Guidebook to the Geology of Central and Southern Idaho

Paul Karl Link
and
William R. Hackett
Editors
Advisory Board

Office of the Governor, Cecil D. Andrus.................................................................Governor
  David Hawk, representative
Stanley Hamilton..................................................................................................Director, Idaho Department of Lands
Paul Karl Link.....................................................................................................Chairman, Department of Geology, Idaho State University
Clayton R. Nichols...............................................................................................Idaho National Engineering Laboratory
Deborah J. Parliman.........................................................................................President, Idaho Association of Professional Geologists
Jack Peterson......................................................................................................President, Idaho Mining Association
Ray Stark..............................................................................................................Legislative Budget Office
Craig M. White....................................................................................................Chairman, Department of Geology and Geophysics, Boise State University
Robert W. Bartlett..............................................................................................Director and State Geologist, ex officio

Administrative and Support Staff

Robert W. Bartlett..............................................................................................Director
Earl H. Bennett.....................................................................................................Associate Director
Roger C. Stewart.................................................................................................Editor
Jennifer Pattison Hall.........................................................................................Publications Assistant
Charlotte D. Fullerton.......................................................................................Secretary/Records Manager
Dreaucine Bonner...............................................................................................Clerical Specialist

Research Staff

Earl H. Bennett.....................................................................................................Associate Director
Bill Bonnichsen.................................................................................................Supervisory Geologist
Roy M. Breckenridge.........................................................................................Supervisory Geologist
Charles R. Knowles........................................................................................Senior Geologist
Kurt L. Othberg....................................................................................................Senior Geologist

Research Associates

Victoria E. Mitchell...............................................................................................Geologist II
Margaret D. Jenks...............................................................................................Research Geologist

Production Staff

Editor....................................................................................................................Roger C. Stewart
Editorial Assistant.............................................................................................Jennifer Pattison Hall
Design and Layout..............................................................................................Jennifer Pattison Hall
Typography..........................................................................................................William R. Hackett and Paul Karl Link,
  Department of Geology, Idaho State University
  Eileen A. Cline, University of Idaho
Primer..................................................................................................................Capitol Lithograph and Printing, Boise
Guidebook to the Geology of Central and Southern Idaho
The "highway flow," a black-lava flow of ferrolattite from Craters of the Moon lava field. The flow is crossed by U.S. Highway 20-26 about 22 miles southwest of Arco, Idaho. Photograph by W. R. Hackett.
Guidebook to the Geology of Central and Southern Idaho

Edited by
Paul Karl Link
and
William R. Hackett

Idaho Geological Survey
University of Idaho
Moscow, Idaho 83843
CONTENTS

CONTENTS BY AUTHOR xi
COVER CREDITS xiii
PREFACE xv
INTRODUCTION 1

CHAPTER ONE
Geology of Central Idaho 3

Structural and Stratigraphic Transect of South-Central Idaho: A Field Guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains
Paul Karl Link, Betty Skipp, M. H. Hait, Jr., Susanne Janecke, and Bradford R. Burton 5

Field Guide to the Pioneer Mountains Core Complex, South-Central Idaho
Stephen L. Wust and Paul Karl Link 43

Cretaceous and Tertiary Intrusive Rocks of South-Central Idaho
Kathleen M. Johnson, Reed S. Lewis, Earl H. Bennett, and Thor H. Kjølsgaard 55

Regional Geologic Setting and Volcanic Stratigraphy of the Challis Volcanic Field, Central Idaho
Falma J. Moye, William R. Hackett, John D. Blakley, and Larry G. Snider 87

A Transect Across an Island Arc-Continent Boundary in West-Central Idaho
Elaine A. Aliberti and Cathryn Allen Manduca 99

Hydrothermal Systems of the Wood River Area, Idaho
Duncan Foley and Leah Street 109
CHAPTERTWO
Paleozoic Stratigraphy
127

Early Paleozoic Continental Margin Development, Central Idaho
Mark D. McFadden, Elizabeth A. Measures, and Peter E. Isaacson

Stratigraphy and Structure of the Milligen Formation, Sun Valley Area, Idaho
Robert J. W. Turner and Bruce R. Otto

Stratigraphy of the Lower Permian Grand Prize Formation, South-Central Idaho
J. Brian Mahoney and R. M. Sengebush
CHAPTER THREE
Economic Geology

181

Geology and Geochemistry of Jasperoid Near Mackay, Idaho
Anna B. Wilson, Sandra J. Soulliere, Betty Skipp, Ronald G. Worl, and
Keith P. Rhea

Ore Deposits of the Carrietown Silver-Lead-Zinc District, Blaine and Camas
Counties, Idaho
Robert S. Darling
CHAPTER FIVE
Geology of the Snake River Plain
245

Geologic Field Trip Guide to the Central and Western Snake River Plain, Idaho,
Emphasizing the Silicic Volcanic Rocks
Bill Bonnichsen, William P. Leeman, Margaret D. Jenks, and Norio Honjo

Explosive Basaltic and Rhyolitic Volcanism of the Eastern Snake River Plain, Idaho
William R. Hackett and Lisa A. Morgan

247
283
CHAPTER SIX
Geology of Southwest Montana
303

Southwest Montana Thrust Belt: Bannack to Melrose
James W. Sears, Larry M. Johnson, Beth C. Geiger, and William C. Brandon
305
CONTENTS BY AUTHOR

Aliberti, E. A., 99

Bennett, E. H., 55
Blakley, J. D., 87
Bloomfield, J. M., 209
Bonnichsen, Bill, 247
Brandon, W. C., 305
Breckenridge, R. M., 201, 209
Burton, B. R., 5

Cotter, J. F. P., 209
Crone, A. J., 227

Darling, R. S., 193

Evenson, E. B., 201, 203, 209
Foley, Duncan, 109

Geiger, B. C., 305

Hackett, W. R., 87, 283
Ilait, M. H. Jr., 5
Honjo, Norio, 247

Isaacson, P. E., 129

Janecke, Susanne, 5
Jenks, M. D., 247
Johnson, K. M, 55
Johnson, L. M., 305

Kiilsgaard, T H., 55

Leeman, W. P., 247
Lewis, R. S., 55
Link, P. K., 5, 43

Mahoney, J. B., 169
Manduca, C.A., 99
McFadden, M. D., 129
Measures, E. A., 129
Morgan, L. A., 283
Moye, F. J., 87

Otto, B. R., 153

Pearce, Suzanne, 203
Pierce, K. L., 233

Rhea, K. P., 183

Schlieder, Gunnar, 203
Sears, J. W., 305
Sengebush, R. M., 169
Skipp, Betty, 5, 183
Snider, L. G., 87
Soulliere, S. J., 183
Stanford, L. R., 209
Stephens, G. C., 201, 223, 241
Street, Leah, 109

Turner, R. J. W., 153

Wilson, A. B., 183
Worl, R. G., 183
Wust, S. L., 43
CREDITS FOR COVER PHOTOGRAPHS

Front cover, clockwise from top:

*Cougar Point Tuff, Bruneau Canyon, Owyhee County, Idaho. Photograph by Bill Bonnichsen.*

*Mt. McCaleb and the west face of the Lost River Range south of Mackay. Photograph by P. K. Link.*

*The Menan Buttes, Pleistocene tuff cones west of Rexburg. Photograph by W. R. Hackett.*

*The Boulder Mountains and the Big Wood River north of Ketchum. Photograph by P. K. Link.*

Back cover:

*Scenic geology in the Frank Church River of No Return Wilderness Area. Dikes in the Idaho batholith. Photograph by R. M. Breckenridge.*
PREFACE

Although guidebooks of this sort are part of the geological literature of most western states, field guides and topical papers about Idaho's geology are sparse. This book presents a comprehensive summary of geologic research in central and southern Idaho, and joins Cenozoic Geology of Idaho (Bonnichsen and Breckenridge, editors, 1982) and the recent guidebook to north-central Utah and southeastern Idaho (Kerns and Kerns, editors, 1985) as modern compilations of geologic work in Idaho.

The last geologic guidebook for central Idaho was written by Clyde Ross (1963). Since that time, the earth sciences have experienced a true scientific revolution with the introduction of plate tectonics as a unifying paradigm. We now recognize the margin of the Paleozoic North American continent in western Idaho, and we have a much clearer understanding of the nature of fold-thrust belts, sedimentary environments, and volcanic processes. Although classical field methods have changed little since 1963, few of the laboratory methods that are now routinely employed in constructing tectonic hypotheses were available then. Such methods include the analysis of trace elements and isotopes in rocks and minerals, and precise geochronometry and paleomagnetic measurements. All of these methods have been applied either directly or indirectly by the authors in this guidebook. The production of this book is therefore timely, perhaps overdue.

This book was produced for the 41st annual meeting of the Rocky Mountain Section, Geological Society of America, held in Sun Valley in May 1988 and hosted by the Geology Department of Idaho State University. The book was written by authors across the United States, was edited and typeset in Pocatello, and was produced and published in Moscow. Its production thus represents a level of statewide geoscience cooperation that has not been achieved before, and its contents reflect current efforts in cooperative research among the Idaho universities and with the Idaho Geological Survey and the U.S. Geological Survey.

In many ways, the production of this book has been an experiment, and one that we consider to have been successful. It represents the arrival of "desktop publishing" to the Idaho geoscience community and was largely produced with computers. Authors submitted manuscripts on computer disks, using prescribed software and formatting; most articles were originally produced on IBM PC™-compatible hardware and software. We then "translated" the manuscripts into Microsoft Word™ for the Apple Macintosh™ computer using MacLinkPlus™ software. We then did preliminary editing and sent the articles out to external reviewers and back to the authors. After our final editing with technical assistance from Roger C. Stewart, we produced typeset copy using Apple Macintosh Plus™ and IBM XTV™ computers, respectively interfaced with Apple LaserWriter™ and Hewlett Packard LaserJet Series II™ printers. (Products are mentioned only for descriptive purposes and this does not
necessarily constitute endorsement.) The editorial staff of the Idaho Geological Survey then prepared the copy and illustrations for printing.

We are grateful to Roger C. Stewart of the Idaho Geological Survey, who agreed to produce the book, and to Jennifer Pattison Hall for page design and layout. In Pocatello we received invaluable drafting, photographic and typing aid from Carla J. Griffith, L. Marie Hollist, Vicky Davidson, and Julie Vanek. Paul Kidd of the Idaho State University Graphic Arts Department assisted with illustrations, and Tim Frazier of the Idaho State University Mass Communications Department gave valuable advice on photographic methods.

The articles were edited by us and by one or more external reviewers. We thank all those who reviewed manuscripts, including Mike Blaskowski, Bradford R. Burton, Paul Castelin, Valerie E. Chamberlain, James P. Evans, Donald W. Fiesinger, Richard F. Hardyman, Mel A. Kuntz, Gerry Lindholm, S. D. Luddington, J. Brian Mahoney, Falma J. Moye, Robert Q. Oaks, Jr., H. Thomas Ore, Forrest G. Poole, David W. Rodgers, Rick Sanford, Clyde Smith, Richard P. Smith, Walter S. Snyder, Sandra J. Soulliere, Claude Spinosa, Charles Waag, Craig M. White, Dick Whitehead, Monte D. Wilson, Spencer H. Wood, Ronald G. Worl and Bill Young.

We hope that this guidebook is only the first of many statewide cooperative publications by the Idaho Geological Survey.

Paul Link
Bill Hackett
Pocatello, Idaho
May 1988

REFERENCES CITED


xvi
Central and southern Idaho contain the western edge of the North American Precambrian craton with sutured oceanic terranes along its western border. This continental margin experienced miogeoclinal sedimentation during the early Paleozoic, followed by a series of compressional orogenies starting with the late Paleozoic Antler orogeny. The Cordilleran orogeny, manifested by a fold-thrust belt on the east and the Idaho batholith on the west, was the last major compressive event. Early Tertiary extensional tectonics and associated magmatism produced the Challis volcanic field and the rise of the Pioneer Mountains core complex. The dominant north-south Mesozoic structural grain of central Idaho was cut in Neogene time by the arcuate east-west Snake River Plain, a bimodal volcanic belt, and basin and range extension which produced the present northwest-trending mountain ranges. This book contains fifteen field trip guides and seven short papers that discuss this complex geologic history. The areas addressed by each of the articles are shown on Figure 1.

The book is organized geographically and chronologically. Chapter 1 concerns the geology of central Idaho near Sun Valley and covers a cross section in space and time including early Paleozoic sedimentation, Mesozoic intrusion of the Idaho batholith, the formation of a fold-thrust belt, and the suturing of oceanic terranes on the western edge of the state. Tertiary geologic events discussed in the first chapter include the Challis magmatic episode, the formation of the Pioneer Mountains core complex, and the present-day hydrothermal systems of the Wood River Valley. Chapter 2 addresses Paleozoic stratigraphy and discusses new work in early Paleozoic continental margin strata near Challis, the Devonian Milligen Formation in the Wood River Valley, and the Permian Grand Prize Formation at the headwaters of the Salmon River. The economic geology of south-central Idaho is the topic of Chapter 3, with articles on the Carrietyown mining district west of Ketchum and jasperoid occurrences near Mackay. One field trip, composed of six individual articles about the Quaternary geology of central Idaho, makes up Chapter 4. Topics covered include glacial geology, the history of normal faulting on the western margin of the Lost River Range, and Holocene volcanic activity at Craters of the Moon National Monument. Two field trips in Chapter 5 discuss important new work on the physical volcanology and volcanic stratigraphy of the western and eastern Snake River Plain. A lone field trip in the southwest Montana thrust belt occupies the last chapter.

The book is written in full geological jargon. We hope that readers who are not steeped in this terminology will nonetheless find the maps, photographs and roadlogs useful.

1 Department of Geology, Idaho State University, Pocatello ID 83209
Guidebook to the Geology of Central and Southern Idaho

1. Link and others, p. 5-42.
2. Wust and Link, p. 43-54.
3. Johnson and others, p. 55-86.
7. McFadden and others, p. 129-152.
17. Pierce, p. 233-240.

Figure 1. Map showing areas discussed in articles of this volume.
Chapter One
Geology of Central Idaho

Folded strata of the upper Paleozoic Wood River Formation in the head of Basin Gulch, Boulder Mountains, northeast of Sun Valley. The rugged mountains on the skyline are the Pioneer Mountains core complex. Photograph by B. R. Burton.
Structural and Stratigraphic Transect of South-Central Idaho:
A Field Guide to the Lost River, White Knob, Pioneer, Boulder, and Smoky Mountains

INTRODUCTION

The mountains of south-central Idaho expose a cross section of the Cordilleran orogenic system including a Mesozoic fold-thrust belt on the east and Cretaceous and Tertiary intrusive complexes on the west. The orogenic belt is overprinted by Basin and Range extensional faulting and bimodal volcanism of the Snake River Plain. Major tectonic events in the development of the Cordilleran orogenic system in south-central Idaho include:

1. The late Paleozoic Antler orogeny: uplift of the Antler highland was followed by flysch deposition in a foreland basin to the east.

2. The late Mesozoic Cordilleran orogeny: east-directed thrusting of the Sevier orogeny in east-central Idaho was accompanied by intrusion of the Idaho batholith farther west and deep-seated ductile deformation and metamorphism in rocks now exposed in the core of the Pioneer Mountains.

3. Paleogene extensional tectonism and the Eocene Challis magmatic episode: these two tectonic events overlapped in time and space. Paleogene formation of the Wildhorse detachment system of the Pioneer Mountains core complex was accompanied by low-angle normal faulting in areas to the west, and overlapped the volcanism, intrusion, faulting, and sedimentation of the Challis magmatic episode. Challis Volcanics include voluminous ash-flow tuffs, lavas, and hypabyssal intrusives interbedded with continental fluvial and lacustrine sedimentary rocks.

4. Neogene basin and range extension and Snake River Plain bimodal volcanism: the eastward migration of rhyolitic and basaltic volcanism on the Snake River Plain during the last 15 million years overlapped with basin and range extensional faulting. The faulting continues to the present in south-central Idaho.

DEDICATION

This article is dedicated to the late Wayne E. Hall (1920-1986) upon whose pioneering geologic studies of...
the central Idaho black-shale mineral belt much of our work is based.

PURPOSE

This field trip examines the stratigraphic, structural, and igneous record of some of these tectonic events as reflected in lateral changes in Paleozoic sedimentary units, Mesozoic deformational style, and Paleogene and Neogene volcanism and extensional tectonism of the Lost River, White Knob, Pioneer, and Smoky Mountain Ranges. Field trip stops and their purposes are listed in Table 1. The route of the field trip is shown in Figures 1, 2, and 5. Figure 4 is a geologic map of the eastern part of the area.

The purposes of this field trip are:
(1) To demonstrate the contrasts in Paleozoic stratigraphy from east to west across the Mesozoic thrust belt in south-central Idaho from the Lost River Range to the Smoky Mountains on the west, where the thrust belt is intruded by Cretaceous rocks of the Idaho batholith. The stratigraphic and structural framework is from the work of many people including Betty Skipp, Wayne Hall, Jim Dover and others over the last 30 years, and from work by Link, Burton, and others in the rocks of the Wood River basin over the last few years. Devonian and Mississippian strata record a westward progression from an outer carbonate platform (Stop 1-6) to the Antler flysch trough (Stop 1-6), and then west to the Antler highland (Stop 2-2). Pennsylvanian-Permian strata record a similar westward progression from a carbonate bank to the Copper Basin highland, bordered on the west by the Wood River basin (Stops 1-7, 2-3, 2-7).
(2) To discuss recent work by Skipp near Fish Creek Reservoir (Stops 1-7 and 1-8) and by Skipp and Link in the Wood River Valley in which faults formerly thought to be Mesozoic thrusts are reinterpreted as Tertiary normal faults (Stops 1-8, 2-1, 2-2, 2-4, 2-7, 2-8, 2-9). This interpretation bears on the larger issues of regional Paleogene tectonics including the relations between the rise of the Pioneer Mountains core complex (Stop 2-6) and tectonism associated with the Eocene Challis magmatic episode (Stop 2-9).
(3) To discuss concepts developed by Hait and now being tested by Janecke about the timing and style of Tertiary extension in the Lost River Range (Stops 1-1 to 1-3). Important questions remain about the relation of extensional faulting to the Eocene Challis magmatic episode, Basin and Range deformation, and downwarp of the Snake River Plain.

STRATIGRAPHY OF THE FIELD TRIP AREA

The area of the field trip contains rocks of several Mesozoic thrust plates (Figs. 2 and 4). These thrust plates, as modified from Skipp (1987), include from east to west: Lost River-Arco Hills; Grouse; White Knob; Copper Basin; Devonian, Silurian and Ordovician undivided; Milligen-Wood River; and Dollarhide and Grand Prize. Rocks of the Grand Prize Formation are included with the Milligen-Wood River plate on Figure 2. Stratigraphic columns for five of these thrust plates are shown on Figure 3 (also see Isaacson, 1983).

<table>
<thead>
<tr>
<th>DAY ONE: ARCO TO FISH CREEK RESERVOIR</th>
<th>Stop</th>
<th>Purpose</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-1 View from mouth of Antelope Creek: stratigraphy, Tertiary structure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1-2 Pass Creek detachment fault: Tertiary structure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1-3 Pass Creek view: Tertiary structure, stratigraphy</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1-4 Arco Park: Paleozoic stratigraphy</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1-5 Copper Basin thrust: Mesozoic structure, Paleozoic stratigraphy, Neogene Snake River Plain volcanism</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1-6 Copper Basin Formation turbidite: sedimentology and stratigraphy</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1-7 Wood River Formation at Fish Creek Reservoir: stratigraphy</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1-8 Roberts Mountains Formation and Carey Dolomite: stratigraphy, structure, mineral deposits</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>DAY TWO: WOOD RIVER VALLEY</th>
<th>Stop</th>
<th>Purpose</th>
</tr>
</thead>
<tbody>
<tr>
<td>2-1 Overview of Hailey area: Tertiary structure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-2 Colorado Gulch: Milligen Formation, Tertiary structure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-3 Hailey Conglomerate Member type area: stratigraphy, sedimentology</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-4 Quigley Creek: Wood River thrust, reactivated?</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-5 Deer Creek: Dollarhide Formation contact with Deer Creek stock: structure, magmatism</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-6 East Fork, Wood River: view of Pioneer Mountains Tertiary structure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-7 Wood River Formation in Trail Creek: stratigraphy</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-8 Wood River thrust in Trail Creek: structure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-9 Boulder Mountain view: Tertiary structure, magmatism, mineral deposits</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Precambrian and Lower Paleozoic Rocks

Precambrian gneisses are found only in the Pioneer Mountains metamorphic core complex (Dover, 1981: 1983; Wust and Link, 1988, this volume) where they may be the basement of either the White Knob thrust plate (Fig. 2) or the Copper Basin thrust plate, as suggested here (Fig. 3). Large (6 meter-long) xenoliths of granitic gneiss are found in the Challis Volcanics above the Grouse plate, suggesting that the Grouse plate also contains crystalline basement.

Metasedimentary rocks believed to be of Middle Proterozoic age include foliated pebbly quartzites in the Borah Peak area of the Lost River-Arco Hills plate and gneissose quartzite and calc-silicate rocks of the Pioneer
Mountains core complex (Dover, 1983) (Fig. 3). These correlations are tentative, especially in the Pioneer Mountains.

In the Lost River-Arco Hills plate, Late Proterozoic and lower Paleozoic quartzose detrital strata are found in the Borah Peak area (Fig. 5) (Ruppel and others, 1975; McCandless, 1982; Ingwell, 1980; Skipp, unpublished mapping, 1987). Silurian and Devonian rocks are continental margin carbonates (McFadden and others, 1988, this volume). Lower Paleozoic outer shelf successions similar to those of the Lost River-Arco Hills plate thin westward but are present at least as far west as Fish Creek Reservoir, where they are interpreted to be part of the Copper Basin thrust plate (Fig. 3) (Stop 1-8).

Thick Devonian argillite, limestone and quartzite of the Milligen Formation are present in the Milligen-Wood River plate (Sandberg and others, 1975; Otto and Turner, 1987; Turner and Otto, 1988, this volume). The undated Paleozoic Carrietown sequence (Darling, 1988, this volume) of the Dollarhide plate may contain Devonian rocks (Geslin, 1986; Darling, 1987).

Mississippian Rocks

Mississippian strata are present in Mesozoic thrust plates from western Montana westward to the Copper Basin plate. These rocks record a westward interfingering of carbonate bank limestones of the outer cratonic platform with flysch of the foreland basin that formed during the Antler orogeny (Nilsen, 1977; Skipp, Sando, and Hall, 1979; Skipp and Hall, 1980; Dover, 1980). In the Lost River Arco Hills and Grouse plates, Mississippian formations of the Upper Mississippian carbonate bank complex of Skipp, Sando, and Hall (1979) (Fig. 3) overlie eastward-thinning Mississippian turbidites of the McGowan Creek Formation in the position of the outer cratonic platform. Rocks of the White Knob plate were also deposited on the outer cratonic platform and consist of Upper Mississippian limestones and interbedded clastic sequences of the White Knob Limestone above thick turbidites of the Lower Mississippian McGowan Creek Formation. Upper and Lower Mississippian conglomeratic flysch of the
Figure 2.
Rocks of the Devonian Milligen Formation are thought to have composed part of the the Mississippian Antler highland in Idaho (Skipp and Hall, 1980; Davis, 1984) and to have been the source for some of the coarse-grained sediment that is present in the Mississippian Copper Basin Formation. The Copper Basin Formation contains up to 3500 meters of conglomerate, sandstone, limestone, and argillite (Paul and others, 1972; Paul and Gruber, 1977). The Lower Mississippian portion is inferred to represent an eastward-prograding submarine fan complex (Nilson, 1977). The Lower Mississippian distal flysch trough is represented in the Lost River-Arc0 Hills, Grouse, and White Knob plates by fine-grained sandstones and shales of the McGowan Creek Formation (Sandberg, 1975). The Upper Mississippian White Knob Limestone of the White Knob plate contains western-derived cobbles and pebbles of chert and quartzite, thus recording the interaction of the carbonate platform and the western Antler highland source area.

Pennsylvaniaian and Permian Rocks

In Pennsylvaniaian time, a second carbonate bank complex, the Bluebird Mountain and Snaky Canyon Formations, developed in central Idaho and is now exposed in the Lost River-Arco Hills, Grouse, and White Knob plates (Skipp, Hoggan, Schleichcr and Douglass, 1979). To the west, Mississippian rocks of what is now the Copper Basin plate were uplifted to form the Copper Basin highland, which was flanked on the west by the Middle Pennsylvaniaian-Early Permian Wood River basin (Skipp and Hall, 1980). Rocks of the Wood River basin are found in the Milligen-Wood River plate, the Dollarhide plate, and the Grand Prize plate (Mahoney and Sengebush, 1988, this volume). Strata deposited in the Wood River basin include three partly coeval lithostratigraphic units: the Dollarhide, Grand Prize, and White Knob Formations (Fig. 19). The Wood River basin is discussed more fully in the introduction to the second day of this field guide.

Figure 2. Tectonic sketch map of east- and south-central Idaho, from Skipp (1987). Field trip route is shown in southwest part of map. Map shows interpreted distribution of rocks of the Cabin, Hawley Creek, and State Line thrust sheets. Shaded areas are Clearwater orogenic zone. GH = Grasshopper plateau (Ruppel and Lopez, 1984; Skipp, 1985). W = window through Grasshopper plate; ACP = Ordovician Arnett Creek pluton. BMP = Beaverhead Mountains pluton. KH = Proterozoic gneiss and schist in interpreted kippe of rocks of Cabin plate. Remaining letters designate thrust plates: MWG = Milligen-Wood River thrust complex. This was labelled as the Milligen-Wood River-Grand Prize thrust complex by Skipp (1987) and (Hall, 1985), but recent work suggests that the Grand Prize Formation is not part of the same thrust complex as the Milligen Formation (see Figure 18 of this paper). DH = Dollarhide thrust complex west of MWG plate (Hall, 1985); DO = Devonian, Silurian, and Ordovician rocks undivided (Dover, 1981); CB = Copper Basin plate (Skipp and Halt, 1977); G = Grouse plate; LRAH = Lost River-Arco Hills plate (Skipp and Hait, 1977); FC = Fritz Creek plateau (Skipp, 1985); ML = Medicine Lodge plate (Scholten and others, 1955; Skipp, 1985); FEC = Four Eyes Canyon plateau (Perry and Sando, 1983); T = Tendoy plate (Scholten and others, 1955); TF = Footwall of Tendoy thrust; SP = Sapphire plateau (Hyndman, 1980). A-B-C = line of diagrammatic cross sections shown in Figure 3 of Skipp (1987). Sketch map based on geologic map compiled at 1:1,000,000, then reduced and simplified. Map sources include reports cited in Skipp (1987) and geologic maps of Idaho (Bond, 1978), and Montana (Ross and others, 1955).

TECTONICS OF SOUTH-CENTRAL IDAHO

Mesozoic Thrusting and Intrusion

South-central Idaho is composed of several Cordilleran thrust plates (Skipp and Hait, 1977; Skipp, 1987; Figs. 2 and 4). The thrust plates have distinctive Paleozoic stratigraphic sequences and are recognized primarily on that basis, although segments of the actual thrust faults are locally exposed. Paleozoic strata have been telescoped eastward and their present distribution represents a compressed paleogeography.

The western part of south-central Idaho was intruded by Cretaceous biotite granodioritic of the Idaho batholith, from 100 Ma to 75 Ma (Bennett, 1980; Kilsgaard and Lewis, 1985; Bennett and Knowles, 1983; Johnson and others, 1988, this volume). Three eastern outliers of the batholith occur on the west side of the Wood River Valley (Stop 2-5). In this area the batholith passively intruded country rocks of the Wood River basin, which are contact-metamorphosed over tens of square kilometers (Wavra, 1985). Hydrothermal fluids associated with Cretaceous magmatism are thought to have remobilized metals in carbonaceous Paleozoic rocks, forming the silver-lead-zinc mineral deposits of the Wood River Valley and Smoky Mountains (Hall, Rye, and Doe, 1978; Hall, 1985; Darling, 1988, this volume).

Paleogene Extension and the Challis Magmatic Episode

Paleogene crustal extension affected much of east-central Idaho and southwest Montana (Hait, 1984) and formed the Wildhorse detachment system between upper and lower plates in the Pioneer Mountains core complex (Wust, 1986a, b; Wust and Link, 1987; Wust and Link, 1988, this volume). Extension related to the Wildhorse detachment may have produced low-angle normal faults and reactivated the Wood River thrust in the Wood River Valley to the west (Skipp and others, 1986; Link and Mahoney, 1987; Turner and Otto, 1988, this volume).
Figure 3. Generalized stratigraphic columns showing Precambrian and Paleozoic formations and their thicknesses (in meters) that make up five Cordilleran thrust plates (Skipp, 1987, Fig. 2, this paper) in south-central Idaho in the area of the field trip. Vertical lines indicate a hiatus caused by either nondeposition, erosion, or faulting. A wavy line indicates a disconformity. Question marks indicate systems that may be present but are outside the field trip area, or are not exposed. Columns are revised from Figure 4 of Skipp and Hait (1977). References used to construct Figure 3 that postdate Skipp and Hait (1977) include Skipp, Hoggan, Schleicher, and Douglass (1979); Skipp, Sando, and Hall (1979); Hays and others (1980); Ingwell (1980); McCandless (1982); Davis (1983); and Hobbs (1985).
Figure 4. Geologic map and cross section of the southern Lost River, White Knob, and Pioneer Mountains, from Skipp and Hait (1977, plate 1). Three new interpretations of the structural relations are proposed in the present paper, and not shown on the map and cross section, which reflect concepts developed over 10 years ago. These changes are: a) Rocks shown as belonging to the White Knob plate on this map are now thought to comprise the Grouse plate and the White Knob plate. For location of boundaries of thrust plates see Figure 2. b) This map shows rocks of the Arco Hills east of Arco belonging to the Lost River-Arco Hills plate rather than the Grouse plate. The new interpretation is based on condont alteration index data, and is discussed in text at Stop 1-1. c) This map shows a Fish Creek thrust. This interpretation has been changed as discussed at in the text before Stop 1-7. Also, the cross section shown is not balanced.
Paleogene extension is also manifested in the trans-Challis fault system (Bennett, 1986) in the Challis 1 x 2 degree quadrangle (Fisher and others, 1983). This northeast-trending fault system is thought to have controlled the migration of ore-forming fluids and the locations of calderas in the northern Challis volcanic field. In the southern Challis volcanic field, eruptive centers may have been localized along northwest-trending normal faults (Moye and others, 1988, this volume).

Magmatism and deformation associated with the Challis magmatic episode (51-40 Ma) was probably coeval with some extensional tectonics and produced a large field of extrusive rocks (the Challis Volcanic Group, defined by Fisher and others, 1987) and several generations of plutons in south-central Idaho (McIntyre and others, 1982; Bennett and Knowles, 1985; Moye and others, 1988, this volume).

Eocene tectonism just preceding and coeval with Challis volcanism is documented by the basal Challis conglomerate (Paull, 1974). Recent work demonstrates that this unit represents a conformable transition from proximal alluvial fan sedimentation lacking any volcanic component to volcanic lithic sandstone and heterolithic tuff breccia. Fine-grained inter-beds both above and below the tuff breccia have yielded sparse palynologic assemblages (Burton and Blakley, 1988).

Basin and Range and Snake River Plain

Basin and Range faulting and Snake River Plain bimodal volcanism began in Idaho about 15 million years ago (Armstrong and others, 1975; Kuntz and others, 1986). The rhyolitic caldera-forming volcanism and locus of high-angle normal faulting have migrated northeastward toward Yellowstone Park at a rate of about 3 cm per year (Christiansen and McKee, 1978; Allmendinger, 1982). Basin and range extension has been oriented in a dominantly west-southwest to east-northeast direction in response to regional oblique stress, forming the present northwest-trending mountain ranges. As shown in Figure 5, in many places the basin and range faults juxtapose rocks of different Mesozoic thrust plates.

DAY ONE: GEOLOGY OF THE LOST RIVER, WHITE KNOB, AND SOUTHERN PIONEER MOUNTAINS

Roadlog from Arco to Antelope Creek

The field trip route is shown in Figures 1, 2, and 5, and mileage begins in Arco, Idaho, at the intersection of U. S. Highways 20-26 and 93. A geologic map of the area traversed on the first day of the trip is shown in Figure 4. Head north on Highway 93. Regional compilation geologic maps of the area are available in Rember and Bennett (1979a, b).

The Lost River Range lies to the east of the highway and is bounded by the historically active Lost River range-front fault system (Crone, 1988, this volume; Pierce, 1988, this volume). To the west of the highway are the foothills of Appendicitis Hill and the White Knob Mountains.

At mile 11.1 turn left (west) on Antelope Creek Road and stop on right. In the late 1800s Antelope Creek was part of the route used by miners and cattlemen to get from the Wood River Valley to the Big Lost River Valley on their way to mining areas along the Salmon River.

Stop 1-1: View of Lost River and White Knob Mountains From Mouth of Antelope Creek

Stratigraphy and Thrust Plates

This stop introduces the stratigraphy and structure of the Lost River Range. Stratigraphic differences between Upper Mississippian rocks in the Lost River Range to the east and the White Knob Mountains to the west suggest that they are parts of at least three separate thrust plates (from east to west: the Lost River-Arco Hills, White Knob, and Grouse plates, Figs. 2 and 3) (Skipp and Hait, 1977; Skipp, Sando, and Hall, 1979; Skipp, 1987). The White Knob plate (WK on Fig. 2) is structurally highest, and the Upper Mississippian rocks are White Knob Limestone consisting of limestone and interbedded chert- and quartzite-pebble conglomerate and sandstone. Upper Mississippian rocks of the underlying Grouse (G on Fig. 2) and Lost River-Arco Hills (LRAH on Fig. 2) plates are carbonate bank limestones with no interbedded clastics. Though Upper Mississippian rocks of the Grouse plate are similar to rocks of the Lost River-Arco Hills plate, they are thicker and have lower conodont color alteration index (CAI) values than those of the Lost River-Arco Hills plate. Rocks of the LRAH plate have CAI values of 5 to 5.5 (John Repetski, written communication, 1976), whereas rocks of the Grouse plate have values of 2 to 3.5 (Kirk Denkler, Anita Harris, and Bruce Wardlaw, written communications, 1982).

Extensional Faulting

The Lost River Range has been profoundly affected by Tertiary and Quaternary extension, including low-angle detachment faulting (Hait, 1984) (Steps 1-2 and 1-3) and Neogene to Holocene basin and range faulting. The west flank of the range is bounded by a moderately west-dipping normal fault, which last slipped in October 1983 and produced a magnitude 7.3 earthquake (Crone and Machette, 1984; Crone and others, 1985; Crone, 1988, this volume). Ongoing studies by Janecke at the
Figure 5. Diagrammatic map of eastern Idaho and adjacent Montana by M. H. Hait, Jr. showing Basin and Range normal faults and segments of normal faults above which are thrust plates.
University of Utah are aimed at defining the timing, kinematics and precise geometry of Tertiary extension in the Lost River Range.

**View from the Mouth of Antelope Creek**

The topographic low in the Lost River Range directly to the north marks Pass Creek, which runs north-south (Fig. 1). The view will be discussed in a clockwise manner, starting from the northwest. Figure 4 shows the geology of the area. Figure 6 shows Hait's interpretation of the geologic structure from Mt. McCaleb on the northwest to the Pass Creek and Elbow Canyon areas on the southeast. Figure 7 is a photographic panorama of the same area.

To the far northwest in the Lost River Range, the white peak is composed of Ordovician quartzite within the Dorah Peak horst (Fig. 5) (Ross, 1947, Baldwin, 1951, Skipp and Harding, 1985). Mississippian rocks are exposed on the crest of the range to the south, with the prominent blocky peak (Mt. McCaleb) underlain by resistant Mississippian Scott Peak Formation (Fig. 6, left side). Below Mt. McCaleb is orange talus of the Middle Canyon Formation, which overlies a smooth slope on the McGowan Creek Formation. Halfway down the mountain are juniper-covered slopes of the Devonian Three Forks and Jefferson Formations. Southeast (right) of Mt. McCaleb are complex north-northwest-trending folds in Mississippian and Pennsylvanian limestones, such as the tight syncline-anticline pair at Franklin Canyon (Fig. 6).

The dark, blocky cliffs northwest (left) of Pass Creek Canyon are east-dipping Eocene Challis Volcanics. Pass Creek will be the site of field trip Stops 1-2 and 1-3. A prominent landslide at the mouth of Crows Nest Canyon (Fig. 6) forms the fan-shaped landform below the volcanic cliffs.

The main structure of the Pass Creek area is the north-trending, west-dipping Pass Creek normal fault system (shown on Fig. 5), which here is composed of two low-angle normal faults cutting Paleozoic strata and Challis Volcanics (Fig. 6). In the Pass Creek canyon area Challis Volcanic Group strata are rotated eastward into these faults. Hait (1984) suggests that the extension structures in the Pass Creek area were originally continuous with even larger extension structures of the central Lemhi Range to the east (Hait, 1987). Those structures are overlapped by lower Miocene vertebrate-bearing beds that provide an upper age limit for the large-scale extension.

---

**Figure 6.** Sketch by M. H. Hait Jr. of panorama looking north and east to the Pass Creek normal fault system from the mouth of Antelope Creek (Stop 1-1). Units are as follows: Q--Quaternary; Tct--Challis tuffaceous units; Tea--Challis andesite (?); Ps--Snaky Canyon Formation (Penn.); PMb--Bluebird Mountain Formation, Msw--Surrey Canyon Formation; Mso--South Creek Formation; Msp--Scott Peak Formation; Mmm--Middle Canyon and McGowan Creek Formations; Dl--Three Forks Formation; Djb--Birdbear Member of Jefferson Formation; Dj--Jefferson Formation; Of--Fish Haven Formation and Laketown Dolomite; Ok--Kinnikinic Quartzite.

---

**Figure 7.** Panorama of the Pass Creek area from the mouth of Antelope Creek (Stop 1-1). Photo shows nearly the same area as diagrammed in Figure 6.
The upper fault of the Pass Creek system is exposed in Pass Creek Canyon (Stop 1-2). On this fault, the rocks west of the canyon have been moved westward from on top of the rocks to the southeast (right) of the canyon. The hanging wall of the upper normal fault contains upper Paleozoic, mostly Mississippian strata as well as Eocene Challis Volcanic Group and, perhaps, Tertiary gravels near Wet Creek (Mapel and Shropshire, 1973). The footwall of the lower fault contains Ordovician through Mississippian strata (Fig. 6). To the north, the footwall extends into Precambrian strata (Mapel and Shropshire, 1973). The structurally complex horse between the two faults is made up of Devonian through Mississippian strata.

King Mountain (Fig. 8) to the southeast of the mouth of Antelope Creek is composed of gently dipping Devonian Jefferson Formation (dolomite) at the base, overlain by thin Devonian Three Forks Formation, Lower Mississippian McGowan Creek Formation, orange talus of the sandy Upper Mississippian Middle Canyon Formation, and thick-bedded pure limestones of the Upper Mississippian Scott Peak Formation at the top.

On the south flank of King Mountain is Beaverland Pass (Fig. 8) which occupies a fault valley separating King Mountain (part of the Lost River-Arco Hills plate) from the Arco Hills block (part of the Grouse plate) to the south (Fig. 1). The part of the Arco Hills that lies south-southeast of Beaverland Pass is underlain by limestone and sandstone beds ranging in age from Late Mississippian to Late Pennsylvanian. These rocks have a cooler thermal overprint (CAI of 3 to 3.5) than rocks of the remainder of the Arco Hills farther south and east (CAI of 5 to 5.5). Skipp thus believes that this block has affinities with rocks of the Grouse thrust sheet to the west, which have similar low CAI values, rather than with the remainder of the Arco Hills (Fig. 2).

View to Southwest

To the southwest are the foothills of Appendicitis Hill. To the west and north are the White Knob Mountains, bisected by northeast-flowing Antelope Creek (Figs. 1 and 4). Appendicitis Hill belongs to the Grouse plate, and contains Upper Mississippian and Pennsylvanian carbonate bank facies rocks similar to those found in the Arco Hills (Skipp, Hoggan, Schleicher and Douglass, 1979; Davis, 1983). On the east margin of Appendicitis Hill the limestones are folded to tight east-vergent folds with several-hundred-meter wavelengths (Fig. 9). Lavas of the Challis Volcanic Group overlie the limestones and dip east about 30 degrees to form the brown flatirons on the east face of the hills. Hait believes that the eastward dip of the volcanic rocks suggests that they have been rotated to the east on the west-dipping Lost River range-front fault system.

Sheep Mountain, north of Antelope Creek, is composed of a thick, nearly flat-lying section of Challis Volcanic Group including silicic ash-flow tuffs, and dacitic and andesitic lavas (Moye and others, 1988, this volume).

In the distant northwest are peaks of the White Knob Mountains, underlain by Upper Mississippian White Knob Limestone which is intruded and bleached by the Eocene Mackay stock (Nelson and Ross, 1968, 1969).

Continue north on Highway 93. Pass through

Figure 8. King Mountain, looking east from Stop 1-1. King Mountain is made up of gently dipping Devonian Jefferson Formation (dolomite) at the base, overlain by thin Devonian Three Forks Formation, Lower Mississippian McGowan Creek Formation (bulk of tree-covered slope), orange talus of the sandy Upper Mississippian Middle Canyon Formation, and thick-bedded pure limestones of the Scott Peak Formation at the top. Beaverland Pass is on the right side of the photo.

Figure 9. Folded Pennsylvanian limestones of the Grouse plate on the east side of Appendicitis Hill, just southwest of the mouth of Antelope Creek. View is from Highway 93 about a mile south of Stop 1-1.
Darlington and Leslie and at mile 18.9, turn north (right) on a gravel road to Pass Creek. The turnoff is not obvious, but it is marked by a white trailer house southwest (left) of the road and a small brown sign to Pass Creek about 50 meters north of the highway.

Proceed north toward the Lost River Range. At mile 20.1 take the middle fork of three gravel roads, continuing northeast toward Pass Creek. Rocks to the northwest (left) are folded Upper Mississippian limestones overlain unconformably by east-dipping Challis Volcanic Group. A prominent landslide deposit, cut by a strand of the Lost River fault system, empties from Crows Nest Canyon in the Challis Volcanic Group north of the road (mile 22.0) (Fig. 6).

Preliminary results of work by Janecke suggest that angular unconformities may be present in the Challis Volcanic Group between Crows Nest Canyon and Pass Creek. If these unconformities are caused by syn-eruptive faulting, their dating may bracket the timing of Eocene extension.

Stop 1-2: Exposure of Upper Strand of Pass Creek Detachment Fault System in Pass Creek

Park on the right in small turnout at mile 23.9 just past the sign for entering the Challis National Forest. This location is on the Methodist Creek 7.5-minute quadrangle. Walk east across the diversion dam and up the slope to the southeast.

In a small gully just east of the limestone knob at elevation 6320 is an exposure (discovered by Janecke in 1987) of the upper strand of the Pass Creek normal fault system seen from Stop 1-1. Here the fault dips 35 degrees to the southwest. The fault has sections of shallow dip (15 degrees), separated by segments with steep dip (45 to 50 degrees). The normal fault here separates medium gray limestone of the Upper Mississippian Surrett Canyon (?) Formation above from darker, banded dolomites of the Devonian Jefferson Formation below. The fault is iron-stained and has two prominent sets of striae (trending 290° and 225° azimuth) on a slickenside surface (150, 35W.), demonstrating two episodes of westward transport. Upper plate rocks are locally brecciated but not disrupted on a large scale. Map-scale folds in the upper plate are truncated by the fault. The low-angle normal fault can be seen by climbing to the saddle at elevation 6440 and looking at the cliff to the north.

Return to the cars and proceed northward up the Pass Creek canyon road. The narrow canyon is cut in folded Mississippian limestones of the Surrett Canyon, South Creek, and Scott Peak Formations. The cliff forming unit is the Scott Peak Formation. North of Bluejay Canyon (mile 25.3), a prominent upright crest of an anticline can be seen straight ahead in the cliff.

Stop at corral on right at mile 26.2, just after leaving narrow canyon.

Stop 1-3: View of Syncline in Challis Volcanic Group Strata

As shown in Figure 10, east-dipping white tuffs of the Challis Volcanic Group are present on the west side of the basin. Challis beds in the sage-covered hills to the north (prominent tan-colored cliff) dip west. This suggests to Hait a syncline which formed over a ramp in an underlying detachment fault. To the east are cliffs of folded Mississippian limestone that comprise a “gravity glide” block overlying Tertiary strata (Mapel and Shropshire, 1973). Tertiary gravels near Wet Creek, exposed several kilometers to the northeast of this locality, dip eastward. Hait believes that these tuffs and younger gravels were rotated above a low-angle normal fault in the subsurface. An alternate interpretation of the map relations, proposed after preliminary work by

Figure 10. Sketch by M. H. Hait, Jr. of view looking north from Stop 1-3, Pass Creek narrows, Methodist Creek quadrangle. Unit designations are given in caption for Figure 6.
Janecke, is that the tilt may have been produced by north-trending high-angle normal faults.

The area to the north was mapped by Mapel and Shropshire (1973), who defined a general stratigraphic succession in the Challis Volcanic Group. Basal andesite and basalt are overlain by a middle interval of sediments and tuffs, with an upper unit of basalts.

Return to cars, turn around, and proceed back to Arco. Reset odometers at intersection of Highways 93 and 20-26 in Arco, and proceed west. Stop for lunch at Atoms for Peace Park on the left at mile 0.4.

Stop 1-4: View of Upper Mississippian Carbonate Bank of the Arco Hills

On warm spring evenings, juniors from Arco High School steal up the steep cliffs of the Surrett Canyon Formation at the south end of the Arco Hills (visible directly to the east) and paint their graduation year on the Upper Mississippian bioclastic grainstones of the Surrett Canyon Formation (Fig. 11).

All of the units present in the Arco Hills (from oldest to youngest: Middle Canyon, Scott Peak, South Creek, Surrett Canyon, Arco Hills and Bluebird Mountain Formations) (Fig. 3) are of the Upper Mississippian carbonate facies (Skipp, Sando, and Hall, 1979). In the Arco Hills, the Upper Mississippian Surrett Canyon Formation is overlain by the Arco Hills Formation (type section at south end of outcrop east of town) and the Mississippian/Pennsylvanian Bluebird Mountain Formation (Skipp, Sando, and Hall, 1979). The letter "B" for Butte County is in the sandstones of the Bluebird Mountain Formation.

Roadlog from Arco to Blizzard Mountain Road

Proceed west on Highway 26 from Arco.

At mile 2.3 the road climbs onto a terrace of the Big Lost River. The Arco airport is on the left, and to the north is Appendicitis Hill. To the south is Big Southern Butte, a compound rhyolite dome that intruded and tilted basalt lavas along the axis of the Snake River Plain about 300,000 years ago. The Butte stands about 600 meters above the surrounding basalts.

At mile 7.6, by an irrigation storage pond on the north side of the road, is a view to the northwest of tree-topped Timbered Dome. The dome is capped by locally metaliferous jasperoid, but the flanks are made up of a complete section of Upper Devonian through Upper Pennsylvanian rocks, totalling more than 2960 meters in thickness. A reference section for the international mid-Carboniferous stratotype boundary is located there (Skipp and others, 1985).

Appendicitis Hill is east of Timbered Dome; both are on the Grouse plate. Between them on the skyline ridge of the White Knob Mountains is the craggy tree-covered summit of Sheep Mountain, composed of the Challis Volcanic Group. The sharp, bare ridge west of Sheep Mountain is underlain by Mississippian McGowan Creek Formation and White Knob Limestone of the White Knob Plate. The sharp peak with broad shoulders to the west of the White Knob Mountains is Smiley Peak in the Pioneer Mountains, underlain by Challis Volcanics. The Pioneer Mountains lie to the west and are underlain dominantly by folded Mississippian Copper Basin Formation of the Copper Basin plate.

Rocks of the Grouse thrust plate make up the mountains between Arco and this stop (Figs. 2 and 3) (Skipp, 1987). The Grouse plate contains the Upper Mississippian carbonate bank and overlying Pennsylvanian/Permian carbonate bank. Within the Grouse plate, the Mississippian-Pennsylvanian carbonate rocks stratigraphically overlie the Lower Mississippian McGowan Creek Formation, an eastern, finer-grained facies of the Antler flysch.

At mile 12.2 the road crests a hill just beyond an abandoned log cabin on the north. Straight ahead to the southwest are the basaltic cones of Craters of the Moon lava field (Stephens, 1988, this volume). To the north is an outcrop of folded Ordovician quartzites and dolomites (Summerhouse (?) Formation) surrounded by basalt. This possibly is a block along the down-dropped margin of the Snake River Plain. Pull off at mile 16.6 at the Blizzard Mountain Road to the north.

Stop 1-5: View of Copper Basin Thrust at Blizzard Mountain

Blizzard Mountain is the prominent peak to the
northwest (left) (Fig. 4). Figure 12 is a geologic map of the area. Directly north just below the old ski lift tower is the nearly vertical Copper Basin thrust that separates the upper part of the Copper Basin Formation (Copper Basin plate) on the west from the underlying folded Middle Canyon and McGowan Creek Formations (Grouse plate). Thus, at this locality, coarse-grained Antler-derived clastics were thrust to the east over finer-grained Antler clastics and overlying limestones of the Upper Mississippian carbonate bank (Skipp, 1987).

Above the Copper Basin thrust is the Copper Basin plate, which forms the mountains north of the highway from here west to Fish Creek Reservoir, and which contains the main part of the Antler flysch. The Copper Basin thrust plate is divided into two subplates: an upper Glide Mountain subplate and a lower Copper Basin subplate (Dover, 1981). Nilsen (1977) referred to these subplates as the Brockie and Scorpion subplates, respectively. The Copper Basin plate contains thick Mississippian strata of the Copper Basin Formation underlain by thin Devonian outer shelf facies rocks (Fig. 3, Stop 1-8). The lower part of the Copper Basin Formation is a coarse-grained detrital sequence of turbidite deposits that constitute proximal and distal submarine fan deposits derived from a western Antler highland (Nilsen, 1977; Paull and others, 1972).

Just north of here is Lava Creek, where basaltic vents breach the Paleozoic rocks (Fig. 12). These vents lie on a northwest continuation of the Great Rift, a northwest-trending 85-km-long volcanic rift zone in the Snake River Plain which extends south to Craters of the Moon National Monument and the Kings Bowl area west of American Falls (Fig. 1) (Kuntz and others, 1982; 1986; Skipp and Kunz, unpublished mapping in the SE 1/4 Grouse 15-minute quadrangle). Radiocarbon dates of organic material between basalt flows along Lava Creek provide the following ages: basalt flow of Lava Creek (Qbl) 12.7 ± 0.15 Ka, and basalt flow of Sunset crater (Qbs) 12.0 ± 0.15 Ka (Kuntz and others, 1986).

Just north of Lava Creek, the Copper Basin thrust is offset to the northeast along a tear fault or lateral ramp. Return to Highway 26, heading west. The entrance to Craters of the Moon National Monument (mile 19.0) is on the left. Volcanism at Craters of the Moon occurred along the Great Rift. The youngest flows in the monument are 2 Ka (Kuntz and others, 1982; 1986; Stephens, 1988, this volume).

Continue west to mile 26.5; pull off on right, and disembark carefully on right side. WARNING: this is a blind curve; watch carefully for vehicles.

Stop 1-6: Turbidites of the Mississippian Copper Basin Formation

Rocks exposed in this roadcut are part of the Glide Mountain subplate of Dover (1981) (Fig. 4). The Glide Mountain subplate contains coarse-grained turbidites, deposited in the Antler flysch trough proximal to the western Antler highland. Bedding in the Copper Basin Formation here dips steeply eastward. Rocks are siltstone, fine to coarse sandstone, and both matrix-supported and clast-supported conglomerates displaying complete and partial Bouma sequences (Fig. 13). This outcrop contains most of the structures of the ideal Bouma turbidite sequence shown in the inset of Figure 13. Sedimentary structures visible here include normal and reverse graded bedding (on far western end), load casts, cross beds, cross and parallel laminations, ripple marks, soft-sediment folds, and floating mudstone clasts in matrix-supported conglomerate. *Nereites*-ichnofacies trace fossils or bedding trails are found in some of the mudstones. Their presence indicates a deep water, nonturbulent environment between episodes of coarse turbidite sedimentation.

The basal 10 meters of outcrop displays a coarsening-and thickening-upward cycle interpreted to represent a depositional lobe of a submarine fan. Two fining- and thinning-upward cycles are present in the upper 30 meters of outcrop. These contain matrix-supported conglomerates with soft-sediment clasts and are interpreted to represent submarine fan channel deposits.

Continue west on Highway 26. At mile 34.3, the fault between the overlying Milligen-Wood River plate and underlying Copper Basin plate (WR on Fig. 4) is exposed in the saddle just off the road to the north. Here the Pennsylvanian-Permian Wood River Formation and underlying Devonian Milligen Formation are juxtaposed against Mississippian turbidites of the Copper Basin Formation. Mississippian rocks are not present in the Milligen-Wood River thrust plate, and Devonian rocks of the thrust plate are thought to have been part of the Antler highland during Early Mississippian time.

Reset odometers at the Fish Creek Reservoir Road (mile 36.3). Turn right (north) to Fish Creek Reservoir. The road crosses a Pleistocene basalt flow erupted from a vent just south of Fish Creek Reservoir.

Geologic Setting of Fish Creek Reservoir

The Fish Creek Reservoir area is geologically important because it exposes the stratigraphically lowest rocks of the Copper Basin plate, several thrust faults, and may contain Carlin-type silver and gold mineralization. In the sagebrush-covered hills to the east, the Pennsylvanian-Permian Wood River Formation is thrust over the Copper Basin Formation. The hills to the west contain the Wood River Formation faulted over the Devonian Milligen Formation.

The Fish Creek area was formerly thought to contain a structural window to rocks below the Copper Basin plate (Skipp and Hall, 1975); the map of Figure 4 shows this interpretation. A new interpretation of the geology is presented in Fig. 14, a geologic sketch map and cross sections of the Fish Creek Reservoir area, revised from Skipp and Hall (1975) by Betty Skipp using modern...
Figure 12. Geologic sketch map of southeastern part of the Grouse 15-minute quadrangle showing the Copper Basin thrust in relation to Blizzard Mountain ski lift tower and locations of vent craters and cinder cones in Lava Creek at the northern end of the Great Rift volcanic rift zone. Geology is by Betty Skipp, M.A. Kuntz, and L.A. Morgan. Map units: Qac--Alluvium, colluvium, landslide deposits, some basaltic ash (Holocene to Pleistocene); Qbs--Basalt flow of Sunset crater (latest Pleistocene); Qbm--Basalt cone (Pleistocene); Qbd--Basalt flow of Dry Creek (Pleistocene); Qbl--Basalt flow of Lava Creek (Pleistocene); Qb--Unnamed basalt flow (Pleistocene); Th--Rhyolitic ash-flow tuffs of Heise Group (Pliocene and Miocene); Td--Rhyolite dike (Eocene); Ts--Quartz monzonite stock (Eocene); Tc--Challis Volcanic Group (Eocene); Msu--Surrett Canyon Formation (Upper Mississippian); Ms--Scott Peak Formation (Upper Mississippian); Mm--Middle Canyon Formation (Upper Mississippian); Mmg--McGowan Creek Formation (Lower Mississippian); Mc--Copper Basin Formation (Mississippian); Os--Summerhouse(?) Formation (Ordovician).
thrust belt structural concepts (Dahlstrom, 1970; Royse and others, 1975).

In the new map interpretation (Fig. 14), the concept of the Fish Creek thrust (FC on Fig. 4), which placed Mississippian Copper Basin Formation on older rocks (Skipp and Hall, 1975), is abandoned; faults formerly attributed to the Fish Creek thrust are reinterpreted to be normal faults. The contact between the uppermost Devonian Picabo Formation and the Mississippian Copper Basin Formation, formerly thought to be part of the Fish Creek thrust, is redefined as a disconformable sedimentary contact within the Copper Basin plate. New mapping southeast of the reservoir suggests that large segments of the Wood River thrust that place Wood River Formation against the Copper Basin Formation (Fig. 4 of Skipp and Hall, 1975) are also normal faults. In addition, siltstone, argillite, and quartzite formerly assigned to the uppermost Wood River Formation on the west side of Fish Creek are now thought to correlate with the Milligen Formation, although no age has yet been established for these beds.

At mile 3.0 from the highway cross Fish Creek. At mile 3.8 take a dirt road on the right for 0.2 mile to Stop 1-7.

Stop 1-7: Lower Part of the Wood River Formation

This stop involves a short climb through an unusually coarse eastern facies of the lower part of the Wood River Formation which was deposited proximal to the Copper Basin highland source area (Skipp and Hall, 1975) (Fig. 15). The rocks are sandy, chert pebble-bearing, conglomeratic limestone and purple siltstone. In exposures of the Wood River Formation in the Wood River Valley to the west, such conglomerate-bearing strata are confined mainly to the basal Hailey Conglomerate Member (Fig. 19). Here, however, they are present through the entire exposed section. The bulk of the clasts in the Wood River Formation here were derived from the Mississippian Copper Basin Formation of the Copper Basin plate, which now forms the hills to the east.

Return to main Fish Creek Road (mile 4.2) and continue north. The basalt vent and road to Fish Creek Dam are on the left at mile 4.4. Continue to mile 5.0, turn left on road to the reservoir and park.

Stop 1-8: Silurian-Devonian Roberts Mountains Formation Thrust Over Devonian Carey Dolomite

This stop affords discussion of the structural setting and mineral deposits of the Fish Creek Reservoir area and illustrates outer carbonate platform sedimentary facies of the Silurian-Devonian Roberts Mountains Formation. Hike east from the parking spot, across a thrust fault within the Copper Basin plate that places Upper Silurian and Lower Devonian Roberts Mountains Formation over Middle Devonian Carey Dolomite (Fig. 14).

The Roberts Mountains Formation here (Fig. 16) was first studied by Skipp and Sandberg (1975). It contains stromatoporoid boundstone, laminated siltstone and rip-up clast conglomerate deposited in proximal and marginal carbonate buildup environments.

Figure 17 is an annotated photograph of the hill west of Fish Creek Reservoir, modified from Figure 3 of Skipp and Hall (1975) and taken near this locality. The photograph shows the Milligen-Wood River plate thrust onto the Copper Basin plate which contains the Copper Basin Formation and underlying Devonian strata.
Figure 14. Geologic sketch map and cross sections (no vertical exaggeration) of the Fish Creek Reservoir area revised from Skipp and Hall (1975, Fig. 4). Locations of field trip Stops 1-7 and 1-8 are shown. Map units: QTa--Alluvium, landslide deposits, colluvium, terrace deposits, and jasperoid undifferentiated (Quaternary and Tertiary); Qy--Younger basalt (Quaternary); Qs--Snake River Group (Quaternary); Tc--Challis Volcanic Group (Eocene); P/Pw--Wood River Formation (Permian and Pennsylvanian); MC--Copper Basin Formation (Mississippian); Dp--Picabo Formation (Upper Devonian); Dj--Jefferson Formation (Upper and Middle Devonian); Dc--Carey Dolomite (Middle and Lower Devonian); Dm--Milligen Formation (Upper and Middle Devonian); Dm?--Milligen (?) Formation (Upper and Middle Devonian?); DSR--Roberts Mountains Formation (Lower Devonian and Upper Silurian); SOR--Silurian and Ordovician rocks undivided-unit used only on cross section A-A'.

Sea Level
Guidebook to the Geology of Central and Southern Idaho

Figure 15. Measured section of lower part of Wood River Formation in NW 1/4 sec. 23 and NE 1/4 sec. 15 (approx.), T. 1 N., R. 22 E., Blaine County, Idaho. From Skipp and Hall (1975, Fig. 8).

Figure 16. Diagrammatic (partly restored) section of the Roberts Mountains Formation showing conodont zonation of faulted sequence along ridge on east side of Fish Creek Reservoir in NW 1/4 sec. 14 (unsurveyed), T. 1 N., R. 22 E. From Skipp and Sandberg (1975, Fig. 3).

Mineral Deposits at Fish Creek Reservoir

The major normal fault that drops Copper Basin Formation down against thrust-faulted Silurian and Devonian rocks (Fig. 14) can be observed on the ridge to the southeast. The normal fault recently has been prospected for metal deposits.

Mineral exploration in the Fish Creek area was initiated in 1977 as a result of publications by Skipp and Hall (1975) and Skipp and Sandberg (1975). Wayne Hall and Betty Skipp recognized the similarity of the structural and stratigraphic setting of the Fish Creek area to that of the Carlin gold deposit in Nevada. Cordex, founded by John Livermore, a co-discoverer of the Carlin ore body, was the first company to test the prospect in the Fish Creek area. Samples from three rotary holes drilled just east of the fault that drops Copper Basin Formation down against Silurian-Devonian carbonate rocks yielded as much as 0.61 ppm gold and 11 ppm silver. More exploratory drilling for deeper targets has
been proposed for the area by James D. Loghry, a consultant from Tucson, Arizona, who kindly furnished the historical and sample information on the prospect.

Return to the intersection of Fish Creek Road and Highway 26-93. Reset odometers to zero. Continue west toward Carey.

