounts yielding temperatures of ore formation of 394°C±25°C, and 465°C±30°C, respectively. The temperatures were estimated by the intersection of the pyrite-pyrrhotite univariant and the atomic percent arsenic isopleths in figure 7 of Kretschmer and Scott (1976). Both ore samples from the Alma and Tyrannis mines contained the necessary S_2 buffering assemblage of pyrite-pyrrhotite required for use of the arsenopyrite geothermometer (Kretschmer and Scott, 1976).

A summary of all available thermal information from the Carrietown mineralized area is listed in table 3. Thermal constraints imposed by fixed-point geothermometers agree well with temperatures calculated from sulfur isotopes and arsenopyrite and are consistent with a mesothermal to hypothermal (250°C–450°C) environment. Thermal data collected from ore deposits in the Carrietown mineralized area are generally consistent with those for similar deposits in the Wood River district, 30 km to the east (Hall and Czamanske, 1972; Hall and others, 1978).

OXYGEN AND SULFUR FUGACITIES OF ORE FLUIDS

Estimation of O_2 and S_2 fugacity in ore-forming fluids of the Carrietown area is illustrated in figure 10. Stability fields for pyrite, pyrrhotite, magnetite, and hematite were calculated at 350°C (average ore-forming temperature) by use of thermodynamic data of Robie and others (1978) and methods outlined in Garrels and Christ (1965) and Nordstrom and Munoz (1985). Because ores hosted by banded quartzite commonly contain pyrite and pyrrhotite, ore-forming fluids have sulfur fugacities in the 10^{-11}–10^{-9} bar range at 350°C, whereas ore hosted by carbonaceous marble (pyrite, without pyrrhotite) has higher sulfur fugacities (about 10^{-6}–10^{-4} bars). The lack of iron oxide minerals (magnetite, hematite) in samples from any mine in the area indicates low oxygen fugacities (about 10^{-35} bars), reflective of strongly reducing conditions.

SOURCE OF SULFUR AND LEAD

In their isotopic study of the black shale belt, Howe and Hall (1985) performed nine sulfur-isotope analyses from four mines in the Carrietown district. The mines and sulfides analyzed from them include (1) the Carrie Leonard mine, galena and sphalerite; (2) the Horn Silver mine, galena and pyrite; (3) the Upper prospect, galena, molybdenite and sphalerite; and (4) an unnamed dump, galena and pyrite. As previously discussed, analyses from the Carrie Leonard mine and the unnamed dump were used to calculate sulfur-isotope fractionation temperatures. The location of the unnamed dump is not known, but the Upper prospect described by Howe and Hall (1985) is most likely the Silver Crown mine because it is the only location in the Carrietown district that has a sphalerite+galena+molybdenite sulfide assemblage. $\delta^{34}S$ values for the analyzed sulfides range from –0.1 to +6.2 per mil (Howe and Hall, 1985, p. 192). These isotope values are depleted in ^{34}S with respect to Pennsylvanian-Permian seawater (that is, Dollarhide Formation) and indicate a more depleted source of sulfur for ore deposits in the Carrietown district. Howe and Hall (1985) indicated that ore deposits having this isotopic signature suggest isotopic mixing between enriched sedimentary sulfur and strongly depleted magmatic sulfur.

Sanford and Wooden (this volume), as part of an isotopic study of the black shale belt, performed lead-isotope analyses on 10 samples from various mines in the Carrietown area (specific mines unknown). The samples included galena from both vein-type and bedding-conformable deposits in the Dollarhide Formation. The galena separates yielded $^{206}Pb/^{204}Pb$ ratios of 18.0–19.2 and $^{208}Pb/^{204}Pb$ ratios of 39.5–40.0. These Pb/Pb ratios are not consistent with a Paleozoic argillite source and, if combined with other lead-isotopic data from similar silver-lead-zinc deposits in central Idaho (Carrietown, Lava Creek, and Minnie Moore lead types), yield a calculated model age of 2,400±100 Ma. These data indicate that the source of lead in ore deposits from the above mines is not the Dollarhide Formation but rather upper crustal Precambrian basement rocks (Sanford and Wooden, this volume). Furthermore, because epigenetic and bedding-conformable galena have almost identical isotopic ratios, Sanford and Wooden (this volume) also infer that bedding-conformable galena has an epigenetic rather than syngenetic origin.

The sulfur- and lead-isotope data suggest that magmas were generated in the lower crust and, as they ascended, transported their inherited (Precambrian) lead and depleted magmatic sulfur. The magma crystallized, differentiated, and may have generated late-stage magmatic (metal-bearing) hydrothermal fluids that circulated through the Dollarhide Formation scavenging enriched sedimentary sulfur. The resulting hydrothermal fluid precipitated sulfide phases having Precambrian lead-isotopic and mixed (igneous+sedimentary) sulfur-isotopic signatures.

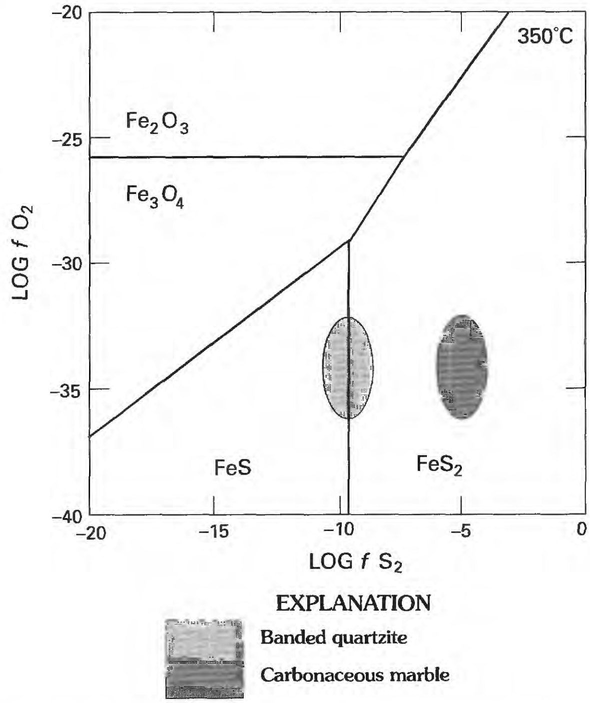


Figure 10. Log fO_2-log fS_2 diagram illustrating volatile fugacity of ore-forming fluids between banded quartzite and carbonaceous marble of the Dollarhide Formation, Carrietown mineralized area, Idaho.

AGE OF MINERALIZATION

Because the ore deposits of the Carrietown mineralized area have an epigenetic origin and are partly hosted by granodiorite at the Silver Crown mine (Darling, 1987) and granite at the Main Moly prospect (Gehlen, 1983), the deposits are not older than Late Cretaceous. Furthermore,

Table 3. Summary of available thermal data from mines in the Carrietown mineralized area, Idaho.

Mine name	Method	Notes	Estimated temperature	Source
Mines near Carrietown	Cubanite exsolution thermometer	Thermal minimum	250°C–300°C	Darling (1987, 1988).
Carrie Leonard mine	Sulfur-isotope thermometer	Sphalerite-galena pair	269°C±45°C	Howe and Hall (1985).
Unnamed dump	Sulfur-isotope thermometer	Pyrite-galena pair	375°C±45°C	Howe and Hall (1985).
Alma mine	Arsenopyrite thermometer	*d*-spacing of FeAsS (131) plane	394°C±25°C	This report.
Tyrannis mine	Arsenopyrite thermometer	*d*-spacing of FeAsS (131) plane	465°C±30°C	This report.
Mines near Carrietown	Pyrite-arsenopyrite thermometer	Thermal maximum	491°C–528°C	Darling (1987, 1988).

the principal mineralized structure in the Dollarhide mine is cut by a dacite porphyry dike (Umpleby, 1915; Darling, 1987), indicating the deposits are no younger than Eocene (47.2 Ma). Because the ore deposits formed at elevated temperatures (250°C–450°C), mineralization is interpreted to be temporally related to either Cretaceous or Eocene intrusive activity.

The thermal data and field relations previously discussed suggest a Cretaceous age of mineralization. Ore deposits in the Dollarhide, Silver Star, Stormy Galore, King of the West, and Buttercup mines are less than 150 m vertically below the unconformable contact between the hosting Dollarhide Formation and overlying Eocene Challis Volcanic Group (figs. 2–4, 6). The ore-forming event is inferred to be prevolcanic because the volcanic rocks (which are highly susceptible to alteration) lack hydrothermal alteration (Gehlen, 1983) and do not host sulfide mineral deposits. An Eocene, prevolcanic ore-forming event would, therefore, require a very shallow depth, an environment that conflicts with observed mesothermal to hypothermal mineral assemblages and the thermal minimum of 250°C–300°C.

Silver-lead-zinc deposits hosted by the Dollarhide Formation lack extensive hydrothermal alteration and therefore are without suitable material for radiometric dating by K-Ar or Ar-Ar methods. The lack of alteration in the Dollarhide Formation is probably a result of its composition. The banded quartzite and marble contain few, if any, feldspar or iron-magnesium silicate minerals, phases typically susceptible to hydrothermal alteration. Similar silver-lead-zinc deposits hosted by granodiorite at the Ontario mine in the Rooks Creek mineralized area, 18 km to the northeast, have been dated, however, at 91.09±0.25 Ma by Ar-Ar methods on hydrothermal muscovite (Park, 1990, in press).

At the Main Moly prospect in the Little Smoky Creek district (fig. 4), muscovite in Cretaceous leucocratic granite is dated at 77.4±2.6 Ma (M. Tschanz, 1982, unpublished report; cited in Gehlen, 1983). This date was cited by Darling (1987) as the age of the leucocratic granite; however, we now infer that muscovite in this rock is hydrothermal in origin because (1) it is present as patchy masses spatially associated with quartz-molybdenite veins, (2) it lacks the primary igneous texture (books and laminated flakes) typical of two-mica granite in the core of the Idaho batholith (Kiilsgaard and Lewis, 1985), and (3) an igneous origin requires a minimum magmatic pressure of 3.7±0.2 kb, a pressure higher than the 2.5–3.5-kb range inferred from metamorphic mineral assemblages (Whitman, 1990, this volume). The 3.7±0.2-kb minimum is derived from the intersection of the upper stability curve of muscovite+quartz (Chatterjee and Johannes, 1974) and the minimum granite (Ab+Or+Qtz+H_2O) melting temperature (Tuttle and Bowen, 1958). For these textural and petrologic reasons, the K-Ar date of 77.4±2.6 Ma is inferred to reflect the age of quartz-molybdenite mineralization in the Dollarhide Mountain quadrangle. This date and the 91.09-Ma date for hydrothermal muscovite in the Rooks Creek district indicate that active ore-forming fluids were present during the Late Cretaceous but suggest that silver-lead-zinc mineralization is temporally related to the older granodiorite phases of the Idaho batholith, and molybdenum mineralization is related to the younger leucocratic granite.

PROPOSED PRESSURE-TEMPERATURE-TIME PATH

A proposed pressure-temperature-time path from the Late Cretaceous to the Eocene is illustrated in figure 11. Rocks in the Carrietown mineralized area underwent regional heating from intrusion of Cretaceous plutons and contact metamorphism (~84 Ma) at 500°C–600°C and 2.5–3.5 kb (Whitman, 1990, this volume). Ore deposits are interpreted to have formed soon after metamorphism (~75–80 Ma) and probably with little change in crustal level (inferred isobaric cooling). Subsequent uplift and erosion resulting in exposure of the Dollarhide Formation and Cretaceous intrusive rocks at the surface during Eocene time (~47 Ma) is indicated by the unconformable contact between the Challis Volcanic Group and both these units. The convex-upward pressure-temperature-time path illustrated in figure 11 is typical for magmatic-hydrothermal ore deposits.

GENERAL MODEL AND COMPARISON TO OTHER SILVER-LEAD-ZINC DEPOSITS IN THE BLACK SHALE BELT OF IDAHO

Our work suggests that the important controls on formation of mineral deposits in the Carrietown area are (1) presence of Fe^{2+}- and carbon-bearing roof pendants to change fluid chemistry from relatively oxidizing to reducing; (2) proximity to the top of the batholith, perhaps on a cupola containing metal-enriched fluids from the last stages of magmatic crystallization; and (3) open structures for fluid migration and ore deposition.

Geologic conditions thought to be common to all silver-lead-zinc vein deposits in the central Idaho black shale mineral belt were discussed by Hall (1985, p. 127). These conditions, as we now understand them, are discussed as they relate to ore deposits in the Carrietown area.

Host rocks.—Host rocks for mineral deposits in the Smoky Mountains were generally described as "black-shales" by Hall (1985). The term black shale is a misnomer in the Carrietown area because banded quartzite, carbonaceous marble, and phyllite of the lower member of the Dollarhide Formation are the principal host rocks. Only the phyllite could have originated as a black shale. More than half the deposits are hosted by banded quartzite, which clearly lacks compositional or textural similarities to black shale.

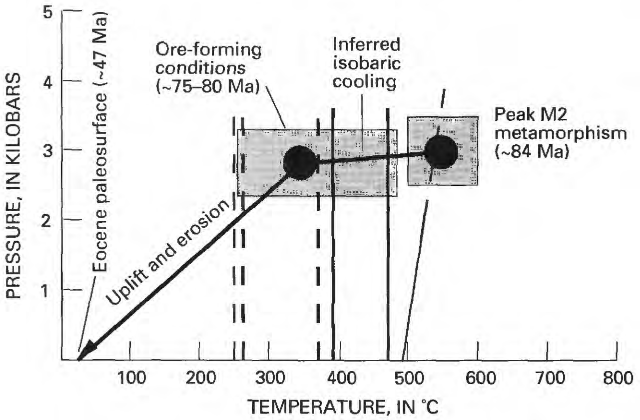


Figure 11. Proposed pressure-temperature-time path for rocks in the Carrietown mineralized area, Idaho. Vertical solid heavy lines are arsenopyrite temperatures; vertical dashed heavy lines are sulfur-isotope temperatures; vertical thin line is cubanite thermal minimum; inclined solid-dashed line is pyrite-arsenopyrite thermal maximum. Square box indicates metamorphic conditions (Whitman, this volume); rectangular box indicates inferred ore-forming conditions. Very heavy line represents inferred pressure-temperature time path from plutonism, metamorphism, and sulfide mineralization during the Late Cretaceous to volcanism of the Challis Volcanic Group and associated hypabyssal plutons during the middle Eocene.

Relationship of mineralization to regional thrust faults.—Hall and others (1978) and Hall (1985) developed a model in which mineralized veins are under but close to regional thrust faults, citing the Triumph mine southeast of Ketchum as an example. As discussed earlier in this paper, the interpretation of a regionally extensive thrust fault separating the Dollarhide Formation from the "Carrietown sequence" has been abandoned and with it the inference that the thrust fault acted as a principal pathway for hypogene ore fluids (Darling, 1987, 1988). Furthermore, Hall's mapping of a regional Wood River thrust fault in the area of the Triumph Mine, south of Ketchum, has been reinterpreted (Link and others, this volume; Turner and Otto, this volume). The permeability barrier in the Triumph area is the sheared unconformity between the Devonian Milligen Formation and the overlying Middle Pennsylvanian and Lower Permian Wood River Formation.

Relationship of mineralization to granitic intrusive rocks.—The general model of Hall (1985) states that mineralized veins are close to contacts of granitic intrusive rocks, predominantly of Cretaceous age but also of Eocene age. Our findings in the Carrietown area suggest that only the Cretaceous intrusive rocks directly influenced formation of mineral deposits. Eocene mineralization is unlikely.

Characteristics of gangue minerals.—The general model of Hall (1985) states that the most productive deposits have siderite-quartz gangue rather than calcite-quartz gangue. This is true in the Carrietown mineralized area. In all three districts of the area, ore grades of deposits having quartz-calcite gangue are generally lower than those of deposits having quartz-siderite gangue. For example, ore in the Carrie Leonard and King of the West mines (quartz-calcite gangue) averaged 150 and 120 ounces of silver per ton, respectively, whereas ore in the Stormy Galore and Isabella mines averaged 400 ounces of silver per ton (Umpleby, 1915). Furthermore, geochemical studies of both host rocks (Darling, 1987) indicate that banded quartzite (veins with siderite-quartz gangue) has higher overall trace-metal contents than does calcareous marble (veins with calcite-quartz gangue), and future exploration in the district should focus on locating deposits in banded quartzite.

REFERENCES CITED

Bennett, E.H., 1986, Relationship of the trans-Challis fault system in central Idaho to Eocene and Basin and Range extensions: Geology, v. 14, p. 481–484.

Chatterjee, N.D., and Johannes, W., 1974, Thermal stability and standard thermodynamic properties of synthetic $2M_1$-muscovite, $KAl_2[AlSi_3O_{10}(OH)_2]$: Contributions to Mineralogy and Petrology, v. 48, p. 89–114.

Clark, L.A., 1960, The Fe-As-S system—Phase relations and applications: Economic Geology, v. 55, p. 1345–1381, 1631–1652.

Craig, J.R., and Vaughan, D.J., 1981, Ore microscopy and ore petrography: New York, John Wiley and Sons, 406 p.

Darling, R.S., 1987, The geology and ore deposits of the Carrietown silver-lead-zinc district, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 168 p.

———1988, Ore deposits of the Carrietown silver-lead-zinc district, Blaine and Camas Counties, Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 193–198.

———1989, Geologic map and cross-sections of the northern half of the Dollarhide Mountain quadrangle: Idaho Geological Survey Technical Report 89–1, 1 plate.

Federspiel, F.E., Gabby, P.N., Neumann, T.N., and Leszcykowski, A.M., 1987, Mineral resources of the South Boise Yuba Study Area and additions, Blaine, Camas, and Elmore Counties, Idaho: U.S. Bureau of Mines Open-File Report MLA 64–87, 117 p.

Federspiel, F.E., Horn, M.E., Campbell, H.W., and Peters, T.J., 1992, Mineral resource investigation of the Smokies Study Area, Blaine, Camas, and Elmore Counties, Idaho: U.S. Bureau of Mines Open-File Report MLA 4–92, 244 p.

Gammons, C.H., Rose, A.W., Snee, L.W., and Lund, K., 1985, Paragenesis, fluid inclusions, and Ar-dating of the Big Creek mining district, Valley County, central Idaho: Geological Society of America Abstracts with Programs, v. 17, p. 588.

Garrels, R.M., and Christ, C.L., 1965, Solutions, minerals and equilibria: New York, Harper and Row, 380 p.

Gehlen, W.T., 1983, The geology and mineralization of the eastern part of the Little Smoky Creek mining district, Camas County, Idaho: Moscow, University of Idaho, M.S. thesis, 116 p.

Geslin, J.K., 1986, The Permian Dollarhide Formation and Paleozoic Carrietown sequence in the SW¼ of the Buttercup

Mountain quadrangle, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 116 p.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

Hall, W.E., and Czamanske, G.K., 1972, Mineralogy and trace element content of the Wood River lead-silver deposits, Blaine County, Idaho: Economic Geology, v. 67, p. 350–361.

Hall, W.E., Rye, R.O., and Doe, B.R., 1978, Wood River mining district, Idaho—Intrusion-related lead-silver deposits from a country rock source: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 579–592.

Howe, S.S., and Hall, W.E., 1985, Light-stable isotope characteristics of ore-systems in central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–P, p. 183–192.

Kiilsgaard, T.H., Fisher, F.S., and Bennett, E.H., 1986, The trans-Challis fault system and associated precious metal deposits, Idaho: Economic Geology, v. 81, p. 721–724.

Kiilsgaard, T.H., and Lewis, R.S., 1985, Plutonic rocks of Cretaceous age and faults in the Atlanta lobe of the Idaho batholith, Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–B, p. 29–42.

Kretschmer, U., and Scott, S.D., 1976, Phase relations involving arsenopyrite in the system Fe-As-S and their application: Canadian Mineralogist, v. 14, p. 364–386.

Lewis, R.S., in press, Alteration and mineralization in the eastern part of the Soldier Mountains, Camas County, Idaho: U.S. Geological Survey Bulletin 2064–V.

Link, P.K., Skipp, Betty, Hait, M.H.,Jr., Janecke, Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knob, Pioneer, Boulder and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 5–42.

Lund, K., Snee, L.W., and Evans, K.V., 1986, Age and genesis of precious metals deposits, Buffalo Hump district, central Idaho—Implications for depth of emplacement of quartz veins: Economic Geology, v. 81, p. 990–996.

Mahoney, J.B., Link, P.K., Burton, B. R., Geslin, J.K., and O'Brien, J.P., 1991, Pennsylvanian and Permian Sun Valley Group, Wood River Basin, south-central Idaho, *in* Cooper, J., and Stevens, C., eds., Paleozoic paleogeography of the Western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 2, p. 551–580.

Marvin, R.F., and Dobson, S.W., 1979, Radiometric ages—Compilation B, U.S. Geological Survey: Isochron West, no. 26, p. 17.

McIntyre, D.H., Ekren, E.B., and Hardyman, R.F., 1982, Stratigraphic and structural relations of the Challis volcanics in the eastern half of the Challis 1°×2° quadrangle, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 3–22.

Moye, F.J., Hackett, W.R., Blakely, J.D., Snider, L.G., 1988, Regional geologic setting and volcanic stratigraphy of the Challis volcanic field, central Idaho, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 87–98.

Moye, F.J., Hackett, W.R., Snider, L.G., and Mertzman, S.A., in press, Stratigraphy, chronology, and petrology of the southeastern Challis volcanic field, central Idaho: U.S. Geological Survey Bulletin 2064–EE.

Nordstrom, D.K., and Munoz, J.L., 1985, Geochemical thermodynamics: Menlo Park, Calif., Benjamin/Cummings, 477 p.

O'Brien, J.P., 1991, Stratigraphy of the Pennsylvanian-Permian Dollarhide Formation in the Buttercup Mountain quadrangle, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 163 p.

Ohmoto, Hiroshi, and Rye, R.O., 1979, Isotopes of sulfur and carbon, *in* Barnes, H.L., ed., Geochemistry of hydrothermal ore deposits (2nd ed.): New York, Wiley-Interscience, p. 509–567.

Park, W.A., 1990, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: Pocatello, Idaho State University, M.S. thesis, 135 p.

Park, W.A., in press, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–CC.

Picot, P., and Johan, Z., 1982, Atlas of ore minerals: Amsterdam, Netherlands, Elsevier, 458 p.

Ramdohr, P., 1980, The ore minerals and their intergrowths (2nd ed.): Oxford, England, Pergamon, 1,207 p.

Robie, R.A., Hemingway, B.R., and Fisher, J.R., 1978, Thermodynamic properties of minerals and related substances at 298.15 K and 1 bar (10^5 pascals) pressure and at higher temperatures: U.S. Geological Survey Bulletin 1452, 456 p.

Ross, C.P., 1930, Geology and ore deposits of the Seafoam, Alder Creek, Little Smoky Creek, and Willow Creek mining districts, Custer and Camas Counties, Idaho: Idaho Bureau of Mines and Geology Pamphlet 33, p. 7–18.

Skipp, B., and Hall, W.E., 1980, Upper Paleozoic paleotectonics and paleogeography of Idaho, *in* Fouch, T.D., and Magathan, E.R., eds., Paleogeography of the West-Central United States: Rocky Mountain Section, Society of Economic Geologists and Mineralogists, West-Central Paleogeography Symposium 1, p. 387–422.

Stewart, D.E., Schmidt, K.L, Moye, F.J., Mahoney, J.B., and Link, P.K., in press, Eocene intrusive complexes in the Smoky and Boulder Mountains, south-central Idaho: U.S. Geological Survey Bulletin 2064–HH.

Streckeisen, A., 1976, To each plutonic rock its proper name: Earth Science Reviews, v. 12. p. 1–33.

Tuttle, O.F., and Bowen, N.L., 1958, Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O: Geological Society of America Memoir 74, 153 p.

U.S. Bureau of Mines, 1990, Mineral commodity summaries: 201 p.

Umpleby, J.B., 1915, Ore deposits in the Sawtooth quadrangle, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 580–K, p. 221–249.

Wavra, C.S., 1985, Structural geology and petrology of the SW¼ of the Mahoney Butte quadrangle, Blaine County, Idaho: Moscow, University of Idaho, M.S. thesis, 103 p.

Wavra, C.S., Issacson, P.E., and Hall, W.E., 1986, Studies of the Idaho black shale belt—Stratigraphy, depositional environment and economic geology of the Permian Dollarhide Formation: Geological Society of America Bulletin, v. 97, p. 1504–1511.

Whitman, S.K., 1990, Metamorphic petrology and structural geology of the Pennsylvanian-Permian Dollarhide Formation, Blaine and Camas Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 108 p.

Metamorphism and Deformation on the Eastern Margin of the Idaho Batholith, Blaine and Camas Counties, South-Central Idaho

By Shelly K. Whitman

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–Q

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

FIGURES

TABLE

Metamorphism and Deformation on the Eastern Margin of the Idaho Batholith, Blaine and Camas Counties, South-Central Idaho

By Shelly K. Whitman[1]

ABSTRACT

Rocks of the Middle Pennsylvanian to Lower Permian Dollarhide Formation that crop out on the southeastern edge of the Cretaceous Idaho batholith, south-central Idaho, have undergone two Late Cretaceous pulses of heating and strain related to intrusion of the batholith. During the first metamorphic event (M_1), andalusite, tremolite, diopside, and staurolite porphyroblasts formed pre- or syn-kinematically in the Dollarhide Formation along with a poorly developed biotite foliation (S_1). Retrogression after M_1 resulted in formation of elliptical mica spots and poikiloblastic porphyroblasts and pseudomorphs of quartz and potassium feldspar after andalusite and staurolite.

Features of a second metamorphic event (M_2) overprint the earlier M_1 prograde and retrograde features. During the M_2 event, porphyroblasts of sillimanite, staurolite, andalusite, tremolite, diopside, and wollastonite formed pre- or syn-kinematically along with a well-developed penetrative biotite foliation (S_2). These parageneses constrain the temperature of M_2 metamorphism to 525°C–600°C, the pressure to 2.5–3.5 kb, and the fluid composition to P_{CO_2}<0.18. Whole-rock K-Ar analysis of schist containing well-preserved M_2 minerals yields an age of 83.9±3.4 Ma (R.L. Armstrong, written commun., 1990), interpreted to be the minimum age of the M_2 metamorphic event. This age, and the synkinematic nature of the porphyroblasts, indicates that both metamorphism and deformation are Late Cretaceous in age.

Emplacement of the batholith caused strain in the surrounding Dollarhide Formation. Deformation manifested by map-scale folds began before peak metamorphism (M_2) and continued after late-stage Cretaceous leucocratic dike emplacement, although most folding probably occurred during development of S_2 foliation. Fold hinges and mineral, boudin, and intersection lineations are all parallel and have an average orientation of 45°, N. 70° E., perpendicular to the south-southeast trend of fold hinges measured throughout south-central Idaho. The unusual orientation of fold axes in the study area is thought to be due to high shear strain in rocks along the margin of the batholith. Prolonged strain in the margin area could have caused the rotation of fold axes to their present east-northeast direction.

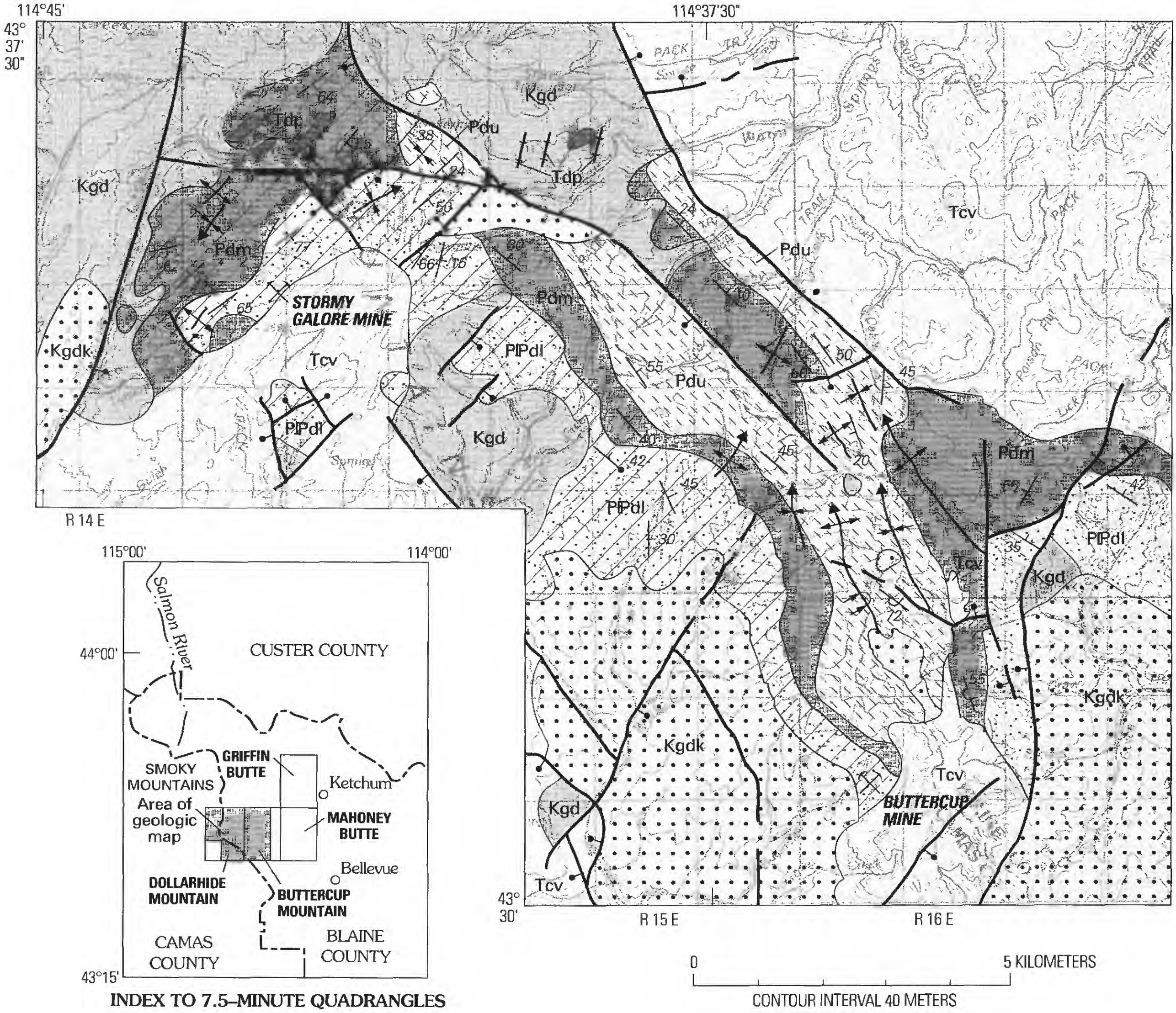
A previously mapped regionally extensive thrust fault separating the Paleozoic so-called Carrietown sequence from the overlying Pennsylvanian and Permian Dollarhide Formation has been reinterpreted as a stratigraphic contact between phyllitic quartzite and siliceous marble. Both rock types are now included in the Dollarhide Formation; the name Carrietown sequence was previously abandoned. Mineralized shear zones along the contact most likely formed due to rheologic differences between quartzite and marble during deformation and are not the result of regional thrusting.

INTRODUCTION

The study area discussed herein is in the Smoky Mountains along the southeastern edge of the Atlanta lobe of the Idaho batholith in Blaine and Camas Counties, south-central Idaho (fig. 1). The Middle Pennsylvanian to Lower Permian Dollarhide Formation in this area is part of the central Idaho black shale mineral belt, a northwest-trending belt 145 km long and 15–45 km wide composed of black shale, argillite, and carbonate rocks that crop out on the eastern side of the Idaho batholith from the Salmon River in the south-central Challis 1°×2° quadrangle south to Bellevue in the Hailey 1°×2° quadrangle (Hall, 1985) (fig. 1). Mineral deposits hosted by sedimentary rocks in the black shale belt include syngenetic and epigenetic silver-lead-zinc deposits, as well as tungsten skarn deposits (Hall, 1985; Darling and others, this volume; Link and others, this volume; Park, in press).

Umpleby (1915) and Ross (1930) were the first to map and study the mineral deposits of the Smoky Mountains. The contact between Cretaceous batholith and Paleozoic sedimentary rocks in the Buttercup Mountain and Dollarhide Mountain 7.5-minute quadrangles was mapped in detail by Hall (1985), who delineated and named the Dollarhide Formation, which had previously been included in the Wood

[1]Matney Frantz Engineering, Bozeman, Montana 59771.



EXPLANATION

- Tcv — **Challis Volcanic Group (Eocene)**—Mainly dacite and andesite flow rocks
- Tdp — **Dacite porphyry (Eocene)**—Hypabyssal intrusive rock
- Tdp — **Dacite porphyry dike (Eocene)**
- Kgd — **Biotite granodiorite (Cretaceous)**—Includes leucogranite and quartz diorite
- •Kgdk — **Hornblende-biotite granodiorite (Cretaceous)**
- Pdu — **Upper member of the Dollarhide Formation (Lower Permian)**
- Pdm — **Middle member of the Dollarhide Formation (Lower Permian)**
- PIPdl — **Lower member of the Dollarhide Formation (Lower Permian to Middle Pennsylvanian)**

- **Contact**
- **Area of strong metamorphic overprint**
- 30 **Strike and dip**
- 30 **Strike and dip where top of bed is known**
- 30 **Overturned bedding and dip**
- **Vertical bedding**
- **Normal fault**—Bar and ball on downthrown side; dashed where approximately located; dotted where concealed
- **Anticline**—Showing direction of plunge
- **Overturned anticline**—Showing direction of plunge
- **Syncline**—Showing direction of plunge
- **Overturned syncline**—Showing direction of plunge

Figure 1. Map showing geology of the eastern margin of the Idaho batholith and adjacent Dollarhide Formation in the Buttercup Mountain and Dollarhide Mountain 7.5-minute quadrangles, Blaine and Camas Counties, south-central Idaho. Modified from Worl and others (1991) and P.K. Link (unpublished mapping).

River Formation (Middle Pennsylvanian to Lower Permian), and mapped the Carrietown sequence–Dollarhide Formation contact as a thrust fault.

In the Smoky Mountains, the eastern edge of the Atlanta lobe of the Idaho batholith intrudes phyllite, quartzite, marble, and argillite of the Dollarhide Formation (Hall, 1985; Wavra, 1985; Geslin, 1986; Wavra and others, 1986; Darling, 1987; Whitman, 1990; Mahoney and others, 1991; O'Brien, 1991). The lower part of the Dollarhide Formation was divided into three stratigraphic units by Whitman (1990). Rocks of unit 1 primarily consist of phyllite, phyllitic quartzite dominates unit 2, and unit 3 is composed of marble and argillite. The contact between units 2 and 3 was formerly interpreted as a thrust fault (Hall, 1985; Geslin, 1986; Darling, 1987). As described later, structural and metamorphic similarities between units 2 and 3 instead indicate a continuous stratigraphic section.