Roadlog from Fish Creek Road to Hailey

Between the Fish Creek Reservoir Road and Carey, Quaternary Snake River Group basalts are south of the road. North of the road, the lower hills are composed of Eocene Challis Volcanic Group. The conspicuous southward-dipping slope that caps these hills is formed on Miocene rhyolite ash-flows of the Idavada Volcanics. These silicic pyroclastic rocks were erupted from multiple vents on the Snake River Plain (F. J. Moye and W. P. Leeman, unpublished mapping, 1987). The term Idavada Volcanics (Malde and Powers, 1962) is used loosely for Miocene rhyolitic and basaltic units on the central Snake River Plain (Bonnichsen, 1982). For more discussion of these rocks see Bonnichsen and others (1988, this volume).

At mile 3.6, the view to the northwest is of the Little Wood River Valley. The mountains on the skyline to the northwest are underlain by Pennsylvanian-Permian Wood River Formation and Devonian Milligen Formation. Eocene Challis Volcanic Group and Miocene Idavada Volcanics underlie much of the middle ground. To the west is the Queen’s Crown, capped by a rhyolite ash-flow tuff of the Idavada Volcanics. This ash flow dips southward, toward the Snake River Plain, and is truncated on the north by a normal fault.

The road to Little Wood River Reservoir intersects the highway at mile 6.8. Pass through Carey and at mile 7.8 turn right (west) on U. S. Highways 20 and 26 toward Fairfield and Sun Valley.

The highway bends to the southwest at mile 9.0 and ascends the Queen’s Crown. The Idavada Volcanics in this area are being studied by W. P. Leeman (Rice University), F. J. Moye and W. R. Hackett (Idaho State University) and include several composite sheets from several vents.

The summit is reached at mile 9.9. From here the road descends through more volcanic rocks, including white tuff of the Idavada Volcanics and Pleistocene olivine basalt of the Bellevue Formation of Schmidt (1961).

At mile 12.4 cross Silver Creek; this spring-fed creek is an internationally famous trout stream. At mile 14.3 enter Picabo. The Challis Volcanic Group forms the main part of the hills to the north (right). To the south, the Timmerman Hills are composed of Wood River Formation intruded by Cretaceous rocks of the Idaho batholith and overlain by Idavada Volcanics. To the west of the valley of the Wood River are the Smoky Mountains, which are composed of Pennsylvanian-Permian strata of the Wood River and Dollarhide Formations intruded by Cretaceous and Tertiary granodiorites, and over lain in places by the Challis Volcanic Group.

At mile 16.6 cross Silver Creek again. At mile 17.7 turn right (north) on county road toward Gannett. The hills to the north are composed of Wood River Formation overlain by andesite and tuff of the Challis Volcanic Group and Idavada Volcanics.

Pass through the town of Gannett (mile 21.6). At mile 25.2, a black-colored tailings pile marks the mine dump of a silver prospect pit in Devonian Milligen Formation. The Wood River thrust, separating the Milligen from overlying Wood River Formation, is about 100 m above the valley floor in the hills to the north.
DAY TWO: GEOLOGY OF THE WOOD RIVER VALLEY

Introduction

Figure 18 is a geologic map and tectonostratigraphic diagram of the Wood River Valley area. Field trip stops focus on three geologic problems:

1. Distinguishing the carbonaceous siltstones and limestones of the Devonian Milligen Formation (Stops 2-2 and 2-4) from similar strata of the Permian Dollarhide Formation (Stop 2-5) is important for two reasons. The black mudrocks in question are host to silver-lead mineralization near Bellevue and Hailey, and the distinction is vital toward reconstructing late Paleozoic facies of the Wood River basin and the geometry of Mesozoic and Paleogene faults.

2. Reconstruction of basin geometry and stratigraphy of the late Paleozoic Wood River basin, its relation to the continental margin, and other coeval basins to the south and east (Stops 2-3, 2-4, 2-7): Figure 19 shows the stratigraphy of the three tectonostratigraphic units containing rocks of the Wood River basin.

3. Faults that place upper Paleozoic rocks of the Wood River basin on older strata have been called the Wood River thrust by previous workers (Hall, Batchelder, and Tschanz, 1978; Dover, 1983). In places they retain what is thought to be their Mesozoic geometry (Stop 2-3), but elsewhere they have the geometry of Tertiary normal faults (Stops 2-2, 2-4, 2-7, 2-8).

The normal faulting may have been related to movement on the Wildhorse detachment system of the Pioneer Mountains core complex (Stop 2-6), denuda-

tion above the rising Boulder Mountains magmatic center (Stop 2-9), or Basin and Range extension.

Dollarhide-Milligen Problem

Black, carbonaceous and metalliferous limestones and mudstones of the Wood River Valley area have traditionally been mapped as Devonian Milligen Formation (Umplesby and others, 1930; Anderson and others, 1950). Sandberg and others (1975) determined that the type Milligen Formation southeast of Ketchum was Devonian in age, and restricted the name Milligen to Devonian age rocks. However, much of the exposed carbonaceous strata near Hailey and Bellevue are sparsely fossiliferous and despite several efforts have not yielded definitive fossil assemblages.

Dark-colored mudrocks west of the Wood River were all mapped as Milligen Formation until 1983, when Wayne Hall and co-workers discovered Permian fusilinids in carbonaceous limestones near the headwaters of Deer Creek, about 6 km west of Stop 2-5 (Hall, 1985; Wavra, 1985). Most dark mudstones west of the Wood River were redesignated as Permian Dollarhide Formation (Hall, 1985).

As seen at Stop 2-2, rocks of the Milligen Formation commonly show penetrative cleavage (Turner and Otto, 1988, this volume), while rocks of the Dollarhide Formation (Stop 2-5) lack a penetrative fabric. However, this criterion for distinguishing the formations has not proven applicable at all locations.

The geologic compilation map of Hall (1985) shows an east-dipping thrust fault with Milligen Formation over Dollarhide Formation, west of the Wood River between Hailey and Bellevue. However, detailed mapping of this contact by M. E. Ratchford reveals the relations shown in Figures 18 and 26, with a west-dipping fault placing Dollarhide above Milligen. The interpretation shown in Figures 26 and 27 implies that this contact is a folded Mesozoic thrust. North of Colorado Gulch (Stop 2-2, Fig. 26), a low-angle normal fault cuts both the Wood River thrust and the Milligen-Dollarhide thrust, and it locally places Dollarhide Formation topographically above the Milligen Formation. Major questions remain about the extent of the Dollarhide Formation and its structural relationship to both the Wood River and Milligen Formations.

Wood River Basin

The Wood River basin developed on the site of the former Antler highland (Devonian Milligen Formation) and was filled with several thousand meters of mixed siliciclastic-carbonate sediment. The Wood River basin was coeval with similar depocenters to the south (Oquirrh, Sublett, and Cassia basins). These epicratonic basins are thought to have formed during transtensional faulting of the Ancestral Rockies orogeny (Kluth, 1986). Wavra and others (1988) propose that the Wood River
Figure 18. Simplified geologic map, correlation chart, and tectonostratigraphic diagram of the central Idaho black shale mineral belt, modified from Hall (1985) based on new mapping by P. K. Link, B. R. Burton, and M. E. Ratchford. Locations of photograph (Fig. 21), cross sections (Figs. 26, 27, 31, 32), and geologic maps (Figs. 28 and 30) are shown in italic numbers.
"basin" was instead a west-facing continental margin. An alternate interpretation, that the Wood River basin had complex paleoslopes, is advocated in this paper.

Strata deposited in the Wood River basin now compose three thrust-bounded tectonostratigraphic units: the Wood River, Dollarhide and Grand Prize Formations (Hall, 1985; Link and others, 1987) (Fig. 19).

The Wood River basin received sediment from three sources: (1) The basal Hailey Conglomerate Member of the Wood River Formation had a nearby northeastern source area consisting of uplifted rocks of the Antler flysch trough (Mississippian Copper Basin Formation) and locally, rocks of the underlying Milligen Formation. (2) The bulk of the quartzose sand was derived from the North American craton, probably from uplifts of the Ancestral Rockies system. 3) The carbonate mud found in the Wood River and Dollarhide Formations may have been derived from the carbonate bank now represented by the Pennsylvanian Snaky Canyon Formation of the Lost River-Arco Hills, White Knob, and Grouse plates.

Wood River Formation

The Wood River Formation in the type area east of Bellevue contains at least 3000 meters of strata (Fig. 19), with a formally defined basal member (the Hailey Conglomerate Member) and overlying informal units 2 through 7 (Hall and others, 1974, Goodman, 1983). Siltstones at the top of the formation were proposed to belong to unit 8 (Hall, Rye, and Doe, 1978; Hall, Batchelder, and Tschanz, 1978), but recent work by Burton has demonstrated that unit 8 is unmappable because it cannot be distinguished from unit 7.

The contact between the Hailey Conglomerate Member and underlying Devonian Milligen Formation has been mapped as the Wood River thrust. Our work shows that this contact is both a Mesozoic thrust and a Tertiary normal fault (Fig. 18). In places (Fig. 30) the Hailey Conglomerate Member contains lithoclasts of the subjacent Devonian Milligen Formation, and the contact is interpreted as an unconformity.

Figure 19. Stratigraphic columns for tectonostratigraphic units of the Wood River basin. Sources of stratigraphic information include: Hall and others (1974), Sengenbush (1984), Wavra and others (1986), Geslin (1986), Mahoney (1987) and B. R. Burton (unpublished). Figure is modified from Link and others (1987).
The Hailey Conglomerate Member is thought to be deposits of a braided or fan delta complex that carried clasts from the uplifted and lithified Copper Basin Formation to the northeast (Winsor, 1981). The Hailey Conglomerate Member coarsens and thickens northeastward toward its source area and contains southwest-directed paleocurrents. In the type area of the Hailey Conglomerate Member (Stop 2-3), conglomerate is only present in the basal few tens of meters. In easternmost exposures of the lower Wood River Formation (Stop 1-7 at Fish Creek Reservoir), conglomeratic debris is present through 200 m of section (Fig. 15).

The Hailey Conglomerate Member contains thin interbeds of coralline limestone and is overlain by fossiliferous Wood River Formation unit 2 (Fig. 19), which contains Middle Pennsylvanian macrofossils that lived on a substrate of conglomerate debris. The Hailey Member and unit 2 are mapped as the lower part of the Wood River Formation on the geologic map of the eastern Boulder Mountains (Fig. 30).

The middle part of the Wood River Formation (units 3 through 6, Late Pennsylvanian to Early Permian) consists of light brown to gray, calcareous (locally siliceous) sandstone and sandy limestone deposited on a carbonate ramp, but dominantly above storm wave base. Unit 3, a sub-wave-base siltstone, abruptly overlies shallow-water bioclastic unit 2, suggesting a rapid deepening of the basin.

Where brittle siliceous sandstone of unit 5 is not present, units 4 and 6 cannot be separated and all rocks above unit 2 and below dark siltstone of unit 7 are mapped as middle part, Wood River Formation (Fig. 30).

The upper part of the Wood River Formation (unit 7) contains distinctive thin-bedded, parallel and convolute laminated, silty and sandy limestone and dolomite, plus black shale containing abundant trace fossils. These lithologies suggest deposition from periodic silty turbidity currents (Burton, 1988).

Dollarhide Formation

The Dollarhide Formation was defined by Hall (1985) for carbonaceous limestone, siltite, quartzite and mudstone near Dollarhide Summit in the Smoky Mountains (Fig. 18). More detailed studies by Wavra (1985), Wavra and others (1986), and Geslin (1986) show the Dollarhide to be about 2300 m thick.

Two members (Fig. 19) are mappable: a lower member of sandy limestones and sandstones and an upper member containing black mudstone and argillite with varying amounts of sandy limestone, calcareous sandstone, and conglomerate. The lower member contains abundant soft-sediment folds and was deposited on a slope. The upper member is thought to be a basinal mudstone (Wavra and others, 1986; Geslin, 1986). The upper member contains much carbonaceous matter and strongly resembles rocks mapped as Devonian Milligen Formation east of the Wood River (Link and others, 1987). The upper member hosts silver-lead mineralization west of the Wood River at the Silver Star Queen and Minnie Moore mines near Bellevue (Hall, 1985) and in the Carrietown mining district west of Ketchum (Darling, 1988, this volume).

Grand Prize Formation

The Grand Prize Formation was named by Hall (1985) for a thick sequence of quartzites and banded siltites exposed along Pole Creek, north of Galena Summit (Fig. 18). The Grand Prize Formation includes much of what was mapped as upper Wood River Formation by Tschanz and others (1986) and all of the Pole Creek formation of Fisher and others (1983). The age of the Grand Prize Formation is poorly constrained as Early Permian (Leonardian) but the rocks could be as old as Pennsylvanian (Wolfcampian). These age assignments are based on stretched and corroded conodonts from the base of the section (Hall, 1985).

The Grand Prize Formation is faulted above the early Paleozoic Salmon River assemblage, and also lies in low-angle thrust contact above the Wood River Formation (Hall, 1985). The Grand Prize Formation is nowhere in contact with the Devonian Milligen Formation.

Mahoney and Sengebush (1988, this volume) argue that the Grand Prize Formation contains deposits of subaqueous mass-gravity flows, and suggest that parts of the unit were deposited in a submarine fan setting. The Grand Prize Formation contains less carbonate mud than the Wood River and Dollarhide Formations and must have accumulated in a part of the basin that was isolated from carbonate input.

Correlation Within Wood River Basin Strata

Each formation of the Wood River basin is defined as a tectonostratigraphic unit, bounded above and below by faults (Fig. 18) (Hall, 1985). There are several striking lithologic similarities among the units. The basal conglomerate of the Grand Prize Formation strongly resembles the Hailey Conglomerate Member of the Wood River Formation. The Lower Permian portions of all three formations indicate deepening sedimentary environments. The upper member of the Grand Prize Formation is similar to both the upper member of the Dollarhide Formation and to unit 7 of the Wood River Formation.

Southwestward paleocurrents and northeastward-coarsening of clasts in the Hailey Conglomerate Member suggest that during Middle Pennsylvanian time the Wood River basin lay southwest of a highland that contained rocks of the Copper Basin Formation. Shelf-facies deposits of the lower and middle Wood River Formation may have passed southwestward to deeper water deposits now represented by the Dollarhide Formation (Wavra and others, 1986; Mahoney and Link, 1987a, b; Mahoney and others, 1987; Link and others, 1987).
Wood River Formation unit 7 (Lower Permian) in the Boulder Mountains contains abundant soft-sediment folds that verge southeast (Burton, 1988; Burton and Link, 1988), indicating a southeast palaeoslope in Early Permian time. Unit 7 in the southwestern allochthon of the Boulder Mountains may represent the western edge of an epicratonic Wood River basin. The documentation of a southeastward palaeoslope in unit 7 suggests that the upper Wood River Formation was deposited in a basin with complex and multidirectional palaeoslope.

Time correlations and a rigorous basin analysis for these rocks await biostratigraphic resolution.

Wood River Thrust

In the central Idaho black shale mineral belt of the Pioneer, Boulder and Smoky Mountains, faults of regional extent place upper Paleozoic rocks of the Wood River basin over older, metamorphosed, lower Paleozoic strata (Milligen Formation and Salmon River assemblage). These faults have been grouped under the term "Wood River thrust" (Hall, Batchelder, and Tschanz, 1978; Dover, 1983). In places this fault system is folded into eastward-overturned folds and is intruded by Cretaceous rocks. These localities include the area west of Hailey (Stops 2-1 to 2-3) (Figs. 26 and 27) and the White Cloud Mountains north of the map of Figure 18 (Tschanz and others, 1986; Sengebush, 1984). Other characteristics thought to represent parts of the thrust preserving Mesozoic geometry include the development of a gouge zone (Stop 2-8 and Fig. 20) and the presence of a full section of Wood River Formation above the fault, as in the type area east of Bellevue.

In large areas of the Wood River Valley the Wood River thrust can be recognized only as pre-Challis Volcanic Group in age. In many places the thrust appears to have been cut by later normal faults (Figs. 27 and 31) (Stops 2-2, 2-4) (Skipp and others, 1986; Otto and Turner, 1987; Turner and Otto, 1988, this volume). Among the criteria used to infer Tertiary normal faults are: attenuation and boudinage of the basal Hailey Conglomerate Member and overlying units 2 and 3 of the Wood River Formation above the fault (Fig. 21) (Stops 2-2, 2-4, 2-8), and the presence of breccia zones and smooth slickensided surfaces along the fault (Stop 2-4).

Roadlog from Ketchum to Hailey

Mileage starts at the intersection of Sun Valley Road and State Highway 75 at the stoplight in downtown Ketchum. Figure 18 is a geologic map of the Wood River Valley. Proceeding south on Highway 75 from Ketchum, we travel through the southward-widening Wood River graben filled with Quaternary and Holocene alluvial deposits of the Wood River. To the west is Bald Mountain (Fig. 24) with the main Sun Valley ski area. Bald Mountain is underlain by folded rocks of the upper part of the Wood River Formation. To the east are dacite lavas of the Eocene Challis Volcanic Group (Fig. 23). This area has been mapped by Hall, Batchelder and Tschanz (1978) and Batchelder and Hall (1978).

At the bridge over the Wood River (mile 2.3) rocks to the east are mapped as rhyodacite of the Challis Volcanic Group. Thick-bedded sandstone of Wood River Formation unit 6 lies to the west. At Cold Springs Road (mile 3.6) the unconformable contact of the Challis Volcanic Group over Wood River Formation is visible in the gully directly to the east.

Proceed south to Hailey (main intersection is at mile 12.1). Where the road bends to the east (mile 12.6), remain on South Main Street by bearing right across from the Blaine County Hospital. Pull off at mile 13.2 for a view of the Hailey area.

Stop 2-1: View of Hailey Area and Discussion of Wood River Thrust

To the west lie the tree-lined course of the Wood River and steep-sided hills on the west side of the Wood River graben (Figs. 22 and 25). To the southwest is Colorado Gulch, which contains the critical contact between the Devonian Milligen and Permian Dollarhide Formations. The hill directly west is Della Mountain (Fig. 26). The lower slopes of Della Mountain are underlain by siltstone, sandstone, and phyllite of the Milligen Formation, and the blocky outcrops along the summit ridge are the basal Hailey Conglomerate Member of the Wood River Formation. The Hailey Conglomerate Member overlies the Milligen Formation on the Wood River thrust, which is flat-lying on the summit of Della Mountain and is offset down to the south at the prominent bench on the south flank of the mountain. This bench is thought by Link to be the topographic expression of a low-angle normal fault which cuts the Wood River thrust and which will be visited at Stop 2-2.

To the northwest is the valley of Croy Creek and Carbonate Mountain north of the creek. As shown in the cross section of Carbonate Mountain (Fig. 27), the Wood River thrust has been folded into an eastward-overturned antiform, and the Hailey Member dips steeply westward, overturned. The type area for the Hailey Conglomerate Member is at the base of Carbonate Mountain and will be visited at Stop 2-3. The Milligen Formation underlies the main part of the hills to the west.

Hills to the east across the Wood River Valley are underlain by the Wood River Formation, with Milligen Formation below the Wood River thrust. The Wood River thrust will be visited in Quigley Creek, directly to the east of Hailey, at Stop 2-4.

Return to the cars and proceed south 0.2 mile (mile 13.4) to a small road to the right (west). Cross a canal and bear left on dirt road. Cross the Wood River and proceed to the mouth of the second gulch on the north (right) side of the road (mile 14.2).
Figure 20. The Wood River thrust in Boulder Basin, Boulder Mountains. Here, the thrust appears to preserve its Mesozoic fabric. With conglomerate of the basal Wood River Formation lying above a gouge zone and phyllite of the Devonian Millgen Formation. The same geologic relation is mapped at the Hailey Conglomerate Member type area (Stop 2-3).

Figure 21. View of a klippe of Pennsylvanian-Permian Wood River Formation unit 6 (sandy limestone) (rounded hill on the skyline) above Permian Dollarhide Formation (upper member, dark mudstone) on Buttercup ridge at the head of Willow Creek, Smoky Mountains. Location of photograph is shown on Figure 18. View is to the southeast. This area was mapped by Geslin (1986). This klippe is thought to overlie a flat, top-to-the-west low-angle normal fault which cuts bedding in both the Wood River and Dollarhide Formations, and which attenuates the Wood River Formation, eliminating at least 1500 meters of strata.

Figure 22. The mouth of Colorado Gulch and site of Stop 2-2, taken looking westward from Highway 75. The prominent jog in the south ridge of Della Mountain marks a low-angle normal fault that cuts the folded Wood River thrust and places boudins of Hailey Conglomerate Member on folded Millgen Formation. Figure 26 is a cross section through this ridge.

Figure 23. Eocene lavas of the Challis Volcanic Group on the east side of the Wood River graben, just southeast of Ketchum.
Stop 2-2: Milligen Formation and Low-Angle Normal Fault in Colorado Gulch

This stop involves a walk of about 1.5 miles and a climb of 600 feet. We will walk up the gulch on the south side of Della Mountain and observe isoclinally folded Devonian Milligen Formation with Pennsylvanian Hailey Conglomerate Member in low-angle normal fault contact above it (Fig. 22). Figure 26 is a cross section of the ridge here, north of Colorado Gulch.

The first outcrop on the west side of the gulch exposes isoclinally folded phyllite of the Devonian Milligen Formation. The phyllite has a cleavage (S1), which is axial planar to isoclinal folds, and a later spaced cleavage (S2). Walk up the gulch and around to the right (east) through platy limestone, pink siltite, and gray phyllite of the Devonian Milligen Formation. The Milligen Formation is thought to have made up part of the Antler highland in Mississippian time (Turner and Otto, 1988, this volume).

Bold outcrops of the Hailey Conglomerate Member occur above a prominent bench where there are several prospect trenches. Here the Hailey Conglomerate Member is exposed as boudins along a shallowly dipping normal fault that cuts bedding in both upper and lower plates. The fault extends to the west, up Colorado Gulch. About 2 km to the west, it places brown-weathering Milligen Formation above black-weathering Dollarhide Formation (Fig. 26). This fault is one of several low-angle normal faults that occur on the flanks of the Wood River Valley.

The view from the bench at the top of the jeep road overlooks the Wood River graben, Bald Mountain west of the Wood River, and the high ridges of the Boulder Mountains to the north.

Return to the cars and proceed back to Hailey. Reset odometers at the major intersection at the corner of Main Street (Highway 75) and Bullion Street. Turn west on Bullion Street. Cross the Wood River, and stop at the west side of the bridge (0.4 mile).
Stop 2-3: Type Area of the Hailey Conglomerate Member

Walk upstream along the river for a short distance to view the type area of the basal Hailey Conglomerate Member of the Wood River Formation (Bostwick, 1955; Thomasson, 1959; Winsor, 1981). The rocks dip steeply west and are overturned (Fig. 27). The Hailey Member here consists of sheared pebble to cobble conglomerate and coarse sandstone containing clasts of gray mudstone or argillite, black quartzite, and chert-pebble conglomerate. Sedimentary structures include graded bedding, planar cross bedding, and ripple marks. The Hailey Member lies structurally above the Devonian Milligen Formation here along the folded Wood River thrust. The Wood River thrust here preserves its Mesozoic relations and has not been reactivated as a Tertiary normal fault.

Return to cars and go back to Main Street in Hailey. Reset odometers, head south for one block to Croy Street and proceed east. Croy Street becomes Quigley Gulch Road. In the valley of Quigley Creek (mile 1.0), Wood River Formation underlies the hills to the north, with Milligen Formation at the base of the hills to the south. The Wood River thrust crops out in the steep hills south of the creek about 300 feet above creek level. Bold outcrops of the Wood River Formation lie just above the thrust. Stop at the mouth of Deadman Creek (mile 2.7) to examine the Wood River thrust.

Stop 2-4: Wood River Thrust in Quigley Creek

Tightly folded and silicified, thinly laminated sandy dolomite, siltstone, and purple phyllite make up the Milligen Formation just below the Wood River thrust. Figure 28 is a geologic map of the area. Mesoscopic folds in the Milligen Formation verge east. The Wood River thrust is located at a prominent bench marked by fault breccia and quartz veins and stringers. The overlying Wood River Formation unit 4 is tan- to red-weathering, calcareous sandstone and medium-gray limestone.

This outcrop illustrates the perplexing nature of the Wood River thrust. Here the Hailey Conglomerate Member and overlying units 2 and 3 are missing below...
Figure 26. East-west cross section through Della Mountain on the north side of Colorado Gulch, extreme northwestern Bellevue 7 1/2 minute quadrangle. Stop 2-2 is a foot traverse to the low angle normal fault on the east shoulder of Della Mountain. Cross section is by P.K. Link after geologic mapping by M.E. Ratchford. Dm-Milligen Formation (Devonian); Pw-Hailey Conglomerate Member (Wood River Formation unit 1-Middle Pennsylvanian); Pd-Dollarhide Formation (Permian); Ki-Cretaceous Croesus diorite stock; Qal-Quaternary alluvium. Symbols are shown on Figure 18.

The fault, yet only one km to the east, boudins of Hailey Conglomerate Member are present along the fault. These boudins and the omission of units 2 and 3 suggest that the thrust here was reactivated as a Tertiary normal fault, attenuating the lower part of the Wood River Formation.

Return to the vehicles and drive back to Hailey. Reset odometers at Main Street (Highway 75). Proceed 2.5 miles north to Deer Creek Road and turn west (left).

Mileage for the Deer Creek side trip begins at the intersection of Deer Creek Road and Highway 75. Proceed 2.5 miles north to Deer Creek Road and turn west (left).

Stop 2-5: Deer Creek Stock and Dollarhide Formation near Clarenden Hot Springs

This stop illustrates lithologies of the lower member of the Permian Dollarhide Formation that are in fault contact with biotite granodiorite of the Deer Creek stock. The Deer Creek stock is one of three outliers of the Idaho batholith that intrude strata of the Wood River basin 10 km east of the main batholith (Johnson and others, 1988, this volume). It has been dated at 91 Ma by K-Ar (Hall, cited in Wavra, 1985). The rock is an equigranular biotite granodiorite with trace amounts of hornblende.

The Dollarhide Formation at this locality is folded white quartzite inter-bedded with calcareous, laminated and banded, gray and black siltite. It is thus “typical Dollarhide Formation”, lacking the penetrative cleavage observed in the Devonian Milligen Formation at Stop 2-2.

Return to vehicles and return to Highway 75. Head north for 3.5 miles to East Fork Road. Turn right (east) and proceed up the East Fork of the Wood River for 2.1 miles to an intersection on the right with Canyon Road. Pull out in the clearing on right for a view of the Pioneer Mountains.

Stop 2-6: Pioneer Mountain View

Glacially sculpted peaks of the Pioneer Mountains (including Hyndman Peak on the right) are visible at the head of the East Fork of the Wood River. The peaks are underlain by Proterozoic(?!) schist and quartzite and intruded by Cretaceous and Tertiary plutons (Dover, 1981; Wust and Link, 1988, this volume). The peaks are in the lower plate of the Wildhorse detachment system which separates unmetamorphosed upper plate rocks (including the Wood River and Milligen Formations) from high-grade metamorphic and plutonic rocks of the lower plate. The Wildhorse detachment crops out just below the line of sight in front of the peaks. Movement on the detachment was toward the northwest, and it most probably occurred in Eocene time during and after eruptions of the Challis Volcanic Group (Wust and Link, 1988, this volume). It is probable that some of the low-angle normal faults in the Wood River Valley are related west. Challis Volcanic Group rocks form the craggy summit south of the road at mile 1.2. The faulted contact of Wood River Formation and Dollarhide Formation is crossed at mile 3.9. There are several prospect pits in the Dollarhide Formation north of the road.

Clarenden Hot Springs is to the left of the road at mile 4.5. It is one of several hot springs located on the western margin of the Wood River Valley (Foley and Street, 1988, this volume). Stop in pullout at crest of hill (mile 4.7) to examine intrusive rocks of the Cretaceous Deer Creek stock and beds of the Permian Dollarhide Formation.
to movement on the Wildhorse detachment system.

Return to Highway 75 and north to the stoplight at the intersection of Highway 75 and Sun Valley Road in Ketchum. Set odometers to zero.

Proceed northeast on Sun Valley Road. The road follows the valley of Trail Creek through the stack of thrust plates on the west flank of the Pioneer Mountains. The Boulder Mountains lie northwest of Trail Creek Road (Fig. 18). The Devonian Milligen Formation underlies the hills between Ketchum and Wilson Creek (Fig. 30). The Wood River Formation is complexly folded and imbricat between Wilson Creek and the upper part of the canyon.

Stop at mile 8.2 to discuss Wood River Formation stratigraphy.

**Stop 2-7: Wood River Formation in the Eastern Boulder Mountains**

The ridge northwest of Trail Creek rises from the Wood River Valley to peak 10,458 at the head of Rock Roll Canyon, directly to the west (see geologic map of this area, Fig. 30).

Mapping by Burton and Link northwest of Trail Creek (Fig. 30) shows that here the Wood River Formation is made up of two separate allochthons (Fig. 29). The northeastern allochthon (3000 m thick) contains all the informal units of the stratotype near Bellevue. Sandy limestones of units 4 and 6 are separated by brittle siliceous sandstones of unit 5. In the southwestern allochthon, unit 5 is missing, and the section is only two-thirds as thick (1800 m).

Similar to Stop 2-4, the Wood River thrust here displays complex and variable relations between rocks in the hangingwall and footwall. In the high saddle south of peak 10,458, the upper part of the northeastern allochthon (unit 6) is in fault contact with black argillite of the Devonian Milligen Formation. One kilometer farther west, the entire section of the northeastern allochthon is present above the fault, implying that a lateral ramp is present here. Several kilometers farther west, the Hailey Conglomerate in the northeastern allochthon contains clasts of silicified limestone identical to the subjacent Milligen Formation. Here the contact is mapped as a disconformable sedimentary contact.

The ridge to the south of the high saddle contains facies of the southwestern allochthon (Fig. 29). The stratigraphic differences between the two allochthons require that tens of kilometers of telescoping have occurred between them. A Mesozoic thrust fault is therefore interpreted to be present between the allochthons (Fig. 31). The contact between the Hailey Conglomerate Member and the Milligen Formation in the southwestern allochthon is similar to the depositional contact in the northeastern allochthon, but exhibits some brecciation and slickensided surfaces. It is not clear if this contact is a sheared unconformity or another example of the younger-on-older Wood River thrust.

An admissible cross section that obeys thrust belt rules (Buyer and Elliott, 1982; Woodward and others, 1985) can only be drawn across this area if two stages of deformation are involved: northeast-vergent folding of the Wood River and Milligen Formations, followed by thrusting (Fig. 31). The thrusts are cut by high-angle faults that are pinned by Paleogene dikes. Figure 31 shows an admissible cross section through the eastern Boulder Mountains that reflects the observed position of the Wood River thrust in this area, and accommodates the required displacement between allochthons.

Continue northwest up the Trail Creek road. Stop at mile 10.3 to view a strand of the Wood River thrust system.

**Stop 2-8: Wood River Thrust in Trail Creek**

Here a mineralized and iron-stained gouge zone several meters thick marks the contact between the Wood River
Figure 29. Correlation diagram of the two allochthons of Wood River Formation in the eastern Boulder Mountains. The northeastern allochthon, measured on the ridge east of Stop 2-7, displays a stratigraphic sequence similar to the stratotype of Hall and others (1974), while the southwestern allochthon, measured 2 km to the west, west of Trail Creek, contains a much thinner sequence of rocks in which Unit 5 is missing, and the upper units are thickened at the expense of the middle units. Symbols are shown on Figure 19.

Formation and a fault sliver of the underlying Devonian-Silurian-Ordovician plate. The Wood River Formation is brecciated, thick-bedded, light gray limestone (probably unit 6). The rocks in the footwall are olivine-green to maroon argillite and siltite. The underlying plate contains dark graptolitic shales of the Ordovician and Silurian Phi Kappa Formation and Silurian Trail Creek Formation (Dover, 1983).

The fault geometry is again ambiguous, with the gouge zone suggesting ductile Mesozoic thrusting, but the lower part of the Wood River Formation is missing above the fault, suggesting a normal fault relation.

This contact is mapped as a thrust fault by Dover (1983). In the cross-sections of Dover (1983), the Devonian Milligen Formation of the Milligen-Wood River plate lies structurally above the Devonian-Silurian-Ordovician plate of the footwall to the north. This contact is mapped as a normal fault by Burton (Fig. 30) in the area northwest of here because it exhibits a high angle (58 degrees) younger-on-older relationship. Burton interprets the Wood River thrust to be cut by this normal fault at depth.

Turn around and drive back to the stoplight at Highway 75 in Ketchum.

Road Log from Ketchum to the Baker Creek Road

At the intersection of Sun Valley Road and Highway 75 in Ketchum, reset odometers, turn right, and proceed north on Highway 75. At mile 0.3 the view opens up to the north; high peaks of the Boulder Mountains underlain by the Wood River Formation and the Challis Volcanic Group form the skyline. The lower hills are underlain by Challis Volcanic Group, part of a cauldron hypothesized by Hall and McIntyre (1985). Cross the Wood River at mile 6.7. The highway bends to head westward. The terraced slopes to the east are Challis Volcanic Group, with the Wood River Formation underlying the hills to the north. Challis Volcanic Group rocks crop out to the west. Quaternary glacial moraines crop out north of the road at mile 8.8 (Pearce and others, 1988, this volume).

Turn west on the Baker Creek road at mile 15.1 and stop. Climb about 100 feet up the hill northwest of the road for a view of the Boulder Mountains.

Stop 2-9: View of Boulder Mountains

The spectacular cliff exposure to the north is uplifted along the southwest-dipping Boulder front fault, and displays pink granite of the Eocene Boulder Mountains stock intruding dark Eocene dacite porphyry and light-colored limestone and quartzite of the Wood River Formation. Figure 32 is a cross section of the Smoky Mountains northwest of here, which show identical relations to the west side of the Boulder Mountains, with
Figure 30. Simplified geologic map of the southern half of the Rock Roll Canyon 7.5 minute quadrangle showing location of Stop 2-7 and 2-8. Note the variable nature of the Devonian Milligen Formation (Dm) contact with Pennsylvanian - Permian Wood River Formation. This contact occurs as an unconformity, a locally overturned Mesozoic younger on older low angle fault (the Wood River thrust), and a high angle normal fault. Cross section A-A' is shown on Figure 31. Mapping by Burton and Link, modified in part from Dover (1983) and Hall, unpublished.
Figure 31. Geologic cross section through the eastern Boulder Mountains showing two Wood River Formation allochthons. Location of cross section is shown on Figure 30. A Mesozoic thrust with tens of km of displacement is interpreted to have juxtaposed the northeastern and southwestern allochthons. The thrust is shown to be cut at depth by a high angle normal fault. Brecciated zones at the contact between the lower Wood River Formation and Milligen Formation are in other areas mapped as the Wood River thrust (Dover, 1983). An admissible cross section that obeys thrust belt rules can be drawn if this contact is reinterpreted as flexural slip formed along an unconformity during folding. Cross section by B. R. Burton.

Figure 32. East-west cross section of the Galena quadrangle, from Mahoney (1987). Line of section is shown on Figure 18. Pd--Dollarhide Formation; Pnw--Wood River Formation; Pu--undifferentiated Paleozoic roof pendants; Kgd--Cretaceous biotite granodiorite; Tdp--Eocene dacite porphyry of hypabyssal granodiorite complex; Tpg--Eocene pink granite; Trp--Eocene Rhyodacite porphyry; Td--Eocene dacite lava and breccia (Challis Volcanic Group).

Eocene pink granite intruding only slightly older hypabyssal dacite porphyry and lava.

The Boulder Mountains are the site of the Boulder Basin mining district, studied in detail by M. E. Ratchford of the University of Idaho. Exposed in Boulder Basin (Fig. 18) is the roof zone of an Eocene hypabyssal plutonic-volcanic complex with Wood River and Milligen Formations intruded by dacite, andesite, and rhyolite porphyry, and overlain at shallow levels by the Challis Volcanic Group.

The Boulder Basin mining district produced significant quantities of silver, gold, and lead before 1902 (Tschanz and others, 1986). The Boulder townsite at 9,200 feet elevation in the floor of the Boulder Basin was occupied intermittently from 1879 to 1935. It can be reached by a very rough jeep road that intersects Highway 75 about a mile south of here.

The Smoky Mountains to the west have been mapped recently by several Masters students from Idaho State University and the University of Idaho (Gehlan, 1983; Wavra, 1985; Geslin, 1986; Darling, 1987; Mahoney, 1987; Stewart, 1987). In the headwaters of Baker Creek about 10 km west of here in the Smoky Mountains (Baker Peak and Galena quads), Wood River Formation overlies Dollarhide Formation along a shallow fault that is intruded by Eocene granodiorite (Stewart, 1987). This fault cannot be a top-to-the-east Mesozoic thrust because it places Wood River
Formation (eastern shelf facies of the Wood River basin) over Dollarhide Formation (western basin facies) (Skipp and others, 1986). The fault could be a top-to-the-west low-angle normal fault similar to that seen in Colorado Gulch (Stop 2-2). Such a structural relation may have formed during doming of the Boulder Mountains intrusive center to the northeast, with rocks of higher structural levels (Wood River Formation) displaced southwestward along a normal fault.

ACKNOWLEDGMENTS

Skipp and Hait's work in south-central Idaho has been supported by the U. S. Geological Survey Branches of Central Regional Geology and Central Mineral Resources for many years. Link's work has been supported by the U. S. Geological Survey Branch of Central Mineral Resources and the Idaho State University Faculty Research Committee. Janecke has been supported by an American Chemical Society Petroleum Research Fund grant to R. L. Bruhn. Burton has received support from the Geological Society of America, the Idaho State University Graduate Student Research Council, and the U. S. Geological Survey. We are indebted to J. Evans, F. J. Moye, and in particular to M. A. Kuntz, D. W. Rodgers, and J. B. Mahoney for constructive reviews of this manuscript.

REFERENCES


_____, 1986, Relationship of the Trans-Challis fault system in central Idaho to Eocene basin and range extension: Geology, v. 14, p. 481-484.


Guidebook to the Geology of Central and Southern Idaho


INTRODUCTION

The Pioneer Mountains (Fig. 1) east of Ketchum, Idaho contain a lower plate or core of gneiss, plutonic rocks, and complexly folded metasedimentary strata, and an upper plate composed of a stack of thrust sheets of unmetamorphosed Paleozoic strata overlain by Eocene volcanic and sedimentary rocks.

The area was mapped geologically by Dover (1969, 1981, 1983) who interpreted all low angle faults as thrusts. Current workers reinterpret the geologic relationships (Fig. 2) and suggest that the Pioneer Mountains are an extensional core complex, having evidence of both Mesozoic synkinematic plutonism and Paleogene extensional movement along the Wildhorse detachment system which separates the upper and lower plates (Wust, 1986a; 1986b; 1986c; O'Neal, 1985; Silverberg, 1986; Kim, 1986; Pavlis and O'Neill, 1985; 1987).

The rocks of the Pioneers can be divided into a lower plate (core) and an upper plate, separated everywhere by faults. Precambrian X gneisses, calc-silicates, and marbles are exposed in the central part of the core and are dated at 2 Ga (Dover, 1983). The southwest side of the core exposes a metasedimentary sequence, which includes calc-silicates, metaquartzites, and pelitic schists presumed by Dover (1981, 1983) to be Precambrian Y in age and unconformably overlain by Ordovician quartzites and marbles.

The metamorphic core units are intruded by a Cretaceous-Tertiary granodiorite which is now deformed, a locally deformed quartz monzonite of probable Eocene age, and several minor intrusive bodies, also of probable Eocene age. Geochronologic studies of these plutons and core rocks are in progress by D. Silverberg, K. Hodges and J. Sutter.

Upper-plate units are mostly unmetamorphosed. Paleozoic strata are equivalents of the Antler orogenic belt sequences seen in Nevada. They include: (1) Ordovician, Silurian, and Devonian argillite, siltstone, and chert, with minor sandstone, dolomite, cherty dolomite, and limestone; this is the Antler "oceanic assemblage", of similar lithology to the upper plate of the Roberts Mountain overthrust; (2) the Milligen Formation, a Devonian siliceous argillite that is also part of the Antler overthrust assemblage; (3) the Copper Basin Formation, composed of Mississippian sandstone, conglomerate, siltstone, argillite, and limestone; this formation is part of the Antler flysch; and (4) the Wood River Formation, a Pennsylvanian to Permian unit of calcareous siltstone, calcareous sandstone, and limestone, with a basal chert-pebble conglomerate; this formation represents the Antler overlap sequence.

Overlying the Paleozoic strata are Eocene sedimentary and volcanic rocks. The basal sedimentary unit is only locally present and includes red sandstone and conglomerate; the conglomerate contains clasts of all

1 Chevron USA, Inc., P. O. Box 6056, New Orleans, LA 70174
2 Department of Geology, Idaho State University, Pocatello, ID 83209
upper-plate Paleozoic units. Challis Volcanics overlie the conglomerate and all upper-plate units, but they are not present in the core. The volcanic rocks are mostly andesitic to rhyolitic and in the vicinity of the Pioneer Mountains have radiometric ages of 38-42 Ma (Armstrong, 1975). Northwest of the core, the Summit Creek stock is intruded into upper-plate rocks. It is similar in petrography to the Eocene quartz monzonite in the core.

In the lower plate of the Pioneer Mountains core complex, Mesozoic deformation of the Sevier orogeny is manifested in the folding and metamorphism of Precambrian and Ordovician strata (Silverberg, 1986) and by the intrusion of the Cretaceous-Tertiary granodiorite pluton. Mesozoic deformation in the upper plate is represented by folding and thrusting of Paleozoic strata. Vergence of thrusts and overturned folds is consistently between northeast and east.

As defined by Wust (1986 a and b), the Wildhorse detachment system separates upper plate from lower plate rocks and includes the Wildhorse thrust and part of the Pioneer thrust system of Dover (1983). Trends of striations on the smooth slickensided surface of the detachment consistently cluster around N, 65°W. Mylonitic gneiss, marble, and quartzite are exposed on the northwest side of the core (lower plate), suggesting that the deepest levels of exposure are on the northwest and that during fault movement the upper plate moved northwest relative to the lower plate. Sense of shear derived from S-C relations in mylonites (Berthé and others, 1979) also gives a top-to-the-northwest sense of movement. Separation on the Wildhorse detachment system is estimated to be at least 17 km, based on two observations: a correlation of the quartz monzonite in the core with the Summit Creek stock of the upper plate, and on the width of exposure of lower plate rocks in the core of the range (Wust, 1986a).

As defined by Wust (1986 a and b), the Wildhorse detachment system separates upper plate from lower plate rocks and includes the Wildhorse thrust and part of the Pioneer thrust system of Dover (1983). Trends of striations on the smooth slickensided surface of the detachment consistently cluster around N, 65°W. Mylonitic gneiss, marble, and quartzite are exposed on the northwest side of the core (lower plate), suggesting that the deepest levels of exposure are on the northwest and that during fault movement the upper plate moved northwest relative to the lower plate. Sense of shear derived from S-C relations in mylonites (Berthé and others, 1979) also gives a top-to-the-northwest sense of movement. Separation on the Wildhorse detachment system is estimated to be at least 17 km, based on two observations: a correlation of the quartz monzonite in the core with the Summit Creek stock of the upper plate, and on the width of exposure of lower plate rocks in the core of the range (Wust, 1986a).

The Wildhorse detachment system is interpreted to have formed during Paleogene extensional faulting associated with the development of the Pioneer Mountains core complex. Paleogene crustal extension in central Idaho is manifested not only by deformation in the Pioneer Mountains but also by the widespread and prolonged (51 to 40 Ma) Challis magmatic episode (McIntyre and others, 1982; Fisher and others, 1983; Moye and others, 1988, this volume) and by extension associated with the trans-Challis fault system (Bennett, 1986).

Areas north and west of the Pioneer Mountains also underwent Paleogene extension. Normal faults with shallow dips were mapped by W. E. Hall at several locations in the Wood River Valley to the west, and they
Figure 2. Regional geologic map of the Pioneer Mountains showing field trip stop locations. After Dover (1983) and Wust (1986a).
are also present in the Smoky and Boulder Mountains (Skipp and others, 1986; Otto and Turner, 1987; Ratchford, 1988; Link and others, 1987; 1988). Synchronicity of extensional faulting and the Challis magmatic episode is suggested by relations in the Baker Peak quadrangle of the Smoky Mountains where the Norton Lakes pluton, of similar composition to the Summit Creek stock, intrudes a low angle normal fault (Stewart, 1987; Mahoney, 1987).

The relationship of the plutons in the core to Mesozoic and Paleogene deformation of the Pioneer Mountains is important and controversial. Both plutons are cut by the Wildhorse detachment system. Although there was definitely post-intrusion movement along the Wildhorse detachment system, it is not known whether any of the movement was syn-intrusive. See the comment by Pavlis and O'Neill (1987) and the reply by Wust (1987) for discussion of both pluton geochronology and the timing of movement along the Wildhorse detachment system.

Another controversy concerns the timing of uplift of lower plate rocks. Dover (1981) used two lines of evidence to infer that the core of the Pioneers was uplifted before deposition of the Challis Volcanics. First is a landslide deposit near Pioneer Cabin (Stop 1-1) mapped as unconformably overlain by Challis Volcanics breccia, and containing Ordovician Kinnikinic Quartzite derived from the core. Second is the local conglomerate which underlies the Challis Volcanics and which was interpreted to contain clasts of Kinnikinic Quartzite.

However, uplift of the core before deposition of the Challis Volcanics is difficult to reconcile with recent conclusions that movement on the Wildhorse detachment system occurred in Paleogene time, coeval with eruption of the Challis Volcanics. The geologic relations of the landslide are ambiguous (see discussion in field trip guide), and there are no Challis Volcanics in depositional contact with lower plate rocks. Lower plate clast provenance for the pre-Challis conglomerate is also questionable as Wust (1986a) could find no definitive lower plate clasts and interpreted the quartizes to have come from the upper plate Mississippian Copper Basin Formation.

FIELD TRIP GUIDE

The Pioneer Mountains are both a scenic alpine mountain range and a very well-exposed metamorphic core complex. This guide describes two days of geologic field trips in the Pioneers. Figure 1 shows the general geology and stop locations. The field trips require day-long hikes into the foothills of the Pioneers, up to 9500 feet elevation. Food, water, and appropriate clothing and footwear for high mountain travel are essential. Those using this field guide will find the geologic map of Dover (1983) essential, and the Hyndman Peak and Phi Kappa Mountain 7.5-minute topographic quadrangles very helpful.

The field trips address the stratigraphy of upper plate strata along the Pioneer Cabin trail and the Trail Creek road, the Wildhorse detachment system below Pioneer Cabin, the nature of the core rocks east of Pioneer Cabin at Kane Creek and at Wildhorse campground, and the progressive increase in deformation of the core as the detachment is approached at Kane Creek.

Day 1: Geology of Corral and North Fork Hyndman Creeks near Pioneer Cabin.

This field trip involves a foot traverse of approximately 8 miles and 3000-foot elevation gain from Corral Creek to the North Fork of Hyndman Creek. The one-way traverse requires that a vehicle be spotted at North Fork Hyndman Creek.

The Corral Creek parking area is reached by driving northeast on the Sun Valley-Trail Creek Road from the stoplight in downtown Ketchum. Proceed for 4.9 miles and turn right on the Corral Creek road. Proceed 3.8 miles to parking area at the end of the road.

In the valley of the Big Wood River, Trail Creek, and Corral Creek, the Wood River Formation everywhere overlies dark argillites, limestones, and siltstones of the Milligen Formation along a system of younger-over-older faults that are designated the Wood River thrust (Hall and others, 1978; Dover, 1983). Details of stratigraphy and deformation of the Milligen Formation are discussed by Turner and Otto (1988, this volume). The field trip by Link and others (1988, this volume) discusses strata of the Wood River Formation and relations along the Wood River thrust system.

The Wood River Formation contains at least 3000 meters of strata divided into seven units (Hall and others, 1974). The base of the Wood River Formation is the Hailey Conglomerate which contains quartzite and argillite clasts, some of which were derived from the underlying Milligen Formation. The bulk of the overlying units of the Wood River Formation are sandy limestone, calcareous sandstone, quartzite, and dark mudstone.

In many places in the Wood River Valley, upper units of the Wood River Formation are faulted on Milligen Formation, and synclines in Wood River Formation are truncated along a shallowly dipping fault above the Milligen Formation, suggesting that parts of the Wood River thrust system were reactivated as low angle normal faults (Skipp and others, 1986; Link and others, 1988, this volume).

Geology along the Pioneer Cabin Trail above Corral Creek

The lower slopes of the trail are in Milligen Formation. At 7800 feet elevation the trail crosses the Wood River thrust into Wood River Formation. At the top of the switchbacks (7900 feet) are outcrops of brown-
weathering calcareous quartzite of Wood River Formation unit 6.

Problematic Landslide Deposit

Where the trail leaves the trees at elevation 8600 feet, it crosses onto what is mapped by Dover (1983) as a landslide deposit, of Tertiary or Cretaceous age, and composed of blocks of brecciated Ordovician Kinnikinic Quartzite derived from the lower plate of the Wildhorse detachment system. As mapped, the landslide is overlain by heterolithologic epiclastic breccia of the Challis Volcanics. This implies that the core was elevated during Sevier thrusting or Cretaceous intrusion and is contrary to the idea that uplift of the core occurred during movement on the Wildhorse detachment system. Problems with the mapped relation include the absence of Challis rocks within the core, and the absence within the landslide of core rocks other than quartzite.

An alternate interpretation of the outcrop geometry is that the landslide overlies and surrounds the Challis Volcanics. Problems with this interpretation include the absence of volcanic clasts within the landslide and the lack of clear evidence for superposition.

The simplest explanation may be that the landslide is composed of brecciated and silicified quartzites of the Wood River Formation of the upper plate rather than Kinnikinic Quartzite (R. Turner, pers. comm., 1987), and that it is overlain by Challis Volcanics breccia. This would imply uplift of upper plate strata, but not exposure of lower plate rocks, prior to Challis volcanism.

Critical unanswered questions concerning the mapped landslide are whether it contains Kinnikinic Quartzite or quartzite of the Wood River Formation, and the stratigraphic relationships between the landslide deposits and the Challis Volcanics.

Proceed over the landslide deposit and up the hill. At the trail junction with the Johnstone Creek trail take the left fork to Pioneer Cabin. At the junction the view to the southwest is of Johnstone Peak, which is underlain by folded upper Wood River Formation and Devonian Milligen Formation in the lower slopes. The contact is a steeply dipping fault, mapped by Dover (1983) as the folded Wood River thrust. The ridge north of Johnstone Peak is underlain by Challis Volcanics. The orange talus-strewn hill half a mile to the northwest is composed of brecciated quartzite of the landslide.

The trail crosses heterolithologic breccia of the Challis Volcanics, then back into brecciated quartzite, and up a talus slope before cresting the ridge at Pioneer Cabin (elevation 9440 feet).

Stop 1-1: Panorama at Pioneer Cabin

Pioneer Cabin (Fig. 3) was built in the 1930s for a ski touring basecamp during early development of Sun Valley resort. The view from the outhouse is magnificent.
Figure 3. Panorama from Pioneer Cabin (Stop 1-1). The peaks on the skyline make up the cirque of the North Fork of Hyndman Creek, and are underlain by Proterozoic and Ordovician metasedimentary rocks intruded by a Tertiary-Cretaceous granodiorite pluton which makes up the prominent, jointed face directly over the roof of the cabin. The cabin lies on the upper plate of the Wildhorse detachment. The detachment is located just behind the trees on the left side of the view.

Figure 4. View to the north along ridge north of Pioneer Cabin. Location of Stop 1-2 and the trace of the Wildhorse detachment are shown. White outcrops in the lower plate are Ordovician metasedimentary rocks. The peak on the skyline is underlain by Tertiary-Cretaceous granodiorite.

Figure 5. Cirque of North Fork, Hyndman Creek and location of Stop 1-3. White outcrops in the floor of the valley are Ordovician Kinnikinic Quartzite, grassy areas in the valley floor are Ordovician Ella Marble and Saturday Mountain Formation.
trail about a mile from Stop 1-3. The Wildhorse detachment is immediately above the overhanging cliff.

Stop 1-4: Wildhorse detachment
Below Pioneer Cabin

A spectacular exposure of the Wildhorse detachment (Figs. 6 and 7) can be reached by climbing about 700 feet up a steep talus slope that intersects the trail from the west at elevation 7700 feet. This is the first talus of Wood River Formation limestones.

The detachment is exposed at elevation 8400 feet on the north side of the talus, where it is a polished and striated undulating surface on Ordovician Kinnikinic Quartzite of the lower plate. Lineations on the surface consistently trend to the northwest, but the dip of the surface varies from horizontal to 70 degrees south. Upper plate rocks consist of chaotic and brecciated Wood River limestones and calcareous sandstones which are exposed on the southwest side of the talus slope.

In contrast to areas to the north (e.g., Kane Creek--Stop 2-3), the detachment here contains no mylonite. The Kinnikinic is an unmylonitized quartz arenite. Brittle deformation is demonstrated by local areas of quartzite breccia along the detachment and subsidiary faults in the Kinnikinic Quartzite below the detachment.

Return down the talus slope and continue down the trail. West of the trail, Wood River Formation including the basal Hailey Conglomerate crops out. Where Button Creek enters from the west (about one mile from Stop 1-4) a buttress of Challis Volcanics can be seen to the west. The parking area is a little over two miles from Stop 1-4.

Drive down the Hyndman Creek road, turn right at mile 1.7, turn right again on the East Fork Wood River Road (mile 4.9), pass tailings of the Triumph silver-lead-zinc mine in argillite of the Devonian Milligen Formation (mile 6.6), and turn right on Highway 75 at mile 11.9. The lower reaches of the East Fork drainage are bounded by steep sage-covered hills underlain by Devonian Milligen Formation on the north and Pennsylvanian-Permian Wood River Formation on the south (Hall and others, 1978; Batchelder and Hall, 1978; Hall, 1985).

The Ketchum stoplight is 5.6 miles to the north of the East Fork Road.

Day 2: Field Trip up Trail Creek to Kane and Wildhorse Canyons

From the stoplight in Ketchum, proceed northeast on Sun Valley Road. The road follows the valley of Trail Creek through the stack of allochthons on the west flank of the Pioneer and Boulder Mountains (Fig. 1). The upper allochthons above the Wood River thrust system contain Pennsylvanian-Permian Wood River Formation. Devonian Milligen Formation lies under the Wood River thrust and is exposed south of Wilson Creek.

The Wood River Formation is complexly folded and imbricated between Wilson Creek and the upper part of the canyon.

Stop 2-1: Wood River Thrust on Trail Creek Road

Pull out at wide part of the road 8.2 miles from the stoplight in downtown Ketchum. Just down hill from the inside bend in the road, the Wood River thrust system is unusually well exposed. The thrust separates calcareous sandstone of the upper part of the Wood River Formation from argillites and dark graptolitic shales of the Ordovician and Silurian Phi Kappa Formation, Silurian Trail Creek Formation, and other strata of the western facies oceanic assemblage. The plate of oceanic assemblage rocks is bounded above by the Wood River...
thrust and below by the Pioneer thrust and Mississippian Copper Basin Formation.

Return to the cars and continue up Trail Creek. The road remains in the oceanic assemblage over Trail Creek Summit and to Park Creek (13.0 miles), where it enters the lower allochthons which consist of Mississippian Copper Basin Formation, a thick and folded sequence of carbonaceous mudstones, quartzose flysch, turbidite fan channel conglomerates, and thin limestone marker beds. The Copper Basin Formation is duplicated along the Glide Mountain thrust (Dover, 1983).

Stop 2-2: Little Fall Creek View

This stop is located at mile 13.9 where the road crosses Little Fall Creek. To the north (Fig. 8) are rocks of the Copper Basin allochthon which is overlain at the top of the mountain by the oceanic assemblage along the Pioneer thrust, and intruded by the Eocene Summit Creek stock. A rough 4-wheel drive road up Little Fall Creek affords access to the thrust, which is a smooth surface on silicified Copper Basin Formation and is overlain by tightly folded black argillite of the oceanic plate.

Return to vehicles and proceed north heading down the Summit Creek drainage. The Summit Creek stock crops out on the south side of the road east of Park Creek, and on both sides of the road at mile 14.5 between Little and Big Fall Creeks.

Pass the road to Phi Kappa Creek at mile 16.1 and turn south on Kane Creek Road just after crossing Summit Creek at mile 19.1. Proceed 4.8 miles up the primitive Kane Creek Road. Hills on both sides of the road are Copper Basin Formation. The bare cliffs at the head of the valley are lower plate rocks of the core.

Stop 2-3: Wildhorse detachment in Kane Creek

Mylonitic lower plate rocks and the Wildhorse detachment are exposed between the forks of Kane Creek about 1 mile north of the Devils Bedstead (Phi Kappa Mountain Quadrangle). Park the cars at the fourth crossing of Kane Creek (4.8 miles from the main Trail Creek-Summit Creek road) and follow the trail along the west side of Kane Creek for about half a mile to a footbridge. Cross to east side of creek and follow jeep road until it cuts sharply to the left. Continue straight along the trail that follows the creek. Cross the main fork of Kane Creek at elevation 7900 and proceed up the cliff and talus west of the creek (first prominent bare exposures). The detachment is located in the prominent saddle at the base of the cliff on the ridge to the west at 8950 feet elevation.

This locality exposes a cross section of deformed rocks of the lower plate in what is thought to be the structurally deepest exposed level of the detachment system. Lower plate rocks include foliated Tertiary-Cretaceous granodiorite intimately mixed with Ordovician and Precambrian Y metasedimentary rocks and Precambrian X gneiss. The granodiorite has zones of well-developed foliation as well as unfoliated hornblende-rich zones. Several hundred meters below the detachment the rocks have no lineation but display complex foliation and several orientations of ductile folds formed during Mesozoic or earlier deformation. Farther up the hill, the foliation progressively becomes parallel with the detachment, and a N. 60°W. lineation becomes dominant. At elevation 8800 feet, sills of granodiorite with S-C mylonitic fabrics and interlayered mylonitic quartzite, marble, and granodiorite (Fig. 9) are exposed. Lineations and rotated porphyroblasts support movement to N. 60°W. along mylonitic foliation surfaces that parallel the detachment.

Figure 8. View at Stop 2-2 of Little Fall Creek. Dark rocks on the skyline are Ordovician to Devonian argillites, siltites, and cherts of the "oceanic assemblage" which overlie the Pioneer thrust. The cliff in the middle distance is Mississippian Copper Basin Formation below the thrust. Light-colored dikes of Eocene granodiorite (Summit Creek stock) intrude both plates.

Figure 9. Foliated rocks about 30 meters below the Wildhorse detachment at Kane Creek, Stop 2-3. The thinly layered rock is foliated Ordovician Ella Marble, which is intruded by a granodiorite sill that has been subsequently deformed.
Upper plate black argillite, graded sandstone and conglomerate of the Mississippian Copper Basin Formation can be observed along the ridge to the north.

From the saddle area, the view to the east is of the Wildhorse detachment across Left Fork of Kane Creek (Fig. 10). The view to the southwest along the Right Fork of Kane Creek shows the Pioneer thrust system of Dover (1983) striking into the Wildhorse detachment (Fig. 11). The lithologic change along the western slopes delineates the thrust; the intersection of the two faults is hidden under Quaternary sediments along the creek. Contrary to Dover's interpretation, Wust maps the detachment as cutting the Pioneer thrust: note that the style and orientation of lower-plate structures stay consistent across the intersection of the two faults.

Figure 10. View to the east of the Wildhorse detachment from saddle between Left and Right Forks of Kane Creek. Cliff is made up of both the Cretaceous-Tertiary granodiorite and the Eocene granodiorite of the lower plate. Detachment is at prominent break in slope, and not in the saddle. Upper plate rocks are Mississippian Copper Basin Formation.

Return to cars via trail along the Right Fork of Kane Creek. The trail can be reached by going directly down-slope to the west almost to the valley floor. Walk on the trail to the north and east down switchbacks to the main fork and then downstream along the main trail to the cars. Return to the Summit Creek-Trail Creek road (4.8 miles).

The drive up Wildhorse Creek is scenic (Fig. 12) and affords a look at Precambrian X gneiss of the core. Proceed 3.2 miles northeast down the Trail Creek-Summit Creek Road and turn south on the Copper Basin-Wildhorse Road. Proceed 2.1 miles to the junction to Copper Basin. Take the right fork heading up Wildhorse Creek. The high bare peak visible in the core of the range is Old Hyndman Peak, which is underlain by Eocene quartz monzonite similar in composition to the Summit Creek stock. At 5.5 miles from the Summit Creek Road, just after crossing Wildhorse Creek, bear right toward Wildhorse Campground; the left fork goes to Fall Creek.

Stop 2-4: Wildhorse Campground: Core Rocks

The campground is located 7.8 miles from the Summit Creek-Trail Creek Road and is on the trace of the Wildhorse detachment, which follows Boulder Creek to the west. Up the primitive road past the campground, boulders are seen with varying lithologies of complexly deformed Precambrian X gneiss (Figs. 13 and 14) and foliated Eocene porphyritic quartz monzonite (Fig. 15) of the core. A climb is necessary to see the outcrop but all core lithologies are represented in the glacially polished boulders.

End of field trip.

ACKNOWLEDGMENTS

This manuscript is a revised version of Wust and Link (1987). Early versions of this field guide were reviewed by Charles Waag and David W. Rodgers.
Figure 12. View to the south of the cirque of Wildhorse Creek. Small peak on the right skyline is old Hynaman Peak. Prominent peak in left-center is unnamed and is underlain by early Proterozoic gneiss.

Figure 13. Thinly layered, kink-folded Proterozoic schist boulder from Wildhorse Creek. Note shear zones marked on photograph.

Figure 14. Banded Proterozoic gneiss boulder from Wildhorse Creek.

Figure 15. Orthogneiss boulder from Wildhorse Creek; rock is possibly the foliated Tertiary-Cretaceous pluton, with mafic inclusion at upper left. A shear zone is marked on photograph.
REFERENCES


Berthé, D., Choukroune, P., and Jegouzo, P., 1979, Orthogneiss, mylonite, and non-coaxial deformation of granites: the example of the South Armorican Shear Zone: Journal of Structural Geology, v. 1, p. 31-42.


Ratchford, M. E., 1988, Geology of the Boulder Basin,


Cretaceous and Tertiary Intrusive Rocks of South-Central Idaho

Kathleen M. Johnson¹
Reed S. Lewis²
Earl H. Bennett³
Thor H. Kiilsgaard⁴

INTRODUCTION

The Idaho batholith is one of the great circum-Pacific batholiths. It extends over 300 miles in a north-south direction and is 80 miles wide. Armstrong (1975b) suggests that the batholith be divided into two parts, the Bitterroot lobe to the north and the larger Atlanta lobe to the south. Vast areas of south-central Idaho are underlain by Cretaceous and Tertiary intrusive rocks that are part of the Atlanta lobe. The older plutons were emplaced during Late Cretaceous time, were extensively faulted, and were intruded by Tertiary epizonal plutonic rocks and dike swarms. The Cretaceous batholith was exposed at the surface by early Eocene time and lower extrusive units of the Eocene Challis Volcanic Group were deposited directly on the batholith. The Cretaceous intrusive rocks are exposed in numerous small to large stocks within and to the east of the Atlanta lobe. They intruded to shallow levels between 42 and 50 Ma. Dike swarms and many stocks are concentrated along major regional northeast-trending faults related to Eocene extension.

Previous Work

Early geologic studies in the Atlanta lobe of the Idaho batholith were done by Ross (1928, 1934) and Anderson (1947, 1952), who identified several types of Mesozoic and Tertiary rocks. Ross (1934) mapped and named the Tertiary Casto pluton and several Tertiary diorite bodies. More recently, Schmidt (1964) delineated four Late Cretaceous rock units near Cascade; Reid (1963) and Kiilsgaard and others (1970) mapped Cretaceous and Tertiary rocks in the Sawtooth Range, southwest of Stanley. Lewis and others (1987) provide a comprehensive list of geologic studies in the Cretaceous batholith. Bennett and Knowles (1985) provide a similar list of studies of the Tertiary intrusive rocks.

Geologic mapping and related laboratory studies for projects of the U.S. Geological Survey's Conterminous United States Mineral Assessment Program (CUSMAP) in the Challis and Hailey 1 x 2 degree quadrangles have produced new information about the nature and extent of Cretaceous and Tertiary plutonic rocks of south-central Idaho. Published reports include the Challis quadrangle geologic map (Fisher and others, 1983 and in press), reports of work in progress (Kiilsgaard and Lewis, 1985; Bennett and Knowles, 1985), and the Challis quadrangle mineral resource assessment (Fisher and Johnson, 1987).

GEOLOGIC SETTING

Host rocks along the eastern margin of the Idaho
batholith are Paleozoic sediments that have been metamorphosed to hornfels and calc-hornfels. Rocks of the Eocene Challis Volcanic Group obscure parts of the eastern margin of the batholith (Ross, 1937; McIntyre and others, 1982; Fisher and others, 1983).

The northern part of the Atlanta lobe intrudes a variety of Precambrian metasedimentary and metagneous rocks, most of which are not well studied. Argillaceous quartizes of the Yellowjacket Formation (Ross, 1934; Ruppel, 1975; Lopez, 1982) are intruded by granitic rocks thought to have an approximate age of 1,370 Ma (Evans and Zartman, 1981; Armstrong, 1975a). High-grade gneiss and schist found west of Shoup (Cater and others, 1973) and paragneiss and orthogneiss found immediately north of the main Salmon River (Weis and others, 1972) may be Precambrian crust that is older than the Yellowjacket Formation.

Isolated roof pendants are found in a northwest-trending belt extending across the north-central part of the Atlanta lobe. These metasedimentary rocks are probably Paleozoic in age, but no correlations with known Paleozoic rocks have been made (Hobbs and Cookro, 1987, p. 18). The pendants are exposed on ridge tops and are thought to be at or near the top of the batholith.

The western margin of the Atlanta lobe is juxtaposed against oceanic island-arc rocks of Permian and Triassic age (Hamilton, 1963; Brooks and Vallier, 1978; Myers, 1982; Lund, 1984). Metavolcanic and metasedimentary rocks of the Wallowa-Seven Devils terrane, which have been intruded by Triassic and Jurassic plutons, and the structurally superimposed Riggins Group are now thought to have been accreted to the continent along a right-lateral transcurrent fault during the Cretaceous (Lund, 1984; Sutter and others, 1984; Criss and Fleck, 1987; Vallier and Brooks, 1987, p. 5).

The southwestern and southern margins of the Atlanta lobe are covered by younger volcanic and sedimentary rocks. Extensive flows of Miocene Columbia River Basalt Group mask intrusive relations at the southern end of the western margin. The southern margin is similarly obscured by Miocene and younger volcanic rocks of the Snake River Plain.

The western margin of the Idaho batholith is approximately coincident with the edge of the pre-Mesozoic continental margin. Armstrong and others (1977) recognize a distinct break in the initial 87Sr/86Sr ratios for igneous rocks of this area. East of a north-south line along the western edge of the batholith (Fig. 1), initial Sr ratios are greater than 0.7035. West of the same line they are less than 0.7043. Fleck and Criss (1985) showed that in detail a continuum of 87Sr/86Sr ratios exists across this line, but that the entire transition zone is typically no more than several hundred meters to a few kilometers wide.

Host rocks for Eocene intrusive rocks include all phases of the Cretaceous batholith, Paleozoic metasedimentary rocks east of the batholith, and older Eocene rocks of the Challis Volcanic Group (Fisher and others, in press). Ross (1934) recognized two groups of volcanic rocks in the Casto 30-minute quadrangle. Cater and others (1973) observed that Ross' "Casto Volcanics" were probably altered Challis Volcanic Group and formally abandoned the name "Casto Volcanics." McIntyre and others (1982) and Ekren (1985) have shown that the Challis Volcanic Group resulted from a series of relatively quiet eruptive events and large violent caldera-forming eruptions between 51 and 45 Ma. Thick sections of calc-alkaline intermediate lava and dacitic-rhyolitic tuff are exposed in the eastern half of the Challis and Hailey 1 x 2 degree quadrangles and the western half of the Idaho Falls 1 x 2 degree quadrangle (Moye and others, 1988, this volume).

Cretaceous Granitic Rocks

The Atlanta lobe of the Idaho batholith is 170 miles (275 km) long and 80 miles (130 km) wide and consists of six main rock types: tonalite, hornblende-biotite granodiorite, porphyritic granodiorite, biotite granodiorite, muscovite-biotite granite, and leucocratic granite (Fig. 2). Most of these units are exposed over large
Figure 2. Simplified geologic map of Cretaceous and Tertiary intrusive rocks in and near the Atlanta lobe of the Idaho batholith. Compiled from Fisher and others (in press), Bond (1978), and unpublished mapping by the authors.
areas. The Atlanta lobe lacks the numerous small, discrete granitic plutons that are characteristic of other large batholiths. Instead, the plutons are very large and contacts between them are gradational, even where exposures are good. Except for leucocratic granite, sharp contacts are rare.

Application of radiometric dating techniques in the Cretaceous rocks has met with only limited success. Most ages available were done with potassium-argon methods, and they give only minimum ages of emplacement. Some are too young due to argon loss during Eocene reheating (Armstrong, 1974), and all record only the time since the rock cooled past the argon blocking temperature of the mineral dated. For these reasons, Figure 3 shows both the range of apparent ages and our best estimates, based on both geochronology and field relations, of actual emplacement ages.

Brief summaries of field relations and lithologies of the six major rock types, as they are exposed in the central Atlanta lobe, are presented below. More detailed descriptions can be found in Kiilsgaard and Lewis (1985) and Lewis and others (1987), which form the basis for these summaries. In this report, as in the two cited reports, granitic rocks are classified according to the recommendations of Streckeisen and others (1973).

**EXPLANATION**

<table>
<thead>
<tr>
<th>Mineral dated</th>
<th>Emplacement age</th>
<th>Range of apparent ages</th>
</tr>
</thead>
<tbody>
<tr>
<td>Biotite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Muscovite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hornblende</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Zircon</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Biotite (previous studies)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Figure 3. Radiometric ages for tonalite (Kt), porphyritic granodiorite (Kgdp), biotite granodiorite (Kgd), muscovite-biotite granite (Kg), and leucocratic granite (Klg) in the central part of the Atlanta lobe of the Idaho batholith. Data from Lewis and others (1987) (labeled "this study") and Criss and others (1982, tables 1 and 2). Biotite, muscovite, and hornblende were dated by the potassium-argon method. Zircon age was determined by uranium lead systematics. From Lewis and others (1987).**

**Tonalite**

Tonalite has been mapped on the northwestern margin, in the eastern part, and on the western margin of the Atlanta lobe (Fig. 2). Field and radiometric evidence indicates that tonalite is the oldest of the batholithic rocks (Fig. 3). The rock is typically dark gray, medium-grained, and equigranular; a foliation is defined by biotite and varies from faint to intense. Hornblende generally makes up 2-8 percent of the rock, biotite 5-15 percent; sphene and magnetite are common accessory minerals. Modal plots (Fig. 4) show this unit to be tonalite grading into granodiorite. Mean values for major element oxides are shown in Table 1.

**Hornblende-biotite Granodiorite**

Hornblende-biotite granodiorite is found in small exposures on ridges near the center of the Atlanta lobe as well as in larger masses in the southeastern part of the lobe. This rock is similar in appearance to the tonalite, including a slight foliation, but contains more potassium feldspar and less hornblende. Modally most samples plot in the granodiorite field (Fig. 4). Intrusive relations are not entirely clear, but this unit is thought to represent a slightly older phase than the main mass of biotite granodiorite. No radiometric ages have been obtained for this unit.

**Porphyritic Granodiorite**

Porphyritic granodiorite crops out on ridge tops in a northwest-trending belt about 75 miles (125 km) long in the northeastern Atlanta lobe (included with "tonalite" on Fig. 2). The rock is gray, medium- to coarse-grained, and contains pink or light pinkish-gray microcline megacrysts as much as 4 inches (10 cm) in length. Modal analyses of these rocks plot in the granodiorite and granite fields (Fig. 4). Biotite is the primary mafic constituent, except in the area north and east of Stanley where up to 5 percent hornblende is present in many places. Hornblende-bearing rock resembles the tonalite described above, except for the presence of megacrysts in the granodiorite. Rock lacking hornblende more closely resembles biotite granodiorite. The megacrysts result from potassium metasomatism affecting both tonalite and biotite granodiorite, especially near the roof of the batholith. Numerous veins of potassium-rich feldspar cut the rock and appear to have been conduits for potassium-rich fluid.

**Biotite Granodiorite**

Biotite granodiorite is the most extensive rock type in the Atlanta lobe. Field relations show that it is younger than both tonalite and hornblende-biotite granodiorite. Biotite granodiorite is gradational into younger muscovite-biotite granite over a distance of 1.25 miles (2
km) or more; together these units comprise the main phase of the Atlanta lobe.

Biotite granodiorite is light gray, medium- to coarse-grained, and locally porphyritic with potassium-feldspar phenocrysts up to 1 inch (2.5 cm) in length. Foliation is rare. Biotite is generally 2-8 percent of the rock; hornblende and primary muscovite are rare. Modal plots (Fig. 4) show this unit to consist of granodiorite and granite.

Muscovite-biotite Granite

Muscovite-biotite (two-mica) granite is present as the discontinuous north-south trending core of the Atlanta lobe (Fig. 2). It is massive, light gray, medium- to coarse-grained, and equigranular to slightly porphyritic. The presence of macroscopic plates or books of primary muscovite is the primary criterion for mapping this unit. Muscovite comprises as much as 5 percent of the rock, but typically averages about 2 percent, as does biotite. Modal plots show that the unit ranges from granite to granodiorite (Fig. 4). Radiometric age data (Fig. 3) support the field evidence that this unit is in part slightly younger than biotite granodiorite.

Leucocratic Granite

Leucocratic granite is the youngest phase of the Atlanta lobe. It occurs as dikes, sills, and irregular small stocks that sharply crosscut all earlier phases. Potassium-argon ages on biotite range from 76 to 64

Figure 4. Modal analyses for samples of plutonic rocks of the Atlanta lobe of the Idaho batholith, Challis quadrangle. Analyses normalized to 100 percent quartz (Q) + potassium feldspar (A) + plagioclase (P). Open circles show mean values. A, rock classification from Streckeisen and others (1973); B, tonalite; C, porphyritic granodiorite; D, biotite granodiorite; E, muscovite-biotite (two-mica) granite; F, leucocratic granite. From Kilsgaard and Lewis (1985).

<table>
<thead>
<tr>
<th>Rock type</th>
<th>No. of samples</th>
<th>SiO$_2$</th>
<th>Al$_2$O$_3$</th>
<th>Fe$_2$O$_3$*</th>
<th>MgO</th>
<th>CaO</th>
<th>Na$_2$O</th>
<th>K$_2$O</th>
<th>TiO$_2$</th>
<th>P$_2$O$_5$</th>
<th>MnO</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Cretaceous</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Leucocratic granite</td>
<td>11</td>
<td>75.3</td>
<td>14.0</td>
<td>0.73</td>
<td>0.1</td>
<td>0.78</td>
<td>4.3</td>
<td>4.11</td>
<td>0.05</td>
<td>0.05</td>
<td>0.02</td>
</tr>
<tr>
<td>Musc-biotite granite</td>
<td>12</td>
<td>73.8</td>
<td>14.6</td>
<td>0.92</td>
<td>0.3</td>
<td>1.41</td>
<td>3.8</td>
<td>3.47</td>
<td>0.10</td>
<td>0.04</td>
<td>0.03</td>
</tr>
<tr>
<td>Biotite granodiorite</td>
<td>22</td>
<td>71.9</td>
<td>15.2</td>
<td>1.55</td>
<td>0.4</td>
<td>1.91</td>
<td>4.1</td>
<td>3.30</td>
<td>0.21</td>
<td>0.07</td>
<td>0.03</td>
</tr>
<tr>
<td>Hb-bio granodiorite</td>
<td>3</td>
<td>66.5</td>
<td>15.5</td>
<td>3.79</td>
<td>1.7</td>
<td>3.98</td>
<td>3.3</td>
<td>2.97</td>
<td>0.63</td>
<td>0.21</td>
<td>0.07</td>
</tr>
<tr>
<td>Tonalite</td>
<td>11</td>
<td>64.0</td>
<td>16.9</td>
<td>4.40</td>
<td>1.6</td>
<td>4.64</td>
<td>3.7</td>
<td>2.36</td>
<td>0.77</td>
<td>0.26</td>
<td>0.07</td>
</tr>
<tr>
<td><strong>Eocene</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pink granite</td>
<td>20</td>
<td>75.6</td>
<td>13.1</td>
<td>1.19</td>
<td>0.2</td>
<td>0.58</td>
<td>3.3</td>
<td>4.87</td>
<td>0.11</td>
<td>0.05</td>
<td>0.02</td>
</tr>
<tr>
<td>Diorite complex</td>
<td>25</td>
<td>62.9</td>
<td>16.5</td>
<td>4.83</td>
<td>2.2</td>
<td>4.44</td>
<td>3.5</td>
<td>3.22</td>
<td>0.74</td>
<td>0.26</td>
<td>0.08</td>
</tr>
</tbody>
</table>

* Total iron reported as Fe$_2$O$_3$.

Ma; emplacement age is probably 75-70 Ma (Fig. 3). The granite is light gray, fine-grained and equigranular; it contains 2 percent or less biotite and trace amounts of primary muscovite. Modal analyses plot in the granite field (Fig. 4).

**Tertiary Intrusive Rocks**

Eocene intrusive rocks of south-central Idaho are a bimodal suite of granite and quartz monzodiorite. Each plutonic type has equivalent hypabyssal dikes and volcanic units that are part of the Challis Volcanic Group. Hypabyssal dikes include rhyolites (equivalent to granites) and dacite-rhyodacites (equivalent to quartz monzodiorites). Volcanic rocks include rhyolite flows and ash flow tuffs (equivalent to granites) as well as the preserved in the Custer graben, and dacite-rhyodacite flows and tuffs (equivalent to quartz monzodiorites) such as the tuffs of Black Mountain and Camas Creek. The flows and tuffs were erupted from calderas and calderon complexes including the Twin Peaks caldera, the Thunder Mountain caldron complex, and the Van Horn Peak caldron complex in the Challis 1 x 2 degree quadrangle (McIntyre and others, 1982; Fisher and others, in press) and in the Porphyry Peak and Sheep Mountain areas in the Idaho Falls 1 x 2 degree quadrangle (Moye and others, 1988, this volume). These volcanic packages overlie older Eocene andesite-dacite flows that are much more widespread than the younger units. The older andesite-dacite units apparently came from many small volcanic vents. Plutonic representatives of each group have about equal areal extent; granite plutons are larger and diorite exposures are smaller but more numerous. Detailed descriptions of these rocks can be found in Bennett and Knowles (1985) and Kilsgaard and Bennett (1987a), on which the following summaries are based. Classification follows Streckeisen and others (1973).

**Quartz Monzodiorite**

Rocks collectively called quartz monzodiorite range in composition from gabbro to granite (shown as granodiorite and diorite on Fig. 2). The most prevalent rock types are quartz monzodiorite to porphyritic granodiorite and granite. Modal plots show a range from gabbro to granite (Fig. 5). All are characterized by hornblende, euhedral biotite, magnetite, and complexly zoned plagioclase. The more mafic varieties weather to distinctive chocolate brown soils. Potassium-argon ages for units in the Challis quadrangle range from 45 to 50 Ma (Fisher and others, in press).

The largest exposure of these rocks occurs in the north-central Challis 1 x 2 degree quadrangle where a complex of small dioritic stocks and swarms of associated fine-grained, gray to green, nonporphyritic to porphyritic dacite to rhyodacite dikes are exposed along the canyon of Rapid River. Stocks and dikes both intrude the Cretaceous batholith.

**Granite**

Eocene granite of south-central Idaho is typically coarse-grained and pink. Although many of the approximately forty-five Eocene granite plutons identified in
Idaho have this pink color, neither all phases of any pluton nor all plutons are pink. The pink color is probably due to finely disseminated hematite in the potassium feldspar. Miarolitic cavities, indicative of shallow epizonal crystallization, contain crystals of microcline, smoky quartz, and less commonly beryl, topaz, and fluorite. Average uranium, thorium, and potassium-40 contents, as measured by gamma-ray spectrometry, are 2-3 times higher than in Cretaceous granitic rocks. Modal compositions of these rocks plot entirely within the granite field. Potassium-argon ages from the Challis quadrangle range from 42 to 49 Ma (Fisher and others, in press).

Another distinctive feature of Eocene granite is planar, high-angle jointing. Many spectacular peaks in central Idaho are in steeply jointed Eocene granite, which weathers to sharp peaks and steep topography. This contrasts strongly with the Cretaceous batholith, which has a more subdued topographic expression and concordant elevations.

Trans-Challis Fault System

The trans-Challis fault system (Fig. 6) is described by Kilsgaard and others (1986) as a broad, northeast-trending structural system along which major movement occurred during Eocene extension. The fault system includes a broad set of northeast-trending high-angle faults, aligned grabens, a caldera, and roughly aligned Eocene intrusions and dike swarms. It has been traced for at least 165 miles (270 km) from the vicinity of Idaho City to beyond Leesburg, and is aligned with and appears to be a continuation of the Great Falls tectonic zone described by O'Neill and Lopez (1985) in Montana.

Mineral Deposits

Many mineral deposits in central Idaho are spatially and genetically related to the trans-Challis fault system (Fig. 6). Some of these deposits are hosted in granitic rocks of the Idaho batholith, commonly near or partially within Tertiary dikes. Other host rocks include Eocene plutonic rocks, dikes that cut the plutonic rocks, intrusive Eocene rhyolite, Precambrian sedimentary rocks, roof pendants in the Idaho batholith, and pyroclastic rocks of the Challis Volcanic Group. These deposits typically have more silver than gold, but gold has been the principal metal in terms of value. Other by-products include copper, lead, and zinc. Several molybdenum deposits are known within the fault system, but none have been brought to production.

There are also numerous mineral deposits in central Idaho that are associated with Cretaceous or Tertiary intrusive rocks but are not within the trans-Challis fault system. One such deposit is the low-fluorine granodiorite-hosted molybdenite deposit at Thompson Creek. The Little Boulder Creek deposit in the White Cloud Peaks is a molybdenum stockwork in contact-metamorphosed limy sandstone of the Pennsylvanian-Permian Wood River Formation (Hall, 1985). Most mining in the central Idaho black shale belt, shown as Paleozoic sedimentary rocks on Figure 2, has been from lead-silver-zinc veins (Hall, 1985). These veins are found in black shale near contacts with granitic intrusive rocks of predominantly Cretaceous but also Eocene age. In the southern part of the black shale belt near Hailey and Bellevue the principal sulfide minerals are galena, bournonite, sphalerite, and pyrite. In the northern part of the belt, in the southern part of the Challis 1 x 2 degree quadrangle, the principal ore mineral is jamesonite; galena, sphalerite, chalcopyrite, pyrite, pyrrhotite, and arsenopyrite are also common sulfide minerals. Another significant deposit is at the Ima mine, east of the area of this trip in the Lemhi Range (in the Idaho Falls 1 x 2 degree quadrangle). There large veins of huebnerite with associated molybdenite and silver sulfides occur in quartzite at the contact between quartzite and granite (Anderson, 1948).

Purpose of Trip

Recent work has added much to our understanding of the Cretaceous and Tertiary intrusive history of south-central Idaho. Field trip participants will see the major rock types of the Cretaceous batholith, both felsic and mafic rocks of Eocene age, major structural features crosscutting all these rock types, and Cretaceous and Tertiary molybdenum deposits. In addition, we hope to provide an overview of the geologic setting, other important rock units, and mineral deposits of the area.
ROAD LOG, DAY 1

This trip travels along the east side of the Atlanta lobe of the Idaho batholith to look at Eocene granite, which is rarely exposed along roads, Cretaceous porphyritic granodiorite and biotite granite, and the Thompson Creek molybdenum deposit. Along the way we will discuss the geologic setting for the east side of the batholith, major structural features, and nearby mineral deposits.

The trip route and named localities are shown on Figure 7.

Mileage

Interval mileage is shown in parentheses.

0.0  (0.0) Intersection of Sun Valley Road and Main Street, Ketchum. Leave Ketchum going north on State Highway 75, heading towards Galena Summit. Leaving Ketchum, pass cemetery where Ernest Hemingway is buried (on right).

The rocks visible here are lower parts of Eocene Challis Volcanic Group. Road travels up the valley of the Wood River, which flows along
Figure 7. Map showing route for Day 1 and locations of places and features named in the text.
the west side of the Wood River graben. Graben faulting has preserved the volcanic flows and is related to Miocene basin and range extension. Peak ahead is Glassford Peak, elevation 11,602 feet (3,516 m), exposing Challis Volcanic Group over Pennsylvanian-Permian Wood River Formation and other upper Paleozoic rocks.

3.1 (3.1) Good exposures of lower Challis Volcanic Group on the right (east). Rocks are rhyodacite and dacite flows, as mapped by Hall and others (1978).

3.7 (0.6) More Challis Volcanic Group; dacitic and rhyodacitic flows.

7.9 (4.2) Headquarters of Sawtooth National Recreation Area (SNRA). Maps, information, and public restrooms. SNRA has tapes available for loan describing the drive between here and Stanley.

8.3 (0.4) Valley narrows here and then widens again to the north as a result of side-stepped en echelon graben faulting.

12.4 (4.1) Sign on right pointing out Boulder Mountains. This view of the Boulders shows Eocene pink granite (lower slopes) intruding andesites of the Challis Volcanic Group and Eocene dacite porphyry (Fig. 8).

14.8 (2.4) Exposures of Challis Volcanic Group.

14.9 (0.1) Cross Big Wood River.

16.6 (1.7) Small turnout to right, from which Figure 8 was taken. Fault across front (west side) of Boulder Mountains is the same style of faulting as that responsible for the Wood River graben (basin and range).

20.3 (3.7) Cross Big Wood River. Challis Volcanic Group (dominantly rhyodacite and andesitic flow rocks) in roadcuts from here to Stop 1.

30.0 (9.7) Galena Summit, 8,701 feet (2,652 m), and Milepost 158.

30.9 (0.9) Stop 1. Galena Summit overlook. The view from here is to the north into the Sawtooth Valley (Stanley Basin) and along the east side of the Sawtooth Mountains. The headwaters of the Salmon River are in the drainage immediately west of this overlook. The Sawtooth Valley, another basin and range structure, is a half-graben with the Sawtooth fault running along its west side. The southern end of the Sawtooth Mountains is Cretaceous biotite granodiorite and Eocene pink granite is exposed to the north. Rocks in the range have been mapped by Reid (1963) and Kilsgaard and others (1970).

The Sawtooth batholith is the most prominent of the Eocene pink granites in south-central Idaho. The batholith is exposed in a horst bounded by the Sawtooth fault on the east. The Montezuma fault bounds the western side of the horst. These two northwest-trending faults are good examples of the expression of basin and range structures north of the Snake River Plain. Both faults terminate against major northeast-trending structures that are part of the trans-Challis fault system (Kilsgaard and others, 1986). The trans-Challis fault system underwent major displacement during the Eocene and hosts many precious metal deposits.

Rocks exposed on Galena Summit are Challis Volcanic Group.

Restrooms available here.

31.0 (0.1) Highway Milepost 159.

32.9 (1.9) Turnout to left. The Sawtooth fault and other basin and range structures in Idaho cross-cut older northeast-trending structures south of the trans-Challis fault system. The northeast-trending faults had major movement during Eocene extension but may be older in places. The bleached zone on the north side (west from here) of Smiley Creek may be one of the northeast-trending structures that terminates surface exposures of the Sawtooth batholith. Aeromagnetic data indicate that the batholith is present at depth to the south of Smiley Creek.

36.8 (3.9) Hills ahead and to the right (east) of the
Sawtooth Valley are biotite granodiorite and porphyritic granodiorite of the Cretaceous Idaho batholith.

37.0 (0.2) Sawtooth City limits. High bluff to the east is part of the type section for Permian Grand Prize Formation (Hall, 1985).

37.5 (0.5) Smiley Creek Lodge on left.

38.8 (1.3) Sawtooth City. Discovery of silver in the Vienna district in 1878 led to the establishment of two towns, Vienna on Smiley Creek and Sawtooth City on Beaver Creek. Major production up to 1885 was from the Silver King and Pilgrim mines on Beaver Creek and the Vienna and Mountain King (Webfoot) mines on Smiley Creek. Mills for processing ore were built near each town. Most mining activity was over by 1890 although mines have operated intermittently since, including the Vienna (1915-1917 and 1933-1943) and the Silver King (1937-1941). Several companies have explored the area in recent years but there has been no major mining activity. Partial production records (since 1933) indicate about 4,500 tons of ore yielding almost 60,000 ounces of silver and 88,000 pounds of lead. Earlier production was more substantial; Umpleby (1915) reports $60,000 in production from the Sawtooth City mines in the autumn of 1883 and $200,000 from the Vienna mine in 1884. Estimates of total production are about $500,000 for Vienna and less than half that for Sawtooth City.

The mines are well-known for the beautiful samples of massive and well-crystallized ruby silver that accounted for some fabulous assays of 3,000-5,000 ounces of silver per ton. Other ore minerals include tetrahedrite, galena, sphalerite, and stibnite. The ore occurs in veins in shear zones but mineralization is very spotty and discontinuous. A projection of the Willow Creek-Bear Creek-Emma Creek northeast-trending fault extends right through the Vienna district. [History and production figures are modified from Wells (1983) and Shannon (1971)].

44.0 (5.2) Highway Milepost 172. White Cloud Peaks visible to the northeast. The White Cloud peaks are white calcsilicates resulting from hydrothermal alteration of Paleozoic rocks (consisting mostly of units in the Pennsylvania-Permian Wood River and Permian Grand Prize Formations) near the White Cloud stock. The stock is dated at 84.7 ± 1.9 Ma (Lewis and others, 1987). Alteration is concentrated along and was channeled by several major thrust faults in the Paleozoic rocks that preceded intrusion of the stock. This relation indicates that major thrusting in eastern Idaho was at least partially completed when the batholith crystallized. The Little Boulder Creek molybdenum deposit is located on the east side of the stock and is related to it.

Straight west from here, the intrusive contact between pink (Eocene) granite and gray (Cretaceous) biotite granodiorite is visible in the valley wall. This relation is even better seen from the Fourth of July Ranger Station (east of the highway at Fourth of July Creek).

44.6 (0.6) Cross Salmon River.

46.7 (2.1) Cross Fourth of July Creek. Large moraine to west. Both Pinedale and Bull Lake glaciations affected the Sawtooth Mountains. See Pearce and others (1988, this volume) and Breckenridge and others (1988, this volume) for more information about the glacial deposits and history of this area.

50.2 (3.5) Pass town of Obsidian (mostly on the right).

51.6 (1.4) Cross Williams Creek. First exposures of Cretaceous batholith on right (east). These rocks are biotite granodiorite.

55.6 (4.0) Pass Sawtooth Fish Hatchery on left. Designed to help rebuild the steelhead and salmon populations of the Salmon River.

56.4 (0.8) Cross Salmon River. Begin climbing moraine that forms the east side of Redfish Lake.

57.1 (0.7) Highway Milepost 185. Turn left (west) onto Redfish Lake road. This road travels along the moraine to a large complex of Forest Service campgrounds, lodge, and other tourist facilities.

57.9 (0.8) Sign on right giving name of prominent peak visible straight ahead--Mt. Heyburn, 10,299 feet (3,121 m), composed of pink granite.

58.9 (1.0) Turn right to Redfish Lake Lodge. Road straight ahead goes to campgrounds, where it ends.

59.4 (0.5) Stop 2. Redfish Lake Lodge. Coffee break here. Good opportunity to look at Eocene pink granite exposed in moraine boulders behind the lodge. Good exposures of these rocks do not occur along highways of today's trip, so this is the best opportunity to examine them. Return to State Highway 75 by the same route.
61.7 (2.3) Turn left onto State Highway 75, immediately cross Redfish Lake Creek. Rocks exposed across the Salmon River are Cretaceous biotite granodiorite.

63.6 (1.9) Highway Milepost 187.

64.7 (1.1) Peaks visible on the high skyline ahead are part of the Cabin Creek stock (Cabin Creek Peak). This is a small Tertiary granite stock that shows characteristics of both granite and rhyolite. East of and behind these peaks is the down-dropped block of the Custer graben (McIntyre and others, 1982; Fisher and others, in press).

66.0 (1.3) Intersection with State Highway 21 west. Continue on State Highway 75 toward Lower Stanley.

66.6 (0.6) Cross Valley Creek. Good view of Sawtooth Mountains to the southwest (Fig. 9). The high peaks in this northern part of the range are composed largely of metasedimentary rocks thought by Reid (1963) to be Precambrian in age. They are shown as "roof pendants of uncertain age" by Fisher and others (in press). Traveling downstream (north and east) from here, we will find faulted, altered Cretaceous biotite granodiorite until we pass Mormon Bend.

72.8 (1.4) Mormon Bend Campground. Here we find the first large pink potassium feldspar megacrysts in biotite granodiorite.

74.0 (1.2) Cross Basin Creek. During the uranium boom of the 1950s, uranium was discovered on Basin Creek. Deposits include pitchblende veins in granitic rocks of the Idaho batholith, and coffinite and uraninite in carbonaceous material in the basal conglomerate of the Challis Volcanic Group that was deposited on the rocks of the batholith.

Veins may or may not contain quartz; some contain minor amounts of precious metals, molybdenum, stibnite, and pyrite. Carbonaceous material that had been concentrated in small lakes, swamps and streams fixed uranium as coffinite and uraninite. The carbonaceous deposits also contain some pyrite and marcasite. Choate (1962) suggests that the veins were the source of the uranium that is in the sedimentary deposits.

77.6 (3.6) Stop 3. Highway Milepost 201. Pull off into large parking area after the "Deer Crossing" sign. Please watch for traffic.

Road cuts show metasomatically altered porphyritic granodiorite. There are many exposures of this rock type northwest from here as far as the western edge of the Idaho batholith. Large pink potassium feldspars up to 4 inches (10 cm) are scattered throughout the matrix (Fig. 10). Dating of the tonalitic matrix yields ages of 88 ± 6 Ma (zircon), 84.7 ± 2.9 Ma (hornblende), and 79.8 ± 2.7 Ma (biotite), placing the rock in the group of older plutons in the batholith.

67.2 (0.6) Pass through Lower Stanley.

71.4 (4.2) Little Casino Creek road takes off to the right. Fluorite deposits of Eocene age are found in both Big and Little Casino Creeks. The Giant Spar mine in Little Casino Creek produced only a few truckloads of fluorspar, but the mine is typical of fluorspar deposits in the area. These are vein deposits (quartz-fluorite or quartz-pyrite-fluorite-gold) that are believed to be related to rhyolite dikes and structurally related to the trans-Challis fault system.

77.7 (3.6) Stop 4. Highway Milepost 201. Please watch for traffic.
78.0 (0.4) Sunbeam Hot Springs. This hot springs occurs on a large east-west fault in rocks of the Cretaceous batholith.

79.0 (1.0) Turn left onto the Yankee Fork road. Placer gold was discovered at the confluence of Yankee Fork Salmon River and Jordan Creek (8 miles (13 km) north of here) in 1873. Later lode discoveries gave rise to the Yankee Fork district and the mining communities of Custer and Bonanza. The district produced in excess of $12,000,000. in gold, silver, copper, lead, and zinc, mostly before 1911. Tertiary andesites, pyroclastic rocks, and rhyolites are hosts to vein and disseminated mineral deposits. One of the most famous mines was the General Custer, where the hanging wall was eroded off, leaving the rich vein exposed on the surface. The discoverers made $60,000 in 1876 from the surface showing, and then sold their claims for $285,000.

An 8-cubic-foot dredge operated on the Yankee Fork for a total of eight years in the period 1940-1952. J. R. Simplot, well-known Idaho entrepreneur, was the last to own and operate the dredge. Gold recovery was estimated at 60 percent, and total production was almost 30,000 ounces. The dredge is maintained today as a museum and tourist attraction. See Stephens (1988, this volume) for more discussion of the history of placer mining on the Yankee Fork.

Some properties in the area are still being worked. U. S. Antimony Corporation operates a mill at Preachers Cove, and Sunbeam Mining Company is developing a closed vat, cyanide leach facility at the old Sunbeam mine.

79.7 (0.7) Blind Creek Campground. Bus will turn around here and stop at talus slopes on the west side of the road.

Stop 4. These rocks are Cretaceous biotite granodiorite, dated at 81.3 ± 2.9 Ma by potassium-argon dating of biotite from a sample collected about 0.75 mile (1.2 km) farther in on the Yankee Fork road (Lewis and others, 1987). Although not connected at the surface, these rocks are identical in appearance to the main mass of biotite granodiorite to the west.

80.6 (0.9) Return to State Highway 75, turn left, and cross the Yankee Fork at its confluence with the main Salmon River. Sunbeam Dam was constructed in 1909-1910 to provide power for mining operations at Custer. It is the only dam ever built on the main Salmon River and was dynamited in 1934.

83.4 (2.8) Historical sign about salmon spawning beds (on right). Here we are again passing through approximately 81-million-year-old biotite granodiorite.

89.2 (7.6) Highway Milepost 211. Contact with Paleozoic Salmon River assemblage, a sequence of weakly metamorphosed limestone turbidites interbedded with dark gray argillite, siltite, shale, and fine-grained quartzite estimated to be over 6,000 feet (1,800 m) thick. Units with Cambrian, Devonian, and Mississippian fossil ages are intercalated in the assemblage and relations are complicated by thrust faulting. This poorly understood unit is part of the central Idaho black shale belt (Hall, 1985).

91.2 (2.0) Highway Milepost 213. Slate Creek enters the Salmon from the south. Approximately 100 feet (30 m) of black gouge in rocks of the Salmon River assemblage is visible on the north side of the road. This northeast-trending fault extends up Slate Creek to the south. Cross Salmon River.

92.7 (1.5) At this big bend in the river, Salmon River assemblage is exposed in the road cuts and river gravels are perched on the terrace above. Fisher and May (1983) sampled this part of the Salmon River assemblage for its potential as a vanadium resource. They found 26 percent of the samples contained anomalous vanadium (>1,000 ppm) and most samples enriched in vanadium were also enriched in molybdenum, silver, and zinc.

93.9 (1.2) Thompson Creek enters Salmon River from the north.

95.3 (1.4) Pass Yankee Fork Ranger Station.

96.2 (0.9) Highway Milepost 218. Pass French Creek.

97.8 (1.6) Second bridge to Thompson Creek. Turn left, cross the river, and drive up the mine access road.

101.9 (4.1) Second creek crossing and entrance to Thompson Creek mine (on left).

Stop 5. The Thompson Creek deposit was discovered in 1961 during a regional stream-sediment sampling program conducted by Union Carbide as part of a tungsten exploration program. Geologists followed up a 5 ppm molybdenum anomaly and discovered an outcropping of molybdenite in the poorly exposed Pat Hughes stock. As Carbide had no interest in the property, the claims lapsed and were restaked by Cyprus Minerals Company in 1968. Cyprus explored the property and conducted bulk
sampling for molybdenum from an 8,000-foot(2,400-m)-long exploration drift. Cyprus merged with Standard Oil of Indiana in 1979 and announced plans for a 25,000 ton-per-day open pit mine that would be the largest mine in Idaho and would produce about 10 percent of the world's molybdenum.

Preproduction costs are estimated at $398 million including the following items: the construction of a 25,000 ton-per-day mill, a 7,300-foot(2,200-m)-long conveyer (Fig. 11) and primary crusher to transport ore from mine to mill, and mining equipment along with a fleet of twenty-two 170-ton trucks, four 25-cubic-yard shovels, and four blast-hole drills. Other costs included a tailings impoundment dam that will be over 500 feet (150 m) high if mining continues for the estimated 20 year mine life. The company also built about 260 new homes in Challis for employees and has prepaid taxes for new schools. The population of Custer County swelled from about 3,400 in 1980 to over 6,500 in 1984. Over $30 million was spent on environmental concerns.

End of roadlog for day 1. Bus will return on State Highway 75 to Sun Valley, a distance of 93 miles.

ROAD LOG, DAY 2

The second day of the trip takes us around the south end of the Atlanta lobe. Stops are planned to look at several plutons of uncertain age, the effects of basin and range faulting, a weathered Eocene granite whose miarolitic cavities contain a variety of minerals, and the complex relations between the Cretaceous batholith and the House Mountain metamorphics. In addition we will pass through the rich Wood River mining district and see numerous exposures of Eocene and Miocene volcanic rocks. The structural geology and stratigraphy of the Wood River Valley are described by Link and others (1988, this volume).

The trip route and named localities are shown on Figure 12.

Mileage

Interval mileage is shown in parentheses.

0.0 (0.0) Intersection of Sun Valley Road and Main Street, Ketchum. Leave Ketchum on State Highway 75, heading south toward Hailey and Bellevue.

Ketchum was founded in 1880 and originally called Leadville. It was later named for David Ketchum, early pioneer and prospector who discovered the mineral deposits in the upper Wood River Valley.

0.2 (0.2) Cross Trail Creek.

0.4 (0.2) Highway Milepost 128.

1.5 (1.1) Elkhorn Road. Rhyodacite/dacite flows of Challis Volcanic Group are exposed on the left.

2.0 (0.5) Cross Big Wood River.
4.3 (2.3) Limy sandstones of Pennsylvanian-Permian Wood River Formation, unit 6, exposed on right (west) side of road (Batchelder and Hall, 1978).

6.1 (1.8) Cross Big Wood River.

10.1 (4.0) Hailey Conglomerate (lowermost Wood River Formation) exposed in cliffs on right (west).

11.4 (1.3) Entering Hailey (sign). Town was named for John Hailey, who started a trading center for his Utah, Idaho, and Oregon Stageline, which served the Wood River mines in 1880.

14.3 (2.9) Hills to the left (east) are phyllites and argillites of the Devonian Milligen Formation in thrust contact with overlying limy sandstones of the Pennsylvanian-Permian Wood River Formation (Hall, 1985).

Hills to the right (west) are black argillites of the Permian Dollarhide Formation in thrust contact with the Devonian Milligen Formation (Hall, 1985). (Link and others (1988, this volume) offer a different interpretation of this relation.) Farther west, at the top of the ridge, is the Croesus stock (Lindgren, 1900; Anderson and others, 1950). The stock is pyroxene-bearing quartz monzodiorite of uncertain age. It is intruded on its western side by the Hailey granodiorite unit of Schmidt (1962). Biotite from a sample of the Hailey granodiorite from the McCoy mine was dated by the potassium-argon method at 83.8 ± 2.5 Ma (recalculated from Berry and others, 1976).

Mine dump and cuts in the hills to the right (west) are part of the Silver Star-Queen mine and the adjoining Minnie Moore mine (Umpleby and others, 1930; Anderson and others, 1950). The Minnie Moore vein, which appears to be a mineralized thrust fault, was discovered in 1880. High-grade lead-silver ore was mined for the next nine years. Activity resumed in 1890, but there was no significant production after 1906. The Silver Star-Queen (originally Queen of the Hills) mine was operated in the late 1800s and was later reopened and worked from 1949 until the early 1970s. Both mines are polymetallic vein deposits in black argillites of the Permian Dollarhide Formation (Hall, 1985).

A smelter was built in Hailey in 1880 by the
Wood River Smelting Company. It was dwarfed by the Philadelphia Smelter, built in Ketchum the next year, which installed the first electric light system in Idaho.

The Triumph mine northeast of Hailey was discovered in 1884, developed in 1927 and worked until the 1950s. Also a polymetallic lead-silver vein deposit, it produced more value in ore than all the other Wood River mines combined.

17.3  (3.0) Road to Gannett forks off to the left.

23.6  (6.3) Cross Baseline Road. This is the point of origin for the Idaho land survey system.

26.2  (2.6) Intersection of U. S. Highway 20 and State Highway 75. Turn right (west) onto U. S. 20. Rest area at this intersection.

28.2  (2.0) Final crossing of Big Wood River.

28.6  (0.4) Step 1. Road cut is in rocks mapped by Schmidt (1962) as the Hailey granodiorite unit, a hornblende-biotite granodiorite of probable Cretaceous age. Most of the rocks exposed in this area are Miocene volcanics, related to Snake River Plain volcanism.

31.9  (3.3) Road to Hot Springs Landing (on Magic Reservoir) takes off to left.

32.6  (0.7) Outcrop of Miocene rhyolitic volcanic rocks.

33.6  (1.0) Entrance to Moonstone Ranch on right. Moonstone Mountain to the northwest is a rhyolite dome of Miocene age that contains opalescent feldspars.

35.6  (2.0) Entering Camas Prairie, which is thought to be the surface expression of a graben related to downwarp of the Snake River Plain (Cluer and Cluer, 1986).

37.0  (1.4) The flat-topped mountain to the north is Square Mountain, mapped by Schmidt (1962) as Square Mountain Basalt. More recent work by Leeman (1982) shows the rocks to be ferrolatite. Xenocrysts of quartz and feldspar and xenoliths of granulitic and granitic gneiss are common in this unit.

West of Square Mountain, hornblende-bearing Hailey granodiorite is in contact with biotite granodiorite further to the west. Relative ages of the two units are unknown. The biotite granodiorite is similar to the large masses of biotite granodiorite found elsewhere in the Cretaceous batholith.

39.2  (2.2) Camas County line. Old stage road to Hailey takes off to the northeast.

46.4  (7.2) Highway Milepost 158. Due north from here the highest tree-covered peak is Cannonball Mountain, underlain by Miocene rhyolite and basalt flows.

48.2  (1.8) Highway 46 intersection. Stay on U. S. 20. The low hills visible to the south are the Mount Bennett Hills, mostly Miocene tuffs and basalts of the Idavada Volcanics. Part of this area has been mapped by Smith (1966), and reconnaissance mapping of the region was done by Malde and others (1963).

52.3  (4.1) Turn right (north) at Fairfield.

54.1  (1.8) Old town of Soldier. Turn right (stay on pavement).

54.4  (0.3) Turn left. Prominent range to the northwest is the Soldier Mountains. The core is Eocene pink granite, and the margin is dominantly Eocene hornblende-biotite granodiorite. Bare hills in foreground are deeply weathered, altered Cretaceous biotite and hornblende-biotite granodiorite cut by numerous Eocene dikes, which make the only outcrops. Dark outcrops on low hills to northeast are Miocene basalts resting on Eocene Challis Volcanic Group.

58.5  (4.1) Wells Summit turnoff. Go left, remaining on pavement.

59.9  (1.4) Sheared Cretaceous biotite granodiorite with lots of aplite and pegmatite.

60.3  (0.4) More sheared Cretaceous rock.

60.5  (0.2) Flat-topped hill to right (east) is Miocene ash-flow tuff.

62.2  (1.7) Pavement ends. Take the right fork, heading for Big Smoky Guard Station.

63.3  (1.1) Exposures of lower part of Challis Volcanic Group on left are breccias and dacite/andesite flows. Cross Owens Creek.

63.7  (0.4) Poorly exposed Cretaceous biotite granodiorite on left.

64.4  (0.5) Big turn around. Bus will park here. Smaller vehicles can continue up the road to the stop.
64.4  (0.7)  Andesites and dacites of Challis Volcanic Group exposed on right.

64.7  (0.3)  **Stop 2.** Examine results of basin and range faulting. Large roadcut on right is faulted Cretaceous biotite granodiorite. Prominent exposure up the gulch is Eocene granodiorite. The northwest-trending fault drops Cretaceous granodiorite and overlying Eocene Challis Volcanic Group against Eocene hornblende-biotite granodiorite. The fault zone is at least 300 feet (100 m) wide, but the main fracture comes through the small saddle north of where the road crosses the gulch (see Fig. 13). Sheared Eocene granodiorite is poorly exposed along the road just past the gulch. All the plutonic rocks are cut by rhyodacite and dacite dikes of Eocene age.

The Eocene granodiorite here has very pink potassium feldspar; the plagioclase is white. In general, the Eocene granodiorites have smaller quartz grains than the Cretaceous intrusive rocks. Hornblende in the Eocene rocks is small and acicular in contrast to the large stubby grains found locally in the Cretaceous granodiorites. In addition, the potassium feldspar in the Eocene rocks shows well-developed microcline twins, generally lacking in the Eocene granodiorites. High peaks visible to the west from here are Eocene pink granite. Ridge in foreground is Eocene Challis Volcanic Group.

67.4  (2.7)  Return to pavement and follow it to U. S. Highway 20.

77.3  (9.9)  Intersection with U. S. 20; turn right (west).

82.5  (5.2)  Highway Milepost 147.

85.0  (2.5)  Corral Store. Due north from here is the Iron Mountain fault, a major basin and range fault that passes over Steele Mountain and connects with the Deer Park fault, which terminates at the South Fork Payette River. The area west of the Soldier Mountains is down-dropped and contains hundreds of granophyre dikes and a few rhyolite and rhyodacite dikes, suggesting the presence of an Eocene plutonic complex at depth.

Northwest from here is Chimney Peak, capped by quartzite and other metamorphic rocks (age unknown), which are underlain by foliated aplite and pegmatite dikes and foliated hornblende granodiorite which make up the roof of the Cretaceous batholith. Mylonitic fabric and hornblende lineations in the granodiorite suggest that the Chimney Peak block has moved south on a small decollement. The foreground hills are aplite and pegmatite with exposures of onlapping Snake River Plain basalts at the base. The next hill west from Chimney Peak has exposures of Tertiary granite half way up from the valley floor, and aplite and pegmatite at the top. Gabbroic rocks of probable Eocene age intruded along northeast-trending fractures. Metamorphic rocks occur in the roof.

90.5  (5.5)  Highway Milepost 139. Hill to northwest is Tertiary granite intruded by large Tertiary granophyre dikes. Most of the small hills rising above the prairie floor southwest of here are also pink granite.

93.1  (2.6)  Elmore County line.

93.6  (0.5)  Highway Milepost 136. Turn left onto dirt road.

93.7  (0.1)  **Stop 3.** Park and walk south up a small hill to examine an inactive borrow pit in Eocene granite. Aplites cutting the very decomposed granite have miarolitic cavities containing smoky quartz, potassium feldspar, epidote (?), and axinite (?). The Eocene granites generally contain at least twice the background radiation (uranium and thorium) of the Cretaceous batholith and typically have smoky quartz, due to rearrangement of the crystal lattice by radiation. The lattice is restored by the sun's ultraviolet rays, so these crystals turn clear when exposed to sunlight, for example on talus slopes.

93.8  (0.1)  Return to U. S. 20 and turn left (west).

95.5  (1.7)  Highway Milepost 134. Exposed here are Snake River Plain basalts, whose overall age ranges from 10 Ma to Recent (Craters of the Moon). They are probably Holocene here on the Camas Prairie. See Cluer and Cluer (1986) for more information on the Camas Prairie.

98.6  (3.1)  High Prairie Road takes off to right. On skyline ridge to south is the Blackstone mine. Pit is in Eocene granite; age of mineralization is unknown. The property was discovered in the late 1800s and was reopened in 1984. Exploration and development are continuing. Ores are complex sulfides with precious and base metals.

101.9  (3.3)  Hills in the near distance to the north are Eocene granite; hills to the northwest (farther away and tree-covered) is on the other side of a big northeast-trending fault and exposes aplite and pegmatite cutting metamorphic rocks.
Figure 13. Geologic map for STOP #2 (day 2), showing basin and range faulting in Cretaceous and Tertiary rocks south of Couch Summit.
102.5 (0.6) Pine-Featherville Road turnoff to the right (north). Note the granite pile which was used as a survey station when the road was built.

105.2 (2.7) Stop 4. Cat Creek Summit, 5,527 feet (1,675 m). From the overlook, the skyline ridge to the north-northwest is House Mountain. The far side of House Mountain is made of House Mountain metamorphics, which we will see at Stops 7 and 8.

Tree-covered hill to the north is Wood Creek Mountain. A northeast-trending fault separates Wood Creek Mountain (aplitic, pegmatitic, metamorphic rocks) from the granodiorite of Cat Creek Summit. This is the same fault described at 101.9 miles. There are several prospect pits located along this fault, which is typical of the northeast-trending structures in central Idaho.

Here on the summit are exposures of granodiorite of Cat Creek Summit, whose age is probably Eocene but is not known with certainty. The porphyritic texture seen here is not present in all exposures and is not typical of known Cretaceous granites in this area. The steep, pinnacled weathering style (reflecting high-angle joints) and the crystal habit are also not typical of Cretaceous granites in this area. McDowell and Kulp (1969) obtained a date of 44 Ma from rock near here, but the exact location is not known.

The distinctive topographic expression of this granodiorite can be seen following the road down the hill and around the corner.

Between Cat Creek Summit and Windy Gap, the next hill to the west on the highway, cross Lockman Gulch, just south of Little Camas Reservoir. Lockman Gulch is along another major northeast-trending fault, separating granodiorite of Cat Creek Summit from the older aplite and pegmatite of the batholith and small exposures of Challis Volcanic Group that are present north of Windy Gap. This fault extends along the south side of Anderson Reservoir and exposes Challis Volcanic Group near Wood Creek.

109.3 (4.1) Last outcrop of granodiorite of Cat Creek Summit. The exposures on both sides of the road are believed to be Challis Volcanic Group at the base, overlain just a few meters up the hill on either side of the road by Idavada Volcanics. There are also exposures of granodiorite of Cat Creek Summit on the east side of these hills. Just over the hill on the north side of the highway are exposures of andesite/dacite that are tentatively identified as Challis Volcanic Group. These are the westernmost exposures of Challis rocks north of the Snake River Plain. Historical marker describes Goodale's Cutoff, a hoped-for shortcut to the mining district of Florence from the Oregon Trail (it was not).

109.8 (0.5) Little Camas Reservoir is visible to the right (north).

112.3 (2.5) Stop 5. Windy Gap. Exposures of Miocene Idavada Volcanics occur on both sides of the road and extend to the top of the hill on the left (south). Voluminous eruptions of rhyolitic Idavada Volcanics occurred throughout southwest Idaho and southeast Oregon at approximately the same time as the onset of basin and range faulting in the northern part of the Great Basin and the onset of eruption of the Columbia River Basalt Group in Oregon, Idaho, and Washington. The Danskin Mountains, the low hills visible to the left (south) of House Mountain, expose Idavada Volcanics on top of Cretaceous granodiorite.

The skyline to the north is the Trinity Mountains, underlain by Eocene granodiorite and pink granite.

113.6 (1.3) Turn right to Anderson Ranch Dam and town of Prairie. Pass through old townsite of Dixie. On the skyline, the flat-topped peak to right is Cretaceous granodiorite. Peaks to north on the skyline are the Trinity Mountains.

115.1 (1.5) Highly sheared aplite and pegmatite to the right.

116.5 (1.4) The mine dump to the west (left) is the King shaft of the Golden King mine. The shaft connects to a tunnel in the bottom of the South Fork Boise River canyon almost 900 feet (270 m) below the road. The mine was developed in the 1930s by the Daly Consolidated Mining Company. Metamorphic rocks are exposed in the shaft. The ridge just southwest of the shaft is capped by quartzite and is part of the House Mountain metamorphic suite. The hill in the foreground, just east of the quartzite-covered hill, is a rhyolite dike of uncertain but possible Eocene age. The low hills between the road and the rhyolite are aplite and pegmatite. Hills to the east of the road are sheared aplite and pegmatite at their base and aplite and pegmatite with metamorphic fragments higher up.

116.9 (0.4) End of pavement. Note canyon-filling basalts (age-equivalent to Snake River Plain basalts) across the gorge. Ridge above is called Granite Ridge for the expected reason.

Stop 6. Exit bus halfway down the hill and walk to the dam. Note massive pegmatites and
aplites that make up about 50 percent of the exposures. Most dikes strike east-west and dip north. Gray granodiorite occurs as screens between the dikes. There are also good exposures of young basalt dikes and older andesite/dacite dikes in this cut. A basalt dike near here was dated by Armstrong (1975b) at 12.1 ± 1.2 Ma, which he believes is younger than the true age.

LUNCH at the dam. Restrooms available.

118.6  (1.7) Turn left and drive across Anderson Ranch Dam, on the South Fork Boise River. The Bureau of Reclamation dam was completed in 1950 as part of the Boise Area Irrigation Project. It also generates electricity. Dam is 456 feet (138 m) high and 1,350 feet (409 m) long.

118.8  (0.2) Turn left and go down the hill.

119.3  (0.5) A look back toward the spillway reveals aplite and pegmatite dikes cutting biotite granodiorite and all cut by a basalt dike (Fig. 14).

121.5  (2.2) Road to the right goes to Reclamation Village, home of the crews who operate the dam.

In 1950 1,500 people lived in the area. Population has diminished with automation and remote control.

123.4  (1.9) This is the first outcrop in which we see an appreciable amount of metasedimentary rock of the House Mountain metamorphic suite. Most of the outcrop is Cretaceous granodiorite with aplite and pegmatite, but the small inclusions of metasediments are recognizable.

125.2  (1.8) Increasing amounts of metamorphic rock here below Sproule Flat.

125.5  (0.3) Passing Indian Point (on left). This is close to the eastern edge of the main body of metasedimentary rocks, a large pendant that is elongate in a generally north-south direction. Ridge to the south is capped with quartzite.

127.0  (1.5) Cow Creek bridge.

Side trip

Continue past Cow Creek bridge about 3.5 miles to
where the South Fork road forks. The left fork crosses Danskin Bridge and continues down the river on private property. The right fork is the road to Prairie. Take the right fork and stop near the top of the ridge overlooking the South Fork, at the last outcrops before coming out onto the prairie.

Stop 7. Note sheared and foliated granodiorite that is intruded both along and across the mylonite fabric by nonfoliated aplite and pegmatite dikes (Fig. 15). This is the contact zone between the batholith and the House Mountain metamorphics. The metamorphics make up the high peaks immediately to the east on Pierce Creek. The mylonitized zone extends due north until it and the metamorphics are cut off by a northeast-trending fault just south of the town of Prairie. The foliated granodiorite may have formed by sliding off the House Mountain metamorphic suite, during unloading of the batholith. This tectonic unloading could have triggered emplacement of the aplite and pegmatite that is so prevalent in this southernmost part of the Atlanta lobe. Note canyon filling basalts.

End of side trip

Cross the river on the Cow Creek bridge. The debris fan on Cayuse Creek (to the right after crossing the river) is made up of metasedimentary rock and granite and is an expression of the Cow Creek fault.

128.1 (1.1) Stop 8. A prominent wash comes in from the right. Objectives are to look at the metasedimentary rocks and to discuss whether the felsic layers are aplite and pegmatite or products of partial melting.

128.8 (0.7) Outcrop on right shows aplite, pegmatite, and granodiorite cut by discontinuous basalts. Pods of metamorphic rocks are also discernible. The Cow Creek fault crosses the road about here.

129.3 (0.5) Stop 9. Looking northwest from here we can see a broad zone of white rubble which is the trace of the Cow Creek fault, the largest flat fault mapped to date in the Atlanta lobe of the batholith. The structure extends at least 5 miles (8 km) to the west. Preliminary mapping indicates that the fault separates the House
Mountain metamorphic suite and crosscutting aplite and pegmatite of the younger, upper part of the batholith from older, deeper parts of the batholith (granodiorite and two-mica granite) that are exposed below and to the west of Danskin Peak starting at Cathedral Rocks. The older batholithic rocks form most of the bedrock north of the Snake River Plain from the Danskin Peak lookout to Boise.

The bold outcrops on the skyline ridge are metasediments, with aplite and pegmatite on either side.

129.5 (0.2) Road levels out. Exposures to east (left) are quartzite and to the west are aplite and pegmatite on Long Tom Ridge.

129.9 (0.4) Hill on right is first exposure of Idavada Volcanics.

130.3 (0.4) More Idavada on right, marked by dark rubble.

130.5 (0.2) Summit. Valley ahead is along the trace of a fault, probably a northwest-trending, basin and range structure. Idavada Volcanics to right are on strike with valleys and dip to the south.

131.7 (1.2) Crossing fault in valley bottom.

135.9 (4.2) Cross cattle guard at the drainage divide at the top of Cow Creek. Bennett Mountain is straight ahead. The west end of Bennett Mountain seems to be down-faulted along a northeast fault that extends along U. S. Highway 20 through Dixie Summit. This indicates repeated movement on some Eocene structures.

137.4 (1.5) Turn right onto U. S. Highway 20. (When approaching this turn on the reverse trip, the dirt road has a sign for Prairie.) Idavada Volcanics exposed from here to end of today's log.

141.2 (3.8) Highway Milepost 107. Tollgate Cafe on right.

141.7 (0.5) Immigrant Road enters from right.

End of road log for day 2. U. S. Highway 20 continues for 11 miles to Mountain Home and the intersection with Interstate 84, crossing first Idavada Volcanics and then Snake River Plain basalts. Field trip will continue on I-84 (for 45 miles) to Boise.

ROAD LOG, DAY 3

On the last day of the trip, we travel up the west side of the Atlanta lobe, turn east into the core of the batholith, and return to Boise via Idaho City. We will examine Cretaceous tonalite, two-mica granite, and biotite granodiorite, as well as Eocene granite and granodiorite, and the Little Falls molybdenum deposit. Our route crosses major north- and northeast-trending faults, passes through the Boise Basin, the richest gold producing district in Idaho, and through thick sections of fossiliferous Miocene sediments.

The trip route and named localities are shown on Figure 16.

MILEAGE

Interval mileage is shown in parentheses.

0.0 (0.0) Begin roadlog on the northwest side of Boise at the intersection of State Highways 55 (north) and 44 (west). Take State Highway 55 north, the Payette River Scenic Route.

0.8 (0.8) Gravel pits in this area are in lacustrine sediments. There are many patches of these sediments on top of Snake River Plain basalts. They are part of the Pliocene-Pleistocene Idaho Group and the Pleistocene Ten Mile Gravel. It has been proposed that these are primarily alluvial and lacustrine sediments that accumulated either in a series of lakes or in one big lake (Lake Idaho) (Jenks and Bonnichsen, 1987). Gravity-driven mass flows are also recognized. Silification associated with hot springs has indurated some of the otherwise unconsolidated or loosely consolidated sediments. Stone quarried from these silicified sediments was used to construct many buildings in Boise including the State Capitol.

9.1 (8.3) Boise County line.