The present study concerns the metamorphic and deformational history of the eastern edge of the Atlanta lobe of the Idaho batholith in the Smoky Mountains west of Ketchum, Idaho, and is built on several theses (Wavra, 1985; Geslin, 1986; Darling, 1987; Whitman, 1990; O'Brien, 1991), as well as published and unpublished mapping by R.S. Lewis and P.K. Link. Six structural and metamorphic domains were studied in detail (Whitman, 1990): Tyrannis Creek, Blackhorse Canyon, Little Smoky Creek, West Fork of Willow Creek, Willow Creek, and Buttercup mine (fig. 1). The bulk chemistry of the sedimentary protolith has contributed to the useful and diagnostic metamorphic parageneses observed and helps to tightly constrain the pressure, temperature, and fluid composition at the time of metamorphism. Regional interpretation of the style and timing of metamorphism can be made from comparison of the petrographic elements and pressure-temperature-P_{CO_2} conditions between different areas within the study area.

Acknowledgments.—I would like to thank the many people involved in this project who shared their thoughts and time including Paul Link, David Rodgers, J.P. O'Brien, and Robert Darling, among others. This study was supported by the U.S. Geological Survey U.S. Geological Survey Conterminous United States Mineral Assessment Program (CUSMAP Program), Idaho State Board of Education Grant 89–56 to P.K. Link, and an Idaho State University Student Research Committee grant. The manuscript was reviewed by P.K. Link, D.W. Rodgers, R.G. Worl, and R. Darling.

REGIONAL SETTING

Rocks of the Middle Pennsylvanian to Lower Permian Dollarhide Formation, Cretaceous intrusive rocks of the Idaho batholith, and Eocene dacitic volcanic and intrusive complexes of the Challis Volcanic Group crop out in the study area (fig. 1).

PENNSYLVANIAN AND PERMIAN STRATA OF THE DOLLARHIDE FORMATION

Pennsylvanian and Permian strata of the Dollarhide Formation consist of phyllite, phyllitic quartzite, banded quartzite, calc-silicate hornfels, marble, and argillite derived from calcareous pelitic and psammitic protoliths. Rocks of the Dollarhide Formation have been divided into three members and comprise the southwestern carbonaceous basinal facies of the Middle Pennsylvanian to Lower Permian Sun Valley Group (Mahoney and others, 1991; Link and others, this volume). Contrary to earlier reports (Hall, 1985; Link and others, 1988), the Dollarhide Formation is now known to contain Pennsylvanian as well as Lower Permian rocks (Link and others, this volume; Betty Skipp, oral commun., 1991).

The lower member of the Dollarhide Formation has been divided into three stratigraphic units (Whitman, 1990). The stratigraphic units of the lower part of the Dollarhide Formation can be delineated in the field but, due to the scale of figure 1, are not shown in that figure. Unit 1 consists of brownish-purple phyllite having a golden phyllitic sheen, local black to brown fine- to medium-grained schist, and purple-brown fine-grained phyllitic quartzite. Unit 2 is characterized by purple-brown phyllite and phyllitic quartzite, centimeter-scale banded quartzite, gray to white, calcareous to siliceous, massive to finely laminated marble and calc-silicate hornfels, and moderately thin bedded intervals (<10 m) of black carbonaceous marble. The boundary between units 1 and 2 is defined by the lowest occurrence of pink to white centimeter-scale pods or layers within phyllite or quartzite. Unit 3 consists of black carbonaceous marble, gray to white calc-silicate hornfels, and very fine grained black argillite. The boundary between units 2 and 3 is placed at the lowest occurrence of thick (>10 m) beds of black carbonaceous marble or argillite. Rocks of unit 3 rarely exhibit a foliation in outcrop, but under microscopic examination they exhibit a strong foliation defined by the parallel elongation of recrystallized calcite and quartz. Rocks of unit 3 contain tremolite, diopside, wollastonite and sillimanite, as do rocks of the appropriate bulk composition in units 1 and 2.

The contact between units 2 and 3 was previously mapped as a thrust fault (Hall, 1985; Geslin, 1986; Darling, 1987), but structural and metamorphic similarities of units 2 and 3 instead support the interpretation of a stratigraphically continuous section.

CRETACEOUS IDAHO BATHOLITH AND METAMORPHIC AUREOLE

The Cretaceous Idaho batholith in the southern Smoky Mountains contains hornblende-biotite granodiorite, biotite granodiorite, and leucocratic granite that have yielded radiometric dates of 93–77 Ma (Darling, 1987; Lewis, in press).

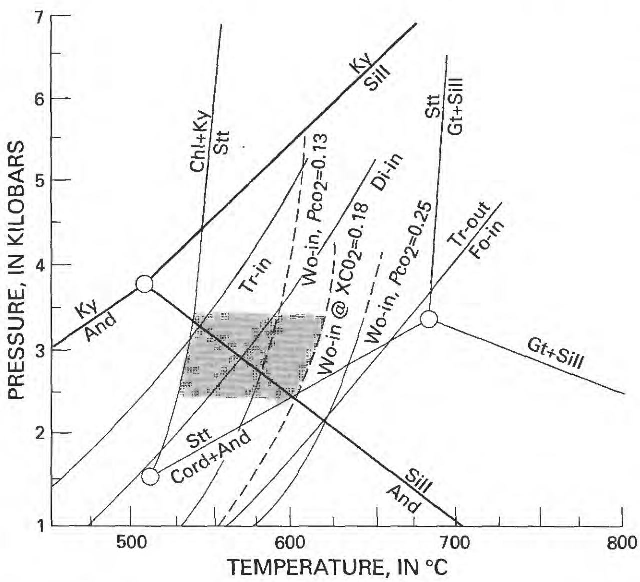


Figure 2. Pressure (P)-temperature (T)-partial pressure (P_{CO_2}) diagram for the metamorphic aureole of the Dollarhide Formation. The experimental curves are discussed in the text. The shaded area represents the P-T-P_{CO_2} field for the Dollarhide Formation adjacent to the Idaho batholith in the southern Smoky Mountains study area; data from Whitman (1990). Minerals: And, andalusite; Chl, chlorite; Cord, cordierite; Di, diopside; Fo, forsterite; Gt, garnet; Ky, kyanite; Sill, sillimanite; Stt, staurolite; Tr, tremolite; Wo, wollastonite.

Satellite plutons east of the main batholith yield Late Cretaceous ages (L.W. Snee, U.S. Geological Survey, unpublished data, 1992; Park, in press).

A broad metamorphic aureole several kilometers wide has developed next to the eastern edge of the batholith, where it appears to be confined to the rocks of the Dollarhide Formation. The exact extent of the aureole is not known at most locations because metamorphic grade cannot always be determined in the field and thin section sampling is sparse. The presence of sillimanite in Dollarhide rocks 2 km from the nearest exposed igneous contact in the Buttercup Mountain quadrangle may suggest, however, that the aureole has affected a large area or that the batholith is extensive in the shallow subsurface.

EOCENE VOLCANIC ROCKS

Tertiary intrusive rocks intruded between 51 and 44 Ma thought to be subvolcanic equivalents of the Challis Volcanic Group (Bennett and Knowles, 1985; Lewis and Kiilsgaard, 1991; Stewart and others, in press) are widespread in the Smoky Mountains.

Eocene (51–40 Ma) volcanic rocks of the Challis Volcanic Group (Johnson and others, 1988; Moye and others, in press) overlie many of the older rocks in central Idaho. Tertiary hypabyssal plutons intrude rocks of the Cretaceous batholith and Tertiary extrusive rocks. Extensive hydrothermal systems having circulation halos 25–50 km wide and 5–7 km deep formed adjacent to some of these plutons (Criss and Taylor, 1983). Tertiary hydrothermal halos were not observed in rocks of the Dollarhide Formation in the study area (Whitman, 1990; Darling and others, this volume).

METAMORPHIC PETROLOGY

METAMORPHIC CONDITIONS

The mixed siliceous and calcareous protolith bulk compositions of the Dollarhide Formation allowed for the development and coexistence of a variety of metamorphic minerals including diopside, tremolite, andalusite, sillimanite, staurolite, and wollastonite. By plotting the stability fields of the coexisting phases on a pressure-temperature diagram with respect to fluid composition, the temperature, pressure, and fluid composition at the time of M_2 metamorphism has been tightly constrained.

Pressures and temperatures of M_2 metamorphism (shaded area, fig. 2) were deduced from the stability fields of the many coexisting minerals using mineral stability curves derived from experimental data. The curves shown in figure 2 are univariant mineral reaction curves shown in pressure-temperature space. For any curve, phases to the left of the curve react during prograde metamorphism to form those minerals indicated to the right. During prograde metamorphism, the unstable phases are "out," and the stable phases are "in." These reactions also occur in reverse during retrograde metamorphism, assuming a closed system.

The Al_2SiO_5 curves used in this study were proposed by Holdaway (1971). These curves best predict mineral assemblages observed in the field and are further verified by independent determinations of pressure and temperature of phases other than the aluminosilicates. For these reasons the Holdaway curves are the most widely used and are most appropriate for this study.

The staurolite-in curve used in this study is from Richardson (1968). This curve assumes pure iron-end member staurolite. If the staurolite has magnesium substituting for iron, the effect is to shift the curve slightly to the left, to lower temperatures; however, the composition of staurolite varies only within narrow limits and there is no significant variation in the replacement of Fe^{+2} by magnesium (Deer and others, 1966, p. 49). The abundance of iron-titanium oxide minerals in rocks in the study area indicates excess iron at the time of metamorphism. In addition, no predominantly magnesium bearing assemblages such as brucite, periclase, or forsterite are present even though temperature and pressure conditions were such that these minerals could have formed. It must follow that magnesium was not a major constituent of the original bulk composition of these rocks. Geochemical data (O'Brien, 1991) indicate equal amounts of Fe_2O_3 and MgO for rocks of the lower part of the Dollarhide Formation collected in the Willow Creek area. Any available magnesium was probably consumed by the formation of tremolite and diopside, with no excess left to form other magnesium-bearing minerals. Staurolite in the study area is most likely close to the iron-end member, and the staurolite-in curve is therefore fixed as shown in figure 2.

The wollastonite reaction curves are from Winkler (1979, p. 130). The wollastonite-in curves are sensitive to fluid composition as shown in figure 2.

The tremolite-in, diopside-in, and forsterite-in curves are from Turner (1980, p. 167–68). These curves assume P_{CO_2} between 0.4 and 1.0. If P_{CO_2} values are less than 0.25, the tremolite-in, diopside-in, and forsterite-in curves shift to slightly lower temperatures. The exact amount of the shift is unknown but is less than 200°C (Turner, 1980). Metamorphic conditions in the study area probably are constrained by the kyanite-andalusite reaction curve and the andalusite-sillimanite reaction curve as suggested by the absence of kyanite and the apparent equilibrium of sillimanite and andalusite. The tremolite-in, diopside-in, and forsterite-in curves cannot have shifted left of the kyanite-out/andalusite-in reaction line or else kyanite would be the stable aluminosilicate. The absence of kyanite restricts the metamorphic conditions to the right of the kyanite-out/andalusite-in reaction curve. Any leftward shift of the tremolite-in, diopside-in, and forsterite-in curves therefore can not exceed approximately 25°C, a value that does not significantly alter the pressure-temperature values reported here. The tremolite-in, diopside-in, and forsterite-in curves are indicated in figure 2.

Rocks in the study area commonly exhibit the assemblage diopside, tremolite, and either andalusite or sillimanite. A sampling traverse along the ridge near the Buttercup mine yielded an M_2 andalusite-bearing rock approximately 210 m from the intrusive contact and a M_2 sillimanite-bearing rock a few meters farther from the batholith. Therefore it is thought that metamorphic conditions lie very near the andalusite-out/sillimanite-in univariant equilibrium curve. The absence of kyanite indicates that the staurolite-in, tremolite-in, and diopside-in curves all are to the right of the kyanite-andalusite reaction curve. Therefore, pressure-temperature conditions are controlled by the kyanite-andalusite and the sillimanite-andalusite univariant reaction curves.

Rocks in this area also contain tremolite, diopside, and staurolite. The tremolite-in curve intersects the andalusite-out/sillimanite-in curve at 3.25 kb and 525°C. The staurolite-in curve also intersects the tremolite-in curve at approximately 3.25 kb and 525°C. If the tremolite-in curve shifts left (if P_{CO_2}<0.25), then the upper pressure value would be approximately 3.5 kb. Any greater increase in pressure would be too far from the sillimanite-in/andalusite-out curve and andalusite would not be observed in the rocks. At 3.5 kb, the temperature must remain above 525°C or chlorite+kyanite would be the stable assemblage rather than staurolite. Therefore, 525°C is the minimum temperature of M_2 metamorphism. The maximum temperature at which staurolite can exist near the andalusite-sillimanite equilibrium is approximately 600°C. At 600°C, staurolite-andalusite-sillimanite can only coexist in a narrow pressure range near 2.5 kb.

The assemblages diopside-tremolite-andalusite/sillimanite-staurolite define the pressure and temperature of M_2 metamorphism to be approximately 2.5–3.5 kb and 525°C–600°C. Metamorphic conditions were similar throughout the study area as shown by table 1.

These temperatures and pressures correspond to conditions of medium-high grade dynamothermal metamorphism, and, if fluid pressure is equal to total pressure, then they indicate a depth of metamorphism and therefore a depth of emplacement of 8.5–12.5 km for the Idaho batholith in the southern Smoky Mountains.

Composition of the fluid phase with respect to CO_2 during M_2 metamorphism can be determined by examining the stability of wollastonite with respect to diopside and to the conspicuous lack of forsterite. The assemblage tremolite, diopside, sillimanite, and wollastonite without forsterite has been observed in the Willow Creek, West Fork of Willow Creek, and Tyrannis Creek areas. As the pressure-temperature diagram (fig. 2) indicates, wollastonite is quite sensitive to fluid composition and the temperature at which it forms is lowered considerably with a lowering of P_{CO_2}. If $0.25<P_{CO_2}<0.40$, then forsterite is the stable phase at the expense of tremolite at 2.5–3.5 kb and 525°C–600°C. Forsterite was not observed in any of the rocks in the study area. Therefore, in order for the observed assemblage diopside-tremolite-wollastonite-sillimanite to have formed, P_{CO_2} must have been less than 0.25. If the lower metamorphic pressure limit is 2.5 kb, as indicated by the coexistence of staurolite-andalusite±sillimanite, and cordierite is absent, then wollastonite is stable at $P_{CO_2}<0.18$ (fig. 2). The wollastonite-in curve at $P_{CO_2}=0.18$ also corresponds to a temperature maximum of 600°C.

During metamorphism of marble, the CO_2 component of the fluid phase usually is from 0.4 to 1.0 P_{CO_2} (Turner, 1980), and the unusually low CO_2 content of the fluid phase of the Dollarhide marbles may be related to dilution by high amounts of water coming from the cooling wet granodiorite or to removal of CO_2 from the system by an unidentified process or processes.

Table 1. Temperature and pressure of M_2 metamorphism in the Dollarhide Formation, south–central Idaho.
[Data from Whitman, 1990]

	Temperature (°C)	Pressure (kilobars)
Buttercup mine	525–600	2.5–3.5
Willow Creek	545–630	2.25–4.5
West Fork of Willow Creek	500–600	2.5–3.5
Tyrannis Creek	500–600	2.5–3.5

DESCRIPTION OF METAMORPHIC FEATURES

Numerous metamorphic features have been observed in rocks in the study area including two sets of porphyroblasts, two foliations, mineral, boudin, and intersection lineations, hornfelsic micaceous spots, and pseudomorphs (fig. 3). The types of features and their crosscutting relationships define the chronology of two metamorphic and deformational episodes.

The lower part of the Dollarhide Formation contains two sets of crosscutting porphyroblasts, two sets of crosscutting foliations, hornfelsic micaceous spots, and pseudomorphs. The nature of these features and their relationships to each other indicate that two prograde metamorphic events have occurred (M_1 and M_2) and that M_1 was followed by a period of retrograde thermal alteration (R_1). Each prograde event was accompanied by strain that formed a foliation within the rocks (S_1 and S_2). Only one event (M_2) was accompanied by folding. Field relationships and a whole-rock K-Ar age of 83.9±3.4 Ma constrain both events to the Late Cretaceous.

Large, well-preserved porphyroblasts of tremolite, diopside, sillimanite, andalusite, staurolite, wollastonite, biotite, muscovite, potassium feldspar, plagioclase, and epidote are present throughout the lower part of the Dollarhide Formation in the study area. The original bulk composition of the protolith controls the metamorphic assemblages: in pelitic beds andalusite or sillimanite formed, in calcareous beds tremolite, diopside, and wollastonite formed, and in quartz-rich beds biotite and recrystallized quartz formed (fig. 3).

The porphyroblasts vary in size from about 0.25 to 1–3 mm and are the same size in any one thin section, except for some samples that have two distinctly different sizes of porphyroblasts. Porphyroblasts are typically euhedral to subhedral and are surrounded by quartz-biotite or quartz-calcite matrices. Sillimanite, tremolite, plagioclase, and biotite were observed in mutual grain-boundary contact, as were diopside, tremolite, and epidote. Wollastonite and scapolite were identified in rocks of the West Fork of Willow Creek and Tyrannis Creek areas. Idocrase was identified in samples from the Tyrannis Creek area near the Stormy Galore mine.

Rocks in the study area exhibit three foliations. Primary bedding (S_0) is defined by compositional and grain-size variations. A poorly developed foliation (S_1) is defined by the parallel alignment of fine-grained biotite and muscovite in the pelitic rocks of units 1 and 2. A well-developed penetrative foliation (S_2) is defined by the parallel alignment of coarse-grained biotite and muscovite in the pelitic rocks and by the parallel alignment of calcite and quartz in the calcareous rocks. The S_2 foliation crosscuts the finer grained S_1 foliation in rocks from the Buttercup mine area. The S_2 foliation is present throughout the lower part of the Dollarhide Formation in the study area, although it is not clearly visible in the calcareous rocks in outcrop. Two mineral lineations were observed in some rocks. Both L_1 and L_2 are defined by elongated biotite grains in the pelitic rocks and by elongated calcite and quartz grains in the calcareous rocks. The lineations are contained within their respective foliations, S_1 and S_2.

Protolith	M_1	R_1	M_2
Pelite	Porphyroblasts Micaceous spots S_1	Pseudomorphs Retrograded porphyroblasts	Quartz-biotite-muscovite-potassium feldspar-andalusite-sillimanite-calcite-diopside-tremolite-staurolite S_2, L_{2x0}, L_{0x2}, L_m, L_b
Limestone	Porphyroblasts Micaceous spots S_1	Pseudomorphs Retrograded porphyroblasts	Calcite-quartz-tremolite-diopside-wollastonite-scapolite-idocrase-plagioclase-epidote S_2, L_{2x0}, L_{0x2}, L_m, L_b

Figure 3. Mineral assemblages and structural elements formed in protoliths of pelite and limestone of the study area during first prograde metamorphism (M_1), first retrograde metamorphism (R_2), and second prograde metamorphism (M_2). Abbreviations: S_1, first foliation; S_2, second foliation; L_{0x2}, intersection lineation of S_2 and bedding measured on bedding surface; L_{2x0}, intersection lineation of bedding and S_2 measured on the S_2 surface; L_m, mineral lineation; L_b, boudin lineation defined by aggregates of mineral grains.

Most rocks in the study area exhibit a spotted or hornfelsic texture in hand sample and in thin section. In hand sample, the spots are white to pink, millimeter- to centimeter-scale pods or bands and are present in phyllitic quartzite and black marble and argillite. Microscopic examination reveals that the spots have an ellipsoidal shape and vary in size from about 0.45 mm by 0.2 mm to 0.75 mm by 0.4 mm. The spots are composed of muscovite, biotite, chlorite, and minor quartz. Although the micas are fine grained in some hornfelsic spots, most have a coarser texture than the surrounding matrix. Most of the hornfelsic spots are oriented with their long axis parallel with the strong biotite foliation (S_2) of the rock; however, some spots have their long axis oblique to the foliation by as much as 20°. In the Buttercup mine area, approximately 200–565 m from the intrusive contact, hornfelsic spots contain a finer grained biotite foliation (S_1) that is subparallel with the coarser grained matrix foliation. Most of the spots in this rock are overprinted by micas of the matrix foliation (S_2).

Throughout the study area rocks of the lower part of the Dollarhide Formation contain pseudomorphs. Tetrahedral pseudomorphs are larger than the ellipsoidal hornfelsic spots and average approximately 1.0 mm long on a side. They are composed of fine-grained polycrystalline quartz and minor potassium feldspar and micas. Because their size and shape is identical to andalusite porphyroblasts observed elsewhere in rocks in the study area, they are interpreted as pseudomorphs after andalusite. Other pseudomorphs in the study area resemble staurolite porphyroblasts and are interpreted to be pseudomorphs after staurolite.

A chronology of the timing of metamorphism and deformation can be determined by examining the nature of, and relationships between, the various petrographic elements observed in the rocks.

Two sets of porphyroblasts were identified in samples from the Tyrannis Creek, Willow Creek, and Buttercup mine areas. The two sets of diopside, tremolite, and andalusite porphyroblasts have two distinctly different sizes and two different degrees of alteration. A sample from the Tyrannis Creek domain near the Stormy Galore mine (fig. 1) exhibits very large (3.2 mm by 1.4 mm), highly retrograded diopside porphyroblasts overgrown by smaller (1.6 mm by 0.52 mm), much fresher diopside porphyroblasts. Numerous samples from the Buttercup mine and Willow Creek areas contain two sets of diopside and andalusite porphyroblasts; one set is thermally altered and shows signs of retrogression, the other set is unaltered and fresh looking. This suggests that two metamorphic events (M_1 and M_2) of similar temperature and pressure have affected these rocks. The events were separated by a period of retrograde metamorphism (R_1) that caused the alteration of the first set of porphyroblasts. The presence of pseudomorphs of quartz after andalusite and staurolite in samples containing fresh andalusite, tremolite, and diopside porphyroblasts also indicates two metamorphic events: andalusite and staurolite growth during M_1 and subsequent replacement of the andalusite and staurolite porphyroblasts with quartz and growth of more andalusite and other minerals during M_2.

Two foliations (S_1 and S_2) were observed in the rocks. The S_1 foliation is finer grained and oriented at an angle of approximately 20° to the second foliation (S_2). This relationship suggests that two deformational events have occurred and that the orientation of the rocks with respect to the principal strain axes accompanying the second event was slightly different from that during the first event.

The ages of the foliation can be tied to the ages of metamorphism through observations of the relationship between the foliations, the M_1 and M_2 porphyroblasts, the M_1 pseudomorphs, and the M_1 and M_2 hornfelsic mica spots. In most rocks in the study area, the strong S_2 foliation wraps around porphyroblasts, and no porphyroblasts were observed cutting this foliation. This relationship indicates that the porphyroblasts grew before or during deformation and foliation development. Clearly, if deformation did not precede contact metamorphism, evidence of porphyroblasts overgrowing and crosscutting the foliation would be observed. Two generations of hornfelsic spots were observed: fine-grained biotite-muscovite spots and coarse-grained spots. In samples from the Buttercup mine area, fine-grained spots are aligned with their long axes 20° to the well-developed S_2 foliation. The S_2 foliation cuts some of the fine-grained spots.

A sample of micaceous calc-silicate from the head of Cabin Gulch (fig. 1) was dated using the whole-rock K-Ar method. An age of 83.9±3.4 Ma was obtained (R.L. Armstrong, written commun., 1990). Based on the metamorphic textures and compositions described above, this age is interpreted as the minimum age of M_2 metamorphism. Therefore, the foliation observed in rocks of the lower part of the Dollarhide Formation on the margin of the Idaho batholith is no older than Late Cretaceous. This Late Cretaceous age is substantiated by other field evidence for Late Cretaceous metamorphism in the study area and by Late Cretaceous ages for satellite plutons east of the main batholith (L.W. Snee, U.S. Geological Survey, unpublished data, 1992; Park, in press).

STRUCTURAL GEOLOGY OF THE DOLLARHIDE FORMATION ADJACENT TO THE IDAHO BATHOLITH

Rocks along the eastern edge of the Idaho batholith in the Smoky Mountains have undergone intense deformation as evidenced by plentiful map- and microscopic-scale geologic structures including northeast- and northwest-striking faults, S_1 and S_2 foliations, open to isoclinal folds, and intersection, mineral, and boudin lineations (L_{2x0}, L_m, and L_b, respectively).

Regionally extensive northeast- and northwest-striking faults (Bennett, 1986; Kiilsgaard and others, 1986) having no more than several kilometers of offset are present in the study area (Darling, 1987; Worl and others, 1991; Darling and others, this volume; Lewis, in press) (fig. 1). Many outcrop-scale faults having less than 10 m displacement were observed in the field but were not mapped.

The regionally extensive metamorphic foliation (S_2) cuts an older, poorly developed foliation (S_1). The S_2 foliation is present in all members of the Dollarhide Formation in the Buttercup Mountain quadrangle and has not been subsequently folded, except at one location on the ridge immediately west of the Buttercup mine ridge where the foliation has been folded with a small, late-stage leucocratic dike; S_2 is axial planar to the folds.

Folds are present throughout the study area and range in wavelength from several kilometers, such as those in the Tyrannis Creek area, to microscopic, such as in samples collected from the West Fork of Willow Creek. The folds are gently to moderately plunging and moderately inclined to recumbent, and they exhibit open to isoclinal apical angles and rounded to kinked hinges. They are present in phyllitic quartzite as well as siliceous and calcareous marble. All fold hinges trend east-northeast (fig. 4). Differences in fold morphology are likely due to rheologic differences between rock types and to differences in distance from the batholith margin rather than being indications of two folding events. Apparent parallel laminations in hand samples of marble are actually highly attenuated limbs of recumbent isoclinal folds.

Mineral lineations, intersection lineations, and boudins are also present in the study area. Mineral lineations (L_m) are defined by 2–3 millimeter-long black euhedral mica pseudomorphs after amphibole in phyllitic quartzite. Intersection lineations (L_{2x0}, L_{0x2}) were formed by the intersection of S_2 cleavage and bedding (S_0), measured on the S_2 and S_0 surfaces, respectively. Boudins of quartz or feldspar crystals are present but are not as widespread as the other structural features.

Lower hemisphere equal-area projections of structural data from the vicinity of Buttercup mine (fig. 4*A*), West Fork of Willow Creek (fig. 4*B*) and Blackhorse Canyon (fig. 4*C*) were constructed. Composite data from the Blackhorse Canyon, West Fork of Willow Creek, Willow Creek, and Buttercup mine areas are shown in figure 4*D*.

The clustering of poles to bedding (S_0) in the southwest quadrant indicates a regional northeast dip of bedding. Although numerous folds are present throughout the area, the poles to bedding do not define a great-circle distribution because of sampling bias; extreme attenuation of most folds in the study area precluded measurement of bedding in the axial regions of folds.

Poles to S_2 also cluster in the southwest quadrant around an average S_2 orientation of N. 20° W., 40° E. In most places S_0 and S_2 are almost parallel. Fold hinges, as well as intersection, boudin, and mineral lineations, have wide scatter but in general cluster in the northeast quadrant around an average orientation of 45°, N. 70° E., for the fold axes. The lineations also cluster in the northeast quadrant. The 45°, N. 70° E., orientation was derived from the best-fit great-circle distribution of poles to S_0 and S_2 obtained from data in the vicinity of Blackhorse Canyon. If folding is cylindrical, or almost so, the fold axis lies 90° from the great circle containing the poles to S_2. The 45°, N. 70° E., orientation of the fold axis lies 90° from the best-fit great circle containing the poles to S_2 (figs. 4*B*, 4*C*). This derived value (45°, N. 70° E.) corresponds well with actual field measurements of fold hinges (figs. 4*B*–*D*) throughout the study area.

All folds in the study area trend east-northeast and plunge moderately away from the batholith. Mapping east of the batholith indicates that regionally folds trend south-southeast and plunge gently (Rodgers and others, this volume). It is herein proposed that anomalous orientations of the fold hinges in this area are due to high shear strain. In areas of high shear stain, the *X*- and *Z*- axes of finite strain rotate about the *Y* axis of finite strain. The *XY*-plane is parallel with the axial plane of the folds, and it rotates from a normal orientation (45°) with respect to the shear zone boundary (batholith margin) in low shear strain environments to a parallel position with respect to the shear zone boundary in areas of high shear strain (figs. 5*A*–*D*). Features such as fold hinges and mineral and boudin lineations generally form parallel with the direction of the intermediate strain axis (*Y*). In areas of high shear strain, where the magnitude of *X* is much greater than that of *Y*, these features can be passively rotated to a position that is parallel with the original direction of least extension (*Z*). This process, originally explained by Hansen (1971), could account for the anomalous orientation of fold axes and mineral lineations in the study area. To confirm this hypothesis, more fold orientation data are needed from the area between the Buttercup Mountain and Griffin Butte quadrangles, where the transition from south-southeast-trending folds to east-northeast-trending folds should be present.

In the northwestern part of the study area, not all map-scale folds (fig. 1) trend east-northeast; this variation in trend may represent broad warping of the foliation. Folding of a late-stage leucocratic dike may also indicate continued deformation after peak metamorphism. Evidence of a second major folding event was not observed in thin section or in the structural analysis. More structural data are needed in order to define the extent and character of structures in this part of the study area.

TIMING OF METAMORPHISM AND DEFORMATION

Petrographic and field evidence suggest that two episodes of metamorphism and deformation occurred during Late Cretaceous time in the study area. The early S_1 foliation is poorly developed and not generally visible in outcrop. M_1 features such as porphyroblasts, hornfelsic texture, and pseudomorphs have been preserved. These features are crosscut by a stronger penetrative foliation (S_2) that wraps around M_2 porphyroblasts. Evidence of postdeformational porphyroblast growth was not observed in any sample. Deformation must have occurred simultaneously with M_2 metamorphism. The minimum age of M_2 metamorphism has been dated by whole-rock K-Ar to be 83.9±3.4 Ma (R.L. Armstrong, written commun., 1990), the time at which the temperature cooled from the peak metamorphic temperature of 525°C–600°C to the blocking temperature of the analyzed micas at approximately 300°C. The age of peak metamorphism is unknown.

Figure 3 summarizes the structural and metamorphic elements produced during each phase of metamorphism and deformation. Field observations, thin sections, and structural analysis (fig. 4) show that fold hinges lie within the plane of the S_2 foliation such that the S_2 foliation is axial planar to the folds. Because of the axial planar nature of the S_2 foliation, folding was synchronous with development of S_2 foliation. Because the S_2 foliation formed during M_2 metamorphism, which was coeval with emplacement of the batholith, folds in the study area are Late Cretaceous in age. There is no evidence of an earlier

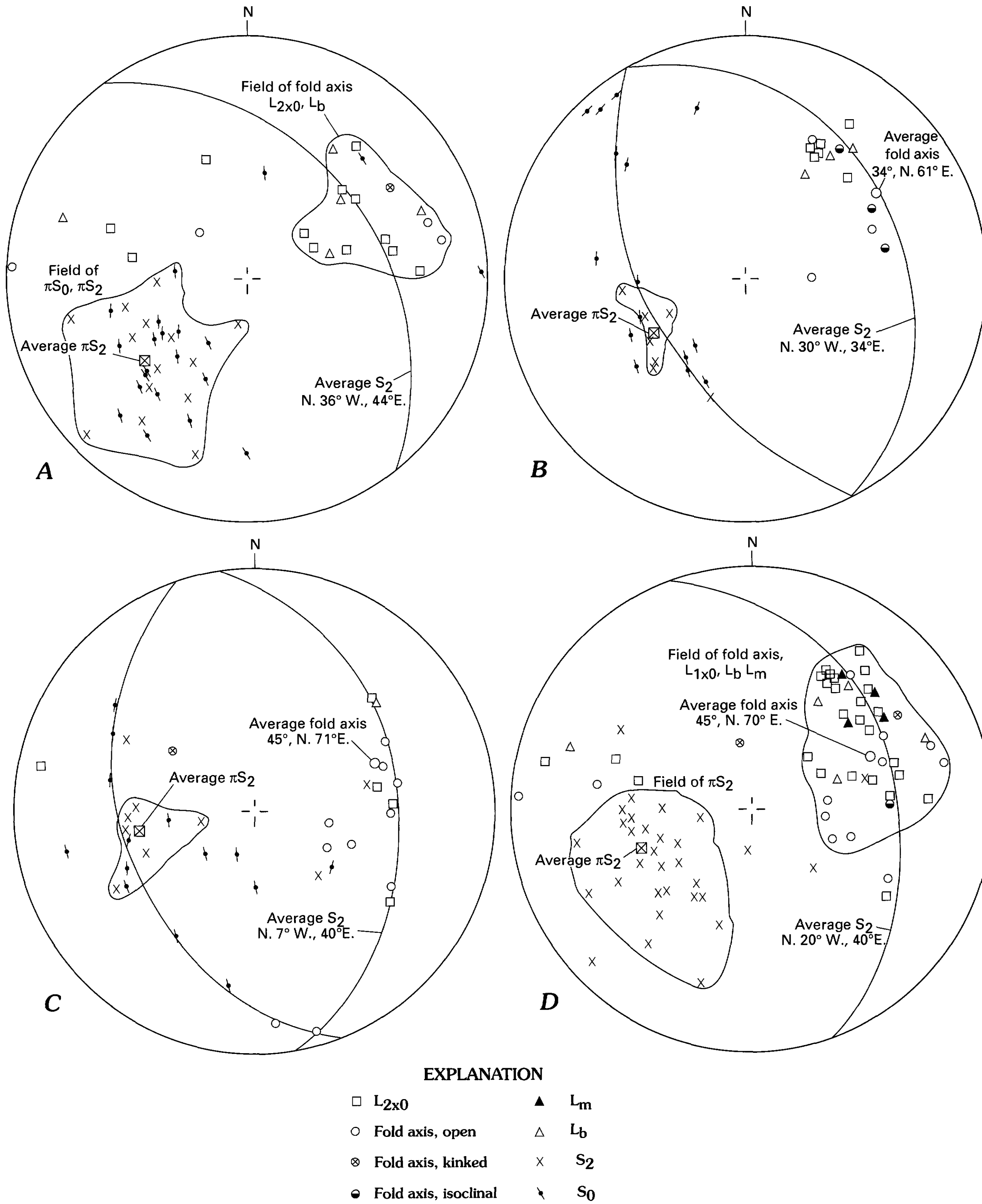


Figure 4. Equal-area Pi diagrams for (*A*) Buttercup mine, (*B*) West Fork of Willow Creek, (*C*) Blackhorse Canyon, and (*D*) composite of Blackhorse Canyon, West Fork of Willow Creek, Willow Creek, and Buttercup mine domains. The Pi diagrams contain plots of poles to S_0 (PiS_0), poles to S_2 (PiS_2), intersection lineation measured on the S_2 surface (L_{2x0}), mineral lineations (L_m), boudin axes (L_b), and fold axes. Data from Whitman (1990).