12.7 (3.6) Highway Milepost 57. Top of hill, 4,242 feet (1,285 m). The big hill to the east with the antennae is Shafer Butte, part of the Bogus Basin ski area. The ridge north of Bogus Basin is the Boise Ridge. The ridge and basin are granodiorite of the Idaho batholith. Just south of the basin the batholith is covered by Idaho Group sediments. The Boise Front fault, which is the south-western most major northwest-trending basin and range fault in the batholith, marks the change from Snake River Plain to Idaho batholith. Hot springs along this fault and related structures are used to heat a number of homes and public buildings in Boise.

14.8 (2.1) Scenic overlook especially positioned to allow view of Bread Loaf Rock (on the right).

15.4 (0.6) Turnout to right allows a view of the
Figure 16. Map showing route for Day 3 and locations of places and features named in the text.
landsides that are so prominent on the slopes of Horseshoe Bend Hill. Note patch of road marked “Sawdust fill.”

Horseshoe Bend Hill is composed of Idaho Group sediments and is crossed by a major northeast-trending fault. The fault has dropped the river valley down, resulting in the Horseshoe Bend in the Payette River. Just north of Horseshoe Bend granitic rocks of the Cretaceous batholith are exposed. The highway department has difficulty maintaining this road along the hill, and the unstable nature of the slopes is shown by the numerous landslides in this area. Sawdust is used as a base for the road in places to facilitate drainage. A new road is planned just east of the present route.

Southwest of here is the Pearl-Horseshoe Bend mining district, where gold was mined beginning in the late 1890s. The district contains numerous Eocene dikes which may mark the southwest extension of the trans-Challis fault system that cuts through Boise Basin.

18.6 (3.2) Highway Milepost 63.
18.9 (0.3) Enter Horseshoe Bend speed trap zone.
19.3 (0.4) Payette River at Horseshoe Bend.
20.0 (0.7) City park on the right (east). Restrooms available. Historical sign describes discovery of gold in Boise Basin (over the ridge to the northeast) in 1862. Traffic to and from Boise Basin passed through here in the early years after the discovery. We will visit Boise Basin later today.

The Boise Ridge, the high ridge east of Horseshoe Bend, is underlain by Cretaceous batholithic rocks and was first mapped by Lindgren (1898) as part of work done in Boise Basin.

21.7 (1.7) Cross Payette River (again). This river is part of the Boise Area Irrigation Project and is a great area for white water enthusiasts as it drops over 1,700 feet (515 m) in about 15 miles (24 km).

22.0 (0.3) Cretaceous tonalite is the dominant rock type from here north to Smiths Ferry, although large areas north and west of here have yet to be mapped and are shown as undifferentiated Cretaceous batholithic rocks on Figure 2. The age of these tonalites is not well known, but they probably were emplaced between 85 and 95 Ma (Lewis and others, 1987).

24.8 (2.8) Entering Gardenia.

25.4 (0.6) Hill to right shows dark brown soils characteristic of Columbia River basalts.

The Columbia River Basalt Group is a series of basalt flows totaling over 15,000 feet (4,500 m) thick and ranging in age from 17 Ma to about 7 Ma. The flows cover parts of Oregon, Idaho, and Washington and were fed from a series of dikes such as the Chief Joseph dike swarm located on the Idaho-Oregon border. The Miocene age of the flows is coincident with the onset of basin and range faulting in the northern part of the Great Basin and voluminous rhyolitic eruptions of the Idavada Volcanics in southwest Idaho and southeast Oregon.

30.7 (5.3) Highway Milepost 75. The trace of the Sr 0.704-0.706 line described by Armstrong and others (1977) lies west of here but is not well constrained in this area. This line separates continental crust to the east from island arc rocks to the west. The western tonalites of the Cretaceous batholith were intruded to the east of this suture line. Cretaceous tonalitic rocks are also present west of the boundary; these rocks typically are deformed and many contain a few percent epidote (Taubenneck, 1971; Bonnichsen, 1987; Aliberti and Manduca, 1988, this volume). The tonalites east of the suture are I-type granitoids and are intruded by younger biotite granodiorite that is gradational to two-mica granite. The granodiorite and two-mica granite together form the bulk of the Atlanta lobe.

34.5 (3.8) Road to Crouch (State Highway 17) takes off to right along South Fork Payette River. We continue north along the North Fork Payette River.

35.7 (1.2) Stop 1. Highway Milepost 80. Roadcut on right (east) shows superb examples of orbicular rock (Fig. 17). It is described by Charles W. Russell (Univ. of Washington, written commun., 1987) as “swirling masses of microgabbro (pargasitic hornblende and sodic labradorite with some larger plagioclase phenocrysts getting up to An60) in a matrix of andesine-rich, biotite-hornblende diorites.” The orbicules are most obvious in the microgabbro but are crudely formed in the diorite as well.

We urge you to be considerate in collecting the orbicules and other interesting textures displayed in these outcrops. There are only limited exposures of these oddities. If you must have large quantities, please try to collect from the riprap along the river or from debris at the base of the outcrop.

36.3 (0.6) Roadcut on right (east) shows a variety of
Johnson and Others—Cretaceous and Tertiary Intrusive Rocks of South-Central Idaho

Figure 17. Orbicules in tonalite at Milepost 80 on State Highway 55, north of Banks (STOP #1, Day 3).

Cretaceous hornblende- and biotite-bearing tonalitic rocks. Potassic alteration is best developed along fractures. The mafic dike at the north end of this exposure is younger than the tonalites. Russell (written commun., 1987) suggests it may be Eocene in age.

Russell (1987) reports that the tonalite in this area occurs at the margins of 0.5-1.5 mile (1-2-km)-long pods of mafic and intermediate rock. The tonalite grades into biotite granodiorite through decimeter-scale intercalations of the two phases, both of which exhibit steeply dipping, subparallel flow foliations.

36.7 (0.4) Highway Milepost 81. Exposures here are of the contact between biotite granodiorite and tonalite. Note the good primary foliation in tonalite south of the small gully across from the south end of the turnout. Biotite granodiorite appears to passively intrude tonalite 0.1 mile south of Milepost 81.

36.8 (0.1) Turn around at railroad bridge across the river. Return to confluence of North Fork and South Fork Payette Rivers.

39.1 (2.3) Turn left (east) onto State Highway 17 to Crouch, Garden Valley, and Lowman. Rocks here are Cretaceous tonalite.

41.2 (2.1) Stop 2. Pull out to right. Exposure of hornblende-biotite tonalite of Idaho batholith. Potassium-argon dating by Robert Fleck on biotite from these rocks yielded an age of 73.6 ± 3.0 Ma (Lewis and others, 1987). This relatively young age probably reflects slow cooling of the batholith; its actual emplacement age is older.

44.3 (3.1) Bridge crossing contact zone between tonalite and biotite granodiorite.

NOTE: The stops from here to mileage 71.7 are a side trip north from Crouch to look at exposures of two-mica granite. The field trip will not make this side trip, but access is over a maintained dirt road and is adequate for passenger vehicles.

47.4 (3.1) Turn left, toward Crouch, following Middle Fork Payette River.
48.3 (0.9) Intersection in city center of Crouch. Turn left.

51.0 (2.7) Pavement ends.

53.8 (2.8) Payette Formation on left. The Payette Formation is a series of loosely consolidated sediments of Miocene age. Recent mapping has shown that this unit and similar lithologies are much more widespread in Idaho than previously known. Many of the sediments are lacustrine deposits, suggesting that Idaho may have had extensive lake coverage in Miocene time as basalts of the Columbia River Group dammed existing drainages. The Latah Formation in north Idaho and the Sucker Creek Formation in southwest Idaho are probably equivalents of the Payette Formation. There are about 5,550 feet (1,680 m) of these sediments in Garden Valley and extensive exposures in the Boise Basin; whether any of the section is repeated by faulting is not known. Early placer operators using hydraulic giants exposed thick sections of these sediments near Idaho City. There are a number of good plant fossil localities in the Payette Formation.

55.0 (1.2) Good exposures of Payette Formation with carbonaceous lenses.

56.3 (1.3) Tie Creek Campground. Restrooms here. Exposures of Payette Formation, including coal seams, in the river.

57.5 (1.2) Exposures of two-mica granite. Emplacement age of the two-mica granite is probably 72-78 Ma (Lewis and others, 1987). This rock type forms the core of the batholith and extends throughout the length of the Atlanta lobe. In the Challis 1 x 2 degree quadrangle the two-mica granite was intruded by fine-grained, leucogranite dikes and small stocks and plugs that represent the last gasp of the Cretaceous batholith and are dated at about 70-75 Ma.

58.3 (0.8) More two-mica granite.

58.6 (0.3) More two-mica granite, with better exposures across the creek.

58.8 (0.2) Better exposures of two-mica granite, although still crumbly.

59.5 (0.7) Hardscrabble Campground. Restrooms and turnaround here.

62.6 (3.1) Tie Creek Campground again.

71.7 (9.1) Return to main road, turn left and cross Middle Fork Payette River.

(End side trip)

73.8 (2.1) Enter Garden Valley. We have just crossed the inferred position of the Boise Ridge fault, a north-trending fault identified by Anderson (1947). This fault is one of the few for which offset can be estimated. Kiilsgaard and Lewis (1985, p. 38) show that basalt of the Columbia River Basalt Group east of the fault has been downdropped a minimum of 1,600 feet (490 m) with respect to the same basalt west of the fault.

74.8 (1.0) Transition zone between two-mica granite and biotite granodiorite. This transition zone is generally mapped as biotite granodiorite.

75.0 (0.2) Stop 3. Alder Creek road. Bus will park here to allow a walk hack down the road (west) to look at the transition zone.

76.5 (1.5) Another outcrop of transition zone between two-mica granite and biotite granodiorite. Here we see even less muscovite than at the last outcrops, as we are closer to the main mass of biotite granodiorite.

Perched gravels across the river (south) were placered for gold at the turn of the century.

77.3 (0.8) Stop 4. Lunch at Hot Springs Campground. Good outcrops of biotite granodiorite to the west.

78.8 (1.5) More placered gravels across the river. These gravels were probably worked near the turn of the century, although placer operations at the head of Alder Creek were active as late as the 1930s.

80.4 (1.6) Pavement ends. A sizeable dam and bridge are visible to the right. The Wharton power plant was built on this site and came on line in 1908. It produced 1400 horsepower and had transmission lines to Centerville, Idaho City, and Quartzburg. It was owned and operated by the Boston and Idaho Gold Dredging Company.

82.0 (1.6) A northeast-trending fault crossing behind the first set of knobs across the river has upthrown the Boise Basin dike swarm and Eocene plutons on its southeast side. This is the westernmost fault of the trans-Challis fault system mapped to date.

84.7 (2.7) The few dikes exposed here are the only dikes that have been recognized in this northwest
85.1 Little Gallagher Creek. The fault described at mileage 82.0 crosses the river and heads up the creek to the northeast.

85.8 Now we have crossed the fault and are going into the upthrown southeast block.

86.2 Here we see the many dikes that are characteristic of this side of the fault. Compositions include andesite, dacite, rhyolite, and lamprophyre.

86.4 Little Falls, on the right. This is the location for which the upcoming Little Falls molybdenite deposit is named.

86.5 Stop 5. Adit at Little Falls deposit. Molybdenite here occurs along fractures and veinlets in a swarm of northeast-trending Eocene dikes, mostly in porcelaneous rhyolite. Molybdenite is exposed over an area 1,000 feet (330 m) wide and 3,000 feet (1,000 m) long. The deposit occurs in a pyritized and silicified zone; oxidation of the pyritized rock has colored the outcrop reddish brown and has created a strong molybdenum anomaly over part of the deposit (Rostad, 1967). Fission-track dating of zircon from a pre-mineral rhyolite dike at the deposit gave an age of 29.3 ± 1.7 Ma (Kiiilsgaard and Bennett, 1987b). The property was discovered in 1960 and was prospected by Abella Resources from 1978 to 1981. Abella's assay data resulted in a weighted average grade of 0.05 percent MoS₂ (Kiiilsgaard and Bennett, 1987b). The rhyolite dikes cross the South Fork Payette River and make the falls in the river.

The swarm of rhyolitic dikes can be traced 2.5 miles (4 km) southwest to the Camo molybdenum deposit. Camo was discovered by AMAX Exploration in 1963 and was evaluated until 1984. Chalcopryte and molybdenite occur in quartz veins and as disseminations in gray quartz monzonite porphyry and other Eocene hypabyssal dikes. Both argillic alteration and silicification are present. Some rhyolitic dikes at Camo postdate the mineralizing event. The deposit is estimated to contain one billion tons of material with an average grade of 0.10 percent MoS₂ (Kiiilsgaard and Bennett, 1987b).

86.8 Road to Deadwood Reservoir. Large parking area. Looking up the road (north) you can see drill roads cut by Abella Resources during exploration of the Little Falls prospect.

87.9 Eocene stock. Although this stock has a pink groundmass and is shown as granite by Fisher and others (1983, in press), it is not typical Eocene granite. It is at the extreme mafic end of granites and it chemically resembles the Eocene diorites and granodiorites. This exposure shows hornblende-biotite granodiorite. Dike has an excellent chilled margin.

88.4 Cross creek.

88.5 Tertiary diorite to quartz monzodiorite.

88.7 Stop 6. Examine contact between Cretaceous biotite granodiorite on east and Eocene quartz monzodiorite on west. This is one of the few places where this contact relation is clearly shown. Similar Cretaceous rocks are exposed from here east as far as Grandjean (about 26 miles (42 km), as the crow flies.)

91.8 Pine Flats Campground.

92.8 A river diversion project was undertaken here between 1910 and 1920. The road is still visible below. The idea was to divert the river through a tunnel in order to work placer deposits in the bend in the river. The project was never completed and the tunnel is now partially caved.

This is the western edge of the Deadwood fault zone, part of a set of north-trending faults that crosses the western part of the Atlanta lobe. The fault zone is mapped for about 50 miles (80 km) north from here along the Deadwood River and Johnson Creek. Foliated hornblende-biotite granodiorite east of the fault zone and near Deadwood mine, about 28 miles (45 km) north of here, is about 1,650 feet (500 m) lower than similar rock on the high ridge west of the fault, suggesting fault displacement of at least that much (Kiiilsgaard and Lewis, 1985).

93.0 Begin paved road.

93.6 Bridge at Deadwood River.

95.7 More biotite granodiorite exposed in a new roadcut. A basalt dike similar to the one exposed here was collected 4 miles (6.5 km) to the west and was dated by Armstrong (1975b) at 15.2 ± 0.5 Ma. The biotite granodiorite exposed here is the typical rock of the Atlanta lobe of the Idaho batholith: there are many, many square miles of this stuff.

97.0 Intersection with State Highway 21. Turn right (south) toward Idaho City and Boise.
97.2  (0.2)  Cross South Fork Payette River.

106.7  (9.5)  Beaver Creek summit, 6,064 feet (1,838 km).

115.2  (8.5)  A major northeast-trending fault, part of the trans-Challis fault system, is exposed here.

117.3  (2.1)  Stop 8. Mores Creek summit, 6,118 feet (1,854 km). Parking available on south side of the summit. Major northeast-trending fault is exposed in the gouge in the roadcut. This fault generally follows Mores Creek to the southwest. It is also exposed in the roadcuts on the northeast side of the summit.

121.6  (4.3)  Bad Bear Creek.

126.3  (4.7)  Extensive placer workings on right (northwest) side of the road, part of the Boise Basin district. Gold was discovered in Boise Basin in 1862, and a major gold rush was soon underway to the rich diggings. Both Pioneerville and Bannock (later renamed Idaho City) were established that year. The early work was in placers; modern dredges arrived in the early 1900s and operated until 1950. Lode deposits were discovered in 1863. The first 10-stamp mill was brought to the district in 1864. Major lode mines included the Golden Age, Comeback, Missouri, Belshazzar, Mountain Chief, and the Gold Hill mines. The Gold Hill mine was worked by the Talache Mining Company until 1938. 

The mines are in epithermal precious-metal vein deposits of Tertiary age (Fisher and others, 1987). The orebodies are shallow; the richer portions are near the surface, and the values diminish with depth. They were mined chiefly for gold and silver but also contain variable amounts of copper, lead, and zinc. Deposits range from a few tens of tons to a few thousand tons, and have grades ranging from trace to 23 ounces of gold and trace to 2,500 ounces of silver per ton. Assay information on ore mined from twelve deposits in Boise Basin indicates an average grade of about 0.5 ounce of gold and 10 ounces of silver per ton. Electrum is the most significant ore mineral at most deposits.

Mines in the Boise Basin district produced more gold than any other district in Idaho, and the area is one of the major producing districts in the country. Boise County gold production from 1863 to 1980 is estimated at 2.8 million ounces, most of which came from the Boise Basin.

127.9  (1.6)  Cretaceous biotite granodiorite.

128.7  (0.8)  Highway Milepost 41. Placered ground on left (southeast) side of the road.

129.0  (0.3)  Outcrops of Columbia River basalt.

130.8  (1.8)  Road to right goes through Idaho City and back to Horseshoe Bend.

End roadlog. Field trip will return to Boise via State Highway 21 and Interstate 84, passing good exposures of Payette Formation at Idaho City, fresh biotite granodiorite south of Idaho City, basalt lava flows of the Snake River Plain, and Lucky Peak Reservoir.

ACKNOWLEDGMENTS

Our research was supported by the CUSMAP program of the U. S. Geological Survey in the Challis and Hailey 2 degree sheets. The manuscript benefited from reviews by S. D. Luddington, F. J. Moye, C. M. White, and R.G. Worl. Photographs are by Earl H. Bennett.

REFERENCES


... 1948, Tungsten mineralization at the Ima mine, Blue Wing District, Lemhi County, Idaho: Economic Geology, v. 43, no. 3, p. 181-206.


Sutter, J. F., Snee, L. W., and Lund, K. W., 1984, Metamorphic, plutonic, and uplift history of a continent-island arc suture zone, west-central Idaho:


Regional Geologic Setting
and Volcanic Stratigraphy of the
Challis Volcanic Field, Central Idaho

Falma J. Moye 1
William R. Hackett 1
John D. Blakley 1
Larry G. Snider 1

INTRODUCTION

Early Tertiary geologic history of the northwestern United States was characterized by a short-lived but intense magmatic episode from 55 Ma to 40 Ma. That episode resulted in an areally extensive and compositionally diverse belt of volcanic and plutonic rocks extending from southern British Columbia across northeastern Washington and central Idaho and into Montana and Wyoming. Although the time equivalence of this belt of rocks has been recognized and some basic geologic, geochemical, and radiometric work has been completed, the Eocene magmatic event remains poorly understood within the context of Eocene tectonics of the northwest.

The Challis volcanic field of central Idaho is the largest and most diverse of the Eocene volcanic fields, both in composition and in variety of volcanic deposits. Because it is dissected to subvolcanic levels, geologists can study the geochemical relationships among cogenetic volcanic and intrusive units and the internal structures of volcanic and hypabyssal complexes.

In this paper we summarize current knowledge of the geology of the Challis volcanic field, including its regional stratigraphy, geochronology, and geochemistry.

REGIONAL GEOLOGIC SETTING

The Challis volcanic field is part of an extensive Eocene volcanic belt in the Pacific Northwest between 42 and 49 degrees north latitude (Fig. 1). The belt extends 1500 km from west to east and includes the British Columbia alkalic province, the Sanpoil field in northeastern Washington and southern British Columbia, the Challis field in central Idaho, the Absaroka field in Montana and Wyoming, the Montana alkalic province, and numerous small, scattered outliers of Eocene volcanic rocks.

The Eocene volcanic belt was originally named the Challis Arc (Vance, 1979), which has led to an acceptance of the volcanics as being subduction-related. However, recent work suggests that this tectonic interpretation is too simplistic (Gest and McBirney, 1979; Moye, 1984; Ekren, 1985; Fox and Beck, 1985; Carlson and Moye, in press). We therefore propose that the name Challis Arc be abandoned for the Eocene volcanic deposits, because the term "arc" implies a subduction-related origin and such a tectonic setting remains to be demonstrated for these igneous rocks.

Two tectonic settings have been proposed for the Eocene volcanism, based on compositional variation and
Figure 1. Distribution of Eocene volcanic rocks in the northwestern United States and southern British Columbia. Major fields are labeled, showing approximate age span of most intense volcanism. Regional structures are shown in relation to distribution of volcanic fields; Trans-Challis Fault Zone is from Bennett (1986) and Great Falls Tectonic Zone is from O'Neill and Lopez (1985).

on spatial and temporal relations. The most commonly accepted model suggests that shallow subduction and imbrication of the Farallon plate beneath the North American plate was responsible for an anomalously wide volcanic arc (Lipman and others, 1971, 1972; Lipman, 1980). This model axiomatically assumes that east-west potash variation in bulk rocks varies systematically as a function of depth to an Eocene Benioff zone. Imbrication of the subducted slab is postulated to explain the apparent repetition of potash-with-depth trends across the arc.

An alternative and opposing model suggests that a collision of the Pacific and North American plates triggered intracontinental rifting and related igneous activity (Ewing, 1980; Fox, 1983; Fox and Beck, 1985). This model is based largely on studies of the alkaline volcanic fields in British Columbia and Washington, and it emphasizes the presence of extensional structures that formed in response to collision. It focuses on studies of Eocene rocks in northeastern Washington and British Columbia and does not incorporate data from the other volcanic fields.

Paleotectonic settings and petrogenesis of ancient volcanic rocks are largely inferred by comparing their geochemical and isotopic data with similar data from active volcanoes in known tectonic settings. Fundamental petrographic and geochemical data from the Sanpoil (Moye, 1984; Carlson and Moye, in press), Absaroka (Gest and McBirney, 1979; Meen, 1987) and Challis (McIntyre and others, 1982; Hackett and others, 1988; Norman, 1988) volcanic fields are still being acquired. When complete, this information will ultimately lead to substantial refinement of Eocene magmatic and tectonic models for the Pacific Northwest. However, Eocene paleotectonics in the Pacific Northwest remains a subject of current debate.
GEOLGY OF THE
CHALLIS VOLCANIC FIELD

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

Figure 3 gives generalized stratigraphic columns for the Challis volcanic field in central Idaho. For descriptive purposes, we divide the Challis volcanic field into a northern portion where first-order mapping and stratigraphic studies are complete, and a southern portion where first-order studies are still in progress. Our discussion of the northern portion of the field is based primarily on work by the U.S. Geological Survey for the Challis quadrangle, and its eruption led to collapse of the

Northern Challis Field

The most voluminous volcanism and plutonism occurred in the interval from 51 to 45 Ma, with minor silicic intrusive activity persisting until 40 Ma. Volcanism involved the construction of composite cone complexes during effusion of intermediate lavas and the formation of calderas associated with explosive ash flow eruptions. Intrusive rocks were emplaced as plugs, domes, dike swarms, and composite stocks. Based on the work of McIntyre and others (1982), volcanic deposits of the northern Challis field can be subdivided into several stratigraphic packages. These are, from oldest to youngest: (1) intermediate rocks, dominantly andesite lava flows; (2) rhyodacite to rhyolite ash-flow tuffs; and (3) rhyodacite to rhyolite domes and plugs.

Intermediate Rocks

Volcanism began about 51 Ma with the effusion of voluminous intermediate to mafic lava flows. The early intermediate to mafic lavas are largely covered by ash-flow tuffs on the northern and western parts of the field, and the early intermediate rocks are therefore best exposed in the southeastern part of the Challis 1 x 2 degree quadrangle (Fig. 2). McIntyre and others (1982) divide the intermediate and mafic rocks into (1) dacite with phenocrysts of plagioclase, pyroxene, and/or amphibole and biotite, and (2) andesite to basalt with phenocrysts of pyroxene or olivine and plagioclase. David McIntyre (written communication, 1986) suggests that early volcanism was widespread, with eruptions from many vents over a large portion of the northern Challis field. Dacites were erupted from small composite volcanoes or dome complexes. Basalts and andesites form lava flows and intrusive masses confined to several northwest-trending zones in the southeastern part of the Challis 1 x 2 degree sheet and along the east flank of the Lost River Range, north of the town of Mackay (Fig. 2).

In addition to the great compositional and lithologic variety of the early volcanic rocks, McIntyre and others (1982) recognize an equally complex assemblage of associated hypabyssal intrusive rocks, including dacite dikes, composite stocks of granite to quartz monzonite, and small syenite intrusions.

Rhyodacite to Rhyolite Ash Flow Tuffs

Early effusive volcanism was followed by a period of explosive rhyodacite to rhyolite volcanism, resulting in subsidence of the Van Horn Peak cauldron complex, the Thunder Mountain caldera, and the Twin Peaks caldera during the brief interval of about 49-45 Ma (Fig. 2; Hardyman, 1981; McIntyre and others, 1982; Leonard and Marvin, 1982; Ekren, 1985). The caldera structures are recognized by the presence of curvilinear faults and by ignimbrite units that record the emplacement of thick intracaldera pyroclastic flows and thin outflow units. Graben subsidence preceded and was synchronous with volcanism, producing co-linear, northeast-trending volcanic-tectonic depressions confined to the trans-Challis fault zone (McIntyre and others, 1982; Bennett, 1986).

Initial explosive volcanism resulted in the widespread emplacement of a rhyodacitic ash-flow tuff: the tuff of Ellis Creek is the most voluminous volcanic unit in the Challis quadrangle, and its eruption led to collapse of the
Figure 2. Distribution of Eocene volcanic and plutonic rocks of the Challis volcanic field, showing known or suggested cauldrons and calderas. Volcanic rocks are subdivided based on primary lithologic type. Volcanic geology of the Challis 1 x 2 degree quadrangle is from McIntyre and others (1982); volcanic geology for the rest of the field is from our reconnaissance field work.
34-by-48-km Van Horn Peak cauldron complex. The tuff of Ellis Creek is of variable thickness, due to deposition on irregular topography. Its maximum thickness outside the cauldron is 400 m, but it reaches 1,500 m within the cauldron complex (Ekren, 1985). The tuff occurs up to 80 km from its source, and McIntyre and others (1982) suggest that its original areal extent was much greater than today, since much of the unit has been removed by uplift and erosion, particularly to the west.

Younger ash-flow tuff units are less voluminous than the tuff of Ellis Creek. These include the tuff of Eightmile Creek, tuff of Pennal Gulch, tuff of Challis Creek, and several unnamed or volumetrically minor ash-flow tuffs (McIntyre and others, 1982; Ekren, 1985). These tuffs are related to the collapse of segments of the Van Horn Peak cauldron complex, culminating with the eruption of the tuff of Challis Creek and the collapse of the Twin Peaks caldera about 45 Ma (Ekren, 1985). The younger tuffs range in composition from quartz latite to alkali rhyolite and are more alkaline than the tuff of Ellis Creek.

In the southeastern portion of the Challis 1 x 2 degree quadrangle, McIntyre and others (1982) have mapped several ash-flow tuff units from unknown sources. The deposits are thought to have erupted from sources to the south, on the Idaho Falls or Hailey 1 x 2 degree quadrangles, but our preliminary mapping has not yet identified any southerly sources for these deposits.

### Rhyolite Domes and Casto Pluton

Waning stages of igneous activity in the northern part of the Challis volcanic field were characterized by the emplacement of rhyolite to alkali rhyolite domes, and intrusion of the granitic Casto pluton into the lower part of the ash flow tuff sequence. Dome emplacement was controlled by cauldron ring faults and by extensional structures. Criss and others (1984, 1985) use oxygen isotopes to map the extent of fossil hydrothermal cells associated with the emplacement of domes and small, shallow intrusions. Epithermal precious-metal mineralization is spatially associated with dome emplacement (Hardyman and Fisher, 1985).

---

**Figure 3. Generalized stratigraphic sections and correlations of volcanic units for the eastern half of the Challis 1 x 2 degree quadrangle (modified from McIntyre and others, 1982) and the northwestern portion of the Idaho Falls 1 x 2 degree quadrangle (from our own reconnaissance mapping).**
Southern Challis Field

Geologic mapping and regional stratigraphic studies of the southern Challis field are in progress. The following descriptions of regional stratigraphy are preliminary, and formal stratigraphic names have not yet been assigned to rock units. Volcanic stratigraphy in the Grouse 15-minute quadrangle is from Betty Skipp (written communication, 1986).

Volcanic and sedimentary deposits of the southern Challis field are subdivided into four stratigraphic packages. From base to top, these are: (1) basal Challis conglomerate and tuff breccia; (2) andesitic lava flows and tuff breccias; (3) dacitic to rhyodacitic ash flows and lava flows; and (4) dacitic to rhyolitic hypabyssal Intrusions.

Basal Challis Conglomerate

Nelson and Ross (1968, 1969) and Dover (1969, 1983) describe a Cretaceous(?)-Tertiary(?), pre-Challis conglomerate in the southern part of the Challis field; the unit is described as being: (1) unconformable on Paleozoic rocks; (2) unconformably overlain by Challis Volcanics; and (3) containing only clasts of Paleozoic sedimentary rocks with no volcanic component. Paull (1974) assigned the name Smiley Creek Conglomerate, described the unit as a post-orogenic conglomerate, and considered the unit to be unconformably overlain by Challis Volcanics. Work in progress (Burton and Blakley, 1988) demonstrates that the conglomerate is locally conformable with Challis Volcanics and has a probable Eocene to Oligocene pollen assemblage. It therefore represents a depositional transition from an orthoconglomerate that lacks volcanic debris into a tuffaceous paraconglomerate whose deposition was coincident with local onset of Challis volcanism. The tuffaceous paraconglomerate in turn grades upward into andesitic tuff breccias and lava flows. The unit is significant in studies of Challis volcanism because it helps to define paleotopography before volcanism and because its volcanic clasts provide a record of compositional diversity during earliest volcanism. In addition, the age constraints on this unit define a minimum age for the initiation of volcanism in this part of the Challis field.

Comparable stratigraphic relations occur in the southeastern Challis 1 x 2 degree quadrangle, where a similar clastic sedimentary package is mapped at the base of the Challis Volcanics in the Sage Creek area (McIntyre, 1982; Fisher and others, 1983).

Andesite Lava Flows and Tuff Breccias

The basal volcanic section in the southern part of the Challis field is dominated by andesite lava flows and tuff breccias of variable thickness. In the Antelope Pass and Sheep Mountain areas, andesite and dacite lavas form sections up to 600 m thick, and the deposits may have been confined to large paleovalleys or syn-volcanic graben structures. In the Porphyry Peak area, the basal volcanic section is 700 m of andesitic lava flows and monolithologic tuff breccias. In the Grouse quadrangle and in the Lake Hills section, the thickness of the basal andesitic unit is much less, with a maximum thickness of 350 m. Individual lava flows and tuff breccia units vary in thickness and lateral extent because they were confined to paleovalleys on the irregular pre-Challis surface.

Source vents for the lavas are difficult to identify because clear volcanic facies patterns are absent in most areas. Abundant dikes, together with the rarity of central-vent facies relationships, suggest that many lava flows may have been eroded from fissures rather than point sources. A K-Ar age of 49.3 ± 0.7 Ma (J.D. Obradovich, written communication to Betty Skipp, 1976; recalculated 1987) was determined for the basal andesite in the Grouse 15-minute quadrangle (Betty Skipp, written communication, 1987).

Phenocrysts in andesite lavas are predominantly olivine and clinopyroxene, with minor plagioclase; hornblende or orthopyroxene are also present in more felsic rocks. Shoshonite lavas contain phenocrysts of olivine and clinopyroxene ± minor plagioclase. Mafic lavas (absarokites and high-K basalts) are dominated by olivine and clinopyroxene phenocrysts; olivine commonly contains inclusions of chrome spinel.

Dacite to Rhyodacitic Ash Flows and Lava Flows

The most voluminous deposits of the southern Challis field are rhyodacite lava flows and ash flow tuffs. Pyroclastic deposits are compound- and multiple-flow units, indicating numerous explosive eruptions. From base to top, this stratigraphic package is generally divided into: (1) lower andesite to rhyodacite ash-flow tuffs and tuff breccias; (2) lower rhyodacite to dacite lava flows; (3) upper rhyodacite ash-flow tuffs; and (4) upper rhyodacitic lavas. Although general compositional and lithologic similarities exist among the sections studied, only the upper ash flows are widespread and offer potential for regional correlation. Deposits lower in the stratigraphic sections are generally more localized and less readily correlated; they are thought to represent parallel eruptive trends from separate eruptive centers.

The Sheep Mountain and Porphyry Peak areas are inferred sources of these voluminous silicic volcanics, but our preliminary mapping has not yet proven any calderas. Aeromagnetic data suggest the presence of buried intrusive bodies in these areas (Dean Kleinkopf, oral communication, 1987). Rhyodacite dikes that are lithologically similar to the lower ash-flow tuffs are present in the Grouse 15-minute quadrangle, suggesting fissure eruptions there. In the Lake Hills area north of Carey, thick sections of ash-flow tuffs were apparently
derived from nearby source(s), but mapping has not yet located them. Further to the west on the Hailey 1x2 degree quadrangle, Hall and McIntyre (1986) suggest a large cauldron complex west of Ketchum (Fig. 2), but our mapping has shown that the area is underlain primarily by intermediate lavas, with little evidence of large-scale explosive volcanism.

Lower ash flows are typically moderately welded, crystal-vitrific dacite to rhyodacite ash-flow tuffs, with subordinate airfall tuffs and lava flows. Several simple and compound cooling units are recognized. The ash-flow tuffs contain 5 to 30 percent phenocrysts of plagioclase, biotite, hornblende, sanidine, and minor but conspicuous pyroxene. Lithic fragments in some units are large and abundant near an inferred source in the Sheep Mountain area. Thickness is 250-300 m in the Grouse and Sheep Mountain areas; the unit is not believed to be present in the Porphry Peak and Lake Hills areas.

Lower dacitic to rhyodacitic lava flows and tuff breccias up to 270 m thick overlie the lower ash flows in the Porphyry Peak and Lake Hills areas. Individual lava flows vary from a few meters to a few tens of meters in thickness and contain varied proportions of plagioclase, hornblende, biotite, with minor pyroxene, sanidine, and quartz phenocrysts. In the Lake Hills area, silicic lava flows are collectively only a few tens of meters thick and are interbedded with ash-flow tuffs. Lava flows are not present at this stratigraphic interval in the Porphyry Peak area. Dacite and Rhyolite Domes

The latest stage of igneous activity in the southern part of the Challis field involved the emplacement of dacite and minor rhyolite domes, together with associated lava flows and localized pyroclastic deposits. As in the northern part of the field, epithermal mineralization is commonly spatially associated with late stage igneous activity. Ore deposits occur along regional structures that focused hydrothermal solutions, altering Paleozoic carbonates to jasperoids or skarns, and forming extensive alteration halos in volcanic rocks (Moye and Hall, 1988). Oxygen isotope studies, similar to those completed on the Challis 1 x 2 degree quadrangle, are now in progress.

Regional Stratigraphic Correlations

Preliminary stratigraphic correlations, based on eleven radiometric age dates from the Challis 1 x 2 degree quadrangle (McIntyre and others, 1982) and three from the Grouse quadrangle (Betty Skipp, written communication, 1987), together with field correlations of ash-flow tuffs in the Sheep Mountain, Porphry Peak and Grouse sections are shown on Figure 3. These correlations suggest that volcanism in the southern part of the Challis field occurred over a similar 5-6 Ma time interval as in the northern part and that volcanism of the southern Challis field was synchronous with middle to latest volcanism of the northern Challis field.

Age dating and detailed lithologic studies are still in progress, and no unequivocal correlations have yet been made between deposits in the Challis 1 x 2 degree quadrangle (McIntyre and others, 1982) and deposits in the northwest Idaho Falls 1 x 2 degree or northeast Hailey 1 x 2 degree quadrangles. A northeast-trending Eocene structural high has been suggested to extend across an area approximately coincident with the map boundary dividing the northern and southern Challis fields at 44 degrees north latitude (Skipp and Harding, 1985). We believe that the abrupt termination of ash-flow sheets both north and south of this inferred paleo-high, and the presence of extensive Eocene conglomerates beneath the Challis Volcanics on the north and south sides of the paleo-high, confirm the presence of this structure.

Geochemistry of Challis Volcanic Rocks

The remarkable geochemical diversity of Challis volcanic and intrusive rocks that formed mostly in a six-million-year period is shown in Figure 4. We emphasize that mafic samples are over-represented in Figure 4, and these data do not represent the relative volumes of lava types in the Challis field. Challis volcanic rocks are dominantly high-K andesites, high-K dacites and latites, but range from basalt to alkali rhyolite in composition. Potassic mafic lavas (including absarokites and shoshonites; Hackett and others, 1988) occur as part of the early intermediate deposits and have compositions similar to rocks...
described from the Absaroka volcanic field (Gest and Mc Birney, 1979; Meen, 1987).

**SUMMARY**

The Eocene Challis volcanic field was active for about 65 Ma, but most volcanism occurred during a period of only 6 Ma. A general evolutionary trend of diverse magma compositions and deposits is evident (Fig. 5). Early volcanism was dominated by the effusion of mafic to intermediate lava flows, beginning at about 51 Ma. Vents are difficult to identify, but general field relations suggest that eruptions occurred from numerous central-vent stratovolcanoes and that fissure-fed lava flows were emplaced into northeast trending graben. The early intermediate volcanic rocks have a broad compositional range, including high-K basalt, shoshonite, absarokite, high-K andesite and latite, but there is no apparent systematic age or spatial relation among these lava types.

Later eruptions of voluminous, intermediate-to-silicic ash-flow tuffs and local lava flows occurred from 48 to 45 Ma in the northern part of the Challis field (McIntyre and others, 1982) and from 49 to about 44 Ma in the southern part (Betty Skipp, written communication, 1987). The general compositional trend with time is from high-K dacitic to rhyodacitic rocks, to more alkali-rich latite, trachyte, and alkali rhyolite. Concurrent subsidence of large cauldron complexes, calderas and northeast-trending volcano-tectonic depressions occurred in the northern part of the field. Similar structures are inferred but not yet proven in the southern part of the field.

Final stages of igneous activity involved the emplacement of dacitic to rhyolitic domes and small intrusive bodies until about 40 Ma. Intrusions were controlled by cauldron structures and northeast-trending regional faults. Epithermal mineral deposits are spatially associated with this late igneous activity.

![Figure 4. Potash versus silica variation diagram, showing compositional variation of bulk rocks from major Eocene volcanic fields in the Pacific Northwest. Compositional fields are from Peccei and Taylor (1976). Filled squares are analyses from the Sanpoil volcanic field (Church, 1963; Moye, 1984). Filled circles are analyses from the Absaroka field (Chadwick, 1970; Gest and Mc Birney, 1979). Crosses are analyses from the Challis volcanic field (32 analyses from McIntyre and others, 1982; 40 new x-ray fluorescence analyses of mafic lavas by S. A. Mertzman, written communication, 1987).](image)

![Figure 5. Photograph showing Challis volcanic stratigraphy of the Porphyry Peak area, in the southeastern Challis field (see Fig. 2 for location). Stratigraphic relations and map units shown are generally representative of the entire Challis field: basal conglomerate is overlain by intermediate lava flows and tuff breccias, in turn overlain by silicic ash flows and lavas. These deposits are intruded by hypabyssal dacite porphyry and small rhyodacite bodies.](image)
ACKNOWLEDGMENTS

We are greatly indebted to Betty Skipp of the U. S. Geological Survey, whose volcanic stratigraphy in the Grouse 15-minute quadrangle is a basis for the rest of the southern Challis field, and who provided us with important field and radiometric data from the Grouse quadrangle. Ron Worl of the U. S. Geological Survey has given valuable logistical support during our field work in central Idaho. Stan Mertzman of Franklin and Marshall College promptly supplied x-ray fluorescence analyses of mafic lavas. We thank R. F. Hardyman for constructively reviewing the manuscript. Our field and analytical studies are funded by an Idaho State Board of Education economic incentives grant to Moye and by a National Science Foundation grant EAR-86-18629 to Hackett and Moye.

REFERENCES


Hackett, W. R., Moye, F. J., and Mertzman, S. A., 1988, Petrology of mafic to intermediate rocks from the Eocene Challis volcanic field, central Idaho:


Meen, J. K., 1987, Formation of shoshonite from calcalkaline basalt magma: geochemical and experimental constraints from the type locality: Contributions to Mineralogy and Petrology, v. 97, p. 333-351.


Peccarillo, A. and Taylor, S. A., 1976, Geochemistry of Eocene calc-alkaline volcanic rocks from the
Kastomonu area, northern Turkey: Contributions to Mineralogy and Petrology, v. 58, p. 63-81.


INTRODUCTION

One of the most striking paleogeographic anomalies of the western U. S. Cordillera is exposed in west-central Idaho where Permian and Triassic island arc volcanics of the Wallowa Terrane are juxtaposed against rocks with cratonic affinities, across a narrow transition zone of metamorphic and plutonic rocks. In this area much or all of the Paleozoic miogeoclinal section present elsewhere along the western edge of the craton is missing, as are commonly intervening belts of melange and eugeoclinal sedimentary rocks of mixed provenance between the oceanic arc terrane and the cratonic margin (Davis and others, 1978; Hamilton, 1978; Lund, 1984). The preserved transition zone records an early to middle Cretaceous period of compressional deformation. This event appears to postdate the truncation of the Paleozoic miogeoclinal section and the removal of transitional assemblages between the arc and cratonic margin.

The original nature of the boundary between accreted arc material and the cratonic margin has been obscured by Cretaceous compressional deformation and by the emplacement of voluminous intrusives. No evidence of the event that truncated the cratonic margin and removed the transitional assemblages is preserved. The boundary is recognized by an abrupt change in the lithology of metamorphic pendants within the plutonic rocks. Pendants of metavolcanic and volcaniclastic schists and gneisses give way to quartzites and metapelites of continental affinity, suggesting that the plutons were emplaced across a steeply dipping boundary between arc and continent (Lund, 1984; Manduca and Kuntz, 1987). Geochemical studies in this area have revealed a rapid change in the strontium and oxygen isotopic ratios of the rocks, which generally corresponds with the change in pendants (Armstrong and others, 1977; Fleck and Criss, 1985; Manduca and others, 1986). The transition zone between rocks of the Wallowa Terrane and the continental margin as preserved within the Cretaceous intrusives of the Idaho batholith is illustrated schematically in Figure 1. The transition zone consists of a series of lithotectonic packages which are, from west to east: the Wallowa Terrane, the Rapid River thrust plate, the Pollock Mountain plate, the Little Goose Creek Complex and the Payette River Complex.

The Wallowa Terrane consists of Permian and Triassic Seven Devils Group arc volcanics, and overlying limestone of the Martin Bridge Formation and slate of the Lucile or Hurwal Formation. These rocks have been relatively unaffected by deformation and metamorphism along the transition zone. The Rapid River thrust system emplaces Riggins Group schists and lesser amounts of Martin Bridge, Lucile and Seven Devils lithologies over the Wallowa Terrane. The Riggins Group is an assemblage of metavolcanic and volcaniclastic schists thought...
to represent a portion of an island arc terrane. These schists have been correlated with either the Wallowa Terrane or the Olds Ferry Terrane and adjacent flysch and melange sequences in eastern Oregon (Hamilton, 1978; Brooks and Vallier, 1978; Lund, 1984, Silberling and others, 1984). The Rapid River thrust plate increases in metamorphic grade from lower greenschist facies in the west to uppermost greenschist or lower amphibolite facies in the east, and is penetratively deformed throughout. The Pollock Mountain fault emplaces the Pollock Mountain Amphibolite and the Hazard Creek Complex over the Rapid River thrust plate. Both the Rapid River thrust system and Pollock Mountain fault are shallowly dipping, northwest-directed thrusts at high structural levels, which steepen into vertical shear zones at depth.

The Hazard Creek Complex, Little Goose Creek Complex and Payette River Complex are all composed primarily of Cretaceous plutonic rocks. The Hazard Creek Complex consists of variably deformed epidote-bearing intrusives and gneissic country rocks emplaced during the early to middle Cretaceous compressional deformation. The westernmost plutons in the Hazard Creek Complex intrude metamorphosed island arc volcanics of the Pollock Mountain Amphibolite. The Little Goose Creek Complex consists primarily of strongly mylonitized porphyritic granodiorite orthogneiss. This mylonitization is the youngest penetrative deformation in the area, postdating the emplacement of all three intrusive complexes. It affects the eastern margin of the Hazard Creek Complex and the western margin of the Payette River Complex, obscuring the original nature of these contacts. The transition in wall rocks from arc-related to continentally derived material and the rapid change in strontium and oxygen isotopic ratios occur within the Little Goose Creek Complex. The Payette River Complex consists primarily of undeformed tonalite, containing abundant screens and inclusions of continentally derived metamorphosed sedimentary rocks.

This field trip traverses the transition zone from west to east, beginning in rocks of the Wallowa Terrane and ending in rocks of the continental margin preserved as pendants within the Cretaceous Payette River Complex (Fig. 2). Although the transition occurs within a 25 mile (40 km) wide zone, roads that are nearly parallel to the strike of the transition zone require a longer trip.

**ROAD LOG**

**Mileage**

0.0 Begin at the Salmon River Inn, U.S. Highway 95, Riggins, Idaho. Drive 4.4 miles south on Highway 95 to Rapid River turnoff. Outcrops along the highway comprise the type section of the Squaw Creek Schist of the Riggins Group. The Squaw Creek Schist is a gray, compositionally layered schist which typically weathers rusty brown. The schists are dominantly composed of sodic plagioclase and quartz with abundant carbonate, biotite, muscovite, and carbon dust with minor amounts of epidote, clinzoisite, garnet, hornblende, and chlorite. Compositional layering of the schists has been transposed by pervasive deformation of this unit. Hamilton (1963) described the schists as marine sediments with a volcanic provenance.

4.1 This ultramafic pod is typical of those found along the contact between Squaw Creek Schist and Lightning Creek Schist.

4.4 Turn right and head west on Rapid River Road. Outcrops along the road expose the Lightning Creek Schist member of the Riggins Group, unconformably overlain by Quaternary gravels.

5.2 Junction; continue straight on dirt road.

6.6 Junction; stay left on main Rapid River Road.

7.0 Stop 1. Park near the Rapid River fish hatchery dam. At this locality we will look at a portion of
the Rapid River thrust system. The Rapid River thrust was originally mapped by Hamilton (1963, 1969) as a major post-metamorphic thrust which emplaces Riggins Group schists over the Seven Devils Group volcanics and the overlying Martin Bridge Formation and Lucile Slate. The Rapid River thrust actually consists of a system of faults which dip shallowly to the east at high structural levels, but steepen rapidly into a vertical shear zone with depth (Aliberti and Wemicke, 1986a). Here the basal thrust of the Rapid River thrust system emplaces Riggins Group schists, as well as blue-grey calcite marbles of the Martin Bridge Formation and black phyllites of the Lucile Slate, over Seven Devils Group volcanics.

Walk up the Rapid River trail to look at the basal thrust of the Rapid River thrust system. Rattlesnakes, poison ivy, and steelhead salmon are seasonally abundant along this trail and stream. Steep outcrops in the canyon consist of pervasively deformed marbles of the Martin Bridge Formation above the basal Rapid River thrust. Approximately 0.25 mile (0.4 km) up the trail on the right side there is a shallow cave in blue marble. The lower slab of marble shows excellent calcite stretching lineations (Fig. 3a). Mineral stretching lineations collected above the Rapid River thrust south of Riggins indicate northwest movement along the Rapid River thrust system (Fig. 3c; Aliberti and Wemicke, 1986a). Martin Bridge Formation and Lucile Slate intercalated with Riggins Group schists above the thrust suggest that all three units were tectonically interleaved before or during thrusting.

Continue west along the Rapid River trail to an unsigned trail junction; bear left on lower trail. Small-scale duplexing in the Martin Bridge Formation is observed in the cliff face across the river. Seven Devils Group metavolcanic tuffs and breccias crop out at the top of the next small hill approximately 0.5 mile (0.8 km) from the car. The mottled green and red colors are characteristic of these volcanics below the thrust. Steep foliations typical of the lower plate volcanics are locally present. Locally, tectonic slivers of Seven Devels Group volcanics are incorporated along the basal thrust. Looking back down river, the low-angle thrust contact between the Martin Bridge Formation and the Seven Devils Group is clearly visible as it crosses the Rapid River.

Return to car. Along the edge of the pool above the dam there are a few boulders of deformed volcanic breccias. These are typical of the Seven Devils Group volcanics, deformed in the root zone of the Rapid River thrust system upstream. Strain analysis of oriented breccias within the root zone shows that they have been flattened by pure shear and pressure solution along vertical foliation planes, which strike approximately N. 15°E. The subparallel attitude of the plane of flattening to the arc-continent boundary at this latitude suggests that the maximum shortening direction was at a high
angle to the continental margin (Aliberti and Wernicke, 1986b). Drive back on Rapid River road toward Highway 95.

7.25 Stop 2. Quaternary river gravels unconformably overlie pyrite-bearing, black phyllite of the Lucile Slate above the Rapid River thrust system. The Lucile Slate is correlated with the Hurwal Formation in Oregon (Brooks and Vallier, 1978), which stratigraphically overlies limestone of the Martin Bridge Formation, in turn resting conformably on Seven Devils Group volcanics of the Wallowa Terrane. At this locality, excellent mineral stretching lineations indicating west-northwest thrust movement are readily observed (Fig. 3b).

7.4 Stop 3. Intersection of Shingle Creek and Rapid River Roads. A silicified talc schist tectonite pod is present on the contact between the Lucile Slate and the Lightning Creek Schist. Such pods are volumetrically very minor, but they are the only material within the transition zone that may represent ocean floor.

9.4 Stop 4. The Lightning Creek Schist of the Riggins Group is a green to grayish green schist composed of chlorite, sodic plagioclase, quartz, actinolite and epidote with local occurrences of biotite, hornblende, garnet, clinozoisite, and carbonate. Here the chlorite schists are dipping gently eastward, with east-plunging mineral stretching lineations. Hamilton (1963) recognized this unit as a sequence of metavolcanic tuffs and flows with local agglomerate horizons. However, specific correlation of these schists with either the Olds Ferry Terrane or the Wallowa Terrane remains unresolved.

9.5 Highway 95; turn right and continue south.

11.8 Stop 5. Rest stop. Riggins Group Squaw Creek Schist is exposed in the roadcuts on the opposite side of the highway. The rock is typically a banded quartz-rich biotite schist or phyllite with variable amounts of carbonate. The metamorphic grade of Riggins Group schists increases eastward from the Rapid River thrust fault, where rocks are metamorphosed to lower greenschist facies, to the Pollock Mountain fault, where rocks reach upper greenschist to lower amphibolite facies. This increase in grade is interpreted to reflect a vertical section of the crust that was brought up along the steeply dipping Rapid River thrust system.

16.9 Turn right on Whitebird Ridge Road.

17.0 Junction; bear right and head up Whitebird Ridge Road.

17.4 Junction; bear left uphill.

19.2 Junction; bear left uphill.

19.7 Stop 6. This stop affords a good view of the Pollock Mountain fault on the hillside across the Little Salmon River. The trace of this low-angle fault cuts below resistant outcrops on the hilltops and down through upper terraces. It is offset slightly by a high-angle fault and then steepens into the river. The Pollock Mountain fault separates high-grade, garnet-bearing amphibolitic gneisses of the Pollock Mountain plate above, from lower grade Riggins Group schists of the the Rapid River thrust plate below. These two plates record radically different metamorphic histories. Two-stage garnets within the amphibolites suggest burial of the Pollock Mountain plate to pressures of 8 to 10 kb and temperatures of 600 to 650° C (Selverstone and others, 1987). In contrast, single-
stage garnets from the Rapid River thrust plate suggest isothermal decompression to 6 kb at 500 to 550°C (Selverstone and others, 1987). The importance of burial metamorphism in the Pollock Mountain plate and the dramatic change in history across the Pollock Mountain fault both suggest that metamorphism is related to regional tectonism and is not primarily the result of heat advected by bathototopic intrusives as suggested by Hamilton (1963). Hornblende stretching lineations along the fault indicate north-northwest movement. To the north, the low rolling hills and grassy slopes are all within the lower plate. Flat-lying lava flows in the distance are Miocene Columbia River Basalt.

20.4 **Stop 7.** Here the typically poorly exposed Pollock Mountain fault cuts through the stream and separates intrusives in the upper plate (outcrops on upper hillside cast of stream) from steeply dipping lower plate, low-grade Riggins Group schists (Squaw Creek Schist). The intrusives at this locality are part of the Whitehorse stock, an epidote-biotite tonalite with minor amounts (about 3%) of muscovite. This composition is similar to that of intrusives in the Hazard Creek Complex to the east. The intrusion shows evidence of pervasive brittle deformation related to movement along the Pollock Mountain fault. A late high-angle fault typical of post-Miocene faults which dissects the area cuts through Ranyhan gulch to the north. Turn around and head back toward Highway 95.

22.7 **Stop 8.** The Pollock Mountain Amphibolite is composed of garnet-bearing amphibolite and interlayered felsic gneisses. Variable microscopic to mesoscopic layering within the mafic amphibolites is characterized by alternating hornblende-rich, plagioclase-rich, and epidote-rich layers. The abundance of epidote associated with layered amphibolite is suggestive of calcic layers in a volcanic pile. Nd and Sr isotope systematics of the Pollock Mountain Amphibolite have been studied by Aliberti and others (1987). Whole rock samples of both mafic and felsic layers are isotopically indistinguishable and show a small range in $^{147}$Sm/$^{144}$Nd from 0.15 to 0.22. Data scatter about a reference isochron of 200 Ma and suggest a roughly Late Triassic crystallization age. Initial epsilon Nd values range from 6.65 to 8.09 and initial epsilon Sr values range from -6 to +22 ($^{87}$Sr/$^{86}$Sr = 0.7041 to 0.7061), suggesting these rocks were island arc volcanics that interacted with sea water either prior to or during metamorphism (Fig. 4). Both compositional characteristics and isotopic constraints indicate that the Pollock Mountain Amphibolite is a metamorphosed pile of Triassic island arc volcanics. A preliminary garnet-whole rock isochron from single-stage synkinematic garnets within the Pollock Mountain Amphibolite near Pollock Mountain gives a metamorphic age of 144 Ma, interpreted to be the earliest stage of metamorphism and deformation associated with the arc-continent collision (Aliberti and others, 1988). Early deformation of the Pollock Mountain Amphibolite was followed by synkinematic intrusion of the Hazard Creek Complex to the east. $^{40}$Ar/$^{39}$Ar ages ranging from 118 to 95 Ma (Snee and others, 1987) may be related to deformation and uplift along the Pollock Mountain fault and Rapid River thrust system, as well as increased magmatic activity.

This locality is typical of high grade injection gneisses of the Pollock Mountain Amphibolite above the Pollock Mountain fault, with garnet-bearing amphibolites pervasively injected by tonalite. Folded concordant sills of hornblende-biotite tonalite are coarse grained and contain xenocrystic garnets from the amphibolites. Deformation is syn- to post-intrusive. Younger but compositionally similar dikes are discordant to foliation, and these intrusives may be correlative with the oldest members of the Hazard Creek Complex. Sets of post-Miocene high-angle faults and joints are superimposed on earlier structures. Return to Highway 95.

23.9 Highway 95; turn right and head south.

25.8 Pinehurst; south of Pinehurst the outcrops along the highway are dominantly Columbia River Basalt.

32.2 Roadcuts in the Hazard Creek Complex begin here. The Hazard Creek Complex is the westernmost unit, composed primarily of Cretaceous intrusive rocks. The major members of the complex are variably deformed and recrystallized epidote-bearing quartz diorite to trondhjemite orthogneisses. These orthogneisses, emplaced during ongoing deformation, are the oldest intrusives along the suture zone. The structural style of the Hazard Creek Complex is very asymmetric. On the east the complex has a steeply dipping, generally north to north-northwest striking foliation. Plutons are elongate bodies with pervasive ductile folding around their margins. On the west, as seen in the following outcrops, dips are generally more shallow with diverse strike directions. Plutons are relatively equant with stoped blocks of wallrocks abundant along their margins. This asymmetry suggests the intrusives were emplaced within and to the west of an active zone of flattening and vertical flow.

32.6 **Stop 9.** These layered mafic gneisses are a minor component of the Hazard Creek Complex and crop

---

`Aliberti and Manduca--Transsect Across an Island Arc-Continent Boundary, West-Central Idaho 103`
The layered mafic gneisses are surrounded by intrusive material of the Hazard Creek Complex and are inferred to be a pendant within the intrusives. Smaller blocks of layered gneiss are common as inclusions within intrusives in this area. The presence of the gneisses as pendants and inclusions, as well as the irregular contact between the Pollock Mountain Amphibolite and the Hazard Creek Complex, suggest that the contact is intrusive rather than tectonic.

The most abundant gneisses are andesitic in composition, containing andesine, hornblende, biotite, quartz ± clinozoisite, epidote, garnet, diopside, and cummingtonite. Amphibolites, calcisilicate and quartzofeldspathic gneisses are interlayered with andesitic gneisses on scales from tens of centimeters to tens of meters. Amphibolites locally contain pyroxene. Calcisilicate gneisses contain varying proportions of hornblende, epidote, tremolite, plagioclase, quartz and sphene. These gneisses, interpreted as metamorphosed volcanics and volcaniclastics of an oceanic volcanic arc, strongly resemble the Pollock Mountain Amphibolite which crops out to the west of the Hazard Creek Complex. Quartzofeldspathic layers, composed of plagioclase, quartz, biotite, clinozoisite, muscovite and garnet, contain large relict, zoned plagioclase grains and are interpreted as metamorphosed tonalite or trondhjemite sills. Compositionally, they strongly resemble early intrusives from the Hazard Creek Complex. Younger cross-cutting biotite granodiorite dikes are probably related to younger intrusives in the Hazard Creek Complex.

34.4 Stop 10. The Hazard Creek Complex contains several generations of intrusives which have been variably deformed into gneisses. At this locality a variety of the older orthogneisses within the Hazard Creek Complex can be observed. The oldest orthogneiss is fine grained, of quartz diorite composition, and contains abundant folded leucocratic veins. It occurs both as blocks and as elongate pods within the epidote-biotite-hornblende tonalite orthogneiss which makes up most of the
outcrop. The tonalite orthogneiss is interlayered and deformed, with minor amounts of epidote-biotite-trondhjemite orthogneiss. Magmas of these compositions were injected throughout an ongoing deformational episode, since these lithologies occur as crystalloblastic gneisses, as gneissic intrusives and as little-deformed intrusives. Tonalite and trondhjemite dikes cross-cutting the orthogneisses represent one of the youngest intrusions in this series.

35.2 Stop 11. Rocks in this outcrop are primarily epidote-biotite tonalite and trondhjemite which form approximately half of the outcrop in the canyon. This body is one of the younger intrusives in the complex and is only slightly deformed. At this locality the intrusive is strongly foliated and contains concordant pods of orthogneiss and layered mafic gneiss. The tonalite is cut by trondhjemite and granite dikes and pegmatites. The latter are related to the youngest intrusive in the area: a leucocratic two-mica granite which crops out in Round Valley to the south.

35.5 Stop 12. Turnout on left; rocks are composed primarily of the epidote-biotite tonalite seen at Stop 11. However, at this locality the tonalite has a weaker foliation and contains stoved blocks of orthogneiss. These blocks can be observed both directly across from and to the south of the turnout. In the large boulder to the south of the turnout, a contact between the epidote-biotite tonalite and the crosscutting garnet-biotite trondhjemite dikes can be examined. Small clots of pure biotite within the granite are interpreted as restite fragments.

The contrast in deformation style between the rocks at Stops 11 and 12 is typical of the western Hazard Creek Complex, where intrusive and deformational styles vary locally. Structural data from the western Hazard Creek Complex show no consistent preferred orientation, in part due to rotation of stoved blocks. However, foliations within a single rock type at a single outcrop also vary significantly, suggesting they formed as undulating surfaces rather than as planar fabrics. This structural style contrasts with the pervasive, steeply dipping foliation observed in the eastern Hazard Creek Complex. Deformation in the eastern Hazard Creek Complex appears to have been largely synplutonic and records east-west compression and vertical flow. The presence of a strong, oriented fabric only on the east side of the complex is interpreted to reflect the emplacement of intrusives within and to the west of a narrow zone of deformation.

51.6 Stop 13. This epidote-biotite hornblende tonalite mylonite crops out along the eastern edge of the Hazard Creek Complex. The outcrop is representative of the larger bodies of older quartz diorite orthogneiss throughout the complex. These gneisses typically have crystalloblastic fabrics, but here the gneiss was deformed by a younger mylonitic event, centered on the Little Goose Creek Complex which crops out several hundred meters to the east. In this area the Little Goose Creek Complex and Hazard Creek Complex appear to be interlayered on a scale of tens to hundreds of meters. Mylonitization has obscured the original relationship between the units.

53.0 Stop 14. The Little Goose Creek Complex is composed primarily of this distinctive mylonitic, porphyritic biotite granodiorite orthogneiss. The porphyritic orthogneiss is tectonically interlayered with other intrusives ranging in composition from granite to gabbro. Epidote-free tonalites similar to those in the western portions of this outcrop are the most abundant interlayers. A hornblende pod, inferred to have been a cumulate, is present on the corner to the east. Layered mafic gneisses similar to those seen in the Hazard Creek Complex are also intercalated with the porphyritic orthogneiss, and can be observed in the upper part of the canyon. In the eastern portion of the Little Goose Creek Complex, metamorphosed continentally derived sediments are interlayered with the porphyritic orthogneiss. Thus, the orthogneiss contains pendants of both arc-related and continentally derived material. Strontium and oxygen isotopes within this unit have been studied by Manduca and others (1987). The isotopic ratios change dramatically from west to east. Calculated initial $^{87}\text{Sr}/^{86}\text{Sr}$ values range from 0.7045 in the west to 0.7097 in the east; $\delta^{18}\text{O}$ values range from 8.0 to 10.9 (Manduca and others, 1986). Ratios generally increase from west to east, with a step in the central part of the complex. This step does not coincide in detail with the change in composition of metamorphic pendants. The gradient is inferred to reflect an eastward-increasing contribution of evolved crustal material (with elevated $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{18}\text{O}$) to the melt which formed the porphyritic orthogneiss. The changes in both pendant compositions and isotopic ratios within the orthogneiss suggest it intruded a steeply dipping lithospheric boundary, along which cratonic and oceanic arc lithologies were juxtaposed.

The Little Goose Creek Complex has been pervasively deformed into a mylonitic gneiss. Mylonitization appears to have occurred under dry conditions at amphibolite facies, since amphibolite facies mineral assemblages have not been retrograded and oxygen isotope variations are
preserved. Mylonitization also affected the western margin of the Payette River Complex to the east, and the eastern margin of the Hazard Creek Complex to the west, as seen at Stop 13. Deformation postdates the emplacement of the Payette River tonalite, and is the youngest penetrative deformational event in the area. Mylonitic foliations strike north-south, and dips are steep to vertical. (Foliation at this outcrop dips more shallowly than average). A well developed mineral lineation plunges down dip, and fold axes generally parallel the lineation. Movement is inferred to have been vertical with a strong flattening component.

58.4 McCall; turn left on Warren Wagon Road and drive north along the west shore of Payette Lake through rocks of the Payette River Complex.

66.1 Junction; stay left on Warren Wagon Road.

72.7 Stop 15. Payette River Complex: the Payette River tonalite is part of a belt of tonalite which crops out on the west side of the Idaho batholith along much of its length. Here the body is approximately 6 miles (10 km) wide and has a strong north-striking igneous flow foliation that dips steeply to the east. A steeply northeast-plunging flow lineation is also commonly present. Mylonitization of the western margin is the only post-emplacement penetrative deformation observed in this body. Concordant screens and pods of metasedimentary rocks, inferred to be metamorphosed continentally derived clastics, are present throughout the complex. In this outcrop they are calcisilicate gneisses and biotite schists. Quartzite and sillimanite-bearing schist are also common. Metasedimentary pendants vary in size from a meter to several kilometers in width, and the largest body is greater than 6 miles (10 km) in length. The metasedimentary material is intimately associated with granitic to tonalitic material, suggesting partial melting and local assimilation occurred during intrusion of the tonalite. Structures and fabrics in the metasedimentary rocks are generally parallel to those in the tonalite, and appear to be due to transposition of the metasedimentary rocks during intrusion of the tonalities. Older structures and fabrics may have been obliterated. The cross-cutting granitic dikes at this outcrop are related to the two-mica granitic core of the Idaho batholith which crops out 4 miles (6.5 km) to the east.

ACKNOWLEDGMENTS

Aliberti's research has been supported by NSF grant EAR 84-51181 to B. P. Wernicke, Geological Society of America research grants 3379-85 and 3531-86, and Harvard University field research grants. Field work by Manduca has been supported by the U.S. Geological Survey in collaboration with M. A. Kuntz. This guide stems from Aliberti's Ph. D. dissertation at Harvard University and Manduca's Ph. D. dissertation at the California Institute of Technology under the direction of L. T. Silver. The manuscript was reviewed by D. W. Rodgers and V. E. Chamberlain.

REFERENCES

Aliberti, E. A., and Wernicke, B. P., 1986a, Late stage processes in the growth of new continental crust: observations from the eastern margin of the Seven Devils terrane, west-central Idaho: Geological Society of America Abstracts with Programs, v. 18, no. 6, p. 524-525.

———, 1986b, Occluded terrane boundaries: an example from an island arc-craton contact in west central Idaho (abst.): Transactions of the American Geophysical Union, v. 67, no. 44, p. 1226-1227.


Manduca, C. A., Silver, L. T., and Taylor, H. P., 1986, Study of an abrupt change in Sr\textsubscript{i} and \delta\textsuperscript{18}O in granodiorite in the western border of the Idaho batholith (abst.): Transactions of the American Geophysical Union, v. 67, no. 44, p. 1268.


Hydrothermal Systems of the Wood River Area, Idaho

Duncan Foley 1
Leah Street 2

INTRODUCTION

This field trip guide describes hydrothermal systems along the southeastern margin of the Idaho batholith in the Wood River area of south-central Idaho (Fig. 1). Our study was undertaken to develop a regional understanding of hydrothermal systems in the Wood River Valley and to evaluate the potential for undiscovered thermal resources in this and similar geologic environments. These systems are of interest geologically for their occurrence in a physiographic transition zone, geochemically for their relatively dilute but high fluorine waters, and economically for their current and potential applications. Guyer Hot Springs, located 3.2 km (2 mi) west of Ketchum, is currently being used to heat more than a hundred homes. Hailey Hot Springs, 3.4 km (2.1 mi) west of Hailey, is the site of a proposed resort development. Magic Hot Springs, at the eastern edge of the Camas Prairie, has been proposed for a variety of industrial applications.

The Wood River area is a transition between the Cretaceous and Tertiary intrusive rocks that compose the Idaho batholith, the Paleozoic thrust belt of the northern Rocky Mountains, and the Cenozoic terrains of the Snake River Plain volcanic province. The batholith terrain north of the study area reaches elevations of nearly 3,000 m (10,000 ft) and has steep-walled valleys that are only a few kilometers wide. The northern Rocky Moun-

1 Department of Earth Sciences, Pacific Lutheran University, Tacoma, WA 98447
2 Idaho Department of Water Resources, 2148 4th Avenue East, Twin Falls, ID 83301

Figure 1. Location map depicting hydrothermal systems and regional heat flow values from Brott and others (1981). RJ - Russian John Hot Spring; E - Easley Hot Spring; W - Worfield (Frenchman's Bend) Hot Springs; G - Guyer Hot Springs; C - Clarendon Hot Springs; HI - Hailey Hot Springs; M - Magic Hot Spring; P - Picabo warm well.

Reconnaissance geologic maps in the areas of several of regional compilations (Rember and Bennett, 1979). Structurally complex than currently published maps mapping suggests that many hot spring areas are more mapped the Magic Reservoir area. Our reconnaissance suggest.

Area around Magic Hot Springs, and Leeman (1982) has Hot Springs. Struhsacker and others (1982) mapped the Camas Prairie area, and it included the Wood River area in a statewide summary of geothermal resources (Mitchell and others, 1980). Individual systems that have been investigated include Magic Hot Springs (Struhsacker and others, 1982) and Guyer Hot Springs (Blackett, 1981). Portions of the work discussed herein have been reported in Zeisloft and others (1983), Foley and others (1983), and Foley and Street (1985; 1986).

Geologic studies of the Wood River area can be divided into the following categories: regional summaries to explain mineral resources (Umpleby and others, 1930; Anderson and Wagner, 1946; Anderson and others, 1950; Hall and others, 1978); studies of individual geologic units (Hall and others, 1974; Sandberg and others, 1975); quadrangle studies (Batchelder and Hall, 1978), and regional compilations (Rember and Bennett, 1979). Reconnaissance geologic maps in the areas of several of the hot springs are contained in Anderson and Bideganeta (1985). Blackett (1981) mapped the area around Guyer Hot Springs. Struhsacker and others (1982) mapped the area around Magic Hot Springs, and Leeman (1982) has mapped the Magic Reservoir area. Our reconnaissance mapping suggests that many hot spring areas are more structurally complex than currently published maps suggest.

Regional Setting

The complex geologic history of the Wood River area can be simplified to three factors that are relevant to the study of hydrothermal systems. These factors are: lithologic units that maintain open fractures regardless of their origin; a tectonic history that has developed a complex interconnected network of such fractures through faulting and intrusion; and topographic, thermal, and hydrologic conditions that allow the development of hot springs. The Wood River region, which lies at the intersection of the three major geologic provinces discussed above, is in a favorable geologic environment for geothermal resources (Foley and Street, 1985). The discussion below is intended to summarize only geologic factors relevant to the numerous hydrothermal systems. More detailed discussions of stratigraphic and structural conditions, and further references to the geologic environment in the Wood River area, can be found in other works in this volume.

Sedimentary, intrusive, and extrusive rocks are found at the Wood River area hot springs. Consolidated sedimentary rocks in the vicinity of the hot springs include the Paleozoic Millingen, Wood River, and Dollarhide Formations (Hall and others, 1974; Hall, 1985; Link and others, 1987; 1988, this volume; Otto and Turner, 1987; Sandberg and others, 1975; Wavra and others, 1986). Mesozoic and Cenozoic intrusions of the Idaho batholith and related rocks (Bennett and Knowles, 1985; Johnson and others, 1988, this volume) created many of the conditions that are currently favorable for the existence of hot springs in the area, including metamorphism, tectonic uplift, possible movement along thrust faults, and the development of extensive fracture networks. Intrusion-related fractures served as channels for ore-forming fluids at the time of emplacement (Hall and others, 1978; Howe and Hall, 1985) and where now open they may be important in modern hydrothermal circulation systems. It is noteworthy that most intrusive rocks, mining districts, and hydrothermal systems are located west of the Big Wood River. Cenozoic Challis Volcanics (Hall and McIntyre, 1986) conceal older sedimentary and intrusive rocks in many areas. Miocene and Pliocene rhyolites, ash-flow tuffs and ferrolatites exposed in the Magic Hot Springs area (Bonnichsem and others, 1988, this volume) are interlayered with and covered by Quaternary basalts from volcanic centers of the Snake River Plain.

Two major trends of fractures and several generations of faults exist near hot springs in the Wood River area. A northeast trend, generally parallel to the Snake River Plain and the Trans-Challis fault system (Bennett, 1986), is perhaps related to the Dillon Lineament of Ruppel (1982). A northwest trend, typified by the strike of the Wood River Valley, is found in local geologic structures at many hot springs. Thrust faults (Skipp and Hait, 1977; Skipp, 1987), high-angle faults related to batholith emplacement (Hall and others, 1978) and current tectonism (Howard and others, 1978), and low-angle normal faults (Wust, 1986; Link and others, 1987; Otto and Turner, 1987) are all found adjacent to hot springs.

Hydrologic and Thermal Conditions

The two most important factors in groundwater
movement in hydrothermal systems of the Wood River area are the differences in elevations between recharge areas and discharge sites, and the thermoartesian head. Topographic relief of more than 1,000 m (3000 ft) between probable recharge areas in mountain ranges and springs and wells in valley bottoms could provide an adequate drive for the hydrothermal systems. Thermoartesian head may make a minor contribution to the upwelling of waters.

Water temperatures measured at hot springs and wells are a function of flow paths, flow rates, thermal flux from the earth, and local and regional stratigraphic and structural conditions. These factors also influence measured thermal gradients. To understand the nature of the systems, subsurface data are required.

Subsurface thermal data for the Wood River area consist only of shallow thermal gradient measurements in wells near selected hot springs, and they must be largely characterized from thermal gradient measurements and heat flow calculations in surrounding terrains. The thermal regime of the Snake River Plain has been studied by Brott and others (1976; 1981), who report heat flow values of approximately 70 to 400 milliwatts/square meter for the Camas Prairie, Mt. Bennett Hills, and Magic Reservoir areas (Fig. 1). They also report a value of 51 milliwatts/square meter from a well northeast of Hailey. These data suggest that heat flow along the margin of the Snake River Plain may be higher than it is north of the plain.

At least one mine in the Wood River District has recorded temperatures that indicate a thermal anomaly. The Croesus mine in Scorpion Gulch, south of Croy Creek and southwest of Hailey Hot Springs, was noted by Umpleby and others (1930) as being unusually warm. They reported a temperature of 13.9°C (57°F) at the 61 m (200 ft) level, and 33.9°C (92°F) at the 244 m (800 ft) level. If these temperatures are accurate, they represent an anomalous thermal gradient of more than 100°C/km.

Most groundwater studies in the Wood River area have emphasized cold, shallow-depth waters from alluvial aquifers or basalt flows (Smith, 1959; Walton, 1962; Castelin and Chapman, 1972) or the impacts of urbanization on water quality (Castelin and Winner, 1975), rather than hydrothermal systems. Smith (1959, p. 15) concludes that the Paleozoic sedimentary and Mesozoic and Tertiary intrusive rocks are "relatively impermeable." The implication of the cold-water studies is that the consolidated sedimentary and intrusive rocks do not contain laterally extensive aquifers. Hot springs, however, are closely associated with the "relatively impermeable" rocks. Interconnected networks of permeable fractures in these rocks are sufficient to allow hydrothermal systems to exist. It is important to note that not all fractures are permeable channels; thrust faults or other faults in the Wood River area may be permeability barriers by being sealed or by placing impermeable rocks against permeable rocks. Faults acting as hydraulic barriers have been noted in Wood River ore deposits by Hall (1985), and in the thrust belt of southeastern Idaho by Ralston and others (1983).

HYDROTHERMAL SYSTEMS

Introduction

Thermal springs and wells in the Wood River area typically occur as individual springs or as closely spaced series of springs and wells, which are usually separated from adjacent hydrothermal systems by at least several kilometers. Hydrothermal systems in the Wood River area, with the exception of the thermal well near Pocatello, are located south and west of the Big Wood River (Fig. 1).

Our conceptual model for geothermal resources in the Wood River area is simple in overall concept but complex in detail. The hypothesis that seems to best explain these systems is that thermal waters are recharged from snowmelt in the mountain ranges. Water flows through intrusive rocks at the margin of the Idaho batholith and affiliated stocks. Thermal waters primarily discharge either from the edges of intrusive rock exposures or from structurally complex terrains that are probably associated with intrusive rock subcrops. Water chemistry, including fluorine concentrations and trilinear plots, provides strong evidence that Wood River thermal waters have primarily flowed through batholith rocks. The tight controls of discharge points suggest that flowpaths for each spring are distinct, and no "common reservoir" exists at depth.

The four hydrothermal systems that we will see on the field trip include, in the order of visitation: Magic Hot Springs, Hailey Hot Springs, Clarendon Hot Springs, and Guyer Hot Springs. These springs are described individually below. Other thermal springs and wells in the Wood River area are discussed after the field trip road log. The geochemistry of the springs is presented at the beginning of this section, to allow easy comparison between sites.

Geochemistry

The chemistry of thermal waters can be used to assess the waters' origin, to indicate subsurface hydrologic and thermal conditions, and to compare different thermal waters in local and regional settings (Ellis and Mahon, 1977). Chemistry can also indicate possible environmental and engineering problems, such as the high fluorine concentrations of Wood River area thermal waters that might accompany development of a thermal resource.

Table 1 presents the results of analyses of major and selected trace elements, and stable isotopes for Wood River area thermal and cold waters. Analytical results of major and trace elements are expressed in milligrams per liter (mg/l). Temperatures are reported in degrees Celsius,
Table 1. Geochemistry of Wood River area waters.

| REF. | SITE | LOCATION | DATE | TEMP (°F) | pH | EC | TDS | Si | PO_4 | Fe | Mg | Ca | Na | K | Cl | F | Mg/Cl | Ca/Cl | Mg/Fe | Mg/Na | Mg/Cl | SI | Fe/Cl | Mg/Cl | Mg/Na |
|------|------|----------|------|-----------|----|----|-----|---|------|----|----|----|----|----|----|----|----|-------|-------|--------|--------|--------|----|-------|--------|--------|
| 1    | Magic #1 | 1S-17E-212AA | 10/15/81 | 75.5 | 36.9 | 20.2 | 1.32 | 78.1 | 1.2 | 1.40 | 0.61 | 0.47 | 0.06 | 0.12 | 0.06 | 0.06 | 0.02 | 0.03 | 0.01 | 0.01 | 0.01 | 0.01 |
| 2    | Magic #2 | 1S-17E-212AA | 10/15/81 | 70.5 | 37.2 | 20.2 | 1.32 | 75.1 | 1.1 | 1.15 | 0.15 | 0.06 | 0.12 | 0.05 | 0.06 | 0.06 | 0.06 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 |
| 3    | Magic #3 | 7S-17E-212AA | 10/15/81 | 75.5 | 37.2 | 20.2 | 1.32 | 75.1 | 1.1 | 1.15 | 0.15 | 0.06 | 0.12 | 0.05 | 0.06 | 0.06 | 0.06 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 |
| 4    | Magic #4 | 7S-17E-212AA | 10/15/81 | 75.5 | 37.2 | 20.2 | 1.32 | 75.1 | 1.1 | 1.15 | 0.15 | 0.06 | 0.12 | 0.05 | 0.06 | 0.06 | 0.06 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 |
| 5    | Magic #5 | 1S-17E-212AA | 11/15/81 | 70.5 | 37.2 | 20.2 | 1.32 | 75.1 | 1.1 | 1.15 | 0.15 | 0.06 | 0.12 | 0.05 | 0.06 | 0.06 | 0.06 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 |
| 6    | Magic #6 | 7S-17E-212AA | 11/15/81 | 75.5 | 37.2 | 20.2 | 1.32 | 75.1 | 1.1 | 1.15 | 0.15 | 0.06 | 0.12 | 0.05 | 0.06 | 0.06 | 0.06 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 |
| 7    | Magic #7 | 7S-17E-212AA | 11/15/81 | 75.5 | 37.2 | 20.2 | 1.32 | 75.1 | 1.1 | 1.15 | 0.15 | 0.06 | 0.12 | 0.05 | 0.06 | 0.06 | 0.06 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 |
| 8    | Magic #8 | 7S-17E-212AA | 11/15/81 | 75.5 | 37.2 | 20.2 | 1.32 | 75.1 | 1.1 | 1.15 | 0.15 | 0.06 | 0.12 | 0.05 | 0.06 | 0.06 | 0.06 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 |
| 9    | Magic #9 | 7S-17E-212AA | 11/15/81 | 75.5 | 37.2 | 20.2 | 1.32 | 75.1 | 1.1 | 1.15 | 0.15 | 0.06 | 0.12 | 0.05 | 0.06 | 0.06 | 0.06 | 0.02 | 0.01 | 0.01 | 0.01 | 0.01 |

Clementi analyses by R. Katesman, University of Utah Research Institute.
<table>
<thead>
<tr>
<th>Source</th>
<th>TDS</th>
<th>EC</th>
<th>pH</th>
<th>Cl</th>
<th>Si</th>
<th>Mg/Fe</th>
<th>Mg/Na</th>
<th>Mg/Cl</th>
<th>Mg/Na</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>2</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>3</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>4</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>5</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>6</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>7</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>8</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>9</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
<td>0.06</td>
</tr>
</tbody>
</table>

Notes:
- **Temperature range**: 32-68°F (0.6/19.8°C)
- **pH range**: 6.9-8.5
- **TDS range**: 0.05-0.3 mg/L
- **Cl range**: 0.05-0.4 mg/L
- **Si range**: 0.05-0.4 mg/L
and flow rates are given in liters per minute. Isotopic analyses are reported in delta notation ("δ"; parts per mil relative to standard mean ocean water (SMOW)). Field sampling was done following the method of Kroneman (1981). Analyses at the University of Utah Earth Science Laboratory for major and trace elements were conducted using the following methods: fluorine was determined by specific ion electrode, chloride by silver nitrate titration, total dissolved solids and sulfate by gravimetric methods, bicarbonate by titration, and all other elements by inductively coupled plasma quartzometer (Christensen and others, 1980). Isotopic analyses are by E. Ripley at Indiana University.

Wood River thermal waters plot in the Na-HCO₃ field on the trilinear diagram of water chemistry (Fig. 2). Cold waters are largely the Ca-HCO₃ type. Waters that plot between the areas of cold and thermal waters may be interpreted as having an origin of deep thermal waters rising and mixing with cooler, near-surface waters.

![Figure 2. Trilinear plot of geochemical data, comparing Wood River area thermal and cold waters with thermal waters from the Idaho batholith.](image)

The Na-HCO₃-SiO₂ chemistry of Wood River area thermal springs compares closely with the chemistry of thermal springs in the Idaho batholith (Young, 1985). Many springs analyzed by Young, however, have a greater SO₄ content than do Wood River area waters. The generally similar chemistry, however, is compatible with the hypothesis that both sets of thermal waters have similar histories: recharge in the batholith, thermal equilibration at depth with intrusive rocks, and discharge along structural zones in the rocks.

Fluorine concentrations in Wood River thermal waters are high. The most likely source of the fluorine is from water-rock interaction with intrusive rocks (Table 2), as other rocks in the Wood River area are low in fluorine. Our data and the studies of Bennett and Knowles (1985) suggest that fluorine may be leached out of fluorine-bearing micas. This implies a significant role for intrusive rocks in the flowpaths of the thermal waters, including chemical equilibration, despite the fact that most thermal springs issue from Paleozoic sedimentary rocks. Fluorine also is an environmental concern, since concentrations in thermal waters are well above standards for drinking water, and problems of fluorosis in wildlife have been noted in the area (J. Shoupe, Utah State University, personal communication, 1986).

Geochemical thermometers suggest possible subsurface equilibration temperatures for the thermal waters (Table 1). Temperatures are calculated using the formulas of Fournier (1981). Two columns of temperatures are presented for silica thermometers: "TSiO₂ (qtz cond)," which assumes that the hydrothermal fluids have equilibrated with quartz in the subsurface and have cooled conductively during their rise to the surface, and "TSiO₂ (chalc)," which assumes that chalcedony is the silica phase controlling the amount of silica in thermal waters. These silica temperatures have not been corrected for pH effects (e.g., Young, 1985), which could lower the calculated values from those reported in Table 1 and bring them into closer agreement with other geothermometers. Cation equilibrium temperatures are presented in the column labeled "TNaKCa." These temperatures have been calculated, where appropriate, using a correc-

### Table 2. Fluorine in rocks and minerals of central Idaho.

<table>
<thead>
<tr>
<th>Geologic material</th>
<th>Fluorine content, ppm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Challis Volcanics, west of Hailey Hot Springs (whole rock) (1)</td>
<td>540</td>
</tr>
<tr>
<td>Paleozoic quartzite, adjacent to Hailey Hot Springs (whole rock) (1)</td>
<td>510</td>
</tr>
<tr>
<td>Paleozoic limestone, adjacent to Hailey Hot Springs (whole rock) (1)</td>
<td>230</td>
</tr>
<tr>
<td>Paleozoic limestone, near Guyer Hot Springs (whole rock) (1)</td>
<td>210</td>
</tr>
<tr>
<td>Tertiary volcanics, near Magic Hot Springs (whole rock) (1)</td>
<td>330</td>
</tr>
<tr>
<td>Eocene granodiorite, Clarendon stock near Clarendon Hot Springs (whole rock) (1)</td>
<td>730</td>
</tr>
<tr>
<td>Biotite from Eocene granodiorite, Clarendon stock near Clarendon Hot Springs (1)</td>
<td>3,700</td>
</tr>
<tr>
<td>Biotite from Cretaceous granodiorite (2)</td>
<td>9,000</td>
</tr>
<tr>
<td>Biotite from Sawtooth batholith (2)</td>
<td>37,000</td>
</tr>
</tbody>
</table>

Data sources: (1) this paper; (2) Bennett and Knowles (1985).
tion for the amount of magnesium in the waters (Fournier and Potter, 1978). The cation temperatures have not, however, been corrected for the effects of carbon dioxide (Paces, 1975).

The geochemical thermometer temperatures must be interpreted with caution. Low-temperature waters may not satisfy the assumptions for reliable interpretation of geothermometers (Fournier and others, 1974). However, the general agreement of silica and cation temperatures suggests that the results are reasonably valid. The thermometers indicate that, with the possible exception of Magic Hot Springs, subsurface thermal waters at any site are not likely to be significantly warmer than known surface waters.

Isotopic analyses of Wood River thermal waters suggest that the waters have a meteoric origin. These analyses, with the exception of Magic Hot Springs, plot near the meteoric water line of Craig (1961) (Fig. 3). The range of isotope values in Wood River thermal waters is, with the exception of δ18O at Magic Hot Springs, within the ranges of δD and δ18O reported by Young (1985) from thermal waters of the Idaho batholith that were -132 to -152 per mil and -17.2 to -19.9 per mil, respectively. The oxygen shift at Magic Hot Springs, which causes deviation of the thermal waters from the meteoric water line, may be due to a shift of δ18O during moderate temperature water-rock interaction (Truesdell and Hulston, 1980), in this case probably with non-batholith rocks. The δD depletion implies that the thermal waters are not derived from current precipitation (Young, 1985).

FIELD TRIP ROAD LOG

Timmerman Hills Rest Area to Magic Hot Springs and Return to Hailey

Mileage

0.0 Rest area; as we head west on State Highway 20 toward Fairfield, the Timmerman Hills are to the south, and the southeastern end of the Hailey Hills is to the north. The Timmerman Hills are mapped by Rember and Bennett (1979) as Tertiary and Quaternary volcanic rocks. The hills are primarily composed of Eocene Challis Volcanics and Miocene Idaho Basalt. The mountains to the north are composed primarily of intrusive rocks. The Timmerman Hills are mantled on their lower slopes by Quaternary glacial and periglacial deposits (Schmidt, 1961). The road in this area is constructed on glacial outwash (Schmidt, 1961). To the west, it is constructed on river terrace deposits.

1.5 The prominent hill to the south is mapped by Schmidt (1961) as Eocene volcanic and Cretaceous intrusive rocks.

2.1 Stanton Crossing; Big Wood River.

2.4 Road cut in Cretaceous granodiorite (see Johnson and others, this volume). The rocks to the south are Quaternary Macon Basalt, which are intercalated with and covered by several sedimentary units. As we drive west, we will continue to see outcrops of granodiorite near the road and basalt flows to the south.

4.2 Moonstone Mountain, which is composed of 3-4 Ma rhyolites (Honjo and others, 1986) and marks the northeastern edge of the Magic Reservoir volcanic center (Leeman, 1982), can be seen to the west. The Hailey Hills to the north are composed primarily of intrusive rocks.

5.2 Quaternary Wind Ridge Basalt, mantled by Macon Basalt and proglacial and periglacial deposits (Schmidt, 1961), is exposed.

5.6 Turn off on gravel road to the south, to Magic Hot Springs.

6.4 Stop 1: Magic Hot Springs. Park in the large open area.

Introduction. Magic Hot Springs is different from the other Wood River area thermal springs in geologic setting, geochemistry, and thermal regime. The geology of the Magic Hot Springs area is discussed by Mitchell.
Numerous proprietary studies have been conducted at Magic Hot Springs and several attempts have been made to develop the springs. The results of work by Hunt Energy Corporation have been released to the Bureau of Land Management, where we have obtained some of the data discussed below.

The main well (Fig. 4) is visible at the north end of the parking area, just east of the buildings of an old resort. It has an artesian flow of about 57 liters/minute (14 gpm) and a depth of 79 m (260 ft) (Anderson and Bideganeta, 1985). Temperatures of this well have been measured at 66 to 74.5°C (151 to 166°F) during the past few years. A second well, which does not flow, is located east of the parking area at the top of a small knoll. Anderson and Bideganeta (1985) report a depth of 117 m (384 ft) and water temperature of 37°C (99°F) for this well. Three thermal seeps occur below the high water line of the reservoir and are most easily seen as snowmelt areas in winter. Two seeps (A and B on Fig. 4) with temperatures of 32 and 39°C (90 and 102°F) are located southwest of the main well on the north side of the reservoir. The third seep ("C" on Fig. 4), which has a temperature of 14°C (57°F), is located on the southwest side of the reservoir.

Geologic Setting. Magic Hot Springs is located in a geologically complex area, with faulted Miocene and Pliocene volcanic rocks cropping out at the junction of the Wood River Valley, Camas Prairie, and the Snake River Plain. The geologic map (Fig. 4) is adapted from Struhsacker and others (1982).

The Magic Hot Springs area has been interpreted as being located at the edge of a volcanic center (Leeman, 1982; Honjo and Leeman, 1987), with a complex sequence of ash flow tuffs and rhyolites probably from local sources (Struhsacker and others, 1982). These rocks overlie both granitic intrusive rocks and older volcanic rocks. Our geologic reconnaissance of the area has identified a previously unmapped granitic intrusive rock outcrop approximately 300 m (1000 ft) south of the hot springs on the western side of the reservoir (below high water), and intrusive rocks are mapped 1.5 km (0.9 mi) southeast of the hot springs (Rember and Bennett, 1979). The older volcanic rocks have been dated at 9-10 Ma, and volcanic rocks of the Magic Reservoir eruptive center have been dated at 3-6 Ma (Struhsacker and others, 1982; Honjo and others, 1986).

Struhsacker and others (1982) map an extensive series of high-angle normal faults intersecting older pre-Quaternary geologic units. The faults cut all pre-Quaternary geologic units. The faults have east-west, northeast, and northwest trends. The northwest trend is similar to major faults mapped by Leeman (1982) in the Mount Bennett Hills and by Rember and Bennett (1979) both northeast and southwest of the hot springs area. Rember and Bennett (1979) also depict offset of Quaternary alluvium by at least one of the northwest-trending faults.

Geochemistry. Thermal waters at Magic Hot Springs have higher concentrations of dissolved solids, higher silica, and a greater isotopic shift than other Wood River area thermal waters. These differences in water chemistry may be due to a number of factors, including higher-temperature waters at depth as suggested by chemical thermometry results of 100 to 150°C (212 to 300°F), and chemical re-equilibration of Magic Hot Springs waters with volcanic and perhaps more reactive vitric units (White and others, 1980) rather than batholith rocks. Chemical re-equilibration of thermal waters at depth with non-batholith rocks might imply the existence of a high-temperature, fractured volcanic rock reservoir in the vicinity of the hot springs; currently, no such reservoir is known.

Thermal seeps A and B (Fig. 4), adjacent to the hot well, may be conductively cooled Magic Hot Spring
waters, since they are chemically indistinguishable from the well. Seep C, on the southeast side of the reservoir, has an intermediate chemical composition compared with other thermal waters and regional cold waters. Seep C is characterized by a 10 x 20 m (30 x 60 ft) area of saturated sediments, with many small outlets that aggregate only about 4 liters/minute (1 gpm) flow. The 4 mg/l fluorine in seep C waters is much greater than the <1 mg/l fluorine that is typically found in cold waters of the area, and indicates that the waters are probably thermal fluids that have mixed with near-surface waters.

**Thermal setting.** The Magic Hot Springs area has been intensively explored for geothermal resources and therefore has the most complete subsurface thermal data of any site in the area. Thermal data from Hunt Energy Corporation and Brott and others (1981) are presented in Figure 5. The regionally high heat flow data point (156 milliwatts/square meter) of Brott and others (1981) is not atypical for heat flow along the edge of the Snake River Plain. The thermal gradient data of Hunt Energy, however, demonstrate that areas with high thermal gradients, which are attractive for further geothermal exploration, are limited in areal extent. Neither Brott and others (1981) nor Hunt Energy report data from wells in the vicinity of the known hot springs and wells. It is likely that high thermal gradients represent areas of upwelling warm waters, and the restricted areas of such gradients may imply structural control of the upwelling along faults such as those mapped by Struhsacker and others (1982).

Return to vehicles, and drive back to the highway along the dirt road.

7.3 Turn west on State Highway 20.
7.8 Quarry in volcanic rocks on south side of highway.

8.2 Stop 2: Road cut at top of hill. Here we will look at the fracture permeability that characterizes ash flow tuffs in the Magic Hot Springs area. The ash-flow tuffs show variations in welding and hydrothermal alteration that can greatly change hydrologic properties of the rock units within a few feet.

Turn around; head east on Highway 20 toward Timmerman Hills rest area.

9.1 Road to Magic Hot Springs.

14.9 Intersection of State Highway 20 with State Highway 75; turn north toward Bellevue, Hailey, and Ketchum.

26.9 Hailey city limits.

28.3 Intersection of Bullion and Main Streets. Reset odometers.

Hailey to Hailey Hot Springs

Mileage

0.0 Turn west on Bullion at intersection with Main.
Drive west up Croy Creek drainage.

0.4 Big Wood River. Basal conglomerates of the Wood River Formation are well exposed at the bridge. The rocks in this area have been steeply folded and have been moved along thrust faults.

1.2 Animal shelter on north side of road.

1.7 Northeastern edge of Democrat Gulch. This area is part of the Mineral Hill mining district, which was originally mined in the late 1800s. Only small mines and exploration pits exist in the area of the hot springs. Umpleby and others (1930) discuss lead and silver production from the Democrat claims at the head of Democrat Creek. The ore occurred along a shear zone in quartz monzonite.

2.1 Stop 3: Hailey Hot Springs. Park by the gate to the field on the west side of Democrat Creek. We will walk to the hot springs from the road.

Introduction. Hailey Hot Springs is located in Democrat Gulch, approximately 3.2 km (2 mi) west of Hailey, north of the valley of Croy Creek. The area has been studied by Blackett (1981, unpublished data, University of Utah Research Institute) and the authors, as well as Batchelder and Hall (1978) and Hall (1985). The hot springs site consists of one major spring and several shallow wells (Fig. 6). The main spring and wells are sited in valley-bottom sediments. Flow of the main spring has been estimated to be 265 liters/minute (68 gpm) by Anderson and Bideganeta (1985); the wells either have low flow or do not flow at all. Surface temperatures of the hot spring and thermal wells range from 42-61°C (108-142°F). Anderson and Bideganeta (1985) report a maximum temperature of approximately 17°C (63°F) at a depth of 34 m (112 ft) in a well, which probably indicates a mixing of thermal and cold waters. Data discussed earlier from the Croesus mine, which is located approximately 3.6 km (2.25 mi) south of the hot springs, indicate that high thermal gradients may exist in other isolated areas in the region of the hot springs. Previous use of the spring to provide heat to Hailey is illustrated by several pieces of wooden pipe that remain in the area. Hailey Hot Springs has been proposed recently for resort development.

Geologic setting. Geology at Hailey Hot Springs is complex. Paleozoic sediments in the area were mapped by Batchelder and Hall (1978) as the Milligen and Wood River Formations. Recognition of the Permian Dollarhide Formation (Hall, 1985) with subsequent reassignment of some rocks mapped by Batchelder and Hall (1978) as Milligen Formation to the Dollarhide Formation, and discovery of a series of northeast-trending, down-to-the-east normal faults (Zeisloft and Foley, unpublished mapping, University of Utah Research Institute) imply that new detailed geologic mapping is required before a comprehensive model of the surface and shallow subsurface geology of this site can be developed. The newly discovered faults have displacements of meters to perhaps tens of meters, trend approximately N.40°E., and have been interpreted primarily from...
topographic expression on the major hill northeast of Hailey Hot Springs. This hill also has low-angle normal faults that may be reactivated thrust faults (e.g., Link and others, 1987).

An outcrop of limestone, mapped by Hall (1985) as part of the Dollarhide Formation, is adjacent to the hot springs. This limestone is highly contorted and fractured, and it may be in fault contact with slitchones and quartzites southwest of the springs. Although Batchelder and Hall (1978) indicate a small outcrop of intrusive rocks southwest of the hot springs, this exposure could not be located with certainty in our preliminary geologic mapping. Intrusive rocks of the Croesus and McCoy mine stocks (Johnson and others, this volume) have been mapped approximately 1.5 km (1 mi) south of the hot springs (Rember and Bennett, 1979).

**Geochemistry.** The low total dissolved solids, Na-HCO₃, and high fluorine (11-16 mg/l) thermal waters at Hailey Hot Springs are chemically similar to most other thermal waters in the Wood River area. The chemical similarity, particularly with thermal waters of the batholith, implies that these waters also have equilibrated with batholith rocks.

**Turn around and drive back to Hailey.**

4.2 Intersection in Hailey. Reset odometers.

**Hailey to Clarendon**

0.0 Intersection of Bullion and Main in Hailey. Drive north on State Highway 75.

2.8 Deer Creek road; turn west.

3.3 Big Wood River.

7.3 **Stop 4:** Clarendon Hot Springs overlook.

**Introduction.** Clarendon Hot Springs is a widespread thermal resource (Fig. 7). Mitchell and others (1980) report a temperature of 47°C (117°F) and a flow rate of 378 liters/minute (100 gpm) from a spring. We have measured a temperature of 54°C (129°F) at a well adjacent to the former hot springs resort. Two additional wells exist in the area. One is located west of the main resort, adjacent to an old swimming pool, and has a temperature of 54°C (129°F); a second well, located on the south side of the drainage, has a temperature of 26°C (79°F). These wells have flow rates of 25 and 22 liters/minute (6.25 and 5.5 gpm) respectively. The resource has been used in the past for space heating a few dwellings and for swimming pools. Plans to redevelop the area have been made by various individuals, but to date no further activity has occurred.

**Geologic setting.** Rocks of the Wood River and Dollarhide Formations are in contact along a thrust fault and have been intruded by the Deer Creek stock in this area (Hall, 1985; Bennett and Knowles, 1985). Thermal wells and springs are located near the intrusive contact, but detailed geologic relations are obscured by alluvium at the well sites.

Anderson and Bideganeta (1985) report near-vertical attitudes of the tightly folded and jointed pre-Tertiary sediments near the contact with the intrusion. Quartz veins associated with the intrusion have been emplaced into the sediments. The veins have been explored for mineral content but have not been exploited. South of the contact, the granite shows minor east-west shearing (Anderson and Bideganeta, 1985). A major northwest-trending fault has been mapped by Anderson and Bideganeta (1985). This fault cuts the sediments between the point of discharge of the spring and the intrusive contact, and could possibly be controlling the circulation of thermal water.

**Geochemistry.** Clarendon Hot Springs thermal waters are chemically similar to the other thermal waters in the Wood River area, particularly in high (16-19 mg/l) fluorine concentrations. Whole rock analyses and biotite mineral grains separated from the intrusion have high concentrations of fluorine (Table 2), suggesting water-rock interaction between the thermal fluids and the intrusion. Geochemical thermometers give calculated temperatures >73°C (163°F) at this site.

**Turn around; drive east toward State Highway 75.**

11.8 Return to highway; turn north on State Highway 75 toward Ketchum. Reset odometers.
Deer Creek Road to Guyer Hot Springs

0.0 Intersection of Deer Creek Road and State Highway 75.
3.0 Big Wood River bridge (south of East Fork).
3.4 East Fork, Wood River.
7.0 Big Wood River.
7.6 Elkhorn Road on the right.
8.8 Trail Creek.
9.1 Traffic light in center of Ketchum (Sun Valley Road and Main Street).
9.3 Warm Springs Road; turn left down small hill. Stay on Warm Springs Road through industrial area.
10.0 Big Wood River.
12.0 Stop 5: Guyer Hot Springs. Park in gravel pit on north side of highway.

Introduction. Guyer Hot Springs is on the west side of Warm Springs Creek (Fig. 8), and we may not be able to get to the springs if the creek is high. However, we will be able to see the springs from the east side of the creek and we will examine the water collection and distribution system for the heating district. Guyer Hot Springs has recently been studied by Blackett (1981) and Anderson and Bideganeta (1985). Temperatures at the main spring range from 62 to 70°C (144 to 158°F). Anderson and Bideganeta (1985) report a greater temperature range of 55 to 70°C (131 to 158°F) and a discharge of 3780 liters/minute (1000 gpm). However, we obtained a measurement of 1710 liters/minute (453 gpm) from the system's collection pipe.

The hot springs have been utilized for space heating of homes and businesses since the 1930s. At the spring vents, the flow of hot water is diverted to a series of concrete basins for collection; it is then gravity fed into town and ultimately disposed into Warm Springs Creek. Much of the spring flow is currently lost through leaks in the collection and pipe system. Additional development of the hot springs for space heating and recreational use has recently been proposed.

Lloyd Hot Springs, east of Guyer Hot Springs on the south side of Warm Springs Creek, is used to heat a private residence. Grayhawk Hot Springs is on the north side of the creek. At one time Grayhawk was used for bathing, but the bath house has been destroyed. There are at least three warm seeps along the north side of Warm Springs Creek that are exposed on the bank near the surface of the creek. Temperatures of 25 to 48.5°C (77 to 119°F) were measured at these seeps.

Many other thermal wells are present in the Guyer Hot Springs area. Foundation test holes and three larger diameter wells were drilled east of Guyer Hot Springs on both sides of Warm Springs Creek. According to the driller's logs, two of the larger diameter wells were cold and the third well had a strong sulfur smell when drilled. The third well originally discharged but now does not flow. A well adjacent to the Warm Springs ski lift at the International Village Condominiums 0.8 km (0.5 mi) downstream from Guyer Hot Springs has 28°C (83°F) water at 21 m (70 ft). Just west of Guyer Hot Springs, also on the west side of the creek, a 160 m (525 ft) well drilled through limestone has a bottom hole temperature of 59°C (138°F) (Anderson and Bideganeta, 1985). This is the only geothermal well currently being used.

Geologic setting. Rocks exposed in the study area are units of the Paleozoic Wood River Formation. The area is structurally complex. Both Blackett (1981) and Anderson and Bideganeta (1985) have mapped an anticline west of the hot springs, but Blackett (1981) locates the axis of the anticline further west. They both show an overturned fold in limestone on the north side of Warm Springs Road; similar overturned rocks may also exist west of the springs. They both map an extensive series of faults, and our preliminary reconnaissance of the area suggests that the structural complexity may be greater than is shown on either of their maps. No intrusive rocks are known near Guyer Hot Springs, and reports of "granite" in well logs have not been confirmed.

Hydrothermal systems. The hot springs at Guyer issue through a series of joints and fractures. Anderson and Bideganeta (1985) conclude that north-south trending faults control the path of the thermal fluids. Blackett (1981) suggests that N.50-60°W fractures at the thermal springs control the flow. He notes that the northwest trend is in agreement with the N.40-58°W. regional trend reported by Umpleby and others (1930).

Calcite deposition is present in the area of the hot springs and on the north side of Warm Springs Road. The northern calcite is topographically higher and is related to earlier hot springs activity. A ripple in Warm Springs Creek, near Lloyd Hot Springs, is caused by cemented alluvium due to mineralization of creek bottom sediments by the thermal waters. Comments in the driller's log of the 160 m (525 ft) well mentioned above include calcite-filled fractures at 37 m (120 ft), which might have been flowpaths of the thermal water. The calcite on the hillside and in the well could also be from hydrothermal fluid movement related to the intrusion of the Idaho batholith. The existence of adjacent thermal seeps strongly suggests that the cementing of river gravels is a current hydrothermal phenomenon.

Six foundation test holes and two 25 cm (10 in.) diameter wells were drilled on the property east of Guyer Hot Springs. All well locations were surveyed, and water
Figure 8. Location map of Guyer Hot Springs, showing hydrothermal springs (large solid dots), seeps (small solid dots), wells (open dots) and foundation test holes (small open dots).
levels and temperature profiles were measured. Temperature profiles from these wells are plotted in Figure 9. Most wells, with the exception of D4, show increasing temperatures with depth. D4 is only 30 m (100 ft) from the creek and is probably cooled by the cold creek waters. The temperature contour map (Fig. 10) and temperature surface map (Fig. 11) illustrate the extensive nature of the hydrothermal system.

Geochemistry. The water chemistry of Guyer Hot Springs is similar to that of the other thermal springs in the Wood River area. Even though the surface rock is limestone, the water chemistry has the typical signature
of batholith rocks. Chemical geothermometers in the Guyer Hot springs area suggest that waters at depth may be slightly warmer than surface discharges. High levels of fluorine have recently been noted in several domestic wells in the Warm Springs Road area. Thermal waters in the Guyer Hot Springs area typically have 15 to 18 ppm fluorine; cold waters, based on our analysis of Warm Springs Creek (Table 1), have <1 ppm fluorine. The high fluorine problem is being studied by the Idaho Department of Health and Welfare, Division of Environment.

The areal extent of the subsurface discharge of thermal waters at the Guyer Hot Springs area has not yet been determined, but it extends at least 0.8 km (0.5 mi) downstream from the major hot springs and may include a much larger area than depicted by Blackett (1981). As Blackett (1981) notes, the relationship of subsurface discharge and other thermal springs, seeps, and wells to Guyer Hot Springs remains enigmatic. From the existing data, it appears that no wells have been drilled into the major structural features supplying Guyer Hot Springs.

The field trip ends at Guyer Hot Springs. Turn around and head back along Warm Springs Road to Sun Valley.

OTHER HYDROTHERMAL SPRINGS AND WELLS

Four other Wood River area geothermal sites are the Picabo warm water well, Warfield (Frenchman's Bend) Hot Springs, Easley Hot Springs, and Russian John Hot Springs. These sites are depicted in Figure 1 but will not be visited on this trip.

Two low-temperature (32°C and 36°C; 90 and 97°F) and low-flow (approximately 60 liters/minute; 15 gpm) thermal wells exist approximately 3.2 km (2 mi) north of Picabo. They are located in a pasture, about 30 m (100 ft) from the range front. Bedrock in this area is Eocene Challis Volcanics overlain by Pliocene tuffs and basalts (Rember and Bennett, 1979). Schmidt (1961) correlated the tuffs and basalts with volcanic deposits at Magic Reservoir. Geothermal waters at Picabo plot isotopically adjacent to the meteoric water line (Fig. 2). The warm wells at Picabo are the only warm waters currently known from east of the Wood River in the study area. These wells are not located adjacent to outcrops or subcrops of intrusive rocks.

Warfield (Frenchman's Bend) Hot Springs, located approximately 13 km (8 mi) west of the center of Ketchum, consists of a bath house, a series of seeps and springs that flow into Warm Springs Creek, and at least two cisterns with thermal water (Fig. 12). Water temperatures of these features range from 41°C to 58°C (106°F to 136°F). Unpublished geologic mapping by Jon Zeisloft, M. C. Adams and R. B. Blackett (University of Utah Research Institute) and the authors shows that these springs flow from a highly faulted and fractured zone at the edge of the Rooks Creek Stock (Johnson and others, this volume), and from the intruded Paleozoic sedimentary rocks. Thermal waters at Warfield have 13-16.5 mg/l fluorine, and chemical geothermometer temperatures are >75°C (167°F).

Easley Hot Spring flows from Challis Volcanics at the edge of the Wood River valley, approximately 23.3 km (14.5 mi) north of the center of Ketchum. The 38°C (100°F) waters are currently used in a private swimming pool. Although Easley Hot Spring flows from fractured volcanic rocks (Rember and Bennett, 1979), it has 16 mg/l fluorine, which is compatible with the hypothesis that the thermal waters have equilibrated with intrusive rocks. Such rocks have been mapped (Rember and Bennett, 1979; Tschanz and others, 1986) approximately 3 km (1.9 mi) north of the springs, on the north side of the valley. Chemical geothermometry of Easley Hot Spring water suggests that the fluids may be hotter than 55°C (130°F) at depth.

Russian John Hot Spring is the northernmost thermal site in the Wood River area, located nearly 29 km (18 mi) north of the center of Ketchum. It consists of one low-temperature (33°C; 91°F), low-flow spring that has been converted into a soaking pool, and several areas of low-temperature thermal seeps, including a second, lower-temperature pool north of the highway. The seeps are marked by a large area of snow melt on both sides of the highway that generally follows the course of Dooley Creek. The snowmelt area extends as much as 30 m (100 ft) on either side of the creek. In April 1986, ground and seep temperatures in the snowmelt area were as high as 28°C (82°F), whereas creek runoff was near 0°C (32°F) but not frozen. Russian John Hot Spring may have a much larger thermal area, and may be a much larger thermal system than is currently recognized. Thermal water at Russian
John Hot Spring issues from alluvial deposits. Eocene volcanic rocks form the hills immediately adjacent to the site (Rember and Bennett, 1979; Tschanz and others, 1986). Tschanz and others (1986) depict high-angle faults cutting the volcanic rocks in the area of the thermal waters, and Tertiary granite at the Boulder Mountains stock 2 km (1.2 mi) north of the hot springs. Thermal waters at Russian John have 13.5 to 16 mg/l fluorine, suggesting equilibration with intrusive rocks. Chemical geothermometer temperatures at Russian John Hot Spring are above 60°C (140°F).

We would like to acknowledge the contributions of Steve Benham, Larry Dee, Robert DeTar, Leslie Foley, Loren Holmes, Allen Merritt, Bruce Sibbett, Deborah Struhsacker and Jon Zeisloft, who directly or indirectly aided in the study or preparation of this report. Reviews by Bill Young, Gerry Lindholm, Dick Whitehead, Paul Castelin, and the editors of this volume greatly improved the text. However, we are responsible for the format and conclusions of this report.

REFERENCES


Truesdell, A. H., and Hulston, J. R., 1980, Isotopic evidence on environments of geothermal systems, in...


Chapter Two
Paleozoic Stratigraphy

The west face of the Lost River Range and the Big Lost River Valley north of Mackay. Lower and middle Paleozoic strata form the mountains. Alluvial fan from Upper Cedar Creek is at the center of the frame. Photograph by B. R. Burton.
Early Paleozoic Continental Margin Development, Central Idaho

Mark D. McFadden ¹
Elizabeth A. Measures ¹
Peter E. Isaacson ¹

ABSTRACT

Ordovician, Silurian, and Devonian stratigraphic units in central Idaho (Custer County) demonstrate a variety of depositional settings, and preliminary synthesis of continental margin behavior is now possible. Late Ordovician carbonates originally named the Saturday Mountain Formation were deposited on a homoclinal to distally steepened ramp. A deepening-upward cycle, from inner to outer ramp, occurs in western sections, and a normal aggradational (shoaling) cycle, from middle to inner ramp, occurs in eastern sections. Accompanied by nonorogenic syenite plutons in the area of the Lemhi Arch, this margin may represent extension and subsequent rapid subsidence.

Silurian Roberts Mountains Formation lithofacies indicate slope deposition of mixed carbonate and siliciclastic debris flows within oxygen-minimum, with shallow platform carbonates to the east. Vertical succession within the Silurian sections suggests shallowing upward to storm wave base in well-oxygenated platform environments.

Early Devonian units are largely unstudied, although they are apparently shallow-water carbonates with few environmentally diagnostic features. Following the Stringocephalus (Givetian) part of the Jefferson Formation, however, a deepening event occurred whereby a Frasnian “Alberta-type” buildup developed on slope units. The buildup lacks talus and “wall” facies. Following subaerial exposure and brief eustatic (?) onlap, the Fammenian witnessed emergence of an Antler-related highland, from which quartz-rich clastics flooded the Jefferson Formation. Therefore, a possible Ordovician extensional regime was followed by Silurian and Early Devonian quiescence and thence Late Devonian compressional tectonics.

INTRODUCTION

Excellent exposures of lower Paleozoic carbonate rocks near Challis provide unique access for study of platform margin evolution during the early Paleozoic. Lithofacies and biofacies changes within Upper Ordovician through Upper Devonian strata document the evolution of the margin through time in response to both internal and external controls. Three exceptionally well-represented units will be studied at two field trip stops: the Saturday Mountain Formation equivalent (Ordovician), the Roberts Mountains Formation (Silurian), and the Jefferson Formation (Devonian) (Fig. 1).

The Upper Ordovician Saturday Mountain Formation equivalent is composed of approximately 250 m (825 ft) of eastern assemblage rocks deposited on an overall
deepening-upward carbonate ramp. It contains interbedded dolomudstones and skeletal dolowackestones with a significant replacement of shallow water organisms by deeper water populations. The Silurian Roberts Mountains Formation consists of over 700 m (2,300 ft) of interbedded impure carbonate, siliciclastic, and bioclastic transitional assemblage rocks representing shallowing-upward deposition along a margin with extensive paleorelief. The Middle and Upper Devonian Jefferson Formation consists of approximately 250 m (800 ft) of deep-ramp laminites, succeeded by a coral-dominated buildup 40 m thick (130 ft), in turn overlain by approximately 230 m (750 ft) of tidal flat sandy carbonates with evidence of incipient Antler uplift to the west.

Location and Access

Two field trip stops demonstrate the lower Paleozoic carbonate sequence (Fig. 1). The Saturday Mountain and Roberts Mountains Formations will be examined at Bradshaw Basin, approximately 15 miles south of Challis (Fig. 2). Best access is to travel southeast from Challis 19 miles on U. S. Highway 93 to the intersection with Spar Canyon Road. Turn west on Spar Canyon Road, and continue 7.2 miles to the intersection with the Bradshaw Basin Road (Fig. 3). Travel 7.0 miles on the Bradshaw Basin Road, keeping left at forks on the more travelled road, until reaching the cattle guard at the summit. Travel 0.6 miles straight ahead to the intersection with Little Bradshaw Basin Road; continue straight for 0.7 mile to a fork in the road. The west fork of the road extends approximately 1 mile to a corral south of the upper portion of the Roberts Mountains Formation section. The north fork of the road is very rough and extends into Bradshaw Gulch near the base of the Saturday Mountain Formation. The measured sections of the Saturday Mountain and Roberts Mountains Formations are located in the W 1/2, Sec. 4, T. 11 N., R. 19 E., Custer County (Lone Pine Peak 7.5-minute quadrangle).

The Jefferson Formation and associated coral-dominated buildup will be studied on a western “spur” of the Lost River Range (northeast of Lone Pine Peak), at the northern end of Grandview Canyon. Travel approximately 12 miles southeast of Challis along U. S. Highway 93 to the north end of Grandview Canyon (Fig. 4). The buildup is best visible as a distinct, massive cliff about halfway up the hillside exposure on the northwest side of Grandview Canyon (Fig. 5). The measured
section is located in the NE 1/4, Section 22, T. 12 N., R. 20 E., Custer County (Antelope Flat 7.5-minute quadrangle).

Time will not permit visits to other lower Paleozoic units including the Laketown Dolomite (Silurian), the Beartooth Butte(?) Formation (Devonian), and Carey Dolomite (Devonian). These rocks lie stratigraphically between the Roberts Mountains Formation and the Jefferson Formation (Fig. 1). Approximately 700 m (2300 ft) of siliciclastic and dolomitic rocks are present within this stratigraphic interval.

LITHOSTRATIGRAPHY OF THE SATURDAY MOUNTAIN FORMATION EQUIVALENT, BRADSHAW BASIN

Previous Work

Upper Ordovician rocks in central Idaho were given the name Saturday Mountain Formation by Ross (1934), with the type area established in lower Squaw Creek, west of Clayton. The formation was originally defined as predominantly shaly dolomite in which the argillaceous beds occur as "banded, carbonaceous, thinly bedded slabs" (Ross, 1934). The Saturday Mountain Formation was later described as interbedded carbonaceous and calcareous shales, shaly limestones (dolomites), banded limestones (dolomites), and massive limestones (dolomites) in the Redbird mine at Squaw Creek (Ross, 1937). Ross (1934, 1937) recognized the unit as time equivalent and lithologically similar to unnamed carbonates in the Lost River and Lemhi Ranges, the Fish Haven Dolomite of Utah and southeastern Idaho, and part of the Bighorn Dolomite of Wyoming; Ross (1947, 1961) eventually extended use of the term "Saturday Mountain Formation" to the Lost River and Lemhi Ranges.

Hays and others (1978) were the first to recognize and map Upper Ordovician carbonates in Bradshaw Basin.
They followed the terminology of Ross (1934, 1937) in the adjacent Germer Basin, called the unit the Saturday Mountain Formation, and later described the brachiopod and conodont biostratigraphy and the lithostratigraphy of the Upper Ordovician through Lower Devonian units of Bradshaw Basin and Spar Canyon (Hays and others, 1980). The carbonate lithofacies of the Upper Ordovician in central Idaho (including Bradshaw Basin) were described by Measures (1986, 1987), and detailed paleogeographic reconstructions were proposed.

The extension of the name Saturday Mountain Formation from its type area at Squaw Creek, eastward to nonargillaceous Upper Ordovician carbonates in Germer Basin, Bradshaw Basin, and the Lost River and Lemhi Ranges is controversial. Sloss (1954) suggested applying the name “Fish Haven Dolomite” to rocks in the Lost River and Lemhi Ranges because of the time equivalence of the sequences in central and southern Idaho and because both sequences were composed of relatively pure carbonates. Churkin (1962) suggested that the use of Saturday Mountain Formation nomenclature be restricted solely to its type area, and he assigned the clean carbonates of Germer Basin and the Lost River and Lemhi Ranges to the Fish Haven Dolomite. He distinguished the transitional assemblage rocks (Saturday Mountain Formation) from the eastern assemblage rocks (Fish Haven Dolomite). However, the Upper Ordovician carbonate lithologies of central Idaho (Germer Basin/Bradshaw Basin) have not been compared with the specific lithologies of the type section of the Fish Haven Dolomite. Therefore, it is suggested that the Upper Ordovician section in Bradshaw Basin (and Germer Basin) temporarily be referred to as either Saturday Mountain Formation equivalent or Fish Haven Dolomite equivalent. In this study the term “Saturday Mountain Formation equivalent” is used.

Nature of the Section

The Saturday Mountain Formation equivalent in Bradshaw Basin is preserved as a composite of three separate stratigraphic sections (Fig. 6) because of structural complications. The basal part is exposed in two adjacent sections (6 m (20 ft) and 45 m (150 ft)) (Fig. 6, sections A and B), separated by a high-angle fault. Displacement and missing section cannot be calculated since no lithologic marker unit is common to both sections. The amount of missing section is assumed to be small since abundant quartz sand is present in the first section and in the base of the second section. The remaining part of the stratigraphic column (Fig. 6, section C) is offset from the other two sections by approximately 0.4 km (0.25 mile) by extensive faulting and deformation at the top of section B. Similar lithological sequences occur in the uppermost part of B and lowermost part of C, but it is uncertain if any section is missing or repeated. The three sections do not follow exactly the line of section of Hays and others (1980) who reported a continuous section along the ridge.

**Lithofacies**

The Upper Ordovician Saturday Mountain Formation equivalent is best described in terms of five lithofacies named for the most dominant lithology. In approximately ascending order, these are: (1) dolomitic quartz arenite, (2) quartzitic dolomudstone and dolomudstone, (3) bioclastic dolowacke/mudstone, (4) intraclastic dolopackstone and laminated dolomudstone, and (5) “flaser” bedded dolomudstone.

**Dolomitic Quartz Arenite Lithofacies. This lithofacies is restricted to and totally comprises section A (Fig. 6). The section is approximately 6 m (20 ft) thick and forms a small outcrop of low relief. The best samples were collected from float.**

The base of the section is composed of a brown weathering quartz arenite (Fig. 7). The grains are medium to coarse sand in size, rounded to subrounded, clear to light gray in color, and cemented by dolomite. The percentage of dolomite in the rock increases upward in the section and about half of the rock is composed of gray dolomudstone with scattered quartz sand; the remaining half contains resistant stringers and laminae of brown quartz arenite.

**Quartzitic Dolomudstone and Dolomudstone Lithofacies. Section B, the basal part of section C, and two intervals at the top of section C (Fig. 6) are composed of this lithofacies. These units are predominantly dolomudstones and are grouped as one lithofacies. However, there are marked differences in quartz sand content, sedimentary structures, and colors of weathered surfaces.**

The lithofacies is generally thin bedded (<0.3 m to 0.5 m; 1 ft to 1.5 ft). It is tan to brownish gray with grayish red interbeds on weathered surfaces and medium gray and red on fresh surfaces. Silicification of beds occurs along faults. Anastomosing stringers of dolomite and silica occur throughout the lithofacies at various angles to bedding.

The quartzitic dolomudstone lacks visible sedimentary structures and allochems. Quartz sand grains are common in the basal units of section B (Fig. 6) and vary from 10 to 40 percent of the rock. Variation in the percentage of quartz may be gradational or may be abrupt, indicating possible hardground surfaces (if relief along the contact is visible). The size of the quartz grains decreases upward (from medium to fine sand) as does the percentage of quartz.

Faint red and pink laminations and poor partings occur in the dolomudstones; the laminations are concentrations of quartz silt, and the partings result from a high concentration of clay-size particles. Laminated siliceous, grayish red or pink, silty dolomudstone interbeds occur
between quartzitic dolomudstone beds. Fossiliferous dolowackestones and dolopackstones occur as minor, thin interbeds with disarticulated brachiopods, pelmatozoan debris, and trilobite fragments (Fig. 8). These beds may be preferentially silicified, with sharp lower contacts. The upper contacts may be sharp or gradational, and they vary from planar to undulose. The dolowackestones have a large content of insoluble quartz silt and finer siliciclastics.

Uncommon chert nodules are first seen 18 m (60 ft) above the base of section B (Fig. 6). They occur scattered throughout the upper 27 m (90 ft) of the section. The dolomudstones of this part of the section are increasingly sheared and deformed and contain no quartz sand or allochems. The base of section C contains similar dolomudstones which have not been tectonically deformed.

The uppermost dolomudstones in section C vary from
light olive-gray (weathered) to medium olive-gray (fresh) to medium to dark gray (weathered and fresh) and may be massive, mottled by bioturbation, or laminated. Laminated units are more abundant upward in the section (Fig. 9). Bioclastic dolowackestones are interbedded in these units but are not as abundant as in the lower portion of the lithofacies.

Bioclastic Dolowacke/Mudstone Lithofacies. The bioclastic dolowacke/mudstones occur exclusively in section C. They are more important in the base of this section where they are interbedded with units of the quartzitic dolomudstone and dolomudstone lithofacies. More commonly they form a very thick, homogeneous, and massive unit approximately in the middle of the section.