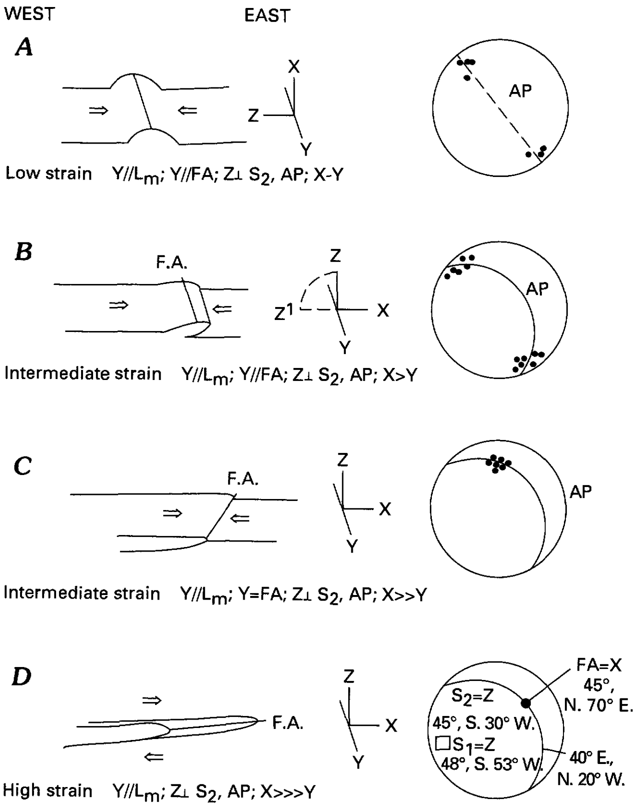


Figure 5. Diagram showing progressive increase in strain for the Dollarhide Formation during M_1 and M_2 metamorphism. Original east-west compression causes rotation of fold axes to 45°, N. 70° E. orientation; mineral lineations and fold axes become parallel. X is direction of most extension; Z is direction of least extension; Y is direction of intermediate extension. L_m is mineral lineation; FA is fold axis; AP is axial plane; S_2 is foliation formed during M_2. Modified from Whitman (1990, fig. 21). *A*, Possible configuration after M_1 and early in M_2. *B*, Possible configuration during intermediate stages of M_2. *C*, Possible configuration during late stages of M_2; the present configuration at distances greater than 2 km from the batholith. *D*, Present configuration near the batholith.

(pre-Cretaceous) deformational event. On the ridge immediately west of the Buttercup mine ridge, a small leucocratic dike has been folded with the S_2 foliation into a small S-fold. The dike is certainly related to late-stage differentiation of the Idaho batholith, and therefore folding occurred on a local scale after peak metamorphism.

The S_1 foliation formed simultaneously with M_1 metamorphism. The S_1 foliation is poorly developed, and structural data were not collected for the M_1 event. Pressure and temperature of metamorphism were such that diopside and andalusite porphyroblasts, as well as hornfelsic texture, developed in the rocks. Although the S_1 foliation and M_1 petrographic features are preserved, there is no evidence of folding related to the M_1 event. The M_1 event probably represents the initial stages of batholith emplacement, whereas the M_2 event represents peak metamorphism and high strain in the margin area.

REGIONAL IMPLICATIONS

MESOZOIC DEFORMATION

The presence of an Antler-age fabric in phyllitic quartzite near Carrietown (fig. 1) was suggested by Geslin (1986) and Darling (1987); however, the present structural and petrographic study reveals no evidence for deformation of the Dollarhide Formation prior to intrusion of the Idaho batholith in the Late Cretaceous. Rotation of fold hinges in regions of high shear strain near the batholith throughout the study area could account for the parallelism between fold axes and mineral and boudin lineations. High shear strain and rotation also account for the disparity between fold-hinge orientations in the Dollarhide Mountain and Buttercup Mountain quadrangles and fold-hinge orientations in surrounding areas. A single generation of folds is documented in the study area. The coeval nature of foliation and fold development with M_2 porphyroblast development prior to the minimum age of metamorphism at 83.9±3.4 Ma indicates that large- and small-scale folds within the study are not much older than Late Cretaceous.

RELATION TO MINERALIZATION

Metasomatic mineral assemblages (idocrase-scapolite-calcite) were observed in the western part of the study area in the vicinity of Tyrannis Creek and the West Fork of Willow Creek (fig. 3). Tungsten minerals are present in phyllite in the West Fork of Willow Creek. Idocrase and scapolite are absent from calcareous rocks in the eastern part of the study area. This pattern may indicate that skarn processes were better developed in the western part of the study area.

Cretaceous mineralization in the study area and in nearby areas is well documented (Darling, 1987; Darling and others, this volume; Park, in press). Silver-lead-zinc mineral deposits in the study area are in northeast-trending shear zones along the lithologic change from quartzite and phyllite to carbonaceous marble and granodiorite and country rocks (Darling and others, this volume). The contact between phyllite of unit 2 and marble of unit 3 previously had been thought to be the thrust fault contact between the Carrietown sequence and the Dollarhide Formation, but the mineral deposits are not localized along one throughgoing structure. Almost half of the mines in the study area are from more than 100 m below to several hundreds of meters above this contact (Darling, 1987; Darling and others, this volume). Rather than being structurally controlled by a thrust fault, the mineralized shear zones most likely formed during Cretaceous compressional deformation as a result of differential stresses along the lithologic and rheologic contact between competent quartzite and less competent marble.

SUMMARY

Rocks of the Pennsylvanian to Permian Dollarhide Formation of south-central Idaho underwent moderate- to high-grade dynamothermal metamorphism during the Late Cretaceous, reaching temperatures of 525°C–600°C and pressures of 2.5–3.5 kb. Metamorphic fluids were dominated by H_2O (P_{CO_2}<0.18) despite the calcareous rock types involved. Depth of emplacement of the southeastern edge of the Atlanta lobe of the Idaho batholith is estimated at 8.5–12.5 km. Although two pulses of heat and strain were documented in the study area, only one generation of folds was observed. Foliation axial planar to the folds indicates that folding was contemporaneous with development of foliation. Fold hinges and mineral, boudin, and intersection lineations are anomalously oriented 45°, N. 70° E., in the study area in contrast to south-southeast-trending folds in areas further from the batholith. It is herein proposed that high shear strain on the margin of the batholith rotated fold hinges and mineral, boudin, and intersection lineations into their present anomalous orientations. Silver-lead-zinc mineralized rock is concentrated in Cretaceous-age northeast-trending shear zones and is related to Cretaceous intrusive activity. Cretaceous mineralization, deformation, and metamorphism were broadly contemporaneous in this area. M_2 metamorphism has been dated by the K-Ar method at 83.9±3.4 Ma, which represents the minimum age of metamorphism. A pre-Cretaceous deformational event is not documented along the eastern edge of the Idaho batholith.

REFERENCES CITED

Bennett, E.H., 1986, Relationship of the trans-Challis fault system in central Idaho to Eocene and Basin and Range extensions: Geology, v. 14, p. 481–484.

Bennett, E.H., and Knowles, C.R., 1985, Tertiary plutons and related rocks in central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–F, p. 81–96.

Criss, R.E., and Taylor, H.P., 1983, An $^{18}O/^{16}O$ and D/H study of Tertiary hydrothermal systems in the southern half of the Idaho batholith: Geological Society of America Bulletin, v. 94, p. 640–663.

Darling, R.S., 1987, The geology and ore deposits of the Carrietown silver-lead-zinc district, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 168 p.

Deer, W.A., Howie, R.A., and Zussman, J., 1966, An introduction to the rock forming minerals: Essex, England, Longman Group Limited, 528 p.

Geslin, J.K., 1986, The Permian Dollarhide Formation and Paleozoic Carrietown sequence in the SW¼ of the Buttercup Mountain quadrangle, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 116 p.

Hall, W.E., 1985, Stratigraphy of and mineral deposits in middle and upper Paleozoic rocks of the black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis 1°×2° quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–132.

Hansen, Edward, 1971, Strain facies: New York, Springer-Verlag, 207 p.

Holdaway, M.J., 1971, Stability of andalusite and the aluminum silicate phase diagram: American Journal of Science, v. 271, p. 97–131.

Johnson, K.M., Lewis, R.S., Bennett, E.H., and Kiilsgaard, T.H., 1988, Cretaceous and Tertiary intrusive rocks of south-central Idaho, *in* Link. P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 55–86.

Kiilsgaard, T.H., Fisher, F.S., and Bennett, E.H., 1986, The trans-Challis fault system and associated precious metal deposits, Idaho: Economic Geology, v. 81, p. 721–724.

Lewis, R.S., in press, Alteration and mineralization in the eastern part of the Soldier Mountains, Camas County, Idaho: U.S. Geological Survey Bulletin 2064–V.

Lewis, R.S., and Kiilsgaard, T.H., 1991, Eocene plutonic rocks in south central Idaho: Journal of Geophysical Research, v. 96, no. B8, p. 13295–13311.

Link, P.K., Skipp, Betty, Hait, M.H., Janecke, Susanne, and Burton, B.R., 1988, Structural and stratigraphic transect of south-central Idaho—A field guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 5–42.

Mahoney, J.B., Link, P.K., Burton, B.R., Geslin, J.K., and O'Brien, J.P., 1991, Pennsylvanian and Permian Sun Valley Group, Wood River Basin, south-central Idaho, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States–II: Pacific Section, Society of Economic Paleontologists and Mineralogists, Publication 67, v. 2, p. 551–579.

Moye, F.J., Hackett, W.R., Snider, L.G., and Mertzman, S.A., in press, Stratigraphy, chronology, and petrology of the southeastern Challis volcanic field, central Idaho: U.S. Geological Survey Bulletin 2064–EE.

O'Brien, J.P., 1991, Stratigraphy of the Pennsylvanian-Permian Dollarhide Formationin the Buttercup Mountain quadrangle, Blaine and Camas Counties, Idaho: Pocatello, Idaho State University, M.S. thesis, 163 p.

Park, W.A., in press, The geology and ore deposits of the Rooks Creek stock, Blaine County, Idaho: U.S. Geological Survey Bulletin 2064–CC.

Richardson, S.W., 1968, Staurolite stability in a part of the system Fe-Al-Si-O-H: Journal of Petrology, v. 9, pt. 3, p. 467–488.

Ross, C.P., 1930, The Little Smoky and Willow Creek mining districts, Camas County: Idaho Bureau of Mines Pamphlet 33, p. 19–26.

Stewart, D.E., Schmidt, K.L., Moye, F.J., Mahoney, J.B., and Link, P.K., in press, Eocene intrusive complexes in the Smoky and Boulder Mountains, south-central Idaho: U.S. Geological Survey Bulletin 2064–HH.

Turner, Francis, 1980, Metamorphic petrology—Mineralogical, field and tectonic aspects (2nd ed.): New York, Hemisphere, 524 p.

Umpleby, J.B., 1915, Ore deposits in the Sawtooth quadrangle, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 580–K, p. 221–249.

Wavra, C.S., 1985, Structural geology and petrology of the SW¼ of the Mahoney Butte quadrangle, Blaine County, Idaho: Moscow, University of Idaho, M.S. thesis, 103 p.

Wavra, C.S., Isaacson, P.E., and Hall, W.E., 1986, Studies of the Idaho black shale belt—Stratigraphy, depositional environment, and economic geology of the Permian Dollarhide Formation: Geological Society of America Bulletin, v. 97, p. 1504–1511.

Whitman, S.K., 1990, Metamorphic petrology and structural geology of the Pennsylvanian-Permian Dollarhide Formation, Blaine and Camas Counties, south-central Idaho: Pocatello, Idaho State University, M.S. thesis, 108 p.

Winkler, H.G.F., 1979, Petrogenesis of metamorphic rocks (5th ed.): New York, Springer-Verlag, 348 p.

Worl, R.G., Kiilsgaard, T.H., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Regional Geophysical Studies Applied to Mineral Resource Assessment of the Hailey 1°×2°Quadrangle and the Western Part of the Idaho Falls 1°×2° Quadrangle, South-Central Idaho

By M. Dean Kleinkopf, Anne E. McCafferty, *and* Gerda A. Abrams

GEOLOGY AND MINERAL RESOURCES OF THE HAILEY AND
IDAHO FALLS QUADRANGLES

U.S. GEOLOGICAL SURVEY BULLETIN 2064–R

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1995

CONTENTS

PLATE

[Plate is in pocket]

FIGURES

Regional Geophysical Studies Applied to Mineral Resource Assessment of the Hailey 1°×2° Quadrangle and the Western Part of the Idaho Falls 1°×2° Quadrangle, South-Central Idaho

By M. Dean Kleinkopf,[1] Anne E. McCafferty,[1] *and* Gerda A. Abrams[1]

ABSTRACT

Interpretations of aeromagnetic and gravity anomaly data for the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle in south-central Idaho were used to delineate buried intrusive complexes, major geologic structures, and areas of possible volcanic subsidence that may relate to hydrothermally altered and mineralized areas.

Three crustal provinces were identified from the aeromagnetic and gravity signatures. The northwest province is composed of Cretaceous granitic terranes of the Idaho batholith that are weakly magnetic and have slowly varying, highly negative Bouguer anomaly values. The northeast province exhibits a variety of weak to strong aeromagnetic anomalies in the Pioneer and White Knob Mountains, which are made up of assorted terranes of thrust- and block-faulted sedimentary rocks, Precambrian metamorphic rocks, Tertiary granitic intrusive rocks, and Eocene and Miocene volcanic rocks. This assemblage has an overall negative mass effect on the gravity field similar to that of the Idaho batholith. The Snake River Plain province consists predominantly of Cenozoic basalt that exhibits complex patterns of high-frequency positive and negative aeromagnetic anomalies that result from high magnetic susceptibility contrasts and reversed magnetizations of the basalt.

The highest aeromagnetic values outside the Snake River Plain province correlate spatially with Tertiary epizonal plutons such as granitic intrusive rocks of the Sawtooth batholith north of Atlanta, the stock at Steel Mountain just north of Rocky Bar, and the Sheep Creek batholith about 24 km west of Rocky Bar. The magnetic anomaly data indicate that each of these granitic masses is a composite intrusive complex of variable composition or state of alteration and is commonly more extensive in the subsurface than in outcrop. Most granitic rocks of the Cretaceous Idaho batholith are biotite granodiorite and two-mica granite. The batholithic terrane exhibits exceptional uniformly low magnetization and low density, except for some diorite and minor border facies of tonalite, both of which are more strongly magnetized and denser than the main mass of the batholith.

The Bouguer gravity anomaly map is dominated by the southern part of a regional negative anomaly of 50–80 mGal in south-central Idaho that shows a strong inverse correlation with regional topography resulting from thickened crust beneath the Idaho batholith. Superimposed on the regional negative gravity anomaly are local positive and negative anomalies that result from mass anomalies in the upper crust. The negative anomalies may be associated with low-density granite stocks or with accumulations of low-density alluvial material in intermontane basins and valleys.

Two buried shear zones that may have influenced the distribution of mineral deposits are interpreted from the aeromagnetic and gravity anomaly patterns. The Atlanta shear zone is defined by a northeast-trending belt of high magnetic gradient that passes through Atlanta. Intrusive rocks of the Sawtooth, Steel Mountain, and Sheep Creek batholiths are along the northwest side of this belt of steep magnetic gradients. The Bouguer gravity anomaly data along the belt show some corroborative evidence for the aeromagnetic interpretation of a buried shear zone. The second proposed shear zone, the Great Rift shear zone, is interpreted from northwesterly trending aeromagnetic and gravity patterns that are projected from similar trends associated with the Great Rift near Craters of the Moon National Monument.

The distribution of mineral deposits commonly correlates with known geologic features or with geologic features postulated from geophysical information. A number of gold deposits are present northeast of Atlanta along the belt of high magnetic gradient. The belt extends to the northeast and to the southwest and may help delineate other areas of undiscovered mineral resources. Four volcanic features defined from geologic mapping in the study area show varying expression in the aeromagnetic and gravity anomaly data.

[1]U.S. Geological Survey, Denver, Colorado 80225.

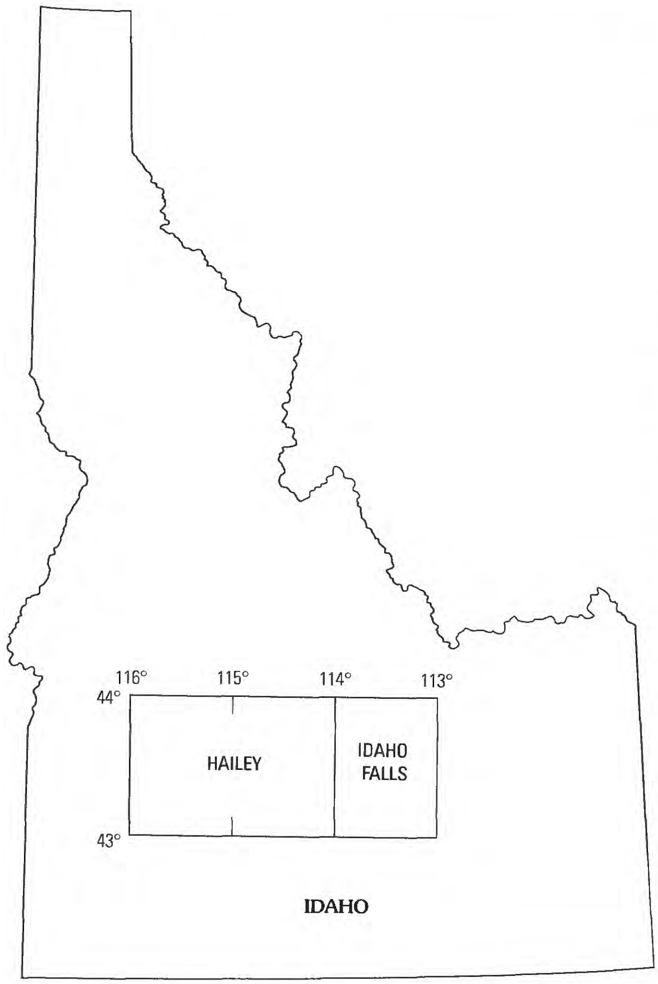


Figure 1. Index map showing the location of the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho.

These are the caldron west of Ketchum, the Magic Reservoir eruptive center east of Fairfield, the Lehman Basin caldron complex west of Mackay, and the Alder Creek eruptive center south of Mackay. The aeromagnetic anomaly data indicate three other possible volcanic features in areas where some mineralized rock is present. Along the postulated Great Rift shear zone, just southeast of the Pioneer Mountains core complex, complex magnetic anomaly patterns suggest a possible volcanic center. Two other possible volcanic features are along the proposed Atlanta shear zone northeast of Atlanta where complex negative magnetic anomaly patterns suggest local fracturing in the granitic rocks and possible volcanic subsidence features.