This lithofacies is very distinctive and easily recognized in the field. It weathers to a medium or dark gray, and the bioclasts are present as white, fine to coarse sand-sized specks scattered throughout the rock (Fig. 10). The poorly preserved bioclasts appear to be predominantly pelmatozoan debris, and the percentage of bioclastic debris is variable throughout the lithofacies. This lithofacies is predominantly thick bedded, 0.8 to 1 m (2.5 to 3 ft) thick. Anastomosing dolomite stringers occur at the base, oriented at various angles to the bedding, and are seen as lighter colored, more resistant wisps. Light yellow to orange crystalline masses visible on outcrop or broken surfaces are a diagenetic replacement of carbonate and are probably saddle dolomite.

Massive and mottled units are predominant throughout Section C; laminated units are uncommon. The dark coloration is the result of abundant organic material. Digestion of this rock in hydrochloric acid produces a black, oily residue.

Intraclastic Dolopackstone and Laminated Dolomudstone Lithofacies. This lithofacies occurs only near the center of section C and is the last unit in the Saturday Mountain Formation equivalent below the Roberts Mountains Formation. It is not extensive but is lithologically distinct with several useful marker beds. Overall it is dark to medium gray on both weathered and fresh surfaces, with laminations a few millimeters thick visible as color variations (Fig. 11). The units may be thin to thickly bedded (0.3 to 1 m (1 to 3 ft) thick) and generally lack bioclasts. The distinct intraclastic dolopackstone occurs near the middle of section C and is an excellent marker unit. The intraclasts are composed of laminated dolomudstones, vary in length from 2 to 30 cm (1 to 12 in), and are tabular in cross-section.

The dolopackstone is underlain by laminated dolomudstones which are lithologically identical to the intraclasts. The uppermost unit in section C is also a laminated dolomudstone which contains an interval of laminated, fossiliferous dolowackestone that is not present elsewhere in the lithofacies. This bioclastic unit is an excellent marker for the top of the section and occurs approximately 1.5 m (5 ft) below the top of the Saturday Mountain Formation equivalent. The bioclastic layer contains poorly silicified, disarticulated brachiopods. Most valves are small (up to 1 cm (0.5 in) in diameter), thin-shelled, and not in current-stable positions.
Figure 9. Polished slab photograph of laminations in
dolomudstone, quartzitic dolomudstone and dolomud-
stone lithofacies. Scale in centimeters.

Hays and others (1980) place the boundary between
the Saturday Mountain Formation equivalent and the
Roberts Mountains Formation 5 m (16 ft) above the
brachiopod layer. Their co-author, R. J. Ross, Jr.,
believes the contact between the formations should be
placed 0.3 m (1 ft) below the brachiopod layer where it
would be closer to the Ordovician-Silurian boundary. We
believe that the contact may be placed either 0.6 m (2 ft)
below the brachiopod layer where chert nodules occur in
a distinct layer (in accordance with the lower boundary of
the Roberts Mountains Formation as designated else-
where) or 1 m (3 ft) above the brachiopod layer where the
lithology changes. The lithologic change consists of the
occurrence of white to pink, resistant, dolomitic and
siliceous laminae approximately 1 mm thick, oriented
parallel to thicker color laminations spaced 0.5 cm to 1
cm (0.2 to 0.4 in) apart.

"Flaser" Bedded Dolomudstone Lithofacies.
This distinct lithofacies occurs only near the top of
section C and can be used as a stratigraphic marker. It
weathers medium to light gray with darker wisps or
"flaser" type structures (Fig. 12). The wisps are finer-
grained than the surrounding matrix, and are parallel to
bedding. The units are thick bedded (0.6 to 1 m; 2 to 3
ft) and are generally faintly laminated in addition to the
wisps. They contain minor poorly preserved pelmatozoan
bioclasts similar to those in the bioclastic dolowacke/
mudstone.

Discussion

Dolomitic Quartz Arenite Lithofacies

This lithofacies appears to be gradational with the
underlying Kinnikinic Quartzite. Hays and others (1980)
infer that the contact between the Saturday Mountain

Figure 10. Polished slab photograph of mottled and
bioclast-rich sample, bioclastic dolowacke/mudstone
lithofacies. Scale in centimeters.

Figure 11. Hand sample photograph of laminated
dolomudstone, intraclastic dolopackstone and laminated
dolomudstone lithofacies. Scale in centimeters.
Formation equivalent and the Kinnikinic Quartzite in Bradshaw Basin is tens of meters below the base of the exposed strata. Interpretation of the environment of deposition is difficult because of the poor exposure and the lack of sedimentary structures.

Oaks and others (1977) and James and Oaks (1977) state that the upper part of the Kinnikinic Quartzite was deposited on a shallow marine shelf. Furthermore, they state that the Kinnikinic Quartzite-Saturday Mountain Formation contact is disconformable, because the uppermost quartzite shows features indicative of subaerial exposure. They suggest that a regression occurred before deposition of the Saturday Mountain Formation and that during the ensuing late Middle Ordovician transgression some of the siliciclastic sediment was incorporated into the carbonate. This reworking of the underlying sands would have occurred in shallow marine conditions, but the exact nature of the environment cannot be determined.

Recent work by Mount (1983, 1984, 1985) and Walker and others (1983) has shown that siliciclastic and carbonate sediment can be time equivalent and can commingle. It is possible that the basal Saturday Mountain Formation equivalent could represent mixing of the two facies. However, this implies time equivalence of the Saturday Mountain Formation and Kinnikinic Quartzite, and this has not been demonstrated biostratigraphically.

Quartzitic Dolomudstone and Dolomudstone Lithofacies

This lithofacies contains many features which indicate shallow subtidal, normal marine conditions. Interpretation of conditions uses the terminology and principles of Wilson (1975). Minor differences in lithologies between the base (section B) and the top (section C) of the sequence point to variations in the depositional environment. The most obvious feature of this lithofacies is the decrease upward in sand content, reflecting a change in source and not a change in environment. The transgression suggested by Oaks and others (1977) and James and Oaks (1977) would eventually cover the siliciclastic source. Deposition of quartz sand-dominated units ceased abruptly during the higher seastand, however, and there was sporadic input of quartz sand during deposition of the overlying dolomudstone, indicating intermittent exposure of the siliciclastic source. After this interval the siliciclastic source was not active. The source of the Middle and Late Ordovician siliciclastics (Kinnikinic Quartzite and basal Saturday Mountain Formation equivalent) has been attributed to erosion of the Proterozoic Swauger Formation and Cambrian Wilbert and Summerhouse Formations exposed on the Lemhi Arch (Oaks and others, 1977; James and Oaks, 1977; Ruppel, 1986). It has been suggested that the Lemhi Arch was located in central Idaho, approximately in the position of the present Lost River, Lemhi, and Beaverhead Ranges (Ruppel, 1986).

The brachiopods, pelmatozoans, trilobites and corals interbedded throughout the quartzite dolomudstone and dolomudstone lithofacies indicate normal marine, open circulation, subtidal inner and middle platform conditions. The term platform is used in the sense of Read (1982) to indicate either a carbonate ramp or shelf. The corals are not in growth position, and the other fossils are disarticulated, which indicates some reworking of the bioclasts although they are not abraded. The more fossiliferous dolowackestones and dolopackstones indicate wave or current concentrations of allochems. These layers may represent storm deposits as described by Kriesa (1981).

The bioturbation indicates an oxygenated substrate on a shallow inner platform. This texture is most abundant in the lower part of the sequence. Laminated textures dominate this lithofacies in the upper part of the sequence. The lamination does not appear to be of algal origin; no stromatolite growth forms occur in the laminated units. Furthermore, the lamination is not crinkly, pustulose, or undulatory and no associated fenestral textures or mudcracks occur. The laminated units are the result of mechanical deposition of carbonate mud and silt in a low-energy environment where infaunal organisms were excluded. Such restriction would occur within oxygen minimum on the middle platform or in protected areas of the inner platform.

Subtidal conditions occurred throughout deposition of this lithofacies. Overall, open inner and middle platform conditions are indicated. Shallower subtidal (inner platform) environments are most common at the base of the sequence, and deeper subtidal (middle platform) are most common in the upper part of the sequence.

Bioclastic Dolowacke/Mudstone Lithofacies

This lithofacies represents deposition in areas and
under conditions similar to those of the quartzitic dolomudstone and dolomudstone lithofacies. Normal marine, open circulation, subtidal inner to middle platform conditions are indicated by the bioclasts and bioturbation. The greater abundance of skeletal grains suggests a more favorable, stable environment overall such as on the middle platform. The laminated units of this lithofacies represent conditions similar to those responsible for the deposition of the laminated quartzitic dolomudstones: open, middle platform within an oxygen minimum zone. The darker color of this lithofacies is a result of diagenesis and does not imply an organic-rich substrate at the time of deposition.

Intraclastic Dolopackstone and Laminated Dolomudstone Lithofacies

Similar to the other laminated units previously discussed, this laminated lithofacies and the laminated intraclasts are not cryptalgal in origin. They represent mechanically deposited carbonate muds and silts undisturbed by burrowing organisms and therefore deposited within oxygen minimum on the middle to outer platform. Somewhat deeper conditions are indicated by the absence of fossils and overall laminated texture. The intraclastic interval represents breakup of a portion of the laminated facies and movement of the resulting allochems. This is the result of slumping of middle to outer platform units which would occur most commonly on the deeper outer platform.

Cook (1983) states that in the Paleozoic there was no abundant source of pelagic carbonate; therefore, thick deposits of deeper water carbonates must have been derived from the middle and inner platform. The carbonate silt and mud of this lithofacies, and of the other laminated outer platform deposits, were moved offshore from the areas of high carbonate production. Hence, the brachiopod dolowackestone in this lithofacies was apparently shoreward derived. The bioclasts within the units of this lithofacies result in a coarse-grained allochem texture in an otherwise fine-grained lithology. The bioclasts were transported but not destroyed by an unspecified mechanism of downslope movement.

"Flaser" Bedded Dolomudstone Lithofacies

The predominance of pelmatozoan debris indicates normal marine, open platform, subtidal conditions, similar to the other lithofacies with abundant pelmatozoan fragments. However, the overall lack of allochems in the unit and laminated texture implies deep subtidal, outer platform conditions below the depth of abundant production of bioclasts. Flaser bedding has been reported in carbonate tidal flats associated with silicilastic (Wilson, 1975; Flugel, 1982) and is also reported in deeper water platform and slope deposits (Wilson, 1975). The flaserlike textures in this lithofacies indicate deeper water, outer platform conditions. The exact mechanism responsible for their formation is unknown.

Paleogeographic Interpretation

The wide facies patterns present in the Upper Ordovician carbonates in central Idaho and the lack of any facies associated with a major change in slope indicates that the carbonates were deposited on a ramp (Measures, 1987). We suggest that the Saturday Mountain Formation equivalent in Bradshaw Basin indicates that the ramp evolved from a homoclinal ramp to a distally steepened ramp and back to a more homoclinal attitude, using the terminology of Read (1985). The vertical sequence of carbonates and their faunal changeover indicate a deepening-upward trend in depositional environments and incipient drowning of the ramp (Schlager, 1981). The basal units indicate shallow inner to middle ramp deposition. The laminated units in the base indicate incursions of conditions within the oxygen minimum (deeper water). The upper units indicate mostly outer ramp deposition with some interbedded middle ramp deposits. An overall rise in sea level with changing rates of rise would create this oscillation between deep water and shallow water conditions.

Schlager (1981) reports that rates of carbonate production can exceed the most rapid rate of sea level rise which is caused by glacio-eustatic processes. Therefore, the laminated lithologies represent an initial phase of carbonate production after a subsidence event. Deepening-upward carbonate platforms may be produced when carbonate production lags behind the sea level rise, or the deeper water causes a decrease in the rate of carbonate production (Schlager, 1981).

GENERAL LITHOSTRATIGRAPHY OF THE ROBERTS MOUNTAINS FORMATION, BRADSHAW BASIN

Previous Work

Silurian strata in the region south and east of Challis were originally designated as Laketown Dolomite by Ross (1934), whose initial reconnaissance mapping included the extension of both Laketown and Jefferson Dolomite nomenclature into the area from southeastern Idaho (Ross, 1934, 1937). Although the stratigraphic boundaries established for Jefferson Dolomite lithologies have proven to be accurate, Ross included within the Laketown at Bradshaw Basin a sequence of dolomitic carbonates distinctly different from typical Laketown exposures in southeastern Idaho and northeastern Utah. A thick package (over 700 m, 2300 ft) of interbedded silty and sandy laminated dolomites, bioclastic dolomites, and quartzites was recognized during more recent mapping by Hays and others (1978) as correlative with outcrops of the Roberts Mountains Formation in Nevada and Utah. Further study of measured sections and fossil collections
firmly established the lithostratigraphic and biostratigraphic similarities between the "platy" dolomites at Bradshaw Basin and typical exposures of the Roberts Mountains Formation elsewhere (Hays and others, 1980). Paleontological data indicate a Wenlock to Ludlow age for the Bradshaw Basin section (Hays and others, 1978; 1980). Recent work indicates a gradational transition from the slope and platform margin facies present in Bradshaw Basin to dissimilar platform carbonate facies present in the Lost River Range to the east (Ingwell, 1980; Isaacson and McFadden, 1984; McFadden, 1985).

Nature of the Section

The stratigraphic section of the Roberts Mountains Formation at Bradshaw Basin is well-exposed and of variable topographic relief. The lower part of the section is rather steep and is characterized by extensive "platy" float interspersed with resistant outcrops of various lithologies. The upper portion of the section contains more gentle slopes, with covered intervals more common than lower in the section. Numerous small-scale folds and high-angle faults complicate the stratigraphy, although marker units can be used to unravel minor structural problems. Extensive jasperoid alteration along faults locally obscures bedding and lithology, but beds can be traced laterally to avoid areas masked by pervasive silicification.

Lithofacies

The Roberts Mountains Formation in Bradshaw Basin is a complex sequence of alternating lithologies, with individual lithofacies generally present in units only a few meters thick. The complexly interbedded nature of the section suggests the description of individual lithofacies as the best method of characterization. Four main lithofacies comprise the section: (1) a laminated dolomudstone facies, (2) a massive dolomudstone/dolowackestone facies, (3) a bioclastic and intraclastic facies, and (4) a massive dolomitic quartz arenite facies. In general, laminated dolomudstones and massive dolomitic quartz arenites are more dominant lower in the section, while massive dolomudstones/dolowackestones are increasingly common in the upper part of the section. Bioclastic and intraclastic units are common throughout the sequence (Fig. 13).

Laminated Dolomudstone Facies. The lithology most characteristic of the Roberts Mountains Formation at Bradshaw Basin is the laminated dolomudstone facies. Laminations are visible on a millimeter scale (commonly 2-5 mm, 0.1-0.2 in), and are typically represented by alternating color banding or thin "stringers" of silt or fine sand (Fig. 14). The laminated dolomudstone varies in color from tan to dark grey, with local alteration resulting in pink to rust-colored outcrops. Bedding ranges from 0.2-2.5 m (0.6-8 ft) in thickness.

The content of terrigenous siliciclastic grains in the laminated dolomudstone facies is extremely variable within the section. Individual beds range from slightly silty or sandy dolomudstone to dolomitic fine- to medium-grained sandstone. Textural or color laminations are generally visible even in the more sandy units. Extensive talus is derived from the laminated lithofacies in the form of platy or flaggy silicic float (Fig. 15). The facies commonly weathers into distinct slabs between 0.5 and 2.0 cm (0.2 and 0.8 in) in average thickness due to bedding plane partings or cleavage at low angles to bedding (Fig. 16). Outcrops of the laminated lithofacies are variably resistant, with increasing sand content resulting in less resistant beds.

Massive Dolomudstone/ Dolowackestone Facies. Resistant outcrops of massive dolomudstone/dolowackestone are scattered throughout the Bradshaw Basin section, generally increasing in frequency higher in the section. Bedding is commonly
medium to thick (0.5-2.0 m, 1.6-6.5 ft), and colors range from medium to dark grey. Outcrops of massive dolomudstone/dolowackestone commonly exhibit extensive mottling visible as both textural and color variations (Fig. 17). Minor lenses of fine skeletal dolopackstone are locally abundant, with common fine-grained crinoid debris. Outcrops of this lithofacies are generally free of terrigenous sand, although minor mounts of silt and fine sand are present in some beds.

**Bioclastic and Intraclastic Facies.** Resistant outcrops of bioclastic and intraclastic rudstones to floatstones comprise a significant portion of the Roberts Mountains Formation at Bradshaw Basin. The skeletal-rich units are typically intercalated with beds of the laminated dolomudstone facies, although successive beds of the bioclastic and intraclastic facies are locally common. Bedding is generally medium to thick (1-2 m; 3-6.5 ft), and lower contacts are sharp and erosional (Fig. 18). Upper contacts commonly grade upward into either laminated or massive dolomudstones, and beds are typically lensoid, tending to pinch out laterally when traced along strike.

Bioclasts present in this lithofacies include rugose and tabulate coral fragments, and bryozoan, crinoid,
stromatoporoid and brachiopod debris ranging from sand to minor cobble sizes. Skeletal fragments are present in both graded and ungraded beds, with a general trend toward grain support near the lower contacts (rudstone) and more mud support higher within the beds (floatstone) (Fig. 19). Abrasion effects are clearly evident on most bioclasts, and extensive transport of shallow-water faunas into deeper waters is indicated by mixed assemblages of brachiopods including perireefal forms (*Pentamerifera*) with deeper faunas (*Dicaelosia*). Elongate intraclasts up to 5 cm (2 in) in length are common in this lithofacies. The intraclasts are commonly intermixed with bioclastic debris, although local concentrations of intraclasts are present. Internal fabric of the intraclasts ranges from laminations on a millimeter scale to structureless dolomudstone or dolowackestone (Fig. 20). Both the composition and lack of extensive transport evident in many intraclasts suggests a local derivation from previously deposited units by erosion.

Massive Dolomitic Quartz Arenite Facies. Massively bedded dolomitic quartz arenites are common within the Bradshaw Basin section. Bedding is medium to thick (0.5-2.0 m, 1.6-6.5 ft) and generally lacks internal fabric, although rare examples of trough and planar cross-stratification have been found. Grain sizes vary between fine to medium sand, and grains are generally well-sorted and well-rounded (Fig. 21). The massively bedded sands are brown in color and commonly laterally discontinuous, occurring in irregular lensoid bodies. Lower bedding contacts are generally sharp, while upper contacts are commonly gradational into overlying sandy massive dolomudstones/dolowackestones. Bioclasts and intraclasts are generally rare within the massive dolomitic quartz arenite facies.

Discussion

The complex interbedding and repetition of lithofacies throughout the Bradshaw Basin section makes subdivision of the section into successive depositional packages difficult. However, close inspection of the components and internal fabrics present within beds of individual lithofacies provides important information on depositional processes. Study of the relationships between adjacent beds of the major lithofacies demonstrates the distinct variability of the depositional events and high relief at the platform margin during Silurian time.

The fine, even laminations present within outcrops of the laminated dolomudstone facies suggest a lack of benthic fauna capable of bioturbation. Benthic fauna could be excluded by rapid sedimentation rates; however, poor oxygenation of bottom waters is cited as the most likely reason for the lack of benthic marine invertebrates in similar lithofacies of the Roberts Mountains Formation in Nevada (Matti and McKee, 1977). The high content of terrigenous sand within many rocks of the laminated dolomudstone lithofacies indicates that pelagic sedimentation is not an adequate mechanism for deposi-
tion of the facies. Laminations on a millimeter to centimeter scale within highly sandy mudstones and examples of graded bedding suggest density current deposition of many parts of the lithofacies. Gradations upward from laminated dolomudstones into massive dolomudstones and dolowackestones suggest sedimentation rate decrease or fluctuation in oxygenation of bottom waters to allow bioturbation.

Internal fabrics and faunal content of beds of the bioclastic and intraclastic facies provide indications of extensive transport of sediments on the platform margin. Grading and sorting of bioclasts generally increases upward within beds, and mud support of skeletal fragments also increases upward. Extensive abrasion of bioclasts and mixing of depth-zoned brachiopod assemblages indicates a significant transport distance or a slope of considerable paleorelief. The lensoid bedding common in outcrops of the bioclastic and intraclastic facies generally has sharp, erosional lower contacts and indicates channels cutting into underlying units. Intraclasts cannibalized from adjacent rocks of the laminated dolomudstone lithofacies, massive dolomudstone/dolowackestone lithofacies, and intraclastic and bioclastic lithofacies are common. Density current deposition of substantial amounts of bioclastic sediment derived from shallow platform areas is proposed for the bioclastic and intraclastic facies.

Mature, well-rounded and well-sorted quartz grains present in rocks of the massive, dolomitic quartz arenite facies and the laminated dolomudstone facies suggest derivation from an emergent area containing lithologically similar source rocks. Outcrops of Kinnikinic Quartzite and other Lower Paleozoic and Upper Proterozoic quartz arenites exposed in east-central Idaho and southwestern Montana are possible sources (Ruppel, 1986).

The increasing abundance of units of well-bioturbated massive dolomudstone/dolowackestone higher in the section suggests shallowing upward or relative deepening of the oxygen-minimum zone. Articulated brachiopods are also relatively common higher in the section, indicating the presence of a well-established benthic fauna. The general trend within the Roberts Mountains Formation at Bradshaw Basin is of shallowing-upward repeated density current deposition on a platform margin with apparently high paleorelief.

ANATOMY OF A FRASNIAN (LATE DEVONIAN) BUILDUP, GRANDVIEW CANYON VICINITY, CUSTER COUNTY

Introduction

Devonian carbonate buildups are common throughout the world and occur in a wide variety of tectonic settings. These buildups vary in scale and type from isolated, deeper water mud mounds to frame-supported, laterally extensive reefs along platform margins. Most of the well-documented Devonian buildups (especially those of Frasnian age) are constructed dominantly by stromatoporoids with corals and other organisms as lesser components (Wilson, 1975).

Scattered coral-stromatoporoid buildups occur in the upper part of the “dark dolomite member” of the Jefferson Formation (Givetian-Famennian) in east-central Idaho. This paper describes the faunal characteristics and regional stratigraphic context of one of the better-exposed buildups near Grandview Canyon, Lost River Range, Idaho (Fig. 2). The buildup is 22 to 40 m (72 to 130 ft) thick (Fig. 5) and is approximately 1.8 km wide (SW-NE, present outcrop trends). Its northwest-southeast length (probable trend of depositional strike) is indeterminate. Lithostratigraphically equivalent bio-
stromes of considerably less thickness are found 66 miles south-southeast (Skipp and Sandberg, 1975).

The buildup may span a critical period of time in the evolution of reefal communities, namely the Frasnian-Famennian boundary. This may be a significant factor towards the overall development of the buildup and, pending further detailed biostratigraphic studies, it may make the buildup an important locality for global studies of mass extinction events in the Late Devonian. The buildup also formed in a critical position along the Frasinian platform-to-basin transition from westernmost Montana to east-central Idaho, although regional lithostratigraphic correlations and facies relationships are still unclear. The buildup therefore is important for defining the nature of the platform-to-basin transition and for regional paleogeographic reconstructions.

History

Benson (1966), Loucks (1977), and Sandberg and others (1983) have discussed regional stratigraphic and depositional relations for the Jefferson Formation in southwestern Montana and east-central Idaho. However, buildup complexes in the Jefferson Formation of east-central Idaho have not been documented in detail. Isaacson and others (1983a, 1988) presented a brief report of vertical biofacies within the biosterome complex, and there have been references to stromatoporoidal biostromes within the formation at Fish Creek Reservoir (approximately 66 miles south-southeast; Skipp and Sandberg, 1975).

Geologic Setting and Stratigraphic Relationships

The Devonian Jefferson Formation (Givetian to Famennian) in east-central Idaho is exposed in a series of north to northwest-trending, elongate, fault-block mountain ranges. Most Devonian exposures are allochthonous, having been transported east or northeast along complex thrust and transcurrent fault systems during Pennsylvanian(?)-Tertiary time (Skipp and Hall, 1975; Skipp and Hait, 1977; Skipp and others, 1979; Reynolds and Kleinkopf, 1979; Ruppel, 1979; Ruppel and Lopez, 1984).

Recent studies in central Idaho (Skipp and Sandberg, 1975; Moser, 1981; and Simpson, 1983) have reported scattered stromatoporoid occurrences within the Jefferson Formation. Johnson and others (1985) have demonstrated, with biostratigraphic information, that in Idaho the Jefferson Formation represents a considerably greater time span than it does in Montana (Fig. 22). In east-central Idaho (the region where the bioherm/biostrome complex occurs), the formation was deposited from Middle through Late Devonian time (lower Givetian through mid-Famennian), while in Montana the formation is Late Devonian (mid-Frasnian) in age.

Isaacson and others (1983a) proposed that three stages of deposition are present in the Jefferson Formation in central Idaho. The lower part ("dark dolomite" member) consists of a thick sequence of laminated to massive dark mudstones and bioturbated wackestones with scattered stromatoporoids, brachiopods, gastropods, and rugose corals (approximately 250 m, 800 ft). A bioherm/biostrome complex (described below) overlies this and contains several growth forms of stromatoporoids, rugose corals, and tabulate corals (22-40 m, 70-130 ft). The upper part of the formation (designated as the Grandview Dolomite by Ross, 1937, and the Grandview Dolomite Member by Isaacson and others, 1983b) consists of detrital quartz-rich carbonates with cryptalgal laminations, LLH stromatolites, pseudomorphs after gypsum, and scattered shelly fossils (Simpson, 1983). Isaacson and others (1983a, b) suggested that the Grandview Dolomite Member is a local unit that changes over nine miles (nonpalinspastic distance) from western tidal flat facies, which were occasionally subaerially exposed, to eastern bryozoan-rich subtidal units. The lateral facies relationships suggest that the Grandview Member represents the earliest stages of an uplift to the west, which was the source area for quartz-rich sands that were shed to the east of this high. Isaacson and others (1983a), therefore, suggested a Late Devonian age for incipient Antler uplift and probable attendant foreland basin development, which permitted deposition of several Mississippian flysch and fan units (Nilsen, 1977).
Figure 23. Measured sections of carbonate buildup, upper part of “dark dolomite” member, Jefferson Formation. Lithofacies, biofacies and suggested development are shown.
Anatomy

Figure 23 contains a lithologic and faunal summary of the Jefferson Formation buildup. The buildup has a limited range of lithologies, namely mudstone, wackestone, floatstone, and bafflestone. Considering this emphasis is placed on the organisms present within the buildup and detailed description of the four distinct biofacies. The buildup was constructed on laminites and downslope breccias of the lower, “dark dolomite” member of the Jefferson Formation. Origin of the buildup may be a response to development of hardgrounds, on which biofacies A is established by stromatoporoids and colonial rugose corals.

Lithofacies

All units are dolomitized, although dolomitization has not destroyed original depositional textures. Mudstones are fine-grained, consistently bioturbated, and contain skeletal debris throughout. They dominate the eastern measured sections (Fig. 23) and are also laterally persistent across the top of the buildup.

Wackestones are common throughout the entire buildup and are characterized by abundant fine-grained skeletal debris. They commonly grade into floatstones, where larger skeletal remains dominate. They are locally bioturbated, and wackestone appears to be the dominant matrix in the coarse-grained floatstones and bafflestones.

Floatstones are dominated by large (up to 5 cm, 2 in) coral and stromatoporoid fragments. In places, crude horizontal alignment of bioclasts is preserved. The floatstones occur throughout the buildup; they are laterally persistent across the top of the buildup and comprise a larger volume of the western measured sections (Fig. 23).

Bafflestones are less common in the buildup than the other lithologies, but they are distinctive and important in the context of buildup development. They are locally abundant where in situ coral heads occur with a mudstone or skeletal wackestone matrix. Uniquely occurring in this lithology are the branching (phaceloid) growth forms of corals and occasional Stachyodes and Amphilora.

Biofacies

Significant organisms playing a role in the development of the buildup include, in decreasing order of abundance: “Thamnopora” with vertical growth (up to 15 cm, 6 in) exceeding horizontal dimensions, Peneckiella in hemispherical colonies, Syringopora in delicate hemispherical colonies, nodular (boulbus), tabular, and encrusting stromatoporoids, Amphipora, pelmatozoans, gastropods, brachiopods, and Stachyodes. Interesting to note, therefore, is the major role played by corals in the buildup development. Biofacies are listed in ascending stratigraphic order.

Biofacies A (Peneckiella and nodular and encrusting stromatoporoids)

This biofacies is laterally persistent and compositionally uniform across the base of the buildup (Fig. 23). Relatively thin (4-7 m, 13-23 ft) when compared with the other biofacies, it changes lithologic character from the base (wackestones) towards the top (floatstones and bafflestones). It is characterized by abundant, hemispherical growth forms of the colonial rugose coral, Peneckiella (Fig. 24), and stromatoporoids of a bulbous shape (according to the classification scheme of Kobluk, 1978). It also contains minor amounts of solitary rugose corals, pelmatozoans, and gastropods. Biofacies A is generally established on a hardground surface which shows numerous borings and irregular relief (up to 10 cm, 4 in).

The lower portion of the biofacies contains bulbous stromatoporoids in growth position (Fig. 25); their sizes range from 2-10 cm (1-4 in) in height and 5-7 cm (2-3 in) in diameter. Stromatoporoids that encrust Peneckiella colonies (Fig. 26), and nodular types also occur. This latter term is applied to those forms that may have been classified as bulbous (Kobluk, 1978, Fig. 5) but that are more concentric and commonly occur with favositid coral fragments as nuclei (Fig. 25).

Hemispherical growth forms of Peneckiella are abundant in the upper part of the biofacies. Colonies can be large (45 cm, 18 in, in diameter and 30 cm, 12 in, in height), and they can occur in growth position or less commonly overturned. Debris from these colonies is scattered throughout the biofacies. Matrix between the colonies’ corallites is fine-grained, unfossiliferous mudstone. Both the matrix and corallites commonly display truncation surfaces, with stromatoporoids subsequently encrusting the colonies.
“Thamnopora” occurs as debris throughout the biofacies. This taxon is more significant in other biofacies. Occurring as small, phaceloid colonies, whose upward growth component exceeds horizontal components, the growth form of this taxon is similar in all four biofacies. The Jefferson buildup has a notable lack of variation of growth forms of favositids, such as those described by Isaacson and Curran (1981) for an Early Devonian buildup. Other skeletal debris in this biofacies is composed of solitary rugose corals, pelmatozoan columnals, and gastropods within indistinct beds. Highest diversity of these organisms is in the eastern sections (Fig. 23).

Biofacies B (“Thamnopora”-dominated)

This biofacies is the distinctive biogenic component of the buildup. It is continuous in all the sections studied, and it thickens conspicuously in the western two sections (10-15 m, 32-50 ft, Fig. 23). Dominant biota are “Thamnopora,” discussed above in biofacies A, whose vertical growth dimension rarely exceeds 15 cm (6 in). Other organisms are not abundant, relative to the tabulate corals. They include Peneckiella, Syringopora (colonial tabulate), stromatoporoids (bulbous and tabular growth forms), and brachiopods. There is a general decrease of organismic diversity in the eastern sections. In the west, the biofacies is comprised of floatstones and bafflestones; in the east, mudstones are interbedded with floatstones. Lithologic changes between biofacies A and B are as conspicuous as the biotic changeovers.

“Thamnopora” colonies occur as the major biota and are closely packed (Fig. 27). They occur as small fragments or debris (2-4 cm, 1-2 in) in the lower part of the biofacies and as large, complete specimens in the middle and upper parts of the biofacies. Poorly developed beds (4-30 cm, 2-12 in, thick) are defined by those specimens that are toppled over, relative to upright occurrences.

Biofacies C (Amphipora, “Thamnopora,” and variable stromatoporoids)

The most variable biofacies, biofacies C is prevalent in the western three measured sections (10-12 m, 32-40 ft, thick), and it is much thinner (1.5 m, 5 ft) and less apparent in the eastern section (Fig. 23). The biofacies is distinguished by laterally variable concentrations of Amphipora, associated with bulbous and encrusting stromatoporoid growth forms, “Thamnopora,” Peneckiella, and Syringopora in skeletal wackestones and floatstones. Figure 28 shows Syringopora utilizing a stromatoporoid as a substrate. Biofacies C has the highest organismic diversity, and no single taxon is dominant. Amphipora occurs as both debris and intact, upright growth forms. It is distributed randomly throughout the biofacies. “Thamnopora” is restricted to the two western sections and occurs only as debris (2 cm, 1 in, fragments). Of the other stromatoporoid growth forms present, the bulbous type is next in rank abundance; tabular (encrusting) forms also occur (Fig. 29). Syringopora, the dominant organism in biofacies D, appears gradually towards the top of biofacies C. Shelly macrofaunas are rare, relative to biofacies A and B. Disarticulated brachiopods, pelmatozoan columnals, and solitary rugose corals are mostly found in the western section.

Biofacies D (Syringopora)

Occurring in bioturbated mudstones and wackestones are hemispherical colonies of Syringopora, in a delicately phaceloid growth form (Fig. 30). Occurrences of the taxon are laterally variable, such that in situ colonies are common in the eastern section and occur with Syringopora bioclasts in the western sections (Fig. 23). Biofacies D maintains an equal thickness in all sections (11-13 m, 36-43 ft). Associated with Syringopora in
biofacies D are rare *Peneckiella* colonies and bioclasts, solitary rugose corals, gastropods, brachiopods, bulbous and tabular stromatoporoids, and rare pelmatozoan debris. Biofacies D is terminated by an erosional surface of up to 2 m (6.5 ft) relief. A coarse-grained, faintly laminated recrystallized grainstone overlies this biofacies in all measured sections.

**Age**

The age of the buildup has not been precisely established. Brachiopods being recovered within the laminites, lower in the section, suggest an early Frasnian age, based on occurrences of *Tenticospirifer* and other taxa. The buildup part of the sequence has been interpreted as lower Frasnian, and the overlying Grandview Member has been dated as lower Famennian by Johnson and others (1985).

**Discussion**

Characteristics of the buildup make it important for analysis of relative sea level fluctuations, regional paleogeographic reconstructions, and study of the mechanisms responsible for overall development of the buildup.

**Relative Sea Level Changes and the Buildup Sequence**

The buildup begins with small (5-10 cm, 2-4 in, diameter), spherical to hemispherical stromatoporoids (biofacies A; Fig. 25) which were initiated above the basal hardground horizon. The basal hardground probably served as the substratum for initial development of the buildup. An increase in stromatoporoid abundance upwards and subsequent development of *Peneckiella* bafflestones and floatstones (biofacies A) represents the beginning of organic colonization on the hardground. The *Peneckiella* stage varies in thickness laterally. The upper part of biofacies A includes a faunal changeover to stromatoporoids which initially encrust *Peneckiella* corallites. Mud infilling, lithification, and then bioerosion(?) occurred before stromatoporoid encrustation (Fig. 26).

A buildup “core” phase (western parts of biofacies B) is dominated by abundant *in situ* branching favositid corals (“Thamnopora”-dominated) with a muddy matrix. This part of biofacies B is the zone of greatest accumulation of *in situ* organic skeletons within the buildup, although relative faunal diversity is low in comparison to other biofacies. The growth forms of the favositids indicate organic growth occurred during moderate sedimentation rates under relatively quiet water, yet turbid conditions (Wilson, 1975). In fact, the dominance of corals over stromatoporoids throughout the buildup and the predominant branching growth morphology of the corals suggest that large parts of the buildup formed in relatively deeper water, turbid environments, with moderate sedimentation rates (Brouwer, 1964; Wilson, 1975; Duretche, 1981). This “core” phase may have been an attempt by reefal communities to aggrade vertically.

Biofacies C contains beds which probably were deposited within wave base. These beds consist of fragmented and abraded allochthonous bioclasts in random orientations and tabular stromatoporoids. Gently undulatory to more planar laminations in coarse-grained bioclastic beds are similar to hummocky cross-stratified beds (Harms and others, 1975) which probably formed as the result of sediment reworking by storm-generated waves. The tabular stromatoporoids also suggest growth in moderate turbulence (Kobluk, 1978; Steam, 1982).
Figure 29. Polished slab photograph of tabular to domal stromatoporoid, showing some dissolution (biofacies C, 1X magnification).

These beds therefore indicate that environmental conditions, namely wave energy, also exerted a control on buildup development.

The upper part of biofacies C may have formed during an apparent relative rise in sea level. Faunal diversity and abundance in the upper part of biofacies C decrease abruptly, although most organic skeletons are still in situ. For whatever reason, buildup communities apparently were unable to keep pace with relative sea level rise. Additionally, the lack of toppled or abraded skeletons and of wave-generated sedimentary structures in the upper part of biofacies C suggests that the sediment-water interface then was below storm wave base.

Biofacies D probably records a postbuildup phase of deposition which formed under relatively quiescent conditions. A continued relative sea level rise is suggested by uniformity of both the lithologic and faunal composition of the biofacies and its constant lateral thickness (Fig. 23). The delicate branching morphology of Syringopora indicates relatively quiet water conditions with moderate sedimentation rates.

However, the contact between Biofacies D and the overlying Grandview Dolomite Member is an erosional surface with substantial local relief. Petrographic, field, and geochemical evidence from the upper 2 m (6.5 ft) of biofacies D also suggest that the erosional surface was exposed subaerially prior to deposition of the overlying Grandview Member (Dorobek and Filby, 1988, in press). This suggests that the deepening event which terminated the buildup was followed in quick succession by a fall in relative sea level. If the sea level fall was a eustatic event, it is difficult to estimate the amplitude of the fall without knowing the thickness of sediment which may have been removed prior to Grandview deposition. Regional tectonic uplift of the buildup (earliest phases of the Antler Orogeny??) also may have played an important role in the relative sea level fall.

Figure 30. Polished slab photograph of intact, in situ Syringopora colony on tabular stromatoporoid, with incipient stylolite (biofacies D, 1.5X magnification).

Other Factors Possibly Affecting the Buildup Sequence

Other factors may have influenced the ultimate vertical sequence of lithofacies and biofacies within the Jefferson buildup. The disappearance of most biotic constituents above biofacies C may not be related to any extrinsic environmental factors (i.e., relative sea level rise) but instead may reflect the global extinction of reef-building organisms at the Frasnian-Famennian boundary. More precise biostratigraphic dating needs to be completed before this can be evaluated as a possible influence on the overall buildup sequence.

Episodic tectonic subsidence/uplift also may have affected the buildup sequence. The buildup formed in a near-platform margin setting, during the transition of the platform from a passive margin, distally steepened ramp (sensu Read, 1985) in early Frasnian time to a ramp which faced an incipient foreland basin in late Frasnian time (Isaacson and others, 1983a; Dorobek, 1987, in press). Therefore, it might be expected that the buildup experienced “atypical” crustal movements, which may have influenced development of the stratigraphic sequence.

Sediment influx from the broad carbonate platform to the east also may have affected development of the buildup. The Givetian-early Frasnian part of the Jefferson Formation which underlies the buildup consists of well-laminated to thin-bedded fine-grained dolostones (probable distal turbidites) which grade upwards into bioturbated dolowackestones and dolopackstones.
Clearly, carbonate sediment was supplied from the eastern carbonate platform in Montana to deeper water environments in east-central Idaho. Basin infilling may have allowed development of the buildup, even though carbonate sediment was still being washed off of the eastern platform. Terminal stages of buildup development (upper biofacies C and biofacies D) may record higher rates of platform-derived carbonate sediment influx during the apparent sea level rise (cf. Schlager, 1981; Kendall and Schlager, 1981).

Comparisons with Other Devonian Buildups

The buildup at Grandview Canyon is different from many age equivalent carbonate buildups around the world. The biostratigraphic equivalency of buildups used in this discussion is based on conodont zonation in Johnson and others (1985).

The well-documented late Frasnian reef complexes of western Canada (including buildups in the upper Leduc Formation, Nisku Formation, and Fairholme Group) commonly formed on top of broad, relatively flat-topped, antecedent carbonate platforms (Klovan, 1974). Leduc reefal facies may have formed near-vertical reef walls which shed their own skeletal debris downslope; shallow water peritidal carbonates then accumulated in interior portions of the isolated platforms (sensu Read, 1985) which were formed by the concentric reef walls (McGillivray and Mountjoy, 1975; Walls, 1983). Fairholme reefal facies similarly consist of relatively narrow reef walls at the margins of isolated platforms with associated peritidal facies in platform interiors (Mountjoy, 1967). By comparison, the Frasnian buildup at Grandview Canyon never developed a rigid reef wall; framestone fabrics are completely absent in the buildup. Skeletal debris from organisms in the buildup only occur as matrix between *in situ* skeletons; discrete flank facies never formed. Also, quiet water, protected facies do not occur because the buildup probably never aggraded to near sea level and a rigid reef wall never formed.

Canadian Nisku buildups, which are closest in age to the Grandview Canyon buildup (Johnson and others, 1985), occur as “pinnacle reefs” (Pounder and others, 1980) and as “bank-edge reefs” (Machel, 1983). The “pinnacle reefs” occur as lenticular biohermal masses in downslope settings; the “bank-edge reefs” occur as biohermal masses along the basinward margin of the Outer Nisku shelf (Pounder and others, 1980; Machel, 1983). Machel (1983) has described the bank-edge reefs as “coral-bearing mud mounds (buildups)”; fasciculate coral growth forms dominate with stromatoporoids as accessory components in a muddy matrix. In terms of biota and paleogeographic setting, the Nisku buildups are more similar to the Grandview Canyon buildup than are other Canadian Frasnian buildups.

Frasnian reefal facies in the Napier Range, Canning Basin, Western Australia occur as laterally continuous, but relatively narrow reef walls which formed along a steep platform margin (Playford and Lowry, 1966). The reef wall shed coarse rubble into slope and basinal environments and also acted as a protective barrier for quieter water backreef environments. The reef wall is largely constructed of massive stromatoporoid/calcareaous algae framestones. Corals are very rare in the reef wall, although they do occur in backreef facies. The Canning Basin reefs differ significantly from the Grandview Canyon buildup in that they were dominated by stromatoporoids which formed rigid, wave-resistant frames along a steep platform margin. As such, they influenced backreef environments, while no evidence exists for the Grandview Canyon buildup as having influence to any great degree surrounding environments.

Upper Frasnian buildups in Belgium occur as mud mounds with large, lamellar or rare branching tabulate corals as the dominant faunal component (Lecompte, 1970; Tsien, 1971; Burchette, 1981). Abundant calcareous algae, rare stromatoporoids, and scattered brachiopod-echinoderm-bryozoan debris are the remaining fauna. Stromatactis cavities also are common. These mud mounds are enclosed in dark shale or argillaceous limestone. The mounds differ from the Grandview Canyon buildup in that they probably formed in deeper water, more basinal settings where more turbid water conditions existed (Lecompte, 1970; Tsien, 1971). However, the mud mounds still had to be within the photic zone in order for algae to exist.

In summary, these comparisons of Frasnian buildups from around the world allow several generalizations to be made about coral-dominated Frasnian buildups. Most buildups similar to the Grandview Canyon buildup probably formed in relatively deeper water environments with moderate sedimentation rates and turbid water conditions. Additionally, they most likely occur along gently sloping carbonate platforms with ramplike profiles (Ahr, 1973; Read, 1985).

Conclusions

The coral-dominated carbonate buildup in the Jefferson Formation of east-central Idaho is important for documenting the faunal sequence of a deeper water carbonate buildup. The buildup probably formed in a tectonically unstable crustal setting following vertical aggradation of deeper water slope facies, or it may have formed slightly downslope along the distal parts of a carbonate ramp. Sea level changes occur within the buildup sequence, which appears to lack a reef wall, a feature many coeval Alberta complexes possess.

The faunal diversity and growth morphologies are like those found in other Frasnian buildups from similar tectonic and paleogeographic settings. The dominance of fasciculoid corals in the buildup is indicative of deeper water, moderate sedimentation rates, and turbid water conditions. The overall buildup sequence may reflect eustatic sea level fluctuations, crustal movements, global
extinction events, or variation in influx of platform-derived carbonate sediment.

ACKNOWLEDGMENTS

The Idaho Mining and Mineral Resources Research Institute provided McFadden, Measures, and Isaacson partial field support for this study. The University of Idaho Research Council also partially supported Isaacson's fieldwork and travel to the Second International Symposium on the Devonian System, Calgary, Alberta. Measures was also supported by an American Association of Petroleum Geologists Grant-In-Aid, a Sigma Xi Grant-In-Aid, and a scholarship from the Dallas Geological Society. The manuscript was reviewed by Claude Spinosa.

REFERENCES


Sloss, L. L., 1954, Lemhi Arch, a mid-Paleozoic positive element in south-central Idaho: Geological


Wilson, J. L., 1975, Carbonate Facies in Geologic History: Springer-Verlag, New York, 471p.
Stratigraphy and Structure of the Milligen Formation, Sun Valley Area, Idaho

Robert J. W. Turner
Bruce R. Otto

ABSTRACT

The Milligen Formation contains black argillite interbedded with limestone and quartz arenite turbidites, chert and minor chert-pebble conglomerate. The formation occurs as a fault-bounded allochthon with an aggregate thickness of greater than 3,500 feet and is subdivided into five lithostratigraphic units. In ascending stratigraphic order they are: lower argillite, Cait quartzite, Lucky Coin limestone, Triumph argillite, and Independence sandstone. Correlation with the conodont localities published by Sandberg and others (1975) indicates a probable early Late Devonian age for the Independence sandstone and a probable early Middle Devonian age for the Lucky Coin limestone. Milligen strata indicate that deposition changed from a deep marine, starved basin environment during Early (?) and Middle Devonian to a complex, deep marine turbidite-dominant environment during Middle (?) and Late Devonian times. The presence of quartzite and shallow water bioclastic calc-turbidites, and the predominance of channeled turbidite depositional systems indicate an outer miogeoclinal (continental slope and rise) setting on the margin of continental North America. Local chert-pebble conglomerates and mafic sills and tuffs suggest intrabasinal syn-depositional faulting as has been recorded in similar coeval strata of the northern Cordillera.

Three major deformational events are recorded in the Milligen strata. D1 structures include tight folds (F1) and a penetrative axial planar cleavage subparallel to bedding; F1 folds commonly display a soft-sediment style of deformation. The D2 event includes open to tight, north-trending, east-vergent folds with variably developed spaced and crenulation cleavage, and northwest-trending, southwest-dipping thrust faults. D3 low-angle west-dipping normal faults with N. 50-70° W. displacements and north- and northeast-trending high-angle faults (D4) offset Eocene volcanic rocks.

D1 structures are correlated with the Early Mississippian Antler orogeny based on the soft-sediment-style of many F1 folds, the folding of D1 structures by D2 folds, and the absence of similar D1 structures in overlying, post-Early Pennsylvanian strata. D2 folds and thrust faults are interpreted to be Mesozoic age, based on similarities to structures in overlying Carboniferous to Permian strata. Tertiary age D3 low-angle faults reflect NW-SE extension of the Milligen Formation related to the formation of the adjacent Pioneer core complex. High-angle D4 normal faults are related to development of the Basin and Range province.

INTRODUCTION

The name Milligen Formation was given to dark-colored argillaceous rocks of supposed Devonian age in

---

1 Westley Mines, Ltd., 900-475 Howe St., Vancouver, B. C., Canada V6C-2B3
2 Mintek Resources, Ltd., (residence: 333 Provident Drive, Boise, ID 83706

the Wood River area of central Idaho (Fig. 1) by Umpleby and others (1930). Recent workers (Roberts and Thomasson, 1964; Sandberg and others, 1975; Poole and others, 1979) interpret the Milligen Formation as continental rise deposits that were thrust eastward onto the continental shelf of North America as a part of the Roberts Mountains allochthon during the Mississippian Antler orogeny. Based on conodont collections reported in Sandberg and others (1975), the age of the Milligen Formation is, at the youngest, Middle and Late Devonian, and these authors inferred that it likely represents much of the Devonian period. Stratiform syngenetic lead-zinc-silver deposits within the Milligen Formation have made it a focus of mineral exploration (Hall, 1985).

![Figure 1. Location map showing the Milligen Formation outcrop in Blaine County, Idaho, upon which the conclusions of this paper are based. The map of Hall (1985) shows the Milligen Formation also outcropping west of the Wood River and just east of Hailey. These areas were not studied in detail for this paper. IM--Independence Mine, LCM--Lucky Coin Mine, PM--Parker Mine, TM--Triumph Mine.](image)

The internal stratigraphy of the Milligen Formation has remained poorly understood due to structural complexity, poor exposure of rock units, and a paucity of fossils. Umpleby and others (1930) named the strata exposed in Milligen Gulch, 6 miles east of Ketchum, as the type section of the Milligen Formation. Kiilsgaard in Anderson and others (1950) mapped discrete limestone units within a sequence of black carbonaceous argillites in the area of the Triumph and Parker mines east of Ketchum. Sandberg and others (1975) divided the Milligen Formation into a thick, lower argillite member with thin quartzite beds of Early (?) and Middle Devonian age, and a thin upper calcareous turbidite member of Late Devonian age. This interpretation was based on conodont collections from several measured sections including one in Milligen Gulch.

This paper discusses the internal stratigraphy and structure of the Milligen Formation and refines and expands upon work originally presented in Otto and Turner (1987). Because much of our data are currently of a proprietary nature, we emphasize general characteristics of the Milligen Formation rather than data presentation. From 1983 to 1987, approximately 33 man-months were spent investigating the Milligen Formation for Westley Mines, Mintek Resources Ltd. and Noranda Exploration Ltd. Our efforts were confined to outcrops of the Milligen Formation east of the Wood River between Eagle Creek to the north and Bellevue to the south (Fig. 1). Much of this work was conducted in the vicinity of the East Fork of the Wood River, Lake Creek and Trail Creek.

Interpreting the stratigraphy of the Milligen Formation is complicated by complex structure, facies changes within stratigraphic units, and in particular by a lack of biostratigraphic control due to the paucity of fossils. Our stratigraphic interpretations are based on detailed studies of better exposed areas and regional scale mapping.

**STRATIGRAPHY OF THE MILLIGEN FORMATION**

The Milligen Formation is exposed in a north-trending belt 30 miles long and up to 6 miles wide within Blaine County (Fig. 1). We have recognized five lithostratigraphic members within the Milligen Formation. These are informally named, in ascending stratigraphic order, lower argillite, Lucky Coin limestone, Triumph argillite, Cait quartzite, and Independence sandstone (Fig. 2). The Milligen Formation occurs as a fault-bounded allochthon (Dover, 1980) and stratigraphic contacts of the Milligen Formation with older or younger strata have not been recognized.

**Lower Argillite**

The lowest part of the Milligen Formation is an argillite-dominated unit here referred to as the lower argillite (Fig. 2). In Otto and Turner (1987), this unit was named the lower sand unit; however, current work indicates that this unit is only locally sand-rich and the term lower argillite is more appropriate.

The lower argillite is predominantly a siliceous black argillite with variable amounts of thin-bedded sandstone and chert that underlies the Lucky Coin limestone. The
base of the lower argillite is not exposed, but the unit is at least 400 feet thick.

The relative amounts of carbonaceous argillite, thin-bedded sandstone and thin-bedded chert vary throughout the lower argillite. A chert-rich facies and a sand-rich facies are recognized, but their relative distribution is poorly understood. The sand-rich facies is characterized by thin-bedded, calcareous to noncalcareous, fine-grained sandstones that constitute up to 40 percent of the strata (Fig. 3). These sandstones are commonly massive and ungraded, and are interpreted to represent deposition from low-energy turbidity currents. The chert-rich facies of the lower argillite is composed of thinly interbedded black to grey chert and carbonaceous argillite (Fig. 4) and represents a starved basin environment dominated by the deposition of pelagic and hemipelagic material.

The lower argillite is separated from the lithologically similar but stratigraphically higher Triumph argillite by the Lucky Coin limestone. Where the Lucky Coin limestone is absent (Fig. 2), the contact of the lower argillite with the overlying Triumph argillite lies approximately at the Cait quartzite.

### Cait Quartzite

Interbedded with the upper portion of the lower argillite and possibly the lower part of the Triumph argillite is a distinctive coarse-grained quartzite unit here named the Cait quartzite, after excellent exposures on lode mining claims of the same name near Sawmill Gulch, a tributary of the East Fork of the Wood River. The Cait quartzite is composed of well-rounded, dark-colored, quartz-cemented, coarse quartz grains that range in diameter from 0.5 to 1 millimeter (Fig. 5). The Cait quartzite varies markedly in thickness laterally within the Milligen Formation. Generally the quartzite occurs in beds a few inches to a few feet thick (Fig. 6), however individual beds locally attain thicknesses up to thirty feet, and the aggregate thickness of stacked quartzite beds may exceed 300 feet (Fig. 2). The thicker beds of quartzite are generally lenticular in shape. Flute casts on the bases of beds generally indicate east-to-west transport directions.

Where the Lucky Coin limestone is absent, the Cait quartzite serves as an important stratigraphic marker of proximity (within 200 feet) to the top of the lower argillite (Fig. 2).

The narrow stratigraphic interval and the widespread occurrence, plus the lack of associated thin-bedded or finer-grained quartzites associated with the coarse-grained beds, suggest that Cait quartzite deposition was a short-lived event that spilled quartz sands as turbidity flows

---

**Figure 2. Schematic figure of internal stratigraphic relationships within the Milligen Formation. No scale is intended.**
from an eastern shelf source into the Milligen basin. Much of this flow was channeled, but widespread thin Cait sands attest to transport by laterally extensive sheetlike flow. Seismic activity may have been responsible for destabilizing these outer shelf sands.

**Lucky Coin Limestone**

The Lucky Coin limestone is named for exposures near a prospect of the same name in lower Independence Canyon 2 miles southeast of Ketchum. This limestone unit is equivalent to the lower and upper tremolite limestone members of the Milligen Formation in the Triumph-Parker mine area as described by Kiilsgaard in Anderson and others (1950). The Lucky Coin limestone is composed of silty limestone turbidites interbedded with carbonaceous siliceous argillites. The thickest exposure of the Lucky Coin limestone is 1,200 feet in Independence Canyon, but this may include structural thickening. The Lucky Coin limestone varies greatly in thickness and is commonly absent, such that the Triumph argillite directly overlies the lower argillite. In many places, the absence of the Lucky Coin limestone coincides with thicker sections of the Cait quartzite (Fig. 2).

Three distinctive limestone bed types occur in the
Lucky Coin limestone: thick-, medium- and thin-bedded. Thick-bedded limestones are dark grey, carbonaceous and fine- to medium-grained; individual beds range from 5 to 20 feet in thickness. Occasional argillite fragments up to 2 feet in dimension are distinctive of these limestone beds. Rare bioclastic conglomerate and pebbly limestone occur in the thick-bedded limestones; fragments of bryozoa (?) and partially recrystallized crinoids (?), and shelly material are mixed with abundant carbonate and argillite fragments. Thick-bedded limestones are Tae turbidites (terminology of Bouma, 1962) interbedded with black carbonaceous argillites and commonly occur as amalgamated Taaa turbidite bed sets several feet thick. In Independence Canyon where the Lucky Coin limestone is thickest, thick-bedded limestones dominate. Dark-colored actinolite, tremolite, and chloritoid are common in thick-bedded limestones that have been contact metamorphosed.

Medium-bedded limestones are calcarenites as much as 2 feet thick rhythmically interbedded with calcareous and noncalcareous shales (Fig. 7). The calcarenites display planar lamination and graded tops, and are Tbde turbidites. Medium-bedded calcarenites also occur with thick-bedded limestones or in monotonous sequences up to 100 feet thick.

Thin-bedded, fine-grained, micritic limestone beds 0.5
Figure 7. Lucky Coin limestone: medium-bedded limestone turbidites interbedded with calcareous to noncalcareous, carbonaceous argillite.

Figure 8. Lucky Coin limestone: thin-bedded amalgamated limestone beds.

to 3 inches thick (Fig. 8) are composed commonly of a basal lamina 1-2 mm thick of medium-grained quartz overlain by ungraded, massive micrite, and interpreted as Tdte turbidites. Thin-bedded limestones commonly occur as bedsets 10 to 50 feet thick of amalgamated turbidites.

Sandberg and others (1975) report early Middle Devonian conodonts from a thin limestone encrinite bed interbedded with argillites in Sawmill Gulch (SE 1/4 NW 1/4 NW 1/4 Sec. 23, T. 4 N., R. 19 E.). It is possible that this bed is part of a thin marginal portion of the Lucky Coin limestone.

The Lucky Coin limestone is interpreted to have been deposited from turbidity flows into a sub-wave base, euxinic environment. Several characteristics of the Lucky Coin suggest it was deposited in a channeled environment: (1) thicker beds dominate where the unit is thickest, and beds thin where the unit thins; (2) both thick-bedded and thin-bedded limestones are commonly amalgamated, suggesting very high sedimentation rates; (3) the occurrence of large argillite fragments in thick beds suggests an erosional channel environment and (4) the variation in thickness and the common absence of the Lucky Coin limestone suggest lenticular beds. Thick and massive limestones represent Tae turbidite deposition within a channel environment and are ungraded due to homogeneous micrite mud composition. Argillite clasts are likely rip-up clasts derived locally from incised channel walls. Rhythmically interbedded, medium- and thin-bedded limestones represent lower energy turbidity flow deposition perhaps peripheral to channels. Associated bioclastic beds dominated by abraded and broken
shallow-water faunas such as bryozoa and crinoids suggest that the Lucky Coin limestone was derived from a carbonate platform source, likely to the east as argued by Sandberg and others (1975).

Mafic Sills and Tuffs

Thin mafic sills and tuffs are interbedded with parts of the lower argillite and Lucky Coin limestone (Figs. 9 and 10). The sills are fine grained, contain phenocrysts of amphibole and biotite, and are weakly amygdular. Locally they exhibit spherulitic devitrification textures of an originally glassy matrix (Figs. 11 and 12). Sills commonly are 1 to 4 feet thick, although they range up to 20 feet thick (Fig. 10). Whole rock analyses of the sills demonstrate they are basalts and basaltic andesites. Thin sills are folded by F2 folds and locally contain a well-developed S1 cleavage.

Thin-bedded tuffs up to a foot thick are interbedded with parts of the lower argillite (Fig. 9). They are pale green in color, banded to laminated, and contain argillite fragments. Tuff beds display a well-developed S1 cleavage and are commonly folded by F1 folds (Fig. 9).

Because the tuff beds are folded by F1 folds, they are clearly of Devonian age. The sills are also likely to be of Devonian age based on their aphyric texture, which suggests intrusion into very shallow (wet?) sediments and deformation by F2 folds. They were likely associated with extensional faulting of the Milligen basin.
Triumph Argillite

The Triumph argillite is a black argillite unit between the Lucky Coin limestone and the Independence sandstone (Fig. 2). The Triumph argillite ranges in thickness from 100 to 500 feet and is composed of black siliceous argillite with minor interbeds of chert and fine-grained sandstone (Fig. 13). Both the underlying Lucky Coin limestone and the overlying Independence sandstone are interbedded with the Triumph argillite over tens of feet.

The black siliceous argillites are composed of fine-grained quartz, carbonaceous matter and minor illitic clay. Micro-laminations of carbonaceous matter and quartz are commonly evident in thin section. Carbonaceous argillites grade compositionally into black cherts with an increasing ratio of fine-grained quartz to organic matter. Cherts are typically thin-bedded and interbedded with argillite.

Fine-grained calcareous sandstones occur as beds up to 20 inches in thickness and locally are an important lithology within the Triumph argillite. Such beds commonly exhibit sharp bases and plane laminated tops. They can be described as Tbde beds (Bouma, 1962). These sands are composed predominantly of 0.1 to 0.25 millimeter quartz grains in a matrix of calcite. Both massive carbonaceous limestone beds and thin-bedded silty limestone beds similar to those within the underlying Lucky Coin limestone occur predominantly within the lower part of the Triumph argillite.

The high organic carbon and quartz content of the
carbonaceous argillites and cherts reflect euxinic conditions starved of detrital input. Interbedded calcareous quartzose sandstones and limestones are interpreted to reflect deposition from mixed quartz sand and micrite turbidity flows, likely from a shelf environment.

**Independence Sandstone**

The Independence sandstone, named for exposures in Independence Canyon, is the uppermost unit in the Milligen Formation. Although the top of the unit has not been observed due to faulting, it is over 1000 feet thick in Independence Canyon. The Independence sandstone is composed of interbedded fine-grained sandstone and argillite, and minor limestone and chert-pebble conglomerate.

The Independence sandstone is divided into two distinctive sandstone lithofacies (Fig. 2). The "interbedded" facies is characterized by thin-bedded sandstone interbedded with argillite and minor limestone.

Sandstones are distinctive in the abundance of sedimentary textures such as planar lamination, graded bedding, ball and pillow structures and climbing ripples (Figs. 14 and 15). Many thin-bedded sandstones contain 1-5 percent disseminated pyrite resulting in a very diagnostic reddish brown weathering color. Interbedded argillites are black near the base of the Independence sandstone but are transitional upwards to a pastel green color.

The second lithofacies of the Independence sandstone is referred to as the thick-bedded facies (channel facies of Otto and Turner, 1987) and is composed of thick bedded sandstones, many with less than 10 percent interbedded argillite (Fig. 16). Beds commonly range from 1 to 4 feet in thickness, and bed forms in order of abundance are planar laminations, climbing ripples, ball and pillow structures, and cross laminations. Beds are Tabce, Tbeo or Tec turbidites (after Bouma, 1962) and can exceed 5 feet in thickness.