Interpretation of the aeromagnetic and gravity anomaly data suggests a number of areas for followup studies to identify possible new mineral deposits. These include (1) areas interpreted to be underlain by igneous intrusive rocks, (2) northeast-trending belts of high magnetic gradient reflecting possible faulting parallel with the Trans-Challis fault system, (3) areas of crossfaulting along the proposed Atlanta and Great Rift shear zones, and (4) aeromagnetic and gravity anomalies that correlate with geochemical anomalies.

INTRODUCTION

Analyses of aeromagnetic and gravity anomaly data were made as part of a multidisciplinary effort, the U.S. Geological Survey Conterminous United States Mineral Assessment Program (CUSMAP), to assess the mineral resource potential of the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle in south-central Idaho (fig. 1). The objectives were (1) to compile and display available aeromagnetic and gravity anomaly data for the area and (2) to interpret the geologic significance of the anomaly fields and evaluate their application to mineral resource studies. Herein we provide results of geophysical interpretations that we previously have presented both orally and as poster papers at conferences and public meetings during the course of the CUSMAP investigations (Kleinkopf and others, 1988a, b, 1989; Kleinkopf, 1989). Total-intensity magnetic and complete Bouguer gravity anomaly maps of the study area have been published at a scale of 1:250,000 (Abrams, 1991; McCafferty and Abrams, 1991). These are companion maps to geological (Worl and others, 1991), geochemical, and radioelement maps prepared at the same scale by other members of the Hailey CUSMAP team.

Radioelement data were compiled for the study area by M.G. Medberry (written commun., 1989) from aerial gamma-ray spectrometer surveys flown as part of the National Uranium Resource Evaluation (NURE) program of the U.S. Department of Energy (Geometrics, Inc., 1979a, b). Radioelement maps were prepared that show surface concentration values for potassium, thorium, and uranium.

Interpretation of aeromagnetic and gravity anomaly data was an integral part of the mineral assessment of the study area. The anomaly data were used in conjunction with geological and geochemical investigations to predict locations of buried intrusive complexes, to delineate major fault and shear zones, and to identify locations that might be hydrothermally altered or mineralized. The aeromagnetic anomaly data were more useful in the direct detection of mineralized features, whereas the gravity anomaly data helped to define regional structure and discriminate low-density intrusive bodies in terranes of high-density rocks.

A map of generalized geologic terranes and major structures of the study area is shown in figure 2 (Worl and Johnson, 1989, this volume). A geologic terrane is defined herein as an area in which one or more assemblages of rock types are present in surface outcrops. Comparison of the aeromagnetic and gravity anomaly data with the geology illustrates signatures that correlate with major structures and lithology, as well as other signatures that correlate with similar structures and lithology projected in the subsurface.

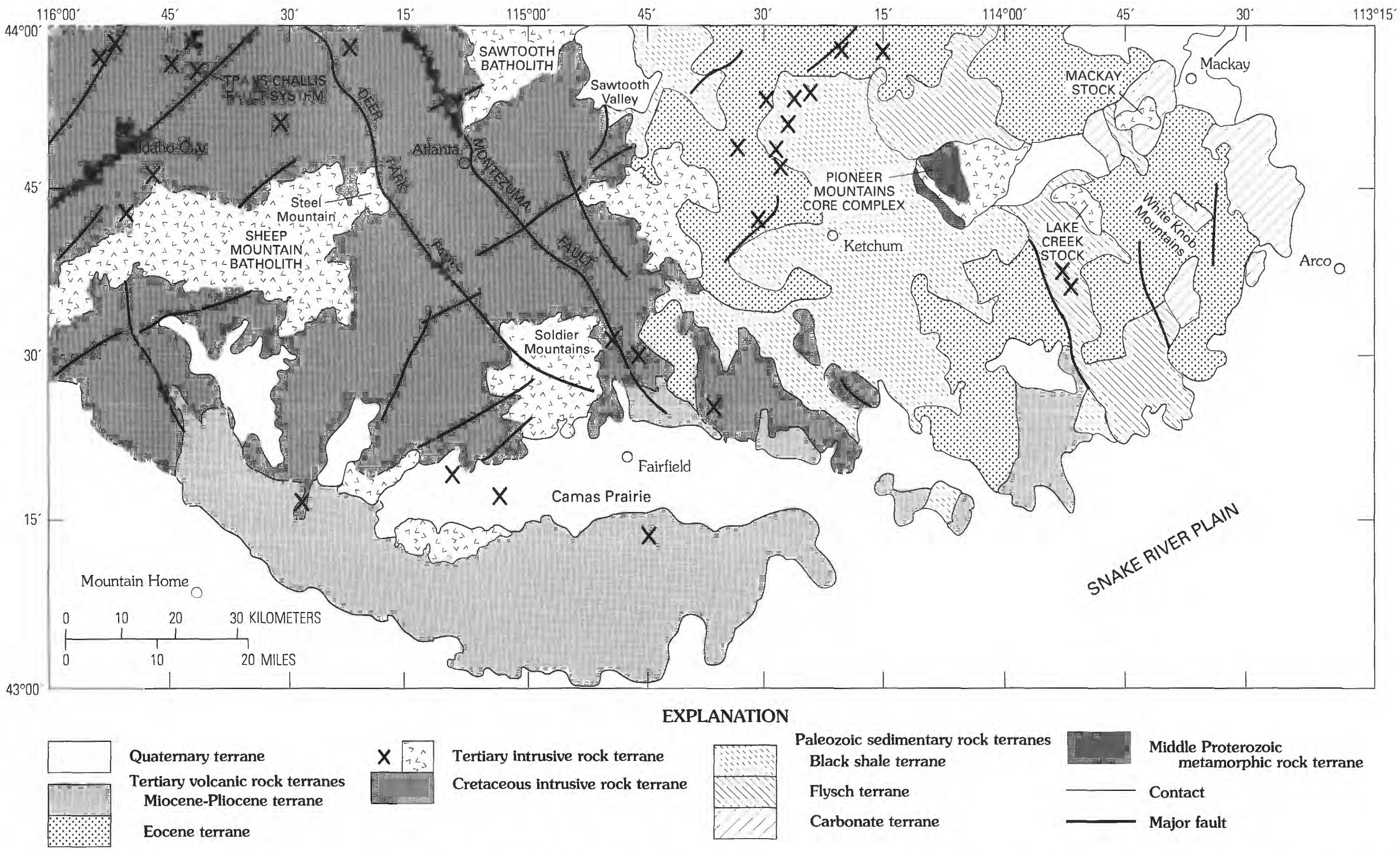


Figure 2. Map showing geologic terranes and major faults of the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho. Modified from Worl and Johnson (1989, this volume).

Mines and prospects containing precious metals are widely scattered in the study area. Many deposits are associated spatially with granitic intrusive rocks and known or inferred fault zones, particularly the northeast-trending Trans-Challis fault system and subparallel systems (Bennett and Lewis, 1989). Geophysical studies of the Trans-Challis fault system by Mabey and Webring (1985) during mineral resource assessment of the adjacent Challis 1°×2° quadrangle provide information applicable to the study area. The northeast-trending Hailey mineral belt is of particular interest in the mineral resource studies because of its parallelism and proximity to northeast-trending mineralized faults of the Trans-Challis fault system in the northwestern part of the Hailey quadrangle.

In this part of Idaho, numerous magnetic and gravity surveys and regional compilations for mineral resource exploration and regional framework studies have been conducted during the past 30 years. Published reconnaissance magnetic and gravity anomaly maps are particularly useful for describing the general geophysical setting of the region (Bonini, 1963; Mabey and others, 1974; Eaton and others, 1978; Mabey and others, 1978; Zietz and others, 1978; Geometrics, Inc., 1979a, b; Hildenbrand and others, 1982; Bankey and others, 1985; Jachens and others, 1985; Hill, 1986; Bankey and Kleinkopf, 1988; Cady and others, 1991; Bankey, 1992; McCafferty, 1992). Various workers have made geophysical studies applied to mineral resource assessment and to interpretation of structural framework of the southern Idaho batholith and the Snake River Plain (Hill and others, 1961; La Fehr and Pakiser, 1962; Hill, 1963; Hill and Pakiser, 1966; Mabey, 1966, 1976, 1978, 1981, 1982a, b, 1986; Kleinkopf, 1970, 1989; Zohdy and Stanley, 1973; Zoback and Thompson, 1978; Bhattacharyya and Mabey, 1980; Pankratz and Ackermann, 1982; Kuntz and others, 1983; Criss and Champion, 1984; Mabey and Webring, 1985; Mabey and Tschanz, 1986; Whitehead, 1986; Cluer, 1987; Toth and others, 1987; Webring and Mabey, 1987; Cluer and others, 1988; Kleinkopf and others, 1988a, b, 1989; Sanford and others, 1989; Wilson and others, 1990; Abrams, 1991; McCafferty and Abrams, 1991).

Acknowledgments.—Viki Bankey and Robert Kucks assisted in computer processing of the data. Robert Kucks also helped with the fieldwork.

MAGNETIC ANOMALY DATA

The residual total-intensity aeromagnetic map was compiled using data from nine surveys flown at different line spacings and altitudes (pl. 1*A*). Through a series of standard analytical techniques, the data were reduced to a common surface of 1,000 ft (305 m) above ground (pl. 1*A*). The description of the computational procedures and the contoured aeromagnetic anomaly map at a scale of 1:250,000 have been published previously (McCafferty and Abrams, 1991). Because of the large variation of flightline spacings, from 0.5 to 3 mi (0.8–4.8 km), an index of the magnetic surveys used in the compilation is included on plate 1 to aid the reader in evaluating the data in specific areas. The aeromagnetic data for this study are included in a previous compilation done for a regional mineral resource assessment of the Idaho batholith and adjacent areas (McCafferty, 1992). The data are in digital form (McCafferty and others, 1990) and can be obtained through the Earth Resources Observation Systems (EROS) Data Center, U.S. Geological Survey, Sioux Falls, South Dakota 57198, or the National Geophysical Data Center (NGDC), National Oceanic and Atmospheric Administration (NOAA), Boulder, Colorado 80303.

Other reconaissance aeromagnetic maps of the Hailey and Idaho Falls 1°×2° quadrangles, from surveys flown as part of the National Uranium Resource Evaluation (NURE) program of the Department of Energy, are available from the U.S. Geological Survey (Geometrics, Inc., 1979a, b; Hill, 1986).

GRAVITY ANOMALY DATA

The complete Bouguer gravity anomaly map (pl. 1*B*) is controlled by more than 2,500 unevenly spaced gravity stations. Most of this control is from a data set compiled for the recently published Bouguer gravity anomaly map of Idaho (Bankey and others, 1985). These data are from surveys conducted by the U.S. Geological Survey, U.S. Department of Defense, and various universities. Other data compiled for the map are from almost 300 observations made by B.L. Cluer collected as part of a Master of Science thesis study of Camas Prairie in the area around Fairfield (Cluer, 1987). Additional data are from 120 new stations collected by the U.S. Geological Survey during 1989 and 1990. The Bouguer gravity anomaly values for the map are estimated to be accurate to 1 milligal (mGal).

Description of the computational procedures and an earlier version of contoured gravity anomaly map at a scale of 1:250,000 were published previously (Abrams, 1991). The data are for this study are included in an earlier compilation done for a regional mineral resource assessment of the Idaho batholith and adjacent areas (Bankey, 1992). The data in digital form (McCafferty and others, 1990) may be obtained from the Earth Resources Observation Systems (EROS) Data Center, Sioux Falls, South Dakota 57198, or the National Geophysical Data Center (NGDC), National Oceanic and Atmospheric Administration (NOAA), Boulder, Colorado 80303.

PHYSICAL PROPERTIES OF THE ROCKS

Only qualitative generalizations are made herein about values of density and magnetic susceptibility because of the wide variety of rock types in the study area and the variable lithology within rock unit. Estimates judged to be approximately representative of the major lithologic units were made from a consensus of three sources: (1) laboratory and field measurements, (2) values obtained by Mabey and Webring (1985) from their studies of the adjacent Challis quadrangle, and (3) values obtained by Criss and Champion (1984), who made analyses of more than 600 samples of granitic rocks from the southern part of the Idaho batholith to study effects of hydrothermal alteration on magnetic properties. Cretaceous granodiorite and two-mica granite that make up the main part of the Idaho batholith have an average density of less than 2.6 g/cm^3, and diorite intrusive rocks have an average density of about 2.7 g/cm^3. Densities of Cenozoic sedimentary and Tertiary volcanic rocks are less than 2.5 g/cm^3; however, the volcanic rocks show large variations in density and magnetization. Tertiary granite has an average density of about 2.56 g/cm^3, less than that of Cretaceous granite. Paleozoic sedimentary rocks comprising the allochtons in the eastern part of the study area have an average density of about 2.6 g/cm^3. The low density of granitic rocks of the Cretaceous Idaho batholith and Tertiary batholiths and stocks is possibly related to extensive hydrothermal alteration. Criss and Champion (1984) concluded that pervasive hydrothermal alteration may cause minor lowering of density, on the order of 0.02–0.04 g/cm^3, of most rock types.

Granitic rocks of the Tertiary batholiths and stocks are considerably more magnetic than granitic rocks of the Cretaceous Idaho batholith. In general, granitic rocks of the Idaho batholith, excluding the border facies, are granodiorite and two-mica granite that are weakly magnetic and exhibit few positive anomalies. Exceptions are dioritic intrusive rocks, which give positive magnetic anomalies similiar to expressions of the Tertiary intrusive rocks. Tertiary volcanic rocks, in particular extrusive rocks of the Snake River Plain, are moderately to strongly magnetic. The extrusive volcanic rocks commonly exhibit remanent magnetization greater than induced magnetization, by several orders of magnitude. Criss and Champion (1984) found that induced magnetization is predominant in Cretaceous and Tertiary intrusive rocks and concluded that effects of remanent magnetization are unimportant to the interpretation of magnetic maps of these intrusive rock terranes. The Paleozoic allochthons composed of Devonian to Permian sedimentary rocks are nonmagnetic, except for areas of local alteration or areas of magnetite enrichment around Tertiary intrusions; such differences could be important to the small-scale prospector who might wish to conduct detailed ground magnetometer traverses.

CRUSTAL PROVINCES

Three crustal provinces were distinguished in the study area on the basis of contrasting aeromagnetic and gravity signatures (pls. *1A*, *B*). These crustal provinces can be used to provide a framework for the application of geophysical interpretations to mineral resource evaluation. (1) The northwest province, composed of Cretaceous granitic terranes of the Idaho batholith, is weakly magnetic and has slowly varying, highly negative Bouguer anomaly values. (2) The northeast province in the area of the Pioneer and White Knob Mountains comprises assorted terranes of thrust and block-faulted Paleozoic sedimentary rocks, Precambrian metamorphic rocks of the Pioneer Mountains core complex, Tertiary granitic intrusive rocks, and Eocene and Miocene volcanic rocks that exhibit a variety of weak to strong magnetic anomalies and give an overall negative mass effect on the gravity field similar to that of the Idaho batholith (Bankey and Kleinkopf, 1988). (3) The Snake River Plain province of predominantly Cenozoic basalt exhibits complex patterns of high-frequency positive and negative magnetic anomalies that reflect high magnetic susceptibilities and reversed magnetizations.

Most of the granitic rocks of the Cretaceous Idaho batholith (northwest province) are biotite granodiorite and two-mica granite (Kiilsgaard and Lewis, 1985; Lewis and Kiilsgaard, 1991). The batholithic terrane exhibits exceptional uniformly low magnetization and low density, except for a few diorite intrusive bodies and minor border facies of tonalite, which are more strongly magnetized and denser than the main mass of the batholith. The highest magnetic values in the study area, except for volcanic rocks of the Snake River Plain province, correlate spatially with Tertiary epizonal plutons such as granitic intrusive rocks of the Sawtooth batholith north of Atlanta, the stock at Steel Mountain just north of Rocky Bar, and the Sheep Creek batholith about 24 km west of Rocky Bar. The magnetic anomaly data indicate that each of these granitic masses is a composite intrusive complex of variable composition or state of alteration and is more extensive in the subsurface than in outcrop.

The series of prominent gradients on the Bouguer gravity anomaly map (pl. *1B*) that decrease from the Snake River Plain province to the negative values in the north-central part of the study area are part of a regional negative mass anomaly in central Idaho, 50–80 mGals in amplitude, resulting from a thickened crust beneath the Idaho batholith. This large negative anomaly shows a strong inverse correlation with regional topography. Bonini (1963) calculated the isostatic anomalies to be near zero; thus the region is in approximate isostatic equilibrium (Jachens and others, 1985; Mabey and Webring, 1985). Superimposed on the regional negative gravity anomaly are local positive and negative anomalies, in the form of nosings or closures in the contours, that result from mass anomalies in the upper crust. The negative anomalies may be associated with low-density granite stocks or with accumulations of low-density alluvial material in intermontane basins and valleys such as Sawtooth valley and southeast and northwest of Mackay (Wilson and others, 1990).

In order to remove the broad negative anomaly of 50–80 mGal that dominates the Bouguer gravity anomaly map (pl. 1*B*), a residual map (pl. 1*C*) was prepared. The residual map also enhances local gravity anomalies assumed to be associated with shallow sources of interest. To prepare the residual map the Bouguer gravity anomaly data was transformed to the frequency domain, then the data were filtered (Hildenbrand, 1983) to remove long-wavelength anomalies greater than 50 km, in particular, the regional negative anomaly. The resulting residual map is composed mainly of short-wavelength anomalies; this procedure assumes that the cut-off wavelength of the filter and the maximum depth of source are related (Hildenbrand and others, 1982). On the basis of empirical relationships (Hildenbrand, 1983), the resulting residual gravity anomalies consisting of wavelengths of 50 km and less likely are related to sources above depths of 4–8 km.