Silty limestone beds up to 20 feet thick occur within the Independence sandstone. They are predominantly associated with the thick-bedded lithofacies, and some units appear to be amalgamated, 0.5 to 1 foot-thick limestone beds. Crinoid-rich bioclastic limestone beds within the Independence sandstone in Milligen Gulch have yielded early Late Devonian conodonts (Sandberg and others, 1975).

Chert-pebble conglomerate beds up to 75 feet thick occur in the lower part of the Independence sandstone and upper part of the Triumph argillite (Fig. 17). These thick-bedded conglomerates are composed predominantly of 1 to 3 cm white and grey chert clasts with subordinate amounts of black argillite and sandstone clasts mixed with up to 10 percent calcareous fine- to medium-grained quartz sand.

The Independence sandstone represents the progradation of a major turbidite fan complex into the Milligen basin. The lower part of the Independence sandstone is characterized by upward-thickening sequences of interbedded facies sands overlain by the thick-bedded facies; these deposits are interpreted as the successive progradation of distal fan, fan lobe, and fan channel environments. The change from black to green argillite interbeds up section within the Independence sandstone is interpreted as a change from anoxic to more oxidizing bottom water conditions. The relationship of the chert-pebble conglomerates to
the Independence sandstone depositional system is unclear. It is likely that the chert pebbles were derived from an intrabasinal source, possibly cherts within the Milligen or older Lower Paleozoic strata, and reflect intrabasinal extensional faulting.

**STRUCTURE OF THE MILLIGEN FORMATION**

Strata of the Milligen Formation are complexly deformed. Structural elements include early D1 folding and thrusting(?) possibly associated with the Early Mississippian Antler orogeny, later D2 folding and thrusting associated with the Sevier orogeny of Mesozoic age, D3 extensional faults associated with Paleogene extension related to the Pioneer Mountains core complex, and D4 high angle faults related to Neogene basin and range deformation.

**D1 Deformation event**

Tight to isoclinal folds (F1), associated penetrative axial planar cleavage (S1), and cataclastic zones are the earliest deformation recognized within the Milligen strata. In argillites and phyllitic argillites, the S1 cleavage is well-developed (Fig. 13) due to the preferred orientation of clays, fine-grained muscovite and chlorite; in limestones, S1 is a well-developed spaced cleavage. S1 cleavage is poorly developed in sandstones (Fig. 14), siliceous argillites and cherts, and it is locally obscured by later contact metamorphism, as in the area of the Triumph mine. S1 cleavage is shallowly oblique to bedding, commonly 10 to 30 degrees, and is axial planar to F1 folds (Fig. 18).

F1 folds are typically tight to isoclinal with S1 axial planar cleavage (Fig. 18). Fold amplitudes average 1-20 feet. Map-scale F1 folds have not been documented; however, the entire Milligen section in Independence Canyon is the overturned limb of what may be an F1 fold. In many places, F1 folds appear to have involved only partially lithified strata based on argillite injection along cleavage (Fig. 18) and “soupy” disaggregation of folded sandstone beds (Fig. 19). F1 folds are east-vergent with generally north-trending fold axes, but they have been refolded by younger Mesozoic F2 folding (Fig. 21). F1 folds are distinctive from F2 folds by their
association with a penetrative cleavage (S1).

Microstructural study by Davis (1984) reveals a pervasive chloritic cleavage in Milligen rocks. Fragments of Milligen rock in conglomerates of the Mississippian Copper Basin Formation also contain this foliation, and Davis argues that this dates the cleavage as pre-Mississippian and hence related to the Antler orogeny. This cleavage correlates with our S1 cleavage. Davis also notes that in Milligen rocks this S1 cleavage is cut at a high angle by a crenulation cleavage, presumably our S2 cleavage.

**D2 Deformation Event**

Open to tight, north-trending, east vergent folds, a variably developed spaced and crenulation cleavage axial to the folds and northwest-trending, southwest-dipping thrust faults comprise the D2 deformation event.

Northeast- and northwest-trending folds with northward or eastward vergence are common on outcrop and map scales in the Milligen strata (Fig. 20). They are generally tight and locally overturned, but include open folds. Outcrop-scale folds tend to be strongly disharmonic, and many are spatially associated with small-displacement thrust faults. Map-scale folds have western limbs that generally dip 30-45 degrees west; the eastern limbs are commonly overturned and dip 65-75 degrees west. Core zones of the larger folds typically contain abundant parasitic folds. Axial planar spaced cleavage is typically very weakly developed or absent.

In one well-studied area, there is evidence for multiple folding events (F2-a and F2-b) within D2 deformation. A northeast-trending fold set includes both open and tight folds. Slightly northwest-vergent open folds (F2-a) have a well-developed axial planar crenulation cleavage (S2) at a high angle to bedding and S1; a distinctive crenulation lineation (L1x2) formed by the intersection of the S1 and S2 surfaces occurs on the S1 surface (Fig. 22). Tight northeast-trending northwest-vergent folds (F2-a) appear to be related to the open folds but lack a cleavage. A north-trending map-scale fold (F2-b) bends the S1 cleavage and L1x2 lineation and clearly postdates the northeast-trending folds. Dover (1981) also noted two different fold sets of inferred Mesozoic age within the Pennsylvanian-Permian Wood River Formation.

Sandberg and others (1975) refer to a distinctive crinkly surface and phyllitic sheen that distinguish the Milligen from other argillaceous units; the former is likely the above-mentioned L1x2 crenulation lineation, and the latter refers to the S1 foliation.

Thrust faults trending northwest and dipping 30 to 40 degrees southwest locally repeat the Milligen stratigraphy. Displacement on thrust faults appears modest; where constrained, displacement exceeds 1,500 feet but is probably less than several thousand feet. Thrust faults host lead-silver vein deposits such as at the Independence mine east of Ketchum (Anderson and others, 1950). A Mesozoic age of movement is interpreted for these thrust faults, and they are referred to here as D2 thrusts. The northwest strike of the D2 faults parallels the majority of the regional F2 fold axes, and we tentatively suggest that these northwest-trending folds may correlate with the F2-b fold event described above. D2 thrust faults compare closely to similarly oriented faults (N. 70°W., 30°SW) near Bellevue and Hailey that host major lead-silver deposits (e.g., the Minnie Moore mine). The lead-silver mineralization is interpreted to be related to intrusion of the Idaho batholith of Cretaceous age (Hall and others, 1978) and this requires the faults to be of Late Cretaceous age or older. However, the sheared nature of these lead-silver ores attests to reactivation of the faults after ore formation.

**D3 and D4 Deformation Events**

Two normal fault sets offset D2 structures in the Milligen strata as well as rocks of the Eocene Challis Volcanics. These two fault sets are distinguished by dip orientation and are referred to as low-angle normal faults (D3) and steep normal faults (D4).

Low-angle normal faults (D3) strike north to
northwest and dip 5 to 30 degrees to the west. Fault zones are a composite of polished and slickensided planar surfaces, silicified breccias, coarse calcite marble breccias, and graphitic or mud gouge. Slickensides and fault striae commonly show a N. 50-70°W. transport direction (Fig. 23). Low-angle normal faults are areally extensive, and one such fault has been mapped over an area of several square miles; this fault is recognized by relative east-to-west, normal displacements of crestal portions of F2 anticlines and by similar offsets of stratigraphic units.

In a well-studied area where drill core is available, the Milligen stratigraphy is repeated by several southwest-dipping D2 thrusts, which are in turn offset to the west by three subparallel stacked D3 low-angle normal faults. The lower low-angle fault offsets Milligen strata and a D2 thrust approximately 2,000 feet in a northwesterly direction; the upper flat fault juxtaposes steeply dipping Pennsylvanian-Permian Wood River Formation strata against Devonian Milligen strata (Wood River thrust of Hall and others, 1974). Similar flat faults have been documented in the underground workings of the Minnie Moore mine near Bellevue (Umpleby and others, 1930; plates 28 and 30 and p. 235) and elsewhere in the Wood River area (Skipp and others, 1986; Link and others, 1987; Link and others 1988, this volume). These faults offset, and therefore postdate, Cretaceous intrusive rocks of the Idaho batholith.

The displacement of Eocene Challis volcanic rocks on a 20 degree west-dipping low-angle normal fault is documented in one area based on underground mine exposures, surface trenching, and drill core intersections. Flow banding within the volcanics overlying the low-angle fault is nearly vertical. How much of the Challis volcanic rocks in the Wood River area are underlain by these low-angle, extensional faults is unknown.

High-angle faults trend northeast and northwest and displace strata a few tens to a few hundreds of feet. Northeast-trending faults both cut and are cut by low angle normal faults. Northwest-trending faults (D4) cut all other faults and appear to be related to the present stress regime that has formed the northwest-trending Wood River graben.

**DISCUSSION**

Depositional Environment

The lower part of the Milligen Formation (i.e., lower argillite) is characterized by pelagic and hemipelagic
deposition in a deep marine environment starved of coarse clastic input. The middle and upper part of the Milligen Formation (i.e., Cait quartzite, Lucky Coin limestone, Triumph argillite and Independence sandstone) is a complex array of facies reflecting the progradation into the deep basin of various terrigenous and carbonate sediments. The lower Milligen may reflect a period of tectonic stability or sea level highstand; conversely the clastic-rich upper Milligen may reflect tectonic instability or sea level lowstand.

The clastic sediments within the Milligen (silty limestone, quartzite, calcareous quartz sand) were probably derived from an adjacent continental shelf. Certainly, the presence of the very mature, well-sorted sands of the Cait quartzite demand a continental shelf source. The transported shallow-water bioclastic material in the Lucky Coin limestone and Independence sandstone is also compatible with a shelf source. The channeled nature of the Cait quartzite and the Independence sandstone suggests distributary systems for shelf-derived sediments. Thus, we interpret the Milligen strata to have been deposited in a continental-slope and continental-rise (outer miogeoclinal) environment as was advocated by Sandberg and others (1975) and Poole and others (1977).

Bottom waters of the Milligen basin were euxinic up through the lower part of Independence sandstone time; subsequent dominance of stratigraphically higher green argillites may argue for oxygenation of the bottom waters. The high silica content of the argillites in the lower and Triumph argillites suggests elevated concentrations of silica in the bottom waters; such conditions have been inferred to develop in stratified water masses due to continued input of biogenic silica from overlying waters and the lack of recycling of this silica to shallower waters via upwelling (Fisher and Arthur, 1977).

Thick-bedded chert-pebble conglomerates of the Milligen compare to other Middle and Late Devonian strata in British Columbia and the Yukon. In the northern Cordillera, chert-pebble conglomerates are inferred to have been derived from intrabasinal source areas during extensional or transcurrent faulting of the outer miogeocline (Gordey and others, 1987). The presence of mafic sills and tuffs in the Milligen supports the existence of such an extensional environment. Chert detritus in the Milligen was probably derived from uplifted Milligen or older basinal strata.

Correlation with Previously Published Milligen Stratigraphy

Sandberg and others (1975) proposed a two-fold stratigraphy for the Milligen. Based on lithologic similarities, we suggest that the upper member of Sandberg and others (1975) from which they collected early Late Devonian conodonts, correlates with our Independence sandstone. A part of the lower argillaceous member proposed by Sandberg and others (1975) may correlate with our lower argillite. Sandberg and others (1975) report conodonts of early Middle Devonian age from a limestone bed in Sawmill Gulch which may correlate with part of the Lucky Coin limestone.

Evidence for an Antler-age Deformation Event

The presence of an Antler-age deformation in the Milligen Formation has been inferred by Sandberg and others (1975), Hall and others (1974), Skipp and Hall (1975a and b), and Nilsen (1977). Dover (1980) argued that these claims had not been substantiated by detailed fabric studies or rigorous comparison of fabrics between the Milligen and post-Antler age stratigraphic units. Because of this lack of documentation, Dover advocated that such cleavage may be Mesozoic in age. Detailed microstructural studies were conducted in part by Davis (1984) who documented Antler-age deformation.

F1 folds with both soft-sedimentlike textures and a pervasive axial planar cleavage (S1) are distinctive from F2 folds lacking both such characteristics. Folding of the S1 cleavage by F2 folds and intersection lineations of S1 and S2 fabrics argues for a deformational event prior to the D2 event of Mesozoic age. We attribute the D1
deformation to the Antler orogeny for two reasons. The soft-sediment style of some of this F1 deformation suggests deformation prior to lithification; this is compatible with the dating of the upper Milligen as Late Devonian in age and the dating of the Antler orogeny as Early Mississippian in age (see Sandberg and others, 1983; Murphy and others, 1984). An Antler age for D1 is also supported by the general absence of a penetrative cleavage in overlying Pennsylvanian and Permian strata of the Wood River Formation, although a locally penetrative cleavage is recognized. Therefore, D1 must predate Early Pennsylvanian time. The only post-Late Devonian and pre-Early Pennsylvanian orogenic event recognized in the Cordillera is the Antler orogeny.

Flat Faults of Tertiary Age

This study documents the presence of low-angle normal faults within the Paleozoic sedimentary rocks that flank the Pioneer core complex on the eastern margin of the Milligen outcrop belt. The north-northwest direction of displacement on these faults and their Eocene (syn- or post-eruption of Challis volcanics) timing of movement suggest that they are related to northwest-southeast extension in the Pioneer core complex as documented by Wust (1986) and discussed by Wust and Link (1988, this volume) and Link and others (1988, this volume). The fault contact between the Milligen Formation and the overlying Wood River Formation has been called the Wood River thrust (Hall and others, 1974). However, the parallelism of this fault with flat normal faults below it within the Milligen Formation as well as the lack of stratigraphic inversion suggest that it is also a normal fault. It is clear that many low-angle structures interpreted as thrust faults in the Sun Valley area (see Dover, 1980) require scrutiny.

ACKNOWLEDGMENTS

During the last several years a number of people have been involved in field work that provided the basis for this paper. Significant contributions have been made by Ken Loos, Pat Okita, Carl Michelson, Dave Smith and Mary Fitch. We are grateful to Dr. Clyde Smith of Westley Mines and Mintek Resources for discussion, for his continued support of this project, and for his permission to publish. Reviews by Paul Link, Clyde Smith, Sandra Soulliere, and F. G. Poole proved very helpful, but we assume full responsibility for all of our conclusions.

REFERENCES


Turner and Otto--Stratigraphy and Structure of the Milligen Formation, Sun Valley Area, Idaho 167


Murphy, M. A., Power, J. D., and Johnson, J. G., 1984, Evidence for Late Devonian movement within the Roberts Mountains allochthon, Roberts Mountains, Nevada: Geology, v. 12, no. 1, p. 20-23.


Sandberg, C. A., Gutschick, R. C., Johnson, J. G., Poole, F. G., and Sando, W. J., 1983, Middle Devonian to Late Mississippian geologic history of the overthrust belt region, western United States, in Powers, R. B., editor, Geologic studies of the...
Stratigraphy of the Lower Permian Grand Prize Formation, South-Central Idaho

J. Brian Mahoney 1
R. M. Sengebush 2

ABSTRACT

The Lower Permian Grand Prize Formation of south-central Idaho is over 2500 m thick in composite section. It consists of a basal chert-pebble conglomerate overlain by fine-grained sandstone, calcareous siltstone, banded siltite, sandy limestone, carbonaceous siltstone and mudstone. The formation is subdivided into four informal members: (1) the basal conglomerate and overlying bioclastic limestone and sandstone; (2) a massive, structureless sandstone; (3) a sequence of banded, very fine grained sandstone and siltstone; (4) dark carbonaceous siltstone and mudstone. The Grand Prize Formation is interpreted as the deposits of subaqueous sediment gravity flows including turbidity flows, liquified sediment flows and minor debris flows, intercalated with hemipelagic and pelagic deposits. The composite stratigraphic section described here is an expansion of the type section described by Hall (1985), which did not include the basal conglomerate.

INTRODUCTION

Upper Paleozoic rocks in the Smoky and Boulder Mountains of south-central Idaho are contained in three tectonostratigraphic units: the Middle Pennsylvanian-Lower Permian Wood River Formation, the Lower Permian Grand Prize Formation, and the Lower Permian Dollarhide Formation. These Upper Paleozoic formations are allochthonous, are bounded by low-angle faults, and are arranged in imbricate structural stacks in the Smoky, Boulder, and Pioneer Mountains (Figs. 1 and 2). These partially coeval formations are lithologically similar, consisting predominantly of fine-grained calcareous sandstone, sandy limestone, calcareous banded siltstone, and dark carbonaceous siltstone and mudstone, with minor conglomerate and bioclastic limestone (Hall and others, 1974; Sengebush, 1984; Hall, 1985; Wavra and others, 1986; Geslin, 1986; Darling, 1988, this volume). The carbonaceous lithologies within these formations locally host epigenetic lead-silver mineralization and are included in the Idaho Black Shale Mineral Belt of Hall (1985) (Fig. 1).

The Upper Paleozoic stratigraphy of south-central Idaho is reviewed by Link and others (1987; 1988, this volume) and has only been studied in detail during the last five years. The Wood River and Dollarhide Formations have both been the subject of detailed stratigraphic investigations (Hall and others, 1974; Hall, 1985; Wavra and others, 1986; Geslin, 1986), but descriptions of the Grand Prize Formation have been brief or partial (Sengebush, 1984; Hall, 1985; Mahoney, 1987). In this paper we provide the first detailed description of the stratigraphy of the Lower Permian Grand Prize Formation.

The lithologic similarity and poorly constrained temporal overlap among the Wood River, Grand Prize and Dollarhide Formations suggests that they were deposited in laterally contiguous settings in the Middle Pennsylvanian-Permian Wood River basin (Mahoney and Link, 1987a, b; Link and others, 1988, this volume).

**GRAND PRIZE FORMATION**

**Structural Setting**

The Grand Prize Formation is exposed in a linear belt that extends north from Galena Summit to just north of the Salmon River (Figs. 1 and 2). The Grand Prize Formation forms the structurally highest allochthon in the Boulder and Smoky Mountains; the upper stratigraphic boundary of the formation is not recognized due to Eocene volcanic cover or removal by erosion (Hall, 1985). Where exposed, the base of the formation is a low-angle fault that separates the Grand Prize Formation from the underlying Wood River Formation in the southern portion of the outcrop belt, and from the Paleozoic Salmon River assemblage in the northern part of the outcrop belt (Hall, 1985). Hall (1985) interprets this low-angle fault to be a thrust fault.

**Previous Work**

Rocks now designated as the Grand Prize Formation were previously assigned to the Wood River Formation (Tschanz and others, 1986), the Pole Creek sequence (Skipp and Hall, 1980), and the Pole Creek formation (Fisher and others, 1983; Sengebush, 1984).
Hall (1985) named the Grand Prize Formation for a thick sequence of fine-grained quartzite, limy siltstone, banded siltite and dark carbonaceous silty limestone exposed on Pole Creek north of Galena Summit (Fig. 2). Hall (1985) measured a 1450 m partial section of the formation at its type locality near Grand Prize Gulch on the north side of Pole Creek (location 2 on Fig. 2, sections 2a and 2b of Table 1) and informally subdivided the formation into four members. The age of the formation is poorly constrained as Early Permian (Wolfcampian to Leonardian(?) based on conodonts collected from float at the base of the type section and on stretched and corroded conodonts recovered from limestone interbeds at Peach Creek, northeast of Stanley (Hall, 1985) (Fig. 2).

Sengebush (1984) examined the Grand Prize Formation (which was then designated as the Pole Creek formation) near Fourth of July Creek, north of the type section (location 3 on Fig. 2 and Table 1). He compiled a 2500 m composite section from three areas near the headwaters of Fourth of July Creek and subdivided the formation into four members, including a basal conglomerate that was not included in Hall's section. The Fourth of July Creek section is the most complete and is the only section to extend to the basal fault contact (Sengebush, 1984; Hall, 1985).

Mahoney (1987) measured a 1068 m partial section of the Grand Prize Formation near the headwaters of the Salmon River (section 1 of Fig. 2 and Table 1) and subdivided the formation into two units.

Stratigraphy

Geologic mapping, measured sections, and thin section analysis provide the basis for a stratigraphic description of the Grand Prize Formation. Four stratigraphic sections were measured; locations are given in Table 1 and are numerically indicated on Figure 2. Stratigraphic columns are given in Figure 3.

The most complete measured section is found along Fourth of July Creek in the Washington Peak 7.5-minute quadrangle (location 3 on Fig. 2 and Table 1). Partial sections were measured on the north side of Pole Creek, in the Horton Peak 7.5-minute quadrangle (type section; location 2 on Fig. 2 and sections 2a and 2b of Table 1) and to the west of the headwaters of the Salmon River, in the Frenchman Creek 7.5 minute quadrangle (location 1 on Fig. 2 and Table 1). In addition, detailed mapping by Mahoney in the Frenchman Creek and Horton Peak 7.5-minute quadrangles has documented lateral facies changes, such as variations in terrigenous elastic content within the Grand Prize Formation.

During reexamination of the type section, we identified a low-angle structural discordance, a brecciation zone, and abundant calcite-filled fractures approximately 650 meters above the base, suggesting that the section is cut by a low-angle fault that may repeat the section. This new interpretation has led us to treat the type section as two individual partial sections (sections 2a and 2b of Table 1). Work near Fourth of July Creek indicates that the Grand Prize Formation is much thicker than the 1450 meters measured at Pole Creek by Hall, and that the total thickness exceeds 2500 meters.

We therefore propose to establish a supplemental reference section for the Grand Prize Formation near Fourth of July Creek (section 3 of Table 1). This supplemental reference section is composite, with the basal 500 m (including the basal conglomerate recognized by Sengebush, 1984) located in Strawberry Basin north of the headwaters of Fourth of July Creek (Fig. 2). The remainder of the section is located on the ridge north of Fourth of July Creek, this portion of the section is overturned, with its base marked by a granitic intrusion and its top denoted by a low-angle fault contact with the Paleozoic Salmon River assemblage (Fig. 2).

In this report, the Grand Prize Formation is divided into four informal members that include: (1) a basal conglomerate with overlying bioclastic limestone and graded sandstone, (2) a massive structureless sandstone with minor siltstone laminae, (3) a banded very fine grained sandstone and siltstone unit with interbedded massive sandstones, and (4) a dark, carbonaceous

Table 1. Locations of measured sections of the Grand Prize Formation in the Challis and Hailey 1 x 2 degree quadrangles, central Idaho. See Figure 2 for map locations, and Figure 3 for graphic logs.

<table>
<thead>
<tr>
<th>Section</th>
<th>Location Details</th>
</tr>
</thead>
<tbody>
<tr>
<td>Section 1.</td>
<td>Salmon River Headwaters: measured on west side of Salmon River, on the east flank of peak 9423, 7800 to 9423 feet elevation; latitude 43°48'45&quot;N, longitude 114°46'33&quot;W, Frenchman Creek 7.5-minute quadrangle (Section 1 of Figure 3).</td>
</tr>
<tr>
<td>Section 2a.</td>
<td>Pole Creek section I, measured on north side of Pole Creek, near its confluence with Grand Prize Gulch, on south flank of peak 10166; 7970 to 9560 feet elevation; latitude 43°56'27&quot;N, longitude 114°40'50&quot;W; Horton Peak 7.5-minute quadrangle (Section 2a of Figure 3).</td>
</tr>
<tr>
<td>Section 2b.</td>
<td>Pole Creek section II, measured on north side of Pole Creek, near its confluence with Grand Prize Gulch, on south flank of peak 10166; 9600 to 10166 feet elevation; latitude 43°56'27&quot;N, longitude 114°40'50&quot;W; Horton Peak 7.5-minute quadrangle (Section 2b of Figure 3).</td>
</tr>
<tr>
<td>Section 3.</td>
<td>Fourth of July Creek section; composite section; 0-500 m interval measured in Strawberry Basin, north of Blackman Peak, 9800 to 10111 feet elevation; latitude 44°42'28&quot;N, longitude 114°39'32&quot;W, 500-2500 m interval measured from west to east on ridge north of Fourth of July Creek, 7800 to 9290 feet elevation; latitude 44°3'18&quot;N, longitude 114°43'48&quot;W; Obsidian and Washington Peak 7.5-minute quadrangles (Section 3 of Figure 3).</td>
</tr>
</tbody>
</table>
Figure 3. Stratigraphic columns of Grand Prize Formation measured sections. Section locations are given in Table 1.
siltstone and mudstone with interbedded graded sandstone to siltstone sequences. The section described here is composite and individual unit thicknesses vary laterally. A composite stratigraphic section with detailed lithologic descriptions is given in Figure 4.

Member One

Description

The lowest member of the Grand Prize Formation is about 600 m thick and consists of a basal chert pebble conglomerate, a thin black bioclastic limestone, and an overlying cyclic package of thin-bedded normally graded coarse- to fine-grained sandstone (Fig. 4). The stratigraphic section of this member is composite; a continuous section is nowhere exposed. In Strawberry Basin north of Fourth of July Creek, the basal conglomerate, black limestone and graded sandstone are well-exposed. The black limestone and overlying graded sandstone are exposed on the ridge north of Fourth of July Creek approximately 2.5 km to the west of Strawberry Basin, at the base of the overturned section (Fig. 2). The top of the member is truncated by a bedding-parallel fault. However, the bioclastic limestone serves as a marker bed that allows for correlation of the conglomerate and overlying graded sandstone in adjacent areas.

The clast-supported conglomerate varies from 170 to 800 m thick, and contains well-rounded to subangular elongate clasts of gray chert and minor siltite pebbles in a gray, sandy limestone to calcareous sandstone matrix. Lateral variations in thickness are attributed to both the original depositional pattern (i.e., channels or discontinuous sheets) and to structural deformation (Sengebush, 1984). The unit has been strongly deformed by movement along the regional low-angle fault separating the Grand Prize Formation from the underlying Salmon River assemblage (Hall, 1985). In rare undeformed outcrops, the pebbles are slightly imbricated and locally aligned with long axes subparallel to bedding. Minor interbeds of matrix-supported sandy limestone contain normally graded chert-pebble to granule clasts (Fig. 5a).

The basal conglomerate has been the subject of debate among workers in the region. Early workers assigned the conglomerate to the Wood River Formation, based on its similarity to the basal conglomerate (Hailey Conglomerate Member) of the Wood River Formation (Thomasson, 1959; Tchanz and others, 1986). Subsequent investigators believed the unit to be a tectonic breccia, with subsequent deformation along this depositional contact. The striking similarity between the basal conglomerates of the Grand Prize Formation and the Hailey Conglomerate Member suggests a genetic relationship between the two, and this similarity warrants further investigation.

Overlying the conglomerate with a sharp basal contact are approximately 10 m of black, bioclastic limestone. The limestone is structureless, organic-rich carbonate with locally abundant bioclastic debris (crinoid and pelecypod fragments). The limestone is overlain by about 420 m of cyclically graded sandstone beds, each 1 to 1.5 m thick. The vertical transition from a structureless coarse-grained sandstone at the base upward to finer-grained parallel laminated and cross laminated beds in each cycle is believed to correspond to the Tabc divisions of the classic Bouma sequence.

Although the upper contact of member 1 is not exposed, the member is inferred to have a gradational contact with overlying member 2, based on the presence of cyclically graded sandstone beds (member 1) intercalated with the massive sandstone of member 2 on the ridge west of Strawberry Basin, north of Fourth of July Creek (Sengebush, 1984) (Fig. 2).

Depositional Mechanism

Member 1 lithologies are interpreted as submarine mass-sediment gravity flow deposits. The presence of weak long-axis pebble imbrication, graded beds, and associated debris flow and turbidity current deposits (limestone and cyclic sandstones) suggests the thick basal conglomerate is the product of coarse clastic turbulent flow, probably involving bed load transport at the bases of high-concentration turbidity currents. Rapid deceleration following initiation of turbidity flows lowers the competence of the flows and results in deposition of coarse material while finer material is still being transported in turbulent suspension. This process creates thick clast-supported conglomerates with normally graded beds. Individual deposits combine to form thick successions of similar beds (Walker, 1975a; 1978; Howell and Normark, 1982). Such deposits are normally channelized; we believe that deformation of the unit has probably obscured such features.

The black bioclastic limestone of member 1 is interpreted as debris flow deposits, based on the matrix-supported fabric, the presence of disorganized bioclastic debris (fossil hash), and the structureless character of the deposit. Carbonate-bearing debris flows apparently interrupted the normal pattern of clastic deposition (represented by the conglomerate and overlying graded sandstone sequences) and were probably the result of catastrophic influxes of organic-rich carbonate mud and bioclastic debris from an adjacent carbonate depositional environment.
**Member Four:** Dominant lithology is dark gray to black, carbonaceous siltstone to mudstone, locally parallel laminated, containing interbeds of dark gray, calcareous, very fine-grained sandstone. Siltstone weathers reddish-brown and is thin-bedded (2-8 cm). Individual beds locally aggregate to 1-3 m thick units. The interbedded sandstone is thin-bedded (2-6 cm), normally graded, and contains parallel laminae, cross-laminae, and locally is convolute bedded. Sandstone/siltstone contact is undulose to planar; sandstone beds are locally discontinuous.

Subordinate lithologies include: 1) 10-35 m sequences of thin-bedded (5-15 cm) blue-gray, silty to sandy limestones with thin (3-5 cm) interbeds of black mudstone that display a differential weathering profile. Mudstone interbeds are more resistant than the easily weathered limestone. The limestone is parallel or convolute laminated, the mudstone is massive. Slump folds are common. 2) Graded sandstone to siltstone sequences up to 1 m thick, with well-developed ripple cross-beds, convolute laminae and parallel laminae. 3) Minor massive, carbonaceous limestone units up to 20 m thick near top of section.

**Member Three:** Dominant lithologies consist of thin-bedded couplets (15-45 cm) of light gray fine-grained sandstone and dark siltstone which give the unit a distinctly banded appearance. Each couplet is normally graded and contains: 1) a basal parallel laminated to structureless, very fine to fine-grained sandstone (5-20 cm) with a sharp planar to undulose lower contact with local scour features and flame structures, overlain by; 2) thin-bedded (5-15 cm) cross-laminated to convolute laminated very fine-grained sandstone; 3) parallel laminated very fine-grained sandstone to coarse siltstone; 4) thin-bedded (4-10 cm), black, massive to parallel laminated siltstone with a sharp upper contact with the basal sand of the overlying couplet. Each couplet is tabular and laterally continuous, couplets aggregate to 30-70 m banded sequences, with siltstone interbeds gradually increasing in thickness up section.

Intercalated with the banded lithofacies are thick interbeds of thick-bedded to massive, medium to fine-grained calcareous sandstone. The interbeds appear structureless, although faint parallel, convolute, and cross laminae are locally evident. Sandstone displays sharp basal and gradational upper contacts.

**Member Two:** Dominant lithology is thick-bedded to massive, fine to medium-grained calcareous sandstone with minor interbeds of thin (3-7 cm) black siltstone. Sedimentary structures are lacking, although faint parallel and convolute laminae, and dish structures are locally evident. Sparse graded beds occur, with the sandstone fining upward a few centimeters below the intercalated siltstones. Individual beds are tabular and laterally continuous. Siltstone beds are irregularly spaced, separated by 10-25 m of massive sandstone, and are thinly laminated, and locally intercalated with very fine grained sandstone laminae.

**Member One:** Strongly deformed basal granule and pebble conglomerate overlain by thin black limestone and a thick package of cyclic sequences of thin-bedded normally graded sandstone to siltstone. Conglomerate contains elongate clasts of gray chert and minor siltite, well rounded in part, in a gray sandy limestone to calcareous sandstone matrix; weak long-axis imbrication is locally evident. Minor interbeds of matrix-supported sandy limestone contain normally graded pebble to granule clasts.

Conglomerate is overlain by approximately 10 m of black, massive, organic rich limestone with locally abundant bioclastic debris (crinoid, pelecypod, bryozoan fragments); whole body fossils are absent. Limestone is overlain by cyclically graded sandstone beds (1-1.5 m), each containing: 1) a basal structureless coarse-grained bioclastic (fossil hash) sandstone bed (35-50 cm) with an undulose scoured base, overlain by; 2) planar laminated fine to medium-grained sandstone (15-30 cm), and, 3) ripple cross-laminated fine-grained sandstone (15-25 cm), overlain across a sharp contact with the basal sandstone of the overlying cycle.

---

*Figure 4. Composite stratigraphic section of the Grand Prize Formation with detailed lithologic descriptions. Explanation of symbols can be found in Figure 3.*
The cyclically graded coarse- to fine-grained sand beds overlying the limestone were probably deposited by turbidity flows, as indicated by the well-developed Tabc Bouma sequences. The upper portion (Tde) of each sequence is missing, probably as a result of nondeposition or erosion at the bases of overlying flow units.

The vertical succession of lithologies in member 1, primarily the transition from clast-supported conglomerate in the lower portion of the member to well-developed Bouma sequences in the upper portions, represents a general fining-upward sequence.

Member Two

Description

Member 2 consists of more than 750 m of thick-bedded to massive (>1 m bed thickness), fine- to medium-grained, apparently structureless calcareous sandstone with minor, irregularly spaced interbeds of thin (3-7 cm) black siltite (Fig. 4). Black siltite interbeds are commonly separated by 10-25 m of massive sandstone.

The sandstone of member 2 is well-exposed on the ridge north of Fourth of July Creek, and it forms the base of the type section north of Pole Creek (locations 2 and 3 on Fig. 2 and Table 1). The unit is widely exposed west of the headwaters of the Salmon River, although in this area it exhibits extensive fractures and is contact-metamorphosed to a noncalcareous quartzite (Mahoney, 1987; location 1 on Fig. 2). Between Fourth of July Creek and Pole Creek, the member is sporadically exposed and contains rare interbeds of thin-bedded (5-15 cm), parallel- to cross-laminated fine sandstone overlain by dark siltstone.

Member 2 has a conformable, gradational contact with the overlying member. The thin-bedded banded lithology of member 3 becomes intercalated with the massive sandstone of member 2 at the contact. This relation is well-exposed in both the lower Pole Creek and the Salmon River headwaters sections (locations 1 and 2 on Fig. 2).

Depositional Mechanism

The thick-bedded to massive character, apparently structureless texture, and homogeneity of the sandstone of member 2 suggest that the unit is the product of liquefied sediment gravity flows, with the interbedded siltite representing hemipelagic deposition during
episodes of quiescence or clastic nondeposition. Subsequent flows can occur when rapidly deposited, loosely packed sediment undergoes spontaneous liquefaction due to sudden loss of internal strength (Lowe, 1976). Triggering mechanisms include seismic shock or, more commonly, the failure of unstable slopes created by high sediment influx. The resulting flows transport large volumes of sand down submarine slopes as low as 3 degrees (Lowe, 1976). The resulting deposits are massive and structureless and contain dewatering features such as dish structures.

The rare interbedded parallel- to cross-laminated fine sandstone and overlying dark siltstone found in member 2 between Fourth of July Creek and Pole Creek are believed to be partial Bouma sequences (Tcde). These thin-bedded sequences could have formed by the initiation of turbidity flows following the beginning of deposition (settling) of the liquefied flows or by unrelated turbidity flows generated farther upslope (Blatt and others, 1980).

The lateral continuity, the lack of sedimentary structures, the unfossiliferous character and the presence of thin organic-rich siltite interbeds suggest that the sandstone of member 2 was deposited in a marine environment, below wave base, where the reworking of sediment by waves or currents was minimal, and where dysaerobic bottom conditions allowed for the preservation of organic material.

**Member Three**

**Description**

Member 3 consists of approximately 700 m of thin-bedded, fine-grained sandstone and siltstone interbedded with thick-bedded to massive calcareous sandstone (Fig. 4). Most of the unit comprises thin-bedded couplets of light colored fine- to very fine-grained sandstone and overlying dark colored, fine- to coarse-grained siltstone; these rhythmic interbeds give the unit a distinctly banded appearance (Hall, 1985; Mahoney, 1987).

Member 3 is well-exposed in both the type section on Pole Creek (location 2 on Fig. 2 and Table 1) and on the ridge west of the headwaters of the Salmon River (location 1 on Fig. 2 and Table 1). The percentages of massive sandstone and sandy limestone interbeds in the member are variable, with an increase in the amount of sandy interbeds and a corresponding decrease in the thickness of the thin-bedded banded lithology from the type section to north of Fourth of July Creek (Figs. 2 and 3).

Each couplet in the banded lithology is thin-bedded (15-45 cm), tabular, and laterally continuous, with sharp upper and lower boundaries (Fig. 4). In the type section (location 2 on Fig. 2), abundant soft-sediment deformation features, including convolute laminae, wavy to undulose contacts, and lenticular bedding of uncertain origin are evident in the banded lithology near the top of the member.

Intercalated with the thin-bedded banded lithology are thick sequences (5-50 m) of thick-bedded to massive, medium- to fine-grained calcareous sandstone. The sandstones are irregularly spaced and generally separated by 30-70 m of the banded lithology. Between these sandstone sequences, the thickness of the black siltstone interbeds increases upward within the banded lithology, with the siltstone reaching a maximum thickness of 12-15 cm immediately below the next intercalated massive sandstone.

Member 3 has a conformable, gradational contact with the overlying member. This contact is well-exposed in the type section (location 2 on Fig. 2), where the banded lithology of member 3 becomes much finer grained and higher in organic content, and develops abundant soft-sediment features and possible bioturbation features below its contact with the overlying mudstone of member 4. The contact is also well-exposed on the ridge north of Fourth of July Creek (location 3 on Fig. 2), but both members 3 and 4 at this location have higher sand contents than in areas farther south, and the contact is marked by an increase in blue gray sandy limestone and interbedded siltstone, and by a decrease in the amount of calcareous sandstone.

**Depositional Mechanism**

The banded couplets of member 3 apparently correspond to partial Bouma sequences, primarily divisions Td or Tcde and rarely Tbcde. The well-developed Bouma sequences, lateral continuity, and vertical repetition present in the thin-bedded banded lithology of member 3 are believed to be the result of accumulated low concentration turbidity flows, with each couplet representing an individual flow deposit. The intercalated thick-bedded sandstones are probably the result of periodic high-concentration turbidity flows, which deposit their coarse fraction first, while finer material remains in suspension and is transported farther downslope. This suggested mechanism is similar to that of the basal conglomerate, the primary difference being one of grain size.

The gradual upward thickening of black siltstone interbeds in the banded lithology between the intercalated sandstones was probably caused by migration of the primary turbidity flow depocenter away from the site of siltstone deposition. Migration of the depocenter would have resulted in a decrease in sediment supply and an increase in the importance of hemipelagic deposition. The intercalated sandstone beds represent either an increase in sediment supply from the source, or an abrupt return of the primary site of deposition due to topographic modifications elsewhere in the basin (e.g., distributary channel switching) (Mutti, 1977; J. D. Smith, oral communication 1987).
Member Four

Description

Member 4 is estimated to be at least 450 m thick. It consists of thin- to medium-bedded, carbonaceous siltstone and mudstone, interbedded blue-gray sandy limestone and thin black mudstone with abundant soft-sediment deformation features, and graded sandstone to siltstone sequences up to 1 m thick. The upper boundary of the member is a fault.

The amount of fine-grained terrigenous clastic material increases upsag in member 4 in the Fourth of July Creek section and in the upper portions of the sections on Pole Creek (Figs. 3 and 4). Graded, tan-colored, fine-grained sandstone to siltstone sequences up to a meter thick are interbedded with the darker, very fine-grained lithologies that dominate member 4. These sequences consist of a basal fine-grained structureless to planar laminated calcareous sandstone bed (20-40 cm), an overlying very fine-grained, ripple to parallel laminated sandstone bed (10-30 cm), and an upper parallel laminated black siltite bed (2-5 cm). These graded sequences represent well-developed Bouma sequences, including Tabedc divisions.

Depositional Mechanism

The lithologies present in member 4 are believed to be the product of low-concentration turbidity flows and hemipelagic and pelagic sedimentation. The dominant lithology, the carbonaceous siltstone with interbedded fine sandstone, probably represents partial Bouma sequences, primarily Tde, deposited by low-concentration turbidity flows in the waning stages of transport (Mutt, 1977; Howell and Normark, 1982). The sandy limestone contains Tcde divisions and was also probably deposited by low-concentration turbidity flows. The abundant soft-sediment deformation in this unit indicates very high sedimentation rates, probably on a slope. The graded sandstone to siltstone lithology was apparently deposited by turbidity flows and represents periodic resurgence of clastic sediment supply.

Lithologies of member 4 are generally finer grained, thinner bedded, and more carbonaceous than the underlying members. Member 4 is interpreted to represent a gradual cessation of terrigenous clastic influx and an increase in the influence of hemipelagic and pelagic sedimentation, as would occur in the transition from slope to basin depositional environments.

Tentative Correlation with Dollarhide Formation

The carbonaceous lithologies of member 4 are similar to lithologies in the upper portions of the coeval Lower Permian Dollarhide Formation and may represent an originally contiguous facies. Both member 4 and the upper member of the Dollarhide Formation are fine-grained, have high organic contents, and have variable amounts of terrigenous material (Wavra and others, 1986; Geslin, 1986; Link and others, 1987). The decrease in grain size and increase in organic content of member 4 is similar to the decrease in clastic influx and the establishment of basinal conditions that are evident in the upper member of the Dollarhide Formation. We therefore suggest that member 4 may have been deposited in an environment similar to that of the upper member of the Dollarhide Formation and may be correlative with it (Wavra and others, 1986; Link and others, 1987). However, definitive correlation is precluded at this time by the lack of biostratigraphic control.

Paleoslope Data

Member 4 of the Grand Prize Formation contains abundant soft-sediment deformation features, including a number of well-developed slump folds in the thin mudstone beds that are present throughout the member. The slump folds are best developed in the interbedded sandy limestone and thin mudstone lithology where the slump-folded mudstone beds are bracketed by initially horizontal, undeformed beds.

Twenty-five slump folds from both the Pole Creek and Fourth of July Creek sections were measured (locations 2 and 3 on Fig. 2). Figure 6 is a stereonet plot of the restored slump folds. The restored slump folds suggest an east-northeast-striking paleoslope which dipped to the south-southeast, based on the vergence of the folds. Slump fold analysis is best used to establish local paleoslopes in depositional basins. However, the consistency of our measurements at three stratigraphic sections through a vertical range of at least 300 meters suggests a consistent paleoslope during the deposition of member 4.

This paleoslope evidence conflicts with asymmetric megaripple measurements by Sengebush (1984), which indicate westerly current directions. The contradiction between slump fold and ripple measurements requires further investigation.

BASIN EVOLUTION

The Grand Prize Formation is interpreted as the product of combined sediment gravity flows (including both low and high concentration turbidity flows), liquefied sediment flows and minor debris flows, intercalated with hemipelagic and pelagic deposits. The overall vertical transition from basal conglomerate to massive, structureless sandstone to interbedded very fine sandstone and siltstone overlain by mudstone represents a distinct fining-upward sequence.

The predominance of gravity flows, particularly turbidity flows, the lack of fluvial or wave-generated
FUTURE INVESTIGATIONS

The greatest problem in the investigation of the Grand Prize Formation is the present lack of biostratigraphic control. Until accurate age dates are acquired, correlations of Upper Paleozoic strata in south-central Idaho will remain tentative.

Additional stratigraphic investigations of the Grand Prize Formation, particularly in the area north of Fourth of July Creek, are required to document lateral facies changes within the formation and to define the morphology of the depositional basin.

ACKNOWLEDGMENTS

The authors thank Ron Worl of the U. S. G. S. Branch of Central Mineral Resources for logistical support. Mahoney worked partly under the auspices of the U. S. Geological Survey's Hailey 1 x 2 degree CUSMAP project. This investigation is part of Mahoney's M. S. research at Idaho State University under the supervision of Dr. P. K. Link, and is part of Sengebush's M. S. research at the University of Montana under the supervision of Drs. Jim Sears and Johnnie Moore. Mahoney thanks the folks at ISU for their assistance, both donated and coerced. The manuscript was reviewed by B. R. Burton and R. Q. Oaks, Jr.

REFERENCES


Geslin, J. K., 1986, The Permian Dollarhide Formation and Paleozoic Carrietown Sequence in the SW 1/4 of the Buttercup Mountain quadrangle, Blaine and Camas
Mahoney and Sengebush--Stratigraphy of the Lower Pennian Grand Prize Formation, South-Central Idaho 179


Chapter Three
Economic Geology

The Empire Mine near Mackay Peak, west of Mackay, Idaho. The Lost River Range forms the distant skyline. Photograph by W. R. Hackett.
Extensive areas of jasperoid occur in Paleozoic sedimentary rocks near Mackay, Idaho. Outcrops range from massive, gray-pink colored, hard and dense jasperoid with original sedimentary textures preserved to highly brecciated, quartz-healed, gossan-stained jasperoid showing no sign of its protolith. The Mackay jasperoids are significant because they have geological and geochemical similarities to jasperoids associated with known sediment-hosted precious-metal deposits in the western United States. The jasperoids of the Mackay area 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. Quartz and calcite veins cut some of the jasperoids, and they locally contain barite, fluorite, and antimony minerals. Iron-stained outcrops of jasperoid are common and many contain jarosite.

One hundred ten rock chip samples collected from jasperoid outcrops in five geographic areas of the Mackay region were analyzed for eighteen elements: Ag, As, Au, Cu, Hg, Mo, Pb, Sb, Tl, Zn, Bi, Cd, Ga, Pd, Pt, Se, Sn, and Te. Most samples from all five areas contained Ag, As, Au, Cu, Hg, Mo, Pb, Sb, and Zn; Tl was present in most samples from four of the five areas. Element associations determined by correlation analysis centered on As-Sb-Tl-Hg and Cu-Zn. Copper, Mo, Pb, Ag, and Zn may have been locally derived from metalliferous sedimentary rocks and could have been deposited under almost any conditions of jasperoid formation. However, the elements As, Sb, and Hg with Tl and Au suggest jasperoid formation by replacement from hydrothermal solutions, since these elements are characteristically deposited in the upper parts of geothermal systems. Hydrothermal solutions that formed the jasperoids may also have formed large low-grade precious metal deposits within the jasperoid bodies or within altered country rocks associated with the jasperoids.

INTRODUCTION

This report describes the physical and chemical properties of jasperoid rocks near Mackay and is based on reconnaissance sampling during July and August 1987. This investigation is part of a U. S. Geological Survey CUSMAP project to evaluate the mineral resources of the Hailey 1 x 2 degree quadrangle and part of the Idaho Falls 1 x 2 degree quadrangle. An Idaho State University project studying gold and silver mineralization in central Idaho is cooperating with the Hailey project and providing the geochemical analyses of samples. The potential...
for large low-grade precious metal deposits in the volcanic rocks is discussed by Moye and others (1988, this volume).

The study area is near the town of Mackay in the northwest corner of the Idaho Falls 1 x 2 degree quadrangle, west of U. S. Highway 93 and north of the Pioneer Mountains (Fig. 1). One hundred ten rock chip samples from jasperoid outcrops were analyzed to determine the presence and distribution of gold and other trace elements, such as As, Sb, Hg, and Tl. For descriptive purposes, samples were grouped according to location and local geologic features into five areas: Bartlett Point, Lehman Butte, Sheep Canyon, Grouse, and Timbered Dome (Fig. 1).

Jasperoid, as used in this paper, refers to “an epigenetic siliceous replacement of a previously lithified host rock” (Lovering, 1972, p. 3). Jasperoid bodies near Mackay are geologically and geochemically similar to those associated with large low-grade, sediment-hosted, precious metal deposits elsewhere in the western United States; as such, they should be of considerable interest to the mineral exploration industry. These deposits are sometimes termed “invisible gold” or Carlin-type deposits.

GEOLOGIC SETTING

The geology of the Mackay region was previously mapped by Nelson and Ross (1968; 1969). Part of the area is currently being mapped by Betty Skipp of the U.S. Geological Survey, and her unpublished mapping was used to locate jasperoid bodies near Grouse and Timbered Dome. Figure 2 is a generalized geologic map of the area.

Geologic units of the Mackay area are mainly Paleozoic carbonate rocks, intruded by Tertiary granitic stocks and hypabyssal bodies and overlain by the Eocene Challis Volcanics. The area is cut by high-angle faults that trend north-northwest and northeast.

Paleozoic sedimentary rocks that crop out in the area are Early Mississippian to Early Permian in age and consist of the White Knob Limestone and the Copper Basin, McGowan Creek, Middle Canyon, Scott Peak, South Creek, Surret Canyon, Bluebird Mountain, and Snaky Canyon Formations. The Copper Basin Formation (Mississippian) is interpreted to be a foreland-basin flysch, made up of a thick sequence of clastic rocks, ranging from shale and argillaceous siltstone to conglomerate (Paul1 and others, 1972; Nilsen, 1977; Skipp and others, 1979). The McGowan Creek Formation (Lower Mississippian) was deposited in the shallow, eastern part of the foreland basin and consists of distal thin-bedded turbidites, calcareous siltstone, and minor silty limestone. The Mississippian White Knob Limestone consists of limestone with interbedded conglomerate and sandstone in its upper part; it gradationally overlies the McGowan Creek Formation and represents deposition in a subtidal to intertidal turbulent marine environment (Skipp and others, 1979). The Lower and Upper Mississippian Middle Canyon Formation and the Upper Mississippian Scott Peak, South Creek, and Surret Canyon Formations are carbonate bank and forebank deposits across which a flood of fine-grained sand, the Bluebird Mountain Formation, transgressed in latest Mississippian time (Skipp and others, 1979). The Snaky Canyon Formation (Upper Mississippian to Lower Permian), mainly limestone and sandstone, is a shallow-water carbonate bank. For more discussion of these strata and their structural relations see Link and others (1988, this volume).

Extrusive rock units are part of the Eocene Challis Volcanics and include andesitic to rhyolitic flows, breccias, and tuffs. Possible vent areas for the volcanics are near Porphyry Peak and Sheep Mountain where aeromagnetic data suggest buried intrusive bodies. For
Figure 2. Generalized geologic map of the Mackay region, Idaho (adapted from unpublished map of Worl and others). Blackened areas indicate jasperoid. Qa, alluvium; Qb, basalt; Ti, intrusive rocks; Te, extrusive rocks, predominantly Challis Volcanics; PDC, undifferentiated carbonate rocks, including Snaky Canyon, Bluebird Mountain, Surrett Canyon, South Creek, Scott Peak and Middle Canyon Formations, White Knob Limestone and local, unnamed Permian and Devonian carbonate rocks; Mf, flysch deposits including Copper Basin and McGowan Creek Formations, undifferentiated.
more discussion of Challis Volcanics see Moye and others (1988, this volume).

Intrusive rock units are Tertiary plutons and comagmatic hypabyssal bodies. Plutonic bodies, exposed in the White Knob Mountains and along the southeastern and western margins of Copper Basin (Fig. 2), are multiphase granitic intrusions composed mainly of porphyritic granite, granodiorite, quartz diorite, leucogranite porphyry (Nelson and Ross, 1968 and 1969a and b) and quartz monzonite (Dovcr, 1981). Comagmatic hypabyssal bodies and zones of alteration are common in some areas, aligned in part along northeast- and northwest-trending fracture systems. Numerous dikes of quartz latite, rhyolite, and porphyritic rhyolite cut all rock types and are commonly found in swarms parallel to northwest- and northeast-trending fracture systems (Nelson and Ross, 1969a and b).

MINERAL DEPOSITS

Three major mining districts, Alder Creek, Copper Basin, and Lava Creek, and numerous other prospects in the Mackay area (Fig. 2). Most of the deposits were discovered during the late 1800s when rich silver-lead ores were mined throughout central Idaho. Most production of the late 1800s and early 1900s was from oxidized silver-lead ore. The Empire Mine in the Alder Creek district, active intermittently from 1901 into the 1960s, was a major copper producer. Many other properties in the area have been active intermittently since their discovery, but did not produce substantial amounts of ore.

Three types of mineral deposits have been exploited in the Mackay area: skarn deposits, polymetallic veins in Paleozoic sedimentary rocks, and polymetallic veins in volcanic rocks. The skarn deposits are mainly in the White Knob Mountains where Tertiary leucogranite intrudes Paleozoic limestone. Copper was the main commodity mined, but the deposits also contained lead, zinc, silver, gold, tungsten, and molybdenum (Nelson and Ross, 1969a and b). The deposits consist of chalcopyrite, pyrite, pyrrhotite, calcite, quartz, magnetite, fluorite, scheelite, molybdenite, sphalerite, and specularite in addition to skarn silicate minerals (Upleby, 1917).

Polymetallic fissure veins in volcanic rocks are found mainly in the Champagne Creek area of the Lava Creek District. These are epithermal deposits characterized by complex mineralogy and significant amounts of lead, silver, zinc, copper, iron, gold, tungsten, tin, antimony, arsenic, and bismuth (Anderson, 1947). The orebodies are within highly silicified and pyritized zones in a broad area of altered volcanic rock. The veins occupy northeast- and northwest-trending fracture systems.

Mining deposits in the Mackay area, thought to be Tertiary in age, are the result of either contact metamorphism and metasomatism or deposition from hydrothermal solutions. Metals in these deposits may have been in part remobilized out of Paleozoic sedimentary country rocks. During Paleozoic time, there may have been metal-enriched deposit in deep restricted marine basins resulting in metal-enriched sedimentary rocks that would now be part of the Copper Basin and McGowan Creek Formations. The chemistry of these units is not known, but recent biogeochemical and geochemical studies by Erdman and others (1988) suggest that the McGowan Creek Formation is locally metal-enriched. Magmatic and hydrothermal activity during Tertiary time could have remobilized these metals into the present deposits. However, some metals were probably transported in magma along with volatile elements such as fluorine. Hydrothermal systems operating during the late stages of volcanism and intrusion probably formed the polymetallic vein deposits that are common in the area. These hydrothermal systems were widespread, affecting all rock types throughout the region. They may also have formed some of the jasperoid bodies in the area, especially those that are aligned along high-angle fault systems.

JASPEROID LITHOLOGY

Jasperoids of the Mackay region are mainly in Paleozoic limestones near either high-angle faults or volcanic rocks, but a few occur in sandstone, mudstone, conglomerate, and the Challis Volcanics. Major fault systems through the host rocks probably acted as conduits for silica-bearing solutions that replaced the host rocks with silica. These solutions may also have been metal-bearing in which case there could be large low-grade precious-metal deposits within or near the jasperoids. For this reason the physical and chemical properties of both the jasperoids and the original host rocks are important.

Jasperoids and jasperoid breccias at five areas are described (Figs. 1 and 2): (1) Bartlett Point, (2) Lehman Butte, (3) Sheep Canyon, (4) Girrune, and (5) Timbered Dome. Geologic settings and physical properties of the jasperoids are slightly different at each of the five areas.

Bartlett Point is about 20 miles northwest of Mackay at the mouth of Rock Creek. In this area, Paleozoic sedimentary rocks are gently folded along northwest-
trending axes and are overlain by andesitic flows and breccias of the Challis Volcanics. Faults cutting all rock types trend north-northwest and northeast. Detailed maps are not yet available for this area, but general descriptions of the rocks are found in Nelson and Ross (1968, 1969a and b). The White Knob Limestone consists of gray to black, medium-bedded, silty limestone containing chert nodules and lenses. The upper part of the limestone contains interbedded quartzite clastic rocks. The sedimentary rocks are overlain by The Eocene Challis Volcanics, which are mainly brownish red andesite lava flows. Massive, multi-colored jasperoid bodies are widespread at Bartlett Point where they form outcrops of dense, flinty, massive silica and silicified breccias that show slickensides and more than one period of brecciation. Jasperoid bodies in Rock Creek appear to be aligned along northeast-trending fracture zones.

Lehman Butte, approximately 5 miles southeast of Bartlett Point, is a large outcrop of White Knob Limestone and possibly Snaky Canyon Formation, surrounded by the Eocene Challis Volcanics. The White Knob Limestone is massive and light to medium gray in color, and it contains dark gray chert nodules and many calcite veins. The limestone is very fractured and jointed, and holes formed from weathering are common. Silicified brecciated limestone crops out near the summit. Thin-bedded gray to pink limestone with siltstone interbeds occurs at the summit and most of this rock is silicified. Challis Volcanics near Lehman Butte consist of medium gray to reddish brown, porphyritic, andesitic lava flows and breccias. A northwest-trending fault cuts the volcanic rocks on the west side of Lehman Butte.

Sheep Canyon is approximately 25 miles southeast of Lehman Butte. Here, Challis Volcanics were extruded over Mississippian limestones of the carbonate bank sequence. Gray to black jasperoid bodies are locally brecciated or iron stained, and many outcrops contain silicified crinoidal fragments. Some outcrops are only partly replaced by jasperoid. There is no record of production from the mine workings (one shaft and an adit) at the mouth of the canyon.

Samples of jasperoid were collected along ridges to the north of the small community of Grouse, approximately 6 miles northwest of Timbered Dome. Rocks in this area belong to the McGowan Creek, Middle Canyon, Scott Peak, South Creek, Surrett Canyon, Bluebird Mountain, and Snaky Canyon Formations. The McGowan Creek Formation is grayish black argillite and siltite with minor thin- to medium-bedded, dark gray limestone. The Middle Canyon Formation is dark- to medium-gray, thin-bedded silty limestone. Cliffs and ledges are light gray calcareous and fine-grained, dark gray, fossiliferous cherty limestone of the Scott Peak Formation. The Scott Peak is overlain by thin-bedded, dark gray limestone containing incipient chert nodules alternating with clayey, fissile limestone of the South Creek Formation. The Surrett Canyon Formation above the South Creek consists of dark- to medium-gray, fine-grained, thick-bedded fossiliferous limestone. It is overlain by 15 to 20 feet of very thin-bedded, fine-grained, calcareous sandstone of the Bluebird Mountain Formation. The Snaky Canyon Formation at the top of the sequence consists of thick-bedded, sandy limestone overlain by argillaceous silty limestone, argillaceous limestone, siltstone and sandstone.

These formations are cut by high-angle, north- and east-trending faults, and the rocks are locally silicified to jasperoid both along and away from faults. The jasperoid bodies locally retain original depositional features of the carbonate-bank host formations. Beds vary from massive, siliceous limestone with crinoid stems to pure quartzitic sandstone. Most of the outcrops of jasperoid are very fractured and iron stained; some are brecciated locally. Jasperoids that are not fractured are not iron stained and are gray to grayish pink in color.

At Timbered Dome several high-angle, mainly north- and northeast-trending faults cut outcrops of the McGowan Creek, Middle Canyon, and Scott Peak Formations. Challis Volcanics also crop out in the vicinity. At the summit of Timbered Dome, the McGowan Creek and Middle Canyon Formations have been silicified to maroon- to pink-colored jasperoid and jasperoid breccia.

GEOCHEMISTRY

Analytical Methods

One hundred ten rock-chip samples of jasperoid were collected in the Mackay area from outcrops and mine dumps. Ninety-seven samples were analyzed for eighteen elements using inductively coupled plasma emission spectrography and graphite furnace atomic absorption methods by Geochemical Services, Inc., of Sparks, Nevada. Thirteen samples were analyzed by atomic absorption for Ag, As, Au, Hg and Sb by Skyline Labs, Inc., of Denver, Colorado. (Use of brand names in this report is only for descriptive purposes and does not constitute endorsement by the U. S. Geological Survey.)

All of the data were analyzed by correlation analysis using the USGS STATPAC statistical computer package to determine element associations. Because the data are quantitative and more than 50 percent of the values are unqualified, original units were used and no substitutions were made for qualified data. As the data are insufficient, it was not intended that this method produce statistically rigorous correlations. The data are too sparse and incomplete to allow valid groupings based on factor analysis. Rather, the main purpose was to determine element associations specific to each of the five geographical regions.

Results

A summary of analytical results from the entire data
set is given in Table 1. Data for individual geographic areas are given in Table 2. Ag, As, Au, Cu, Hg, Mo, Pb, Sb, and Zn were detected in more than 84 percent and Tl in more than 50 percent of the samples. Only a few samples had detectable amounts of Bi, Cd, Pt, Se, Sn, and Te (Table 2); those that did were anomalously high in at least one of the other metals. Element associations shown by correlation coefficients are As-Sb-Tl, and Cu-Zn (Table 3).

Discussion

The suite of elements As, Au, Hg, Sb and Tl is present in a majority of the samples. These elements are characteristic of the upper parts of geothermal systems, hot-spring-type precious-metal deposits, and sediment-hosted, disseminated precious-metal deposits. (Barium, another element of this suite, was not determined.) It should also be noted that the ranges in values of Ag, As, Cu, Hg, Pb, Sb, Tl and Zn found in this study are similar to those of some known mineralized (>0.30 ppm gold) areas reported by Bagby and Berger (1985, p. 189).

Detailed geochemical results from each of the five sample areas are discussed below.

Bartlett Point

Thirty-four rock samples were collected in the Bartlett Point area. The samples vary widely from nearly unaltered, gray, massive limestone to oxidized, silicified tectonic breccia. Correlation analysis of sample chemistry from the Bartlett Point area shows a very strong As-Sb-Hg-Tl association; Mo has a lower correlation coefficient and may also be included. Zinc and copper also show an association (Table 3). The highest values for arsenic come from the more oxidized breccias. In general, concentrations of As, Cu, Mo, Sb, Tl, Zn, Cd, Se, Sn, and Te are higher at Bartlett Point proper than in samples immediately to the south in Rock Creek. Samples from Bartlett Point proper gave the highest recorded values for As, Cu, and Sb of all five areas. Samples from Rock Creek tend to have greater amounts of Ag, Au and Pb, and gave the highest recorded values for Ag and Au of all the areas. In Rock Creek the samples were from silicified limestone and conglomerate of the White Knob Limestone. Most samples were not iron stained or brecciated, but jarosite, barite and fluorite were identified in one oxidized and brecciated zone.