The southern part of the Idaho batholith comprises a series of rhombic blocks that are bounded by a combination of northeast-oriented faults, parallel with the Trans-Challis fault system, related to Eocene extension and northwest-oriented faults related to Miocene basin and range extension (Bennett and Lewis, 1989). Segments of these faults can be inferred in the geophysical data. In particular, belts of steep magnetic gradient correlate with faults or postulated shear zones.

The northeast-trending belt of high magnetic gradient near Atlanta (pl. 1*A*) is interpreted to reflect a buried shear zone (Kleinkopf, 1993). The proposed shear zone, here called the Atlanta shear zone, may be related to the northeast-trending Trans-Challis fault system. Intrusive rocks of the Sawtooth, Steel Mountain, and Sheep Creek batholiths are along the northwest side of this belt. The Bouguer gravity anomaly data along the belt show some corroborative evidence for the interpretation of a buried shear zone. To illustate this, the trace of the Atlanta shear zone, as interpreted from the magnetic anomaly data, was plotted on the gravity anomaly maps (pls. 1*B*, *C*). The dominantly northwesterly grain of the gravity anomaly data is interrupted intermittently along the trace of the proposed shear zone by northeasterly contour deflections and alignments of low-amplitude negative and positive residual gravity anomalies, expressed in the form of subtle nosings in the contours. Regional basin and range faults, such as the Deer Park and Montezuma faults, and correlative magnetic and gravity trends show subtle changes of strike from south-southeast north of the proposed shear zone to southeast south of the shear zone (pl. 1).

Across the southern boundary of the Idaho batholith, irregular changes in the magnetic and gravity anomaly patterns reflect the complex lithology of the transition zone between the Idaho batholith and young basalts of the Snake River Plain (pls. 1*A*, *B*). Some of these signatures reflect blocks of granitic rocks of the Idaho batholith intermixed with rocks of the volcanic terrane along the edge of the Snake River Plain province. Residual anomalies in the form of nosings of the contours may be the expression of irregular edges of Tertiary intrusive rocks along the transition zone. The magnetic anomaly map shows that the rather elongate low-amplitude anomalies of the batholith change laterally to smaller, high-amplitude, somewhat equidimensional anomalies characteristic of younger basalts of the Snake River Plain province.

The Bouguer gravity anomaly data in the Camas Prairie area show an east-trending gravity trough (pls. 1*B*, *C*) of 5–10 mGal amplitude near the southern boundary of the Idaho batholith. On the basis of gravity modeling, Cluer and others (1988) interpreted Camas Prairie as an intermontane basin containing some 2.5–3 km of alluvium and basalt and bounded by high-angle normal faults. They described the basin as "structurally within the transition zone between the late Cretaceous-early Tertiary granitic batholith and the Miocene-Recent Snake River Plain volcanogenic province." The magnetic anomaly data exhibit negative anomalies along a corresponding east-west axis at Camas Prairie (pl. 1*A*). The Bouguer gravity anomaly contours trend east, then curve north along outcrops of Eocene extrusive rocks and Paleozoic sedimentary rocks at the edge of the Snake River Plain (fig. 2. pl. 1*B*). Several distinctive negative gravity anomalies are present along the plain, especially near Craters of the Moon National Monument and the town of Arco. To the west of Camas Prairie, east trends in the gravity and magnetic anomaly data changing to northwest trends are punctuated by positive and negative anomalies that reflect mixing of lithologies of the Idaho batholith (northwest province) and the Snake River Plain province.

The eastern boundary of the Idaho batholith, which is generally considered to be near the eastern limit of outcrops of Cretaceous granitic rocks (fig. 2), is indistinct in the geophysical data. The Idaho batholith may extend in the subsurface of the northeast province to the east, possibly as far as Mackay, beneath allochthons of Paleozoic black shale and clastic rocks and younger Eocene volcanic cover. Just to the north, in the Challis quadrangle, in the same structural setting, Mabey (1982b) concluded that allochthonous Paleozoic sedimentary rocks (Hall, 1985) may be underlain by a magnetized basement.

A possible structural boundary, herein called the Great Rift shear zone, is suggested by the geophysical data. In the southeastern part of the study area, in the Snake River Plain province, a northwest-trending zone, named the Great Rift by Stearns (1928), is described by Kuntz and others (1983) as consisting of a volcanic rift, volcanic vents, and fissures. On the basis of alignments of gravity and magnetic anomalies (pl. 1), we interpret the Great Rift shear zone as a deep-seated shear zone that projects northwest past the Pioneer Mountains core complex as far as the north-central part of the study area where it is terminated by northeast trends of the proposed Atlanta shear zone, believed to be part of the Trans-Challis fault system. Bennett (1984) noted that the intersection of the Trans-Challis fault system and basin and range structures marks the change from northwest extension during the Eocene (?) to northeast extension during the Oligocene (?).

MINERAL RESOURCES

The aeromagnetic anomaly data provide definitive information for identifying and mapping plutons and mineralized fault zones, whereas the gravity anomaly data provide general constraints on mapping lithology of granitic units, commonly within batholithic complexes, and information for detecting and projecting major structural features in the subsurface. The aeromagnetic and gravity anomaly data are complementary because the two types of data are measures of different physical properties of the rocks. For example, low-density, high-magnetic-susceptibility felsic granitic intrusive rocks of Tertiary age can be distinguished from higher density, low-magnetic-susceptibility granodioritic intrusive rocks of the Cretaceous Idaho batholith.

The distribution of control for the aeromagnetic and gravity data is also different due to different systems of measurements. The magnetic control consists of almost continuous data collected along flightlines that are usually regularly spaced for a given survey at an interval of 1 mi (1.6 km) or more for regional studies, such as this study (pl. 1*A*). Sampling rates are typically several milliseconds, which translate into recording measurement of magnetic field intensity every few hundred feet. The gravity data, on the other hand, consist of isolated and unevenly distributed data points collected on the ground. The observation points, or stations, normally are spaced from 1 to 3 mi (1.6–4.8 km) or more apart, depending on the accessibility of the terrain, field time available, and the method of transport. In detailed studies that might be considered in the future for the study area, closely spaced data could be collected to map segments of faults or granitic intrusive rocks that have been intensely hydrothermally altered or shattered to the extent that contrasts in magnetic susceptibility and density are perceptibly decreased relative to adjacent unaltered or undisturbed rock.

In order to study spatial relationships between geophysical anomalies and the location of precious-metal and polymetallic deposits, a generalized summary plot (Worl and Johnson, 1989) was prepared that combines distribution of precious-metal and polymetallic vein, stockwork, and skarn deposits for the study area (fig. 3). The deposit types shown correspond to mineral deposit models, or modifications of models, described in Cox and Singer (1986).

In general, the distribution of these particular deposit types correlates with known geologic features or those postulated from the geophysical data. Northeast of Atlanta, gold deposits are clustered along the belt of high magnetic gradient (fig. 3, pl. 1*A*), which is extensive to the northeast and to the southwest and has been interpreted as a major northeast-trending shear zone (Atlanta shear zone) in the Idaho batholith that is probably related to the Trans-Challis fault system. If the trend is part of the Trans-Challis fault system, it may have been the focus of extensive epithermal gold mineralization characteristic of the Trans-Challis fault system (Bennett, 1984; Kiilsgaard and Bennett, 1985).

In view of the association of mineral deposits with volcanic centers and graben structure observed in the neighboring Challis quadrangle (Fisher, 1985; Hardyman, 1985; McIntyre and Johnson, 1985), four volcanic features defined from geologic mapping in the study area are shown on the magnetic and gravity anomaly maps (pl. 1). These are the caldron west of Ketchum (Hall and McIntyre, 1986), the Magic Reservoir eruptive center east of Fairfield (Leeman, 1982; Bonnichsen and others, 1989), the Lehman Basin caldron complex west of Mackay (Moye and others, 1989), and the Alder Creek eruptive center south of Mackay (Moye and others, 1989). Just west of Ketchum, Hall and McIntyre (1986) described an area of downdropped rocks bounded by arcuate faults on three sides as a volcanic caldron. Propyllitic alteration and extensive pyritization are present. High-frequency magnetic anomalies on the east, west, and north sides of the area are suggestive of possible ring structure around the volcanic feature (labeled "caldron" on pl. 1*A*). The Lucky Boy and the Ontario mines are within the caldera and reportedly produced silver, lead, zinc, and copper (Hustedde and others, 1981; Park, 1989). Southeast of Fairfield, a postulated volcanic eruptive and subsidence center at Magic Reservoir (Leeman, 1982; Bonnichsen and others, 1989) exhibits a complex magnetic pattern consisting of a deep low surrounded on three sides by high-frequency positive anomalies (pl. 1*A*). Gold production reportedly has exceeded more than 100,000 ounces (2,830 kg) in the neighboring Camas mining district (Hustedde and others, 1981; Bennett and Lewis, 1989), although no direct relationship between the mineral deposits and the Magic Reservoir eruptive center is inferred.

In the Mackay area, the Lehman basin caldron complex and the Alder creek eruptive center (pl. 1) were mapped and interpeted during the geologic studies (Moye and others, 1989). These were sites of eruptions of intermediate lavas and ash-flow tuffs and intrusions of silicic plutonic rocks. In the Lehman basin and the Alder Creek mining district, base- and precious-metals were deposited in porphyry systems and veins controlled by north- to northwest-trending structures (Strowd and others, 1981; Moye and others, 1989; Snider and Moye, 1989; Worl and Johnson, 1989). The magnetic anomaly data exhibit mainly negative values associated with the volcanic centers (pl. 1), but the 3-mi (4.8 km) spacing of the flightlines in this area provides data of insufficient detail to define boundaries of the centers. The reconnaissance-scale gravity control in this area is not sufficient to define the volcanic features; however, the anomaly data show north- to northwest- and northeast-trends (pls. 1*B*, *C*) that are consistent with regional structural trends (Southworth, 1988; Soulliere and others, 1989).

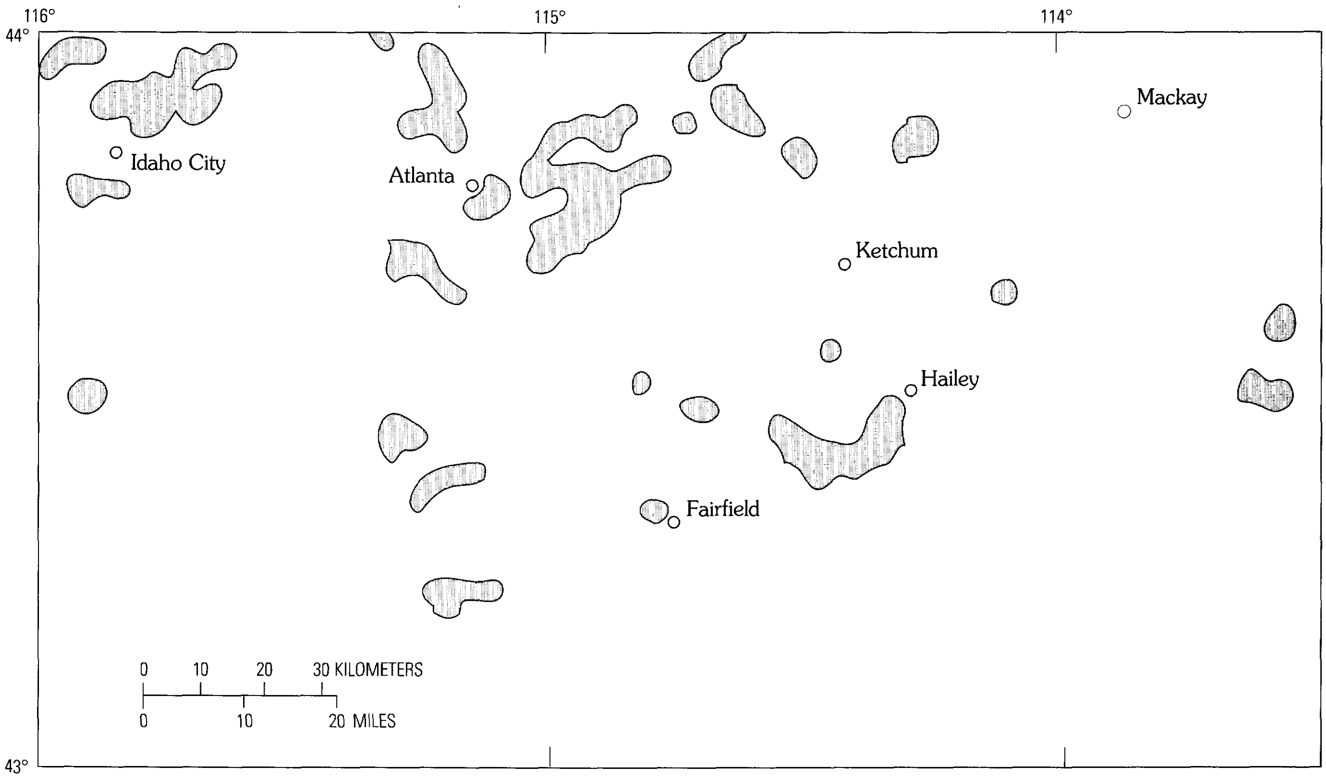


Figure 3. Map showing distribution of precious-metal and polymetallic deposits (shaded areas) in the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle, south-central Idaho. Deposits include complex precious-metal veins, shear-zone-hosted precious-metal veins, shear-zone-hosted polymetallic veins, stockwork deposits containing significant gold, massive gold veins, and massive silver veins. Modified from Worl and Johnson (1989); deposit types following Cox and Singer (1986).

Three volcanic or subsidence features that have not been mapped geologically are suggested by the aeromagnetic anomaly data. One feature is along the postulated Great Rift shear zone just southeast of the Pioneer Mountains core complex, where complicated magnetic anomaly patterns and variable gravity anomaly trends suggest the possibility of a volcanic-intrusive center (pl. 1). The Lake Creek stock is in the eastern part of the feature. The pattern of high-frequency positive and negative anomalies arranged in an annular pattern around the edges of the feature gives a sense of possible ring structure. Base- and precious-metal deposits (fig. 3) that are part of the Little Wood River and Copper Basin mining districts are in the central part of the proposed center (Strowd and others, 1981; Worl and Johnson, 1989).

Two other possible volcanic features (pl. 1*A*) are along the proposed Atlanta shear zone northeast of Atlanta. These are local areas of complex negative magnetic anomaly patterns that may indicate local fracturing in the granitic rocks and possible volcanic subsidence (pl. 1*A*). Other anomalies are in areas that are mineralized, such as the Yuba district near Atlanta (fig. 3), which has produced more than 100,000 ounces (2,830 kg) of gold (Bennett and Lewis, 1985). Additional areas of possible undiscovered mineral resources may extend southwest of Atlanta along the proposed Atlanta shear zone. The Red Warrior mining district, another district in Idaho that exceeded 100,000 ounces (2,830 kg) of gold production, is along this proposed shear zone southwest of Atlanta.

An arbitrarily defined belt, referred to as the Hailey mineral belt, is shown on plate 1. The belt contains variably mineralized rock including precious- and base-metal deposits (fig. 3). It extends northeasterly in a discontinuous fashion for almost 160 km from near Mountain Home, across the southeastern part of the Idaho batholith, past the White Knob and Pioneer Mountains, to as far northeast as Mackay (Johnson and Worl, 1989). Some of the northeast-trending geophysical features within the belt suggest geologic structure approximately parallel with the Trans-Challis fault system.

CONCLUSIONS

Magnetic and gravity anomaly data for the Hailey 1°×2° quadrangle and the western part of the Idaho Falls 1°×2° quadrangle suggest a number of areas that may be favorable for followup studies to identify new occurrences of mineral resources. Detailed studies and collection of additional control can be prioritized on the basis of favorable geological and geochemical areas. The following areas for possible undiscovered mineral resources are suggested from interpretation of the aeromagnetic and gravity anomaly data.

1. Areas interpreted to be underlain by igneous intrusive rocks along the postulated Atlanta and Great Rift shear zones.

2. Areas of interpreted crossfaulting along the Atlanta and Great Rift shear zones including intersections of basin and range faults such as the Montezuma and Deer Park faults.

3. Belts of steep magnetic gradient that may indicate covered border zones and subcrop contacts around Tertiary intrusions such as the Sawtooth, Steele Mountain, Sheep Creek, and Soldier Mountains igneous complexes.

4. Areas of magnetic and gravity anomalies whose sources may be buried intrusions in regions of geochemical anomalies. For example, Smith (1989) reported, from analysis of the nonmagnetic fraction of heavy-mineral concentrates, the possibility of tin granite in the Sheep Creek area of the Twin Springs pluton. The variable magnetic patterns suggest that there probably were multiple intrusions in the Twin Springs area.

5. Southern border zones of the Idaho batholith where magnetic patterns suggest the presence of shallow Tertiary intrusive rocks that may be masked by young basalt of the Snake River Plain.

6. Areas of black shale and calcareous clastic rock terrane marked by positive magnetic anomalies suggestive of granitic intrusive rocks in sedimentary rocks, for example in the Pioneer and White Knob Mountains.

7. Aeromagnetic and gravity signatures that may indicate the presence of mineralized structure or hydrothermal alteration in areas of the mapped and postulated volcanic features discussed in preceding sections (pl. 1). The magnetic and gravity anomaly control are sufficient, in most cases, to corroborate the presence but not the details of the volcanic features, particularly in areas of 3-mile-spaced flightlines (pl. 1*A*). The data indicate, however, that detailed surveys should provide additional subsurface information for a better understanding of the geology and alteration patterns of these volcanic features.

REFERENCES CITED

Abrams, Gerda, 1991, Bouguer gravity anomaly map of the Hailey and western part of the Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF–2160–B, scale 1:250,000.

Bankey, Viki, 1992, Complete Bouguer gravity, isostatic residual gravity, and related geophysical maps centered on the Idaho batholith and Challis volcanic field, lat 42°–47°N., long 110°–118° W.: U.S. Geological Survey Geophysical Investigations Map GP–995, scale 1:1,000,000.

Bankey, Viki, and Kleinkopf, M.D., 1988, Bouguer gravity anomaly map and four derivatives: U.S. Geological Survey Geophysical Investigations Map GP–978, scale 1:1,000,000.

Bankey, Viki, Webring, M.W., Mabey, D.R., Kleinkopf, M.D., and Bennett, E.H., 1985, Complete Bouguer anomaly map of Idaho: U.S. Geological Survey Miscellaneous Field Investigations Map MF–1773, scale 1:500,000.

Bennett, E.H., 1984, The trans-Challis fault zone—A major crustal discontinuity in central Idaho: Geological Society of America Abstracts with Programs, v. 16, no. 6, p. 442.

Bennett, E.H., and Lewis, R. S., 1989, The geology and ore deposits of the Idaho batholith in the Hailey 1°×2° quadrangle, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 26–32.

Bhattacharyya, B.K., and Mabey, D.R., 1980, Interpretation of magnetic anomalies over southern Idaho using generalized multibody models: U.S. Geological Survey Open-File Report 80–475, 48 p.

Bonini, W.E., 1963, Gravity anomalies in Idaho: Idaho Bureau of Mines and Geology Pamphlet 132, 10 p.

Bonnichsen, Bill, Leeman, W.P., Jenks, M.D., and Honjo, Norio, 1989, Geologic field guide to the central and western Snake River Plain, Idaho, emphasizing the siliceous volcanic rocks, *in* Link, P.K., and Hackett, W.R., eds., Guidebook to the geology of central and southern Idaho: Idaho Geological Survey Bulletin 27, p. 183–192.

Cady, J.W., Bankey, Vicki, Lee, Keenan, McCafferty, A.E., Pitkin, J.A., and Stanley, W.R., 1991, Synthesis of geophysical and geological data centered on the Idaho batholith and the Challis volcanic field, Idaho, *in* Good, E.E., Slack, J.F., and Kotra, R.K., eds., USGS Research on Mineral Resources–1991, Program and Abstracts, Seventh Annual V.E. McKelvey Forum on Mineral and Energy Resources: U.S. Geological Survey Circular 1062, p. 6–7.

Cluer, B.L., 1987, A gravity model of basement geometry and resulting hydrogeologic implications of the Camas Prairie, south-central Idaho: Flagstaff, Northern Arizona University, M.S. thesis, 99 p.

Cluer, B.L., Best, D.M., and Agenbroad, L.D., 1988, A hydrogeological interpretation of the Camas Prairie, south-central Idaho, based on gravity anomaly modeling: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 410.

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

Criss, R.E., and Champion, D.E., 1984, Magnetic properties of granitic rocks from the southern half of the Idaho batholith—Influences of hydrothermal alteration and implications for aeromagentic interpretation: Journal of Geophysical Research, v. 89, no. B8, p. 7061–7076.

Eaton, G.P., Wahl, R.R., Prostka, H.J., Mabey, D.R., and Kleinkopf, M.D., 1978, Regional gravity and tectonic patterns, their relation to late Cenozoic epeirogeny and lateral spreading in the western Cordillera in Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 51–91.

Fisher, F.S., 1985, Summary of the geology, mineral deposits, and resource potential for selected commodities in the Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–A, p 1–27.

Geometrics, Inc., 1979a, National Uranium Resource Evaluation aerial gamma-ray and magnetic survey, Idaho Falls quadrangle, Idaho—Final report: U.S. Department of Energy, Grand Junction Office, Open-File Report GJM–310, scale 1:250,000.

———1979b, National Uranium Resource Evaluation aerial gamma-ray and magnetic survey, Hailey quadrangle, Idaho—Final report: U.S. Department of Energy, Grand Junction Office, Open-File Report GJM–338, scale 1:250,000.

Hall, W.E., 1985, Stratigraphy and mineral deposits in middle and upper Paleozoic rocks of black-shale mineral belt, central Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–J, p. 117–131.

Hall, W.E., and McIntyre, D.H., 1986, Eocene cauldron, batholith, and hydrothermal alteration west of Ketchum, Idaho: U.S. Geological Survey Open-File Report 86–0122, 5 p.

Hardyman, R.F., 1985, The Twin Peaks cauldera and associated ore deposits, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–G, p. 97–105.

Hildenbrand, T.G., 1983, FFTFIL—A filtering program based on two-dimensional fourier analysis of geophysical data: U.S. Geological Survey Open-File Report 83–237, 60 p.

Hildenbrand, T.G., Simpson, R.W., Godson, R.H., and Kane, M.F., 1982, Digital colored residual and regional Bouguer gravity maps of the conterminous United States with cut-off wavelengths of 250 km and 1000 km: U.S. Geological Survey Geophysical Investigations Map GP–953–A, scale 1:7,500,000.

Hill, D.P., 1963, Gravity, volcanism and crustal deformation of the eastern Snake River Plain, Idaho: Journal of Geophysical Research, v. 68, no. 20, p. 5897–5819.

Hill, D.P., Baldwin, H.L. Jr, and Pakiser, L.C., 1961, Gravity, volcanism and crustal deformation of the eastern Snake River Plain, Idaho, *in* Short papers in the geologic and hydrologic sciences: U.S. Geological Survey Professional Paper 424–B, p. B248–B250.

Hill, D.P., and Pakiser, L.C., 1966, Crustal structure between the Nevada test site and Boise Idaho, from seismic-refraction measurements, *in* Steinhard, J.S., and Smith, T.J., eds., The earth beneath the continents: American Geophysical Union, Geophysical Monograph 10, p. 391–419.

Hill, P.L., 1986, Lists and location maps of aeromagnetic and aeroradiometric publications from the Department of Energy NURE program: U.S. Geological Survey Open-File Report 86–525–G, 22 p.

Hustedde, G.S., Copeland, J.A., Strowd, W.B., Mitchell, V.E., and Bennett, E.H., 1981, Mines and prospects of the Hailey quadrangle, Idaho: Idaho Bureau of Mines and Geology Mines and Prospects Map Series, 41 p., 1 map, scale, 1:250,000.

Jachens, R.C., Simpson, R.W., Saltus, R.W., and Blakely, R.J., 1985, Isosatic residual gravity anomaly map of the United States (exclusive of Alaska and Hawaii): National Oceanic and Atmospheric Administration Geophysical Data Center Map, scale 1:2,500,000.

Johnson, K.M., and Worl, R.G., 1989, Preliminary assessment of the mineral resources of the Hailey and western part of the Idaho Falls 1°×2° quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 17–22.

Kiilsgaard, T.H., and Bennett, E.H., 1985, Mineral deposits in the southern part of the Atlanta lobe of the Idaho batholith and their genetic relation to Tertiary intrusive rocks and to faults, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–M, p. 153–165.

Kiilsgaard, T.H., Freeman, V.L., and Coffman, J.S., 1970, Mineral resources of the Sawtooth Primitive Area, Idaho: U.S. Geological Survey Bulletin 1319–D, 174 p.

Kiilsgaard, T.H., and Lewis, R.S., 1985, Plutonic rocks of Cretaceous age and faults of the Atlanta lobe of the Idaho batholith, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–B, p. 29–42.

Kleinkopf, M.D., 1970, Aeromagnetic survey and interpretation, *in* Kiilsgaard, T.H., Freeman, V.L., and Coffman, J.S., Mineral resources of the Sawtooth Primitive Area, Idaho: U.S. Geological Survey Bulletin 1319–D, p. D26–D27.

———1989, Gravity and magnetic anomaly patterns in mineral resource exploration, Hailey CUSMAP, Idaho, *in* Schlindler, K.S., ed., USGS Research on Mineral Resources—1989 Program and abstracts: U.S. Geological Survey Circular 1035 p. 40.

———1993, Magnetic interpretation of a mineralized shear in the southern part of the Idaho batholith, *in* Scott, R.W., Detra, P.S., and Berger, B.R., eds., Advances related to United States and international mineral resources—Developing framework and exploration technologies: U.S. Geological Survey Bulletin 2039, p. 31–33.

Kleinkopf, M.D., Bankey, Viki, and McCafferty, A.E., 1988a, Aeromagnetic and gravity anomaly maps of the Hailey and part of the Idaho Falls two-degree quadrangles, Idaho: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 424–425.

———1988b, The application of regional geophysical data in mineral resource assessments in the Hailey and Idaho Falls quadrangles, Idaho: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 425.

———1989, Gravity and magnetic anomaly patterns applied to mineral resource exploration, Hailey 1°×2° quadrangle, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 26–32.

Kuntz, M.A., Mabey, D.R., Champion, D.E., Stanley, D.W., Lefebvre, R.H., Spiker, E.C., McBroome, L.A., and Covington, H.R., 1983, Geologic and geophysical maps of the Great Rift Instant Study Area, Blaine, Butte, Minidoka, and Power counties, Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF–1462–A, scale 1:125,000.

La Fehr, T.R., and Pakiser, L.C., 1962, Gravity, volcanism and crustal deformation of the eastern Snake River Plain, *in* Short papers in the geologic and hydrologic sciences: U.S. Geological Survey Professional Paper 450–D, p. D76–D78.

Leeman, W.P., 1982, Geology of the Magic Reservoir area, Snake River Plain, Idaho, *in* Bonnichsen, Bill, and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 369–376.

Lewis, R.S., and Kiilsgaard, T.H., 1991, Eocene plutonic rocks in south-central Idaho: Journal of Geophysical Research, v. 96, no. B8, p. 13,295–13,311.

Mabey, D.R., 1966, Relation between Bouguer gravity anomalies and regional topography in Nevada and the eastern Snake River Plain, Idaho: U.S. Geological Survey Professional Paper 550–B, p. B108–B110.

———1976, Interpretation of a gravity profile across the western Snake River Plain, Idaho: Geology, v. 4, p. 53–55.

———1978, Regional gravity and magnetic anomalies in the eastern Snake River Plain, Idaho: U.S. Geological Survey Journal of Research, v. 6, no. 5, p. 553–562.

———1981, Aeromagnetic studies of the Boulder-Pioneer Wilderness Study Area, Blaine and Custer Counties, Idaho: U.S. Geological Survey Bulletin 1497–B, p. 81–83.

———1982a, Geophysics and tectonics of the Snake River Plain, Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 139–153.

———1982b, Gravity and magnetic features along the eastern margin of the Idaho batholith in central Idaho: Geological Society of America Abstracts with Programs, v. 14, no. 6, p. 320.

———1986, Gravity and magnetic anomaly maps of the Ten Mile West Roadless Area, Boise and Elmore Counties, Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF–1500–D, scale 1:62,500.

Mabey, D.R., Peterson, D.L., and Wilson, C.W., 1974, Preliminary gravity map of southern Idaho: U.S. Geological Survey Open-File Report 74–78, scale 1:500,000.

Mabey, D.R., and Tschanz, C.M., 1986, Aeromagnetic survey and tentative interpretation, eastern part of the Sawtooth National Recreation Area, Idaho, *in* Mineral resources of the eastern part of the Sawtooth National Recreation Area, Custer and Blaine Counties, Idaho: U.S. Geological Survey Bulletin 1545–B, p. 45–52.

Mabey, D.R., and Webring, M.W., 1985, Regional geophysical studies in the Challis quadrangle, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–E, p. 69–79.

Mabey, D.R., Zietz, Isidore, Eaton, G.P., and Kleinkopf, M.D., 1978, Regional magnetic patterns in part of the Cordillera in the western United States, *in* Smith, R.B.and Eaton, G.P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 93–106.

McCafferty, A.E., 1992, Magnetic maps and terrace magnetization map for the Idaho Mineral Initiative Study, Idaho and vicinity: U.S. Geological Survey Geophysical Investigations Map GP–994, scale 1:1,000,000.

McCafferty, A.E., and Abrams, G.A., 1991, Aeromagnetic anomaly map of the Hailey and western part of the Idaho Falls 1°x2° quadrangles, Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF–2160–A, scale 1:250,000.

McCafferty, A.E., Bankey, V.L., and Cady, J.W., 1990, Digital aeromagnetic, terrace-magnetization, Bouguer gravity, isostatic gravity, terrace-density, and topographic grids for data centered on the Idaho batholith: U.S. Geological Survey data release through EROS data center, Sioux Falls, S.D., 1 9-track magnetic tape.

McIntyre, D.H., and Johnson, K.M., 1985, Epithermal gold-silver mineralization related to volcanic sugsidence in the Custer graben, Custer County, Idaho, *in* McIntyre, D.H., ed., Symposium on the geology and mineral deposits of the Challis quadrangle, Idaho: U.S. Geological Survey Bulletin 1658–I, p. 109–131.

Moye, F. J., Sanford, R.F., Erdman, J.A., and Johnson, K.M., 1989, Volcanic-hosted and intrusion-related mineral deposits in the Hailey and western Idaho Falls quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 70–72.

Pankratz, L.W., and Ackermann, H.D., 1982, Structure along the northwest edge of the Snake River Plain interpreted from seismic refraction: Journal of Geophysical Research, v. 87, no. B4, p. 2676.

Park, W.A., 1989, Geology and ore deposits of the Rook's Creek stock, Blaine County, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 118–120.

Sanford, R.F., Whitney, H.A., McCafferty, A.E., and Gese, D.D., 1989, Mineral resources of the Little Wood River Wilderness study area, Blaine County, Idaho: U.S. Geological Survey Bulletin 1721–D, 15 p.

Smith, C.L., 1989, Geochemical characterization of some mineral terranes in the Hailey 1°x2° quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 23–25.

Snider, L.G., and Moye, F.J., 1989, Regional stratigraphy, physical volcanology and geochemistry of the southeastern Challis volcanic field, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 122–127.

Soulliere, S.J., Wilson, A.B., and Skipp, Betty, 1989, Geologic controls for mineralization in sedimentrary rocks, western Idaho Falls 1°x2° quadrangle, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°x2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 128–135.

Southworth, C.S., 1988, Preliminary lineament map of the Hailey, Idaho and west half of the Idaho Falls, Idaho 1°x2° quadrangles, with inset of regional image mosaic and geologic structures: U.S. Geological Survey Open-File Report 88–640, scale 1:250,000.

Stearns, H.T., 1928, Craters of the Moon National Monument, Idaho: Idaho Bureau of Mines and Geology Bulletin 13, 57p.

Strowd, W.B., Mitchell, V.E., Hustedde, G.S., and Bennett, E.H., 1981, Mines and prospects of the Idaho Falls quadrangle, Idaho: Idaho Bureau of Mines and Geology Mines and Prospects Map Series, 10 p.

Toth, M.I., King, H.D., Kulik, D.M., and Leszcoykowski, A.M., 1987, Mineral resources of the King Hill Creek Wilderness Study Area, Elmore County, Idaho: U.S. Geological Survey Bulletin 1721–B, 88 p.

U.S. Department of Defense, 1974, World relative gravity reference network, part 2, North America (vol. 2): Defense Mapping Agency Aerospace Center Reference Publication 25, with supplement updating gravity values to the International Gravity Standardization Net (IGSN–71), 1,635 p.

U.S. Geological Survey, 1974, Residual magnetic intensity map—Bruneau, Idaho: U.S. Geological Survey Open-File Report 74–149, scale 1:62,500.

———1979a, Aeromagnetic map of the Pioneer area, Idaho: U.S. Geological Survey Open-File Report 79–1450, scale 1:62,500.

———1979b, Aeromagnetic map of the Arco area, Idaho: U.S. Geological Survey Open-File Report 79–1445, scale 1:250,000.

Webring, M.W., and Mabey D.R., 1987, Geophysical anomalies in the Challis quadrangle, *in* Fisher, F.S., and Johnson, K.M., eds., Preliminary manuscript for "Mineral-resource potential and geology of the Challis 1°×2° quadangle, Idaho": U.S. Geological Survey Open-File Report 87–480, p. 66–80.

Whitehead, R.L., 1986, Compilation of selected geophysical references for the Snake River Plain, Idaho and eastern Oregon: U.S. Geological Survey Geophysical Investigations Map GP–969.

Wilson, A.A., Janecke, S.U., Skipp, Betty, Kleinkopf, M.D., McCafferty, A.E., Barton, H.N., and Miller, M.S., 1990, Mineral resources of the Borah Peak Wilderness Study Area, Custer County, Idaho: U.S. Geological Survey Bulletin 1718, p. E15.

Worl, R.G. and Johnson, K.M., 1989, Geologic terranes and mineral deposit types in the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho, *in* Winkler, G.R., Soulliere, S.J., Worl, R.G., and Johnson, K.M., eds., Geology and mineral deposits of the Hailey and western Idaho Falls 1°×2° quadrangles, Idaho: U.S. Geological Survey Open-File Report 89–639, p. 3–16.

Worl, R.G., Kiilsgaard, T.H., Bennett, E.H., Link, P.K., Lewis, R.S., Mitchell, V.E., Johnson, K.M., and Snyder, L.D., 1991, Geologic map of the Hailey 1°×2° quadrangle, Idaho: U.S. Geological Survey Open-File Report 91–340, scale 1:250,000.

Zietz, Isidore, Gilbert, F.P., and Kirby, J.R., 1978, Aeromagnetic map of Idaho—Color coded intensities: U.S. Geological Survey Geophysical Investigations Map GP–920, scale 1:1,000,000.

Zoback, M.L., and Thompson, G.A., 1978, Basin and range rifting in northern Nevada—Clues from a mid-Miocene rift and its subsequent offsets: Geology, v. 6, p. 111–116.

Zohdy, A.A.R., and Stanley, W.D., 1973, Preliminary interpretation of electrical sounding curves obtained across the Snake River Plain from Blackfoot to Arco, Idaho: U.S. Geological Survey Open-File Report 73–370, 5 p.


U.S. DEPARTMENT OF THE INTERIOR
MARCH 3, 1849