Lehman Butte

Fourteen rock samples taken from Lehman Butte are silicified jasperoid breccia from the White Knob Limestone on the east side of a northwest-trending fault. Samples taken on a ridge east of Lehman Creek are from oxidized jasperoid outcrops in highly altered limestone, near a northwest-trending fault. Where this fault cuts the overlying Challis Volcanics, the rocks are brecciated and highly silicified, and the zone has been locally prospected for precious metals. Correlation analysis of the data suggests three general groupings: As-Hg-Sb-Zn, Ag-Au-Cu-Hg, and Cu-Mo-Pb, all with high correlation coefficients (Table 3). The low number of samples (14) dictates a low confidence level for these groupings. However, at Lehman Butte a large number of samples contain the element suite Ag-As-Au-Hg-Sb-Tl in significant amounts (Table 2).

Sheep Canyon

Twenty samples were analyzed from Sheep Canyon. Most samples were massive jasperoid in carbonate rocks with some healed jasperoid breccia with cross-cutting quartz veins. Gold-silver is the only strong geochemical association (Table 3). The data also show associations of

<table>
<thead>
<tr>
<th>Element</th>
<th>Min</th>
<th>Max</th>
<th>Mean</th>
<th>S.D.</th>
<th>N</th>
<th>Lim-1</th>
<th>Lim-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ag</td>
<td>0.013</td>
<td>16.50</td>
<td>0.40</td>
<td>1.68</td>
<td>105</td>
<td>0.015</td>
<td>0.05</td>
</tr>
<tr>
<td>As</td>
<td>1.10</td>
<td>1306.00</td>
<td>134.90</td>
<td>187.60</td>
<td>110</td>
<td>1.0</td>
<td>2.0</td>
</tr>
<tr>
<td>Au</td>
<td>&lt;0.001</td>
<td>0.16</td>
<td>0.01</td>
<td>0.02</td>
<td>99</td>
<td>0.0005</td>
<td>0.002</td>
</tr>
<tr>
<td>Cu</td>
<td>1.24</td>
<td>127.00</td>
<td>10.51</td>
<td>17.46</td>
<td>97*</td>
<td>0.025</td>
<td>-</td>
</tr>
<tr>
<td>Hg</td>
<td>0.04</td>
<td>54.00</td>
<td>1.64</td>
<td>5.19</td>
<td>95</td>
<td>0.10</td>
<td>0.01</td>
</tr>
<tr>
<td>Mo</td>
<td>0.34</td>
<td>121.00</td>
<td>10.40</td>
<td>17.58</td>
<td>97*</td>
<td>0.10</td>
<td>-</td>
</tr>
<tr>
<td>Pb</td>
<td>0.62</td>
<td>452.00</td>
<td>13.42</td>
<td>47.33</td>
<td>97*</td>
<td>0.25</td>
<td>-</td>
</tr>
<tr>
<td>Sb</td>
<td>&lt;0.25</td>
<td>135.00</td>
<td>11.45</td>
<td>16.80</td>
<td>105</td>
<td>0.25</td>
<td>1.0</td>
</tr>
<tr>
<td>Tl</td>
<td>&lt;0.44</td>
<td>17.70</td>
<td>1.08</td>
<td>1.93</td>
<td>58*</td>
<td>0.50</td>
<td>-</td>
</tr>
<tr>
<td>Zn</td>
<td>&lt;0.89</td>
<td>586.00</td>
<td>26.63</td>
<td>62.41</td>
<td>93*</td>
<td>1.0</td>
<td>-</td>
</tr>
</tbody>
</table>

*13 samples from the Grouse area were not analyzed for Cu, Mo, Pb, Tl, or Zn.
As-Mo-Zn and Hg-Mo-Tl, but the importance of the data is the presence of the suite Ag-As-Au-Hg-Sb-Tl in a majority of the samples (Table 2). A small prospect at the eastern end of the canyon contains antimony minerals including stibiconite. A sample collected in the western part of the area had the highest concentration of molybdenum (121 ppm) of any sample in the study.

Grouse

Data from thirty-two samples collected near the town of Grouse show a very strong Ag-Pb association, a strong Au-Hg association, a strong As-Sb-Tl association and a Cu-Mo-Zn association. The last two associations are linked by a moderate correlation between Cu and Tl. The data are not adequate to confidently define these associations because there appear to be three distinct groups of samples based on geography and local geology. Samples with the highest values for Ag-Pb (0.40-2.15 ppm Ag, 100-452 ppm Pb) are generally located at the southern end of a large jasperoid body. These samples also have 0.30-1.35 ppm Bi. Jaspereid outcrops in this area are silicified McGowan Creek, Middle Canyon, and Scott Peak Formations. Samples vary from dense, gray, nearly cryptocrystalline jasperoid to coarse-grained, maroon, iron-oxide stained jasperoid to brecciated jasperoid. Samples from the northern end of the same jasperoid body had the highest values of Cu and Mo in the Grouse area, and Ga was detected in every sample. Mineralization of the northern group was probably controlled by a north- to northwest-trending fault, and the jasperoid appears to have replaced rocks of the Snaky Canyon or Surrret Canyon Formations. Most of the samples were collected from gray, silicified limestone or conglomerate with maroon iron-oxide staining. One sample, from an outcrop of jasperoid breccia, is the least mineralized for this area. The third group of samples for this area was collected near Waddoups Canyon. These samples are mostly gray, brecciated jasperoid that replaced rocks of the Snaky Canyon Formation. Samples collected near Waddoups Canyon yielded the highest gold concentrations (0.01 to 0.03 ppm) for the Grouse area, and one sample yielded the highest value for mercury (54 ppm) in the Mackay area.

Timbered Dome

Correlation analysis of the ten samples collected from jasperoid outcrops near Timbered Dome shows a strong element grouping that includes Ag, As, Au, Cu, Hg, Pb, Sb, and Zn (Table 3). The small number of samples dictates that a low level of confidence be assigned to this association. The importance of the data is that the element suite Ag-As-Au-Hg-Sb is present in most samples. Only one sample contained detectable Tl.

The jasperoid bodies at Timbered Dome are within the McGowan Creek and Middle Canyon Formations and apparently aligned along north- to northeast-trending
Table 3. Correlation coefficients for pairs of significant elements, listed according to the five geographic areas that were sampled near Mackay, Idaho.

<table>
<thead>
<tr>
<th>Element</th>
<th>All 110 Samples</th>
<th>Bartlett Point 34 samples</th>
<th>Lehman Butte 14 samples</th>
<th>Sheep Creek 20 samples</th>
<th>Grouse area (32 samples)</th>
<th>Timbered Dome (10 samples)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ag</td>
<td>1.00</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>As</td>
<td>-0.04</td>
<td>1.00</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Au</td>
<td>0.19</td>
<td>0.11</td>
<td>1.00</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cu</td>
<td>0.08</td>
<td>0.15</td>
<td>0.29</td>
<td>1.00</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hg</td>
<td>0.01</td>
<td>0.11</td>
<td>-0.14</td>
<td>-0.01</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mo</td>
<td>-0.01</td>
<td>0.23</td>
<td>0.07</td>
<td>0.08</td>
<td>0.03</td>
<td>1.00</td>
</tr>
<tr>
<td>Pb</td>
<td>0.11</td>
<td>-0.03</td>
<td>0.05</td>
<td>0.01</td>
<td>-0.03</td>
<td>0.01</td>
</tr>
<tr>
<td>Sb</td>
<td>0.04</td>
<td>0.81</td>
<td>0.12</td>
<td>0.09</td>
<td>0.26</td>
<td>0.00</td>
</tr>
<tr>
<td>Ti</td>
<td>-0.03</td>
<td>0.42</td>
<td>-0.05</td>
<td>0.06</td>
<td>0.25</td>
<td>0.15</td>
</tr>
<tr>
<td>Zn</td>
<td>0.01</td>
<td>0.07</td>
<td>-0.02</td>
<td>0.43</td>
<td>0.07</td>
<td>0.14</td>
</tr>
<tr>
<td>As</td>
<td>0.01</td>
<td>0.17</td>
<td>0.07</td>
<td>0.17</td>
<td>0.27</td>
<td>0.13</td>
</tr>
<tr>
<td>Au</td>
<td>0.09</td>
<td>0.19</td>
<td>0.03</td>
<td>0.05</td>
<td>0.19</td>
<td>0.01</td>
</tr>
<tr>
<td>Cu</td>
<td>0.08</td>
<td>0.19</td>
<td>0.03</td>
<td>0.05</td>
<td>0.21</td>
<td>0.02</td>
</tr>
<tr>
<td>Hg</td>
<td>0.03</td>
<td>0.59</td>
<td>0.08</td>
<td>0.59</td>
<td>0.70</td>
<td>0.00</td>
</tr>
<tr>
<td>Mo</td>
<td>0.09</td>
<td>0.16</td>
<td>0.03</td>
<td>0.16</td>
<td>0.23</td>
<td>0.31</td>
</tr>
<tr>
<td>Pb</td>
<td>0.08</td>
<td>0.16</td>
<td>0.04</td>
<td>0.16</td>
<td>0.33</td>
<td>0.40</td>
</tr>
<tr>
<td>Sb</td>
<td>0.16</td>
<td>0.17</td>
<td>0.25</td>
<td>0.17</td>
<td>0.29</td>
<td>0.24</td>
</tr>
<tr>
<td>Ti</td>
<td>0.10</td>
<td>0.58</td>
<td>0.74</td>
<td>0.58</td>
<td>0.58</td>
<td>0.72</td>
</tr>
<tr>
<td>Zn</td>
<td>0.03</td>
<td>0.64</td>
<td>0.08</td>
<td>0.64</td>
<td>0.83</td>
<td>0.04</td>
</tr>
<tr>
<td>Ag</td>
<td>0.05</td>
<td>0.08</td>
<td>0.05</td>
<td>0.08</td>
<td>0.03</td>
<td>0.01</td>
</tr>
<tr>
<td>As</td>
<td>0.16</td>
<td>0.17</td>
<td>0.04</td>
<td>0.17</td>
<td>0.23</td>
<td>0.01</td>
</tr>
<tr>
<td>Au</td>
<td>0.23</td>
<td>0.28</td>
<td>1.00</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cu</td>
<td>0.99</td>
<td>0.90</td>
<td>0.95</td>
<td>1.00</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hg</td>
<td>0.92</td>
<td>0.90</td>
<td>0.95</td>
<td>1.00</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mo</td>
<td>-0.13</td>
<td>0.57</td>
<td>0.06</td>
<td>0.11</td>
<td>1.00</td>
<td></td>
</tr>
<tr>
<td>Pb</td>
<td>0.85</td>
<td>0.63</td>
<td>0.79</td>
<td>0.87</td>
<td>0.95</td>
<td>1.00</td>
</tr>
<tr>
<td>Sb</td>
<td>0.60</td>
<td>0.61</td>
<td>0.65</td>
<td>0.85</td>
<td>0.40</td>
<td></td>
</tr>
<tr>
<td>Ti</td>
<td>0.04</td>
<td>0.07</td>
<td>0.26</td>
<td>0.08</td>
<td>0.32</td>
<td></td>
</tr>
<tr>
<td>Zn</td>
<td>0.18</td>
<td>0.19</td>
<td>0.07</td>
<td>0.17</td>
<td>0.25</td>
<td></td>
</tr>
<tr>
<td>Ag</td>
<td>0.10</td>
<td>0.17</td>
<td>0.02</td>
<td>0.17</td>
<td>0.25</td>
<td></td>
</tr>
</tbody>
</table>

**Summary and Conclusions**

Many jasperoid bodies in the Mackay area are similar to jasperoids in the sediment-hosted precious-metal deposits of Nevada. Mackay jasperoids are replacements of limestone, shale, sandstone, carbonaceous limestone, and calcareous carbonaceous shales. The degree of silicification varies from patchy, silicified areas to dense, massive jasperoid. Many jasperoid bodies show at least two generations of brecciation and silicification. Quartz and calcite veins are common, and minor barite, fluorite, jarosite, and stibiconite are present.

Jasperoid bodies are commonly aligned along high-angle regional faults and were probably formed from solutions moving along and outward from these faults. The volcanic rocks are brecciated and silicified along faults, and commonly have quartz and carbonate veins. Intermediate to silicic plutonic and hypabyssal intrusive rocks are locally present in the area along high-angle fault systems.

Geochemistry of jasperoids in the Mackay area is characterized by the presence of Ag, As, Au, Cu, Hg, Mo, Pb, Sb and Zn in most samples from all five geographic areas. Thallium is present in most samples from four of the five areas.

The trace-element association As-Au-Sb-Hg-Tl occurs in nearly all sediment hosted, disseminated precious-metal deposits (Bagby and Berger, 1985, p. 194), and these elements also tend to concentrate in the upper parts of geothermal systems (Silberman and Berger, 1983, p. 206). Hot springs type mineral deposits probably all contain the same suite of elements, plus Mo (Berger and Silberman, 1985, p. 246).

Correlation analysis (Table 3) for all samples defines two strong associations: As-Sb-Hg-Tl and Cu-Zn. Associations within each of the five geographic areas are not well-defined, because of inadequate sampling. Most of the within-area associations include As, Sb, Hg and Tl, and commonly Au, Ag, Mo, Cu, or Zn. There is a suggestion that base-metals, especially Cu and Zn, form their own associations, and rarely occur with Ag and Mo. Element associations among the areas are probably due to several factors including lithology and chemistry of the host rocks, source and timing of the mineralizing.
solutions, distance from the source, pressure-temperature evolution of the fluids, and structural setting.

The genesis of jasperoid bodies in the Mackay area is not yet well understood. Nelson and Ross (1968, p. A22) suggest that some jasperoids formed from fluids migrating downward from the Challis Volcanics into underlying limestones. Betty Skipp suggests that some jasperoid may have been formed by meteoric water in karst zones as well as in fault zones. Detailed studies of the geochemistry, alteration, lithology, structure, and isotopic composition of the country rocks near jasperoids and high-angle faults are needed to determine the origin of each jasperoid body.

The suite of elements Ag, Au, As, Hg, Sb and Tl are consistently present, commonly in anomalous concentrations. This suggests that some, if not most, jasperoids of the Mackay area formed through the replacement of sedimentary rocks by silica and associated elements that were carried in hydrothermal solutions.

Trace metals in jasperoid, together with known vein-type metal deposits along high-angle faults indicate that some of the solutions were metal-bearing. These solutions may also have formed precious-metal deposits in altered but nonsilicified country rock near the jasperoids, in silicified country rock, or in veined and brecciated jasperoid.

ACKNOWLEDGMENTS

We thank S. René Evans, Flint Hall, and Falma J. Moye for collecting some of the chip samples of jasperoid outcrops. Cole L. Smith gave valuable suggestions and guidance on sampling and geochemical interpretations, and Mike Allen, Jon Connor and Jim Erdman explained statistical methods and STATPAC. Mike Blaskowski, Rick Sanford, Dick Smith and Bill Hackett gave constructive reviews. Geochemical analyses were funded by an Idaho State Board of Education economic incentives grant to Falma J. Moye for the study of gold and silver mineralization in central Idaho.

REFERENCES


zoic paleogeography of the western United States:
Society of Economic Paleontologists and Mineralogists Pacific Section, Pacific Coast Paleogeography Symposium I, p. 275-299.


Ore Deposits of the Carrietown Silver-Lead-Zinc District, Blaine and Camas Counties, Idaho

Robert S. Darling

ABSTRACT

The Carrietown Ag-Pb-Zn district is located in the northern half of the Dollarhide Mountain 7.5-minute quadrangle, 35 km west of Ketchum and 30 km north of Fairfield. The district covers 25 square kilometers in the southwest part of the central Idaho black-shale mineral belt.

Ore deposits in the district are epigenetic, vein-type bodies containing galena, sphalerite, tetrahedrite, pyrite, arsenopyrite and chalcopyrite. Galena and tetrahedrite are the principal Ag-bearing phases. Localization of deposits is controlled by both structure and lithology. Ag-Pb-Zn mineralization is best developed in: (1) the Carrietown and Dollarhide metasedimentary units, and (2) the northeast-trending fault zones located in proximity to the Carrietown/Dollarhide thrust and close to contacts between the Carrietown and Cretaceous granodiorites. Temperatures of ore formation from several ore mineral pairs suggest a mesothermal environment. Field, thermal and mineralogical relationships are not consistent with Eocene mineralization related to Challis magmatism, but support mineralization related to Cretaceous igneous activity.

INTRODUCTION

The Carrietown silver-lead-zinc (Ag-Pb-Zn) district lies within an oval-shaped area of about 25 square kilometers, in the northern half of the Dollarhide Mountain 7.5-minute quadrangle. The quadrangle is located in the central Smoky Mountains, a north-south 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 the late 1800s. Today, however, the village is abandoned and only a few buildings remain. Carrietown can be reached via light-duty roads from the towns of Ketchum, 35 km to the east, and Fairfield, 30 km to the south (Fig. 1).

Ag-Pb-Zn deposits in the Carrietown area were described first by Umpleby (1915) and then Ross (1930). Umpleby (1915) believed the Carrietown ores were epigenetic and related to Cretaceous igneous activity. Later, the recognition of syngenetic Pb-Zn mineralization in Paleozoic rocks of central Idaho led to a renewed interest in the Carrietown deposits. The principal objectives of this research were to determine if the deposits had a syngenetic or epigenetic origin, and if epigenetic, was the mineralization related to Eocene or Cretaceous igneous activity.

GEOLOGIC SETTING

The Carrietown district is located on the southwest margin of the central Idaho black-shale mineral belt of
Hall (1985), at the junction between allochthonous Paleozoic strata and Cretaceous intrusive rocks of the Idaho batholith.

In the northern half of the Dollarhide Mountain quadrangle, Permian Dollarhide Formation has been thrust over Paleozoic (?) Carrietown sequence (Fig. 2). In most locations, the thrust fault dips gently, but in the area southwest of Carrietown it dips about 65 degrees southeast. The Paleozoic units are preserved as large, almost completely isolated roof pendants in the Cretaceous intrusives (Fig. 2). In the southern part of the district, both Cretaceous intrusives and Paleozoic units are unconformably overlain by Eocene Challis Volcanics, which erupted between 51 Ma and 40 Ma (McIntyre and others, 1982). All of the above rocks are intruded by numerous Eocene hypabyssal dacite porphyry dikes and small stocks. The intrusives are coeval, subvolcanic equivalents of the Challis Volcanics. One dike 0.7 km west of Carrietown has been dated at 47.2 ± 0.9 Ma by K-Ar methods (Wayne Hall, unpublished report, 1978) (Fig. 2).

**Stratigraphy**

The Carrietown sequence is an informal unit consisting of fine-grained, banded quartzites and quartz-biotite phyllites (Geslin, 1986; Darling, 1987). It lies structurally below the Dollarhide Formation and is intruded at its base by Late Cretaceous granodiorites of the Idaho batholith (Fig. 2). In the Carrietown area, the unit is about 1000 meters thick (Skipp and Hall, 1980), but only 176 meters are preserved in the Buttercup Mountain quadrangle, 10 km to the southeast (Geslin, 1986). The quartz-biotite phyllites are strongly foliated and locally contain staurolite, indicating regional metamorphism of the lower amphibolite facies (Geslin, 1986). Banded quartzites of the Carrietown sequence commonly contain fine-grained, locally stratiform pyrrhotite and minor chalcopyrite, presumably of syngenetic or metamorphic origin.

The Permian Dollarhide Formation (2,000 meters thick) is a formal unit containing highly carbonaceous, fine-grained siltstones and argillites (upper member), and sandy limestone, calcareous siltstone and sandstone (lower member) (Hall, 1985; Wavra and others, 1986; Geslin, 1986). This subdivision follows that proposed by Link and others (1988, this volume). The lower member is characterized by synsedimentary folds, convolute bedding, load casts, and graded-, cross- and lenticular bedding (Wavra and others, 1986; Geslin, 1986). In the Carrietown area, the lower member forms the bulk of the outcrop. The Dollarhide Formation commonly contains abundant dark-colored tremolite which formed as a result of contact metamorphism (hornblende-hornfels facies) during intrusion of Cretaceous granodiorites (Darling, 1987).

**HISTORY AND PRODUCTION**

Mining in the Carrietown district began in the summer of 1880, but the activity was short lived and many operations ceased by the turn of the century (Umpleby, 1915). Since then, the district has remained essentially inactive, except for small-scale mining and annual assessment work. In the early 1980s, however, as silver prices soared, both local and major mining companies were exploring the district's potential for undiscovered base and precious metal deposits. Umpleby (1915) reports that mines in the Carrietown area produced about $1,000,000 worth of silver, lead and zinc. Ross (1930) estimates a similar value of $1,200,000 and reports that mines in the Carrietown area produced 1,008,114 ounces of silver, 1,827,337 pounds of lead and 345,313 pounds of zinc; small amounts of gold were locally recovered as well. The average ore grade in the district ranged from 15 to 400 ounces of silver per ton (Umpleby, 1915), but local miners working in the area reported grades as high as 1,000 ounces per ton. Major mines in the Carrietown district and their production values are listed in Table 1. Refer to Ross (1930) for the amounts of metals produced from individual mines.

**ORE DEPOSITS**

Ore deposits in the Carrietown district are tabular, vein-type bodies that generally occupy northeast-trending
fault zones in both Paleozoic units. The mineralized shear zones dip moderately to steeply southeast and northwest, and are commonly positioned parallel to the bedding of their host rocks. Veins are generally located: (1) in or near the thrust fault that separates the Carrietown from the Dollarhide, and (2) near contacts between the Carrietown sequence and Cretaceous granodiorites (Fig. 2). Veins have an average width of 1 to 2 meters and extend into the country rock for several tens of meters. Within the veins, ore minerals occur in a series of irregular, lens-like pods that are positioned parallel to the vein walls. These pods commonly display open-space filling textures and show evidence of intense brittle deformation.

**Mineralogy**

The ore minerals include argentiferous galena, sphalerite, tetrahedrite, pyrite, arsenopyrite and chalcopyrite. Pyrrhotite, cubanite, tetramite, boulangerite and molybdenite occur in minor amounts. Quartz and siderite are the principal gangue minerals where the ore is hosted.
Table 1. Producing mines of the Carrietown district up to 1915. Production values are from Umpleby (1915).

<table>
<thead>
<tr>
<th>Mine</th>
<th>Production (1915)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carrie Leonard</td>
<td>$500,000</td>
</tr>
<tr>
<td>King of the West</td>
<td>$200,000</td>
</tr>
<tr>
<td>Sunny Gabore</td>
<td>$100,000</td>
</tr>
<tr>
<td>Tyrannis</td>
<td>$100,000</td>
</tr>
<tr>
<td>Dollarhide</td>
<td>$75,000</td>
</tr>
<tr>
<td>Silver Star</td>
<td>$75,000</td>
</tr>
<tr>
<td>Margaret</td>
<td>$32,000</td>
</tr>
</tbody>
</table>

(NA) not available

by the Carrietown sequence, but ores hosted by the Dollarhide Formation have a quartz-calcite gangue. A general paragenetic sequence for the mineralization is illustrated in Figure 3. This sequence is based on field observations and petrographic studies of seventy-four polished sections (Darling, 1987).

Many of the above minerals are very coarse-grained and are easily recognized in hand sample; however, cubanite, boulangerite and tennantite are microscopic. Common ore textures are: (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) reports that galena and tetrahedrite are the principal silver-bearing phases in the Carrietown district. Tennantite and boulangerite (although volumetrically insignificant) could be additional silver-bearing phases. The manner in which silver occurs in galena is not known, but microscopic tetrahedrite inclusions are a possible source. The silver-bearing phases (galena-tetrahedrite-tennantite-boulangerite) were deposited relatively late in the mineralization sequence (Fig. 3).

**DISCUSSION**

In recent years, much attention has been focused on the occurrence of stratiform, syngenetic ore deposits in the black-shale belt. Ore deposits of syngenetic origin have been described in the Triumph mine (Triumph-Parker mineral belt) and in the Hoodoo and Livingston mines (Slate Creek district) (Ball, 1985). In the Carrietown district, however, the orebodies appear to have an epigenetic origin. This interpretation is supported by strong structural control of the deposits and open-space filling textures.

**Structural and Lithologic Controls on Deposition**

The localization of deposits is controlled by both structure and lithology, as they are best developed in moderately to steeply dipping, northeast-trending fault zones in both Paleozoic units.

**Structural Controls**

The prominent northeast trend of the veins suggests they may have formed in steep northeast-trending structures related to Eocene extension. However, localization of deposits could also be controlled by bedding-parallel faults. This explanation is supported by: (1) bedding in both host rocks commonly strikes northeast and dips moderately to steeply northwest and southeast, and (2) the veins are commonly positioned parallel to the bedding of their host rocks. Because many of the deposits are located in proximity to the thrust, it is likely that the present orientation was controlled not by Eocene extension but rather by earlier Mesozoic thrusting.

Some of the larger deposits (Carrie Leonard, Silver Star and Horn Silver mines) are located where the thrust fault is steepest (about 65 degrees). This spatial relationship also occurs in the Buttercup Mountain quadrangle. There, the Buttercup mine (the chief producer in the Willow Creek district) is located on the Carrietown/Dollarhide thrust, which dips about 50 degrees eastward (Gaslin, 1986). This relationship suggests that the thrust fault and its imbricated splays acted as a principal pathway for hypogene ore fluids.

**Lithologic Controls**

The deposits are partly controlled by lithology, since most Ag-Pb-Zn mineralization is hosted by Paleozoic metasedimentary rocks. These rocks could have provided a reducing environment for the deposition of metals, an environment not normally expected in Cretaceous intrusive rocks. Also, trace-element signatures of both Paleozoic metasedimentary units are similar to average black shales and would provide an excellent source for metals (Darling, 1987).
Thermal Conditions of Mineralization

Howe and Hall (1985), in an isotopic study of the black-shale belt, performed sulfur isotope analyses on six samples in the Carrietown district, two of which were used to calculate ore-forming temperatures. The two analyses yielded 269 ± 45 degrees C from the sphalerite-galena pair, and 375 ± 45 degrees C from the pyrite-galena pair (Howe and Hall, 1985, p. 192). In polished section, both sulfide pairs commonly exhibit mutual grain boundaries, suggesting that isotopic equilibrium was established and that the calculated temperatures are reliable.

A 250-300 degree C minimum temperature of mineralization is suggested by the presence of cubanite exsolution lamellae in chalcopyrite (Ramdohr, 1980, p. 639). Clark (1960) shows that the equilibrium assemblage of arsenopyrite + pyrite reacts to form pyrrhotite + liquid at 491 ± 12 degrees C (at 1 bar), up to 528 ± 10 degrees C (at 2,070 bars). Thus, the pyrite + arsenopyrite intergrowths establish a minimum mineralization temperature at 491 to 528 degrees C (if P_f is less than 2,070 bars). Thermal constraints imposed by the fixed-point geothermometers agree well with the calculated sulfur isotope temperatures and are generally consistent with a "mesothermal" (200-300 degrees C) environment. The thermal information gathered from ores in the Carrietown district, although scant, is generally consistent with that established for similar deposits in the Wood River District, 30 km to the east (Hall and Czamanske, 1972; Hall and others, 1978).

Age of Mineralization

Because the ore deposits have an epigenetic origin and are partially hosted by granodiorite at the Silver Crown mine, the deposits are no older than late Cretaceous. Furthermore, the principal mineralized structure in the Dollarhide mine is cut by a dacite porphyry dike (Umplesby, 1915; Darling, 1987), indicating the deposits are no younger than Eocene (47.2 Ma). Because the ore deposits formed at elevated temperatures, it seems safe to assume that mineralization is related to either Cretaceous or Eocene intrusive activity.

The spatial relationship of some ore deposits to Cretaceous granodiorites suggests a Cretaceous age. A similar spatial relationship is used by Hall and Czamanske (1972) to argue for Cretaceous mineralization in the Wood River district, 30 km to the east.

The thermal data and field relations previously discussed may help to constrain the age of mineralization. Ore deposits in the Dollarhide, Silver Star, Stormy Galore and King of the West mines are located less than 150 meters vertically below the unconformable contact between the Dollarhide Formation and overlying Eocene Challis Volcanics. Because the volcanics show neither hydrothermal alteration (Gehlen, 1983), nor do they host sulfide mineralization, it is believed that the mineralization is prevolcanic. If the mineralization is prevolcanic, but still related to Eocene intrusive activity, then ore deposits in the above mines must have formed at depths of 150 meters below the Eocene paleosurface. This inferred shallow ore-forming environment conflicts with a number of mineralogical, thermal and field relations, including:

1) The "mesothermal" mineral assemblage is not characteristic of very shallow depths of formation (rather, we would expect an "epithermal" mineral assemblage).

2) Minimum temperatures of 250-300 degrees C would not be expected at 150 meter depth (this would require an unusually high geothermal gradient).

3) Minimum temperatures of 250-300 degrees C at 150 meter depth suggest that water would exist as vapor (however, the observed primary inclusions indicate that the fluid was not boiling).

4) There is no evidence of acid-sulfate (near-surface) alteration occurring in or above the deposits.

It should be noted that Eocene igneous activity lasted for about 10 million years (McIntyre and others, 1982). The deposits could therefore have formed during early Challis magmatism (~51 Ma), could have subsequently been eroded to deep levels, and then covered by younger Challis Volcanics (~40 Ma). However, this interpretation conflicts with the observation that dacite porphyry dikes (dated at 47.2 Ma) intrude many of the volcanics, indicating that the extrusives probably represent early phases of Challis volcanism.

From the observations listed above, mineralization of Eocene age is highly unlikely. The deposits probably formed during or soon after the emplacement of the Cretaceous intrusive rocks.

SUMMARY AND CONCLUSIONS

Ore deposits in the Carrietown Ag-Pb-Zn district are tabular, vein type bodies composed dominantly of Ag-galena, sphalerite, tetrahedrite, pyrite, arsenopyrite and chalcopyrite. The localization of deposits appears to be controlled by both structure and lithology. The deposits are generally restricted to: (1) the Carrietown and Dollarhide metasedimentary rocks; and (2) northeast-trending shear zones located in or near the thrust fault, and near contacts between the Carrietown and Cretaceous intrusive rocks. The strong structural control and open-space filling textures support an epigenetic origin for the deposits. Field relations, geothermometry and mineralogical information are not consistent with an Eocene ore-forming environment, but support a Cretaceous age of mineralization.

ACKNOWLEDGMENTS

The paper presented here is part of a master's thesis completed at Idaho State University under the direction
of Charles W. Blount, Paul Karl Link and Falma J. Moye. Wayne Hall first suggested the area as a possible thesis topic and provided many invaluable ideas in the initial stages of this research. Other individuals who directly or indirectly contributed to this research are: Joyce LoBue-Darling, Doug Dvoracek, Mike Luessen, Fred and Calvin Wolske, Ron Worl, Betty Bailey, J. Brian Mahoney, Jeff Geslin, Dave Stewart, Reed Lewis and Bill Gehlen. R. P. Smith reviewed the manuscript and provided much constructive criticism.

REFERENCES


Glaciated valley of Alturas Creek on the east side of the Sawtooth Mountains. The long tree-covered ridge is a lateral moraine. Alturas (larger) and Perkins (smaller) Lakes are dammed by recessional moraines. Photograph by P. K. Link.
Field Guides to the Quaternary Geology of Central Idaho

Edward B. Evenson¹
Roy M. Breckenridge²
George C. Stephens³

INTRODUCTION TO THE FIELD TRIP GUIDES

Six field guides explain the Quaternary and Recent history of selected areas in central Idaho. The trips for the six guides are shown on Figure 1. Titles for the six trips are as follows:

Part A. Glacial deposits of the Big Wood River Valley, by Suzanne Pearce, Gunnar Schlieder and E. B. Evenson. (Stops 1 and 2)

Part B. Glacial geology of the Stanley Basin, by R. M. Breckenridge, L. R. Stanford, J. F. P. Cotter, J. M. Bloomfield and E. B. Evenson. (Stops 3 through 8)

Part C. History of gold mining on the Yankee Fork River, Custer County, by G. C. Stephens. (Stop 9)

Part D. Surface faulting and groundwater eruptions associated with the 1983 Borah Peak earthquake, by A. J. Crone. (Stops 10 and 11)

Part E. History of Quaternary faulting and scarp degradation studies, southern Lost River Valley, by K. L. Pierce. (Stops 12 and 13)

Part F. Quaternary Volcanism at Craters of the Moon National Monument, by G. C. Stephens. (Stops 14 through 18)

To aid future users who may not wish to make all stops in the order presented in this guide, each article provides detailed locations for individual stops. For closely spaced stops within trips, mile by mile roadlogs are provided.

A map showing the locations of all stops is presented as Figure 1. Other local and regional maps are presented under the introductions to individual articles, or as part of specific stops.

ACKNOWLEDGMENTS

We thank M. D. Wilson and D. W Rodgers for helpful reviews of this guide. We especially thank the Yankee Fork Gold Dredge Association for arranging a special tour of the Yankee Fork Dredge.

¹ Department of Geological Sciences, Lehigh University, Bethlehem, PA 18015-3188
² Idaho Geological Survey, University of Idaho, Moscow, ID 83843
³ Department of Geology, George Washington University, Washington, D.C. 20052

Figure 1. Location map, showing major highways, physiography, field trip route and stop localities.
INTRODUCTION

Detailed mapping indicates that the Big Wood River drainage preserves the deposits of at least two Quaternary glaciations. Some of the evidence supporting this conclusion will be presented at Stops 1 and 2.

From its headwaters near Galena Summit (30 miles north of Hailey) the Big Wood River flows south and receives the drainage of several major tributaries including Prairie, Baker, Boulder, Trail, Warm Springs, and East Fork Creeks. Deposits of the older, more extensive glaciation are not well preserved, but they are present in most valleys and can be used to reconstruct the ice distribution in most areas. Deposits of the smaller, younger glaciation are fresh and well preserved, and allow accurate reconstruction of late Pleistocene ice distribution. In all cases, glaciated deposits are confined to the tributary valleys; no compound ice lobe flowed south down the Wood River Valley. The northern valleys draining the high areas developed ice lobes that extended as far as the Big Wood River (i.e., Prairie Creek and Boulder Creek). Further south, the glaciers originating in lower and drier areas were restricted to the upper reaches of the tributary valleys (Pearce, in prep.; Scott, 1982).

The current lack of radiometric dates for glacial deposits in this area and for Idaho in general makes a determination of absolute age and accurate correlation with other areas impossible. Therefore, age assignments and correlations are based on relative dating techniques (Burke and Birkeland, 1979; Colman and Pierce, 1986; Evenson and others, 1982).

Until the work of Cotter (1980), most glacial deposits in Idaho were named after, and therefore correlated with, the Rocky Mountain Glacial Model (i.e., "Pinedale" and "Bull Lake") which was developed for the Wind River Mountains in Wyoming by Blackwelder (1915). We feel that the extension of the Wyoming nomenclature to Idaho is unjustified. Therefore, we have proposed informal stratigraphic names for the deposits of each glaciation in the study area and have correlated this nomenclature to the Idaho Glacial Model (Evenson and others, 1982), which was developed to allow local and regional correlation within central Idaho. The glaciations of the Idaho Glacial Model can in turn be correlated to the Rocky Mountain Glacial Model. The deposits

mapped in the Big Wood River Valley have been assigned informal local names, which are correlated to the Idaho and Rocky Mountain Glacial Models as shown in Table 1.

Table 1: Nomenclature of the Idaho Glacial Model and correlation with local stratigraphy of the Big Wood River

<table>
<thead>
<tr>
<th>ROCKY MOUNTAIN GLACIAL MODEL</th>
<th>IDAHO GLACIAL MODEL</th>
<th>BIG WOOD RIVER STRATIGRAPHY</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pinedale Glaciation</td>
<td>Potholes</td>
<td>Boulder Creek Advance I-IV</td>
</tr>
<tr>
<td>Bull Lake Glaciation</td>
<td>Copper Basin</td>
<td>Prairie Creek Advance I-III</td>
</tr>
</tbody>
</table>

One of the major problems in mapping Quaternary deposits of the Big Wood River Valley has been the differentiation of Prairie Creek age glacial deposits from older (early Quaternary?) alluvial fan/fanglomerate deposits ("Phantom Hill Gravels" of this study) which were associated with the uplift of the Boulder Mountains. Scott (1982) mapped extensive aprons of poorly sorted alluvial gravel along the Boulder Mountain front. Our detailed mapping confirms many of Scott's (1982) interpretations; however, we think some of his "alluvial fan gravel" is till of the Prairie Creek advance. The two diamictons are similar, having been derived from the same source area. Both may contain boulders and smaller clasts in a fine-grained matrix, and have lobate forms. Striated clasts are undiagnostic because the cobbles and boulders consist mainly of quartzites and volcanics. The quartzites are inherently difficult to striate, and the volcanics weather so rapidly that striations, even in till of Prairie Creek age, are mostly lost. Although striations are occasionally found on fresh clasts of the Prairie Creek till, they are so rare that they cannot be relied upon to differentiate fan gravels from Prairie Creek till.

Other criteria, based on geomorphic and sedimentological evidence, have proven to be much more useful in differentiating these units. In practically all places the fanglomerate sediments (Phantom Hill gravel) are deposited on a bedrock-cored, pedimentlike surface. In some places only a thin veneer of gravel is deposited on bedrock. Also, the gravels are not confined to single valleys, and in many places form divides between tributary valleys. Finally, in the fanglomerate, the larger boulders and cobbles are located close to the mountain front, and grain size decreases downslope. The Prairie Creek till, on the other hand, is restricted to valley sides and bottoms, is not cored by bedrock, and contains very large boulders randomly scattered throughout the deposit. The moraines exhibit a hummocky, commonly kettled topography unlike the alluvial fans which have a smooth, even profile, except where dissected or mass wasted.

Using the geomorphic criteria discussed above and relative dating techniques, the Quaternary deposits of the Wood River Valley have been mapped, subdivided, named and correlated. Stops 1 and 2 show the type areas for the Boulder Creek and Prairie Creek advances.

LOCALITY DESCRIPTIONS

Stop 1: Type Area of the Boulder Creek Advance

Location

This stop is located 13 miles north of Ketchum on Boulder Creek Road (Figure 1; Figure 1 of Evenson and others, 1988, this volume). To reach the section, turn east (right) from Idaho Highway 75 onto Boulder Creek Road. Follow the road approximately 0.25 mile and take the left fork. From the fork drive another 0.8 mile and take another left fork. From the second fork, drive another 0.5 mile and park at the large boulder next to the road. Ascend the moraine to the crest, and the location shown on Figure 1.

Description and Discussion

This stop is the type area of the Boulder Creek I advance. The moraines of the Boulder Creek advance have a fresh, sharp morphology and contain many surface boulders. Drainage development is poor; soils are thin to nonexistent, and kettles are present, although not at this location. En route to this Stop from Highway 75, the Boulder Creek Road ascends the outwash fan graded to the Boulder Creek I terminal moraine. The terminus of the moraine was partially destroyed and reworked by meltwater emanating from the glacier, and small moraine remnants are preserved as bouldery lumps in the otherwise relatively flat outwash deposits beginning about 0.6 mile to the north of Highway 75 (Figure 1). The Boulder Creek glacier never extended far enough to dam the Big Wood River.

Stop 2: Type Area of the Prairie Creek Advance

Location

From the junction of Boulder Creek Road and Idaho Highway 75, turn north (right) onto Highway 75, drive 6.7 miles and then turn west (left) onto a small turn-off (this is not a true road and is not easily visible). Immediately after turning from Highway 75, park in the ample space. Follow the faint path southwest into a wooded area, ascend a small hill, and continue to a
Figure 1. Surficial geologic map of Boulder Creek area and location of Stop 1.
Figure 2. Surficial geologic map of Prairie Creek area and locations of Stops 2 and 2A.
shallow depression (Figure 2; Figure 1 of Evenson and others, 1988, this volume).

**Description and Discussion**

This stop is the type area for deposits of the Prairie Creek I-III advances. Moraines of the older advance have subdued morphology. They lack sharp crests, have fewer surface boulders, and are more dissected than the younger Boulder Creek moraines in this drainage. The Prairie Creek moraines are located about one mile downstream and outside of the deposits of Boulder Creek age (Figure 2). From the pull-off, the path first runs for about 150 yards in a southerly direction on a remnant of Prairie Creek I outwash. The road reaches the head of the outwash and part of the bouldery Prairie Creek I moraine at a curve to the right (west). A relatively well-preserved kettle can be seen along the road about 100 yards to the west of this curve.

**Optional Stop 2A:**
**Ice-streamlined Bedrock Knob**

**Location**

Turn south (right) onto Highway 75 and drive 1 mile. Turn west (right) onto Prairie Creek Road. Drive on Prairie Creek Road for about 1.5 miles and turn into the parking area at the base of a large bedrock knob. Walk to the top of the knob. See Figure 2 for the exact location of Stop 2A.

**Description and Discussion**

The summit of the knob shows ice erosional features (grooves) which are still preserved, although very weathered. The Boulder Creek terminal moraine ends at the base of the knob, showing that during the Boulder Creek advance the ice was not high enough to override the obstacle. Its streamlined shape is therefore a relict of the older and more extensive Prairie Creek glaciation.

**REFERENCES**


Burke, R. M., and Birkeland, P. W., 1979, Reevaluation of multi-parameter relative dating techniques and their application to the glacial sequence along the eastern escarpment of the Sierra Nevada, California: Quaternary Research, v. 11, p. 21-51.
Field Guides to the Quaternary Geology of Central Idaho:
Part B.
Glacial Geology of the Stanley Basin

Roy M. Breckenridge
Loudon R. Stanford
J. F. P. Cotter
J. M. Bloomfield
Edward B. Evenson

LOCATION

This part of the field guide describes the glacial geology of the Stanley Basin (Fig. 1 of Evenson and others, 1988, this volume, Stops 3-8). The field trip route begins at Galena Summit and ends at the town of Stanley. Main access is from Idaho Highway 75, an all season paved road. Much of the trip route is through the Sawtooth National Recreation Area. The road log is keyed to green milepost markers along Highway 75.

INTRODUCTION

The Stanley Basin contains some of the best examples of mountain glaciation in Idaho. Stanley Basin is a north-trending graben formed during late Tertiary time by basin and range style faulting. Major displacement has occurred at the mountain front along the Sawtooth horst on the Sawtooth Fault. Although no fault scarps have been recognized, the area is seismically active (Smith and Sbar, 1974). During the Pleistocene, large glaciers occupied the surrounding ranges: the Sawtooth Range in the west, the White Clouds and Boulder Mountains in the east, and the Salmon River Mountains to the north. The Sawtooth Range consists mainly of granitic crystalline rocks. The mountains to the south and east are composed of Paleozoic sediments and Challis volcanics. The bedrock and structural geology of the area are discussed in other articles of this volume (see Johnson and others, 1988, this volume).

Because the Sawtooths are higher and intercept storms first, they receive more precipitation. Therefore, glaciers from the Sawtooths extended farther out from the mountain front than those from the Boulder and White Cloud mountains to the east. The Sawtooth glaciers poured out on the basin floor and partly coalesced, forming a complex of piedmont moraines. Many of the
Description of Units

A  alluvium
Pt  Pinedale till
Pg  Pinedale outwash
Bt  Bull Lake till (undifferentiated age)
Bt₂  late Bull Lake till
Bt₁  early Bull Lake till
Bu  undifferentiated Bull Lake outwash and till
Bg  Bull Lake outwash (undifferentiated age)
Bg₂  late Bull Lake outwash
Bg₁  early Bull Lake outwash

Figure 1. Surficial geologic map of Stanley Basin (modified from Williams, 1961) and locations of Stops 3-8.
tributary valleys on the west side of the basin have glacial lakes enclosed by end moraines. Outlet glaciers in the larger valleys of the White Cloud Mountains reached the Stanley Basin but did not coalesce. The moraines grade into a sequence of fan and outwash gravels that form much of the basin floor.

The glacial geology of the Stanley Basin area was studied by Williams (1961), who used the glacial terminology that Blackwelder (1915) established in Wyoming. Williams recognized a Pinedale/Bull Lake sequence and mentioned the possibility of a pre-Bull Lake, Buffalo-type glaciation, based on weathered gravels east of the basin. We present his mapping here as a guide to the style and extent of glaciation (Fig. 1) but imply a more complex sequence. From our work at Fourth of July Creek, we infer a three-fold glacial chronology based on weathering and stratigraphic relationships. For the glacial deposits on the east side of the Sawtooth Range, we will continue to use the Pinedale/Bull Lake nomenclature of Williams (1961) until the stratigraphic relationships are better understood and local stratigraphic nomenclature is developed. For the deposits emanating from the Boulder Mountains (Pole Creek and Fourth of July Creek), we will use the local stratigraphic nomenclature described below.

DISCUSSION OF AGE RELATIONSHIPS IN ADJACENT AREAS

On the west side of the Sawtooths, Stanford (1982) identifies at least three glaciations, and informally names them, from oldest to youngest, "Penrod Creek", "Camp Creek" and "Grandjean". Evenson and others (1982) recognize a three-fold sequence in the Boulder-Pioneer Mountains to the southeast, and in adjacent areas of the White Clouds. They propose the use of a two-tiered glacial model for Idaho, similar to the Rocky Mountain Glacial Model of Mears (1974), and use the terms "Copper Basin Glaciation" and "Potholes Glaciation" for deposits similar to the Bull Lake and Pinedale. We use informal, local stratigraphic names at Pole Creek and Fourth of July Creek because of uncertain regional correlation with the type deposits in Wyoming and other regions of Idaho.

Colman and Pierce (1977, 1983) present a summary of Quaternary age-dating techniques often used in the western United States. From newly-developed weathering criteria and other established methods, they distinguish a three-fold glacial sequence in the Long Valley-McCall area (Colman and Pierce, 1986). As in many other areas, few numerical dates exist, and relative age-dating techniques are the best available. Bloomfield (1983) studied volcanic ash in the White Cloud Peaks and Boulder Mountains, and defined a minimum age of the Pinedale retreat at Pole Creek (see Stop 4).

TRIP DESCRIPTION

Refer to Figure 1 for the locations of Stops 3-8. Highway mileposts are included as an aid to location through the Stanley Basin, and are not intended as a precise mileage log.

Milepost

158 Beginning of trip log. Galena Summit, elevation 8701 feet.

159 Galena Summit Overlook turnout. Stop 3.

Stop 3: Galena Summit Overlook.

In the Stanley Basin (Fig. 2) the Salmon River begins its flow to the Pacific Ocean. The Sawtooth Range to the northwest exhibits classic alpine glacial features. Granitic rock of the Sawtooth batholith is carved into the glacial arêtes, cols, and horns on the skyline for which the range is named. The floor of the basin is covered on the west by glacial deposits of a piedmont moraine belt. Most of the moraines are heavily forested by lodgepole pine, in contrast to the sage-covered outwash flats. The tree-covered moraines in the foreground are left lateral moraines of Pole Creek and are Rainbow Creek/Potholes/Pinedale in age (see Table 1). Pole Creek drains a large catchment area in the Boulder Mountains.

Milepost


161 Multiple crests of left lateral moraines of Pole Creek.

162 Roadcut exposures of Pole Creek moraines. Note thin soil development and fresh moraine topography. Junction with Salmon Valley Road on left.

163 Junction Frenchman Creek Road.

164 Terminal moraine of Pole Creek on right, Junction Smiley Creek Road.

165 Junction Valley Road turnoff (Pole Creek access). Turn right on dirt road, cross cattle guard, and proceed across moraine to fork in road. Park here for Stop 4.
Stop 4: The Glacial Geology and Postglacial History of Pole Creek Canyon and Pole Creek Kettle

The deposits of two glacial events – the Boulder Mountain (older) and the Rainbow Creek (younger) – are located and mapped in Pole Creek Canyon (Bloomfield, 1983). The deposits are differentiated on the basis of multiple relative age techniques and are correlated to the Copper Basin and Potholes Glaciations, respectively, of the Idaho Glacial Model (Table 1). Five Rainbow Creek moraines and three related terrace levels are distinguished in Pole Creek Canyon (Fig. 3).

Sediment cores (Fig. 4), taken from an undrained kettle lake (Fig. 5) formed on Rainbow Creek drift, contain a condensed Late Pleistocene and Holocene tephra and pollen record (Cotter and others, 1986). Identification of Glacier Peak Set B tephra near the base of the Pole Creek kettle core places a minimum age of 11,250 Ka on the retreat of valley glaciers from their Late Wisconsinan maxima. Radiocarbon dates of 10,835±0.100 Ka (P.H. 0118) and 8,450±0.085 Ka (SI-5181), and the presence of Mount Mazama ash (6,600 Ka) above, support the Glacier Peak ash identification.

Pollen and sediment analyses (Fig. 4) indicate three

Table 1. Local stratigraphic nomenclature for the North Fork, Pole Creek and Slate Creek drainages, with correlations to the Idaho and Rocky Mountain Glacial Models (from Cotter and others, 1986).

<table>
<thead>
<tr>
<th>IDAHO GLACIAL MODEL</th>
<th>NORTH FORK BIG LOST RIVER</th>
<th>POLE CREEK</th>
<th>SLATE CREEK</th>
<th>WIND RIVER RANGE, WYOMING</th>
</tr>
</thead>
<tbody>
<tr>
<td>EVENSEN et al. 1983</td>
<td>ADVANCE</td>
<td>ADVANCE</td>
<td>ADVANCE</td>
<td>GLACIATION</td>
</tr>
<tr>
<td>POTHOLEs</td>
<td>NORTH FORK</td>
<td>RAINBOW CREEK</td>
<td>SILVER RULE</td>
<td>PINEDALE</td>
</tr>
<tr>
<td>COTTER &amp; EVENSEN 1983</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>COPPER BASIN</td>
<td>KANE</td>
<td>BOULDER MOUNTAIN</td>
<td>SLATE CREEK</td>
<td>BULL LAKE</td>
</tr>
<tr>
<td>PIONEER</td>
<td>NOT RECOGNIZED</td>
<td>NOT RECOGNIZED</td>
<td>NOT RECOGNIZED</td>
<td></td>
</tr>
</tbody>
</table>

intervals of Late Pleistocene and Holocene climatic change. Cool and wet climatic conditions prevailed in the region shortly before and immediately following the deposition of the Glacier Peak-Set B ash (11,250 Ka). Climatic warming occurred from approximately 10,500 to 6,600 Ka, after which dry conditions prevailed.
Sediment accumulation in the kettle ceased by 4.350 Ka (Cotter and others, 1986).

Return to Highway 75. If time and weather permit, take a leisurely one-hour drive to Stop 5 via the Valley Road, which parallels Highway 75. This route crosses the Pole Creek moraines and provides excellent views of the Sawtooth Range without traffic interference. The road is unimproved, and is subject to closure and mudholes. The Valley Road rejoins Highway 75 at milepost 174.

Milepost

166 Pinedale moraine on left. Beaver Creek Lodge, gas and grocery. Junction to old Sawtooth City ghost town (3 miles).

167 Highway cuts through Bull Lake terminal moraine.

168 Alturas Lake junction. Terminal moraine, Pinedale characteristics.

170 Bull Lake outwash on east.

171 Pettit Lake moraines. Pettit Lake junction.

172 Cross Salmon River and the Blaine County-Custer County line. Yellowbelly Lake terminal moraine.

174 Junction with Valley Road. Proceed a few hundred yards further on Highway 75 and turn east on Fourth of July Creek Road. Ascend fan on dirt road and pass through moraines at mouth of canyon.

Note soils in roadcuts. Stop at small unimproved campsite just before road crosses Fourth of July Creek. Park here for Stop 5.

Stop 5: Fourth of July Creek

Figure 6 is a detailed map of the Fourth of July deposits. On the north side of the road are nested right
Figure 4. Sediments, tephas, and pollen percentages of selected taxa of the Pole Kettle core (from Cotter and others, 1986).
lateral moraines of the last glaciation. The two large gullies were formed by breaks in an irrigation ditch above. The eastern gully is about 30 years old, and the western exposure formed in 1985. Take extra caution to avoid loose boulders if you enter these gullies! We have called the youngest deposits “Fourth of July Creek”. The soil is thin (<1 m) and predominantly grayish-brown (10YR 3/3), and granitic clasts are generally fresh although about 15% are weathered. Moraine topography is fresh, with narrow ridge crests and steep slopes. We infer the moraines to be correlative with the Potholes (Pinedale) glaciations.

Scramble up slope to the moraine crest and proceed north along the irrigation ditch to the next moraine ridge. These deposits are more mature in surface morphology, and they are more weathered than those near the road. We have called these deposits “Milky Creek”. Soils are developed to a depth over 1.5 m and are predominantly brown (10YR 5/3). About 50% of the granitic clasts are weathered, in contrast to 15% in the Fourth of July Creek till. These deposits along with Fourth of July Creek deposits were mapped together as Pinedale by Williams (1961).

Cross the irrigation ditch on the crest of the Milky Creek moraine. Follow the dirt road down the crest of the moraine approximately 70 m until a terrace is visible to the north. We call the deposits of this landform “Champion Creek”. These deposits consist of till and outwash that have been highly eroded. Soils are over 2 meters thick, reddish brown (5YR 4/6), and exhibit strong clay development. Nearly all granitic boulders in Champion Creek deposits are rotten. Return to cars at Fourth of July Creek. Drive back to junction at Highway 75 and continue north.

**Milepost**

175 Thousand Springs Road on east. Turn east and STOP at the gate. This exposure is on private land. Access is by permission only.

**Stop 6: Thousand Springs Ranch Landslide**

From here, view the Thousand Springs Ranch landslide area along the terrace on the east side of the road (Fig. 7). A thick section of glaciolacustrine sediments is exposed in the landslide scarp (Fig. 8). We interpret these deposits as having formed in a glacial lake impounded by glacial ice dams on the Pleistocene Salmon River. Here at the mouth of Fisher Creek and Fourth of July Creek they probably represent a deltaic facies. The section is thick and represents a relatively long-lived glacial lake. Champion Creek deposits form the upper surface and appear to grade into the lacustrine section. Figure 8 is a panorama and stratigraphic sketch of the section currently exposed in the landslide scarp. The surface slopes eastward and the deposits coarsen toward the apex, consistent with the delta model.

**Milepost**

176 Champion Creek terraces on east side. Exposures of terraces in landslide scars.

177 On outwash fan from Fisher Creek. Junction of Fisher Creek Road.

178 Town of Obsidian. Lodge, gas and groceries.

179 Williams Creek Trail.

180 Idaho Rocky Mountain Ranch, historic dude ranch.

181 Bull Lake moraine and large boulders to west.

182 Outwash terraces on east, moraines on west.

183 Bull Lake moraine on west. Note reforestation of trees along old sheep runway.
Figure 6. Surficial geologic map of the Fourth of July Creek area (Stop 5).

Breckenridge and Others—Glacial Geology of the Stanley Basin, Idaho
Figure 7. Panorama of the Thousand Springs landslide area (Stop 6).

Figure 8. Glaciolacustrine section at Thousand Springs landslide (Stop 6).

184 Entrance to Fish Hatchery. Bridge across Salmon River. Park beyond bridge at turnout for Stop 7.

Stop 7: Redfish Lake Moraines

The highway is cut through a Bull Lake age moraine near the bridge and through a Pinedale age moraine to the north near the turn to Redfish Lake. Borrow material from the cuts was used for the reconstruction of the new bridge in 1985 and the Fish Hatchery in 1984. The roadcuts have since been graded and reclaimed. Figure 9 shows photographs of the exposures and sketches of the section made in 1984-85. Excavations near roadlevel revealed a glaciofluvial sequence, grading upward into glaciolacustrine sediments overlain by a till. Wells drilled at the fish hatchery intersected similar fine-grained
sediments. We interpret these sections as evidence for a Glacial Lake Stanley in Bull Lake time. Relationships with the glaciolacustrine section at Thousand Springs have not been established. Note that the moraines have pushed the Salmon River southeastward against the valley wall. Coarse boulders in this ungraded reach of the river indicate damming even in Pinedale time (Williams, 1961).

**Milepost**

185 Access road to Redfish Lake. Visitor Center and
tourist facilities. Pinedale terminal moraine exposed in cut slope.

186 Sunny Gulch Campground to right. Note narrows of Salmon River eroded through the terminal moraines of Redfish Lake.

187 Turnoff to ranger station, on left. The ranger station is located on Bull Lake outwash. Visitor information, brochures and maps.

188 Stanley terraces. Park at turnout on east side of highway for Stop 8.

Stop 8: Stanley Outwash Terraces (Fig. 10)

The high terrace on the west side of the highway is capped by Bull Lake outwash. A short sidetrip through Stanley to the airport allows a view from the top of the terrace and examination of the outwash gravels. This exposure in the roadcut shows a weathered horizon (5 YR) that continues through the slope to a reclaimed gravel pit on the south. Above the weathered horizon are cross-bedded sands and coarse outwash gravels. These relationships suggest the upper Stanley terrace is younger and stratigraphically overlies the lower terrace, rather than the opposite interpretation of Williams (1961). Exposures of the weathered horizon are poor but continuous. The fine-grained sediments below the red horizon may represent either a lacustrine or fluvial event.

Junction to town of Stanley. Gas, supplies and lodging.

Continue east on Highway 75 through Lower Stanley along the Salmon River. The terraces across the river are the last large outwash deposits before the Salmon River leaves the basin and flows many miles through a restricted canyon.

REFERENCES


Field Guides to the Quaternary Geology of Central Idaho:

Part C.

History of Gold Mining on the Yankee Fork River, Custer County

George C. Stephens

LOCATION

Refer to Stop 9, Figure 1 of Evenson and others, this volume. The Yankee Fork placer mining area and the Yankee Fork dredge are reached by following Idaho Highway 75 east from Stanley for 13 miles, to Sunbeam. Turn left (north) at Sunbeam toward Bonanza and Custer. This gravel road parallels the Yankee Fork. Approximately 3.5 to 4.0 miles north of Sunbeam, dredge tailings are visible for the next 5 to 6 miles along the banks of the Yankee Fork. The Yankee Fork dredge is located on the left (west) side of the road approximately 8.8 miles north of Highway 75. The old mining town of Custer, now a museum and historical site, is located another 2.2 miles to the north (see Fig. 1 of Evenson and others, 1988, this volume).

HISTORY OF NINETEENTH CENTURY GOLD MINING IN IDAHO

Early settlement of the Idaho Territory (established in 1863) was greatly influenced by the influx of prospectors and miners searching largely for gold and silver in its rugged mountainous regions. Many of these miners were first involved in the great California gold rush of 1848, and in its aftermath they slowly migrated to other promising areas of Nevada, Colorado, British Columbia and Idaho.

In 1860, E. D. Pierce made the first significant gold discovery in what was to become the Idaho Territory, on Oro Fino Creek in Clearwater County (approximately 60 miles east of the present town of Lewiston, Idaho). Subsequently, the discovery of gold on the Salmon River in 1861 inspired a rush to the Salmon in 1862, involving upwards of 10,000 miners. Later that year, placer gold was discovered in the Boise Basin. Lode mining began in earnest, with discoveries near Atlanta, Rocky Bar and Silver City between 1864 and 1869. Placer gold was discovered on Loon Creek, approximately 25 miles northwest of the Yankee Fork, in mid-1869 (Fig. 1).

Discovery of the rich placer and lode deposits of the Yankee Fork region began in 1870 (Idaho State Historical Society, 1976). In the spring of 1871, claims were discovered that produced $8.00 per day. In 1873, placer mining began along Jordan Creek, a tributary to the Yankee Fork located between the future sites of Bonanza and Custer. Still, the Yankee Fork discoveries created little excitement and only modest placer mining took place in the area from 1871 to 1874.
In 1875, a lode deposit was discovered which was the likely source of the Jordan Creek placer deposits. One 2-to-3-inch-wide seam within a gold-quartz vein from this property yielded $11,500 by hand-mortar separation methods in a single month. From 1876 to 1879, the mine produced approximately $133,000, mostly through hand-mortaring. The town of Bonanza sprang up in 1877, and Custer was founded a year or two later.

In the spring of 1879, the gold rush to the Yankee Fork finally began in earnest. By the summer of 1879, the settlement of Bonanza had grown to a population of some 2,000 persons. Gold mining in the Yankee Fork district lasted for some twenty years, and as the more easily mined placer and lode deposits were exhausted, mining activity and population slowly decreased. By 1911, Bonanza and Custer were ghost towns.

Placer Mining and Dredging on the Yankee Fork

The Yankee Fork River has been dredged northward from a point approximately 3.2 miles north of its junction with the Salmon River to its junction with Jordan Creek, about 0.5 mile north of Bonanza (Fig. 2).
The dredged area is approximately 5.2 miles long and 400 feet wide (Choate, 1962). Jordan Creek has likewise been dredged from its confluence with the Yankee Fork to a point approximately 1.2 miles upstream.

The gold-bearing channel of the Yankee Fork has been described by one of the dredge operators as follows:

*The "pay streak" or "pay channel" extends in a meandering fashion, the entire distance from Jordan Creek to the mouth of Yankee Fork. The pay is a very-distinct-appearing, decomposed clay with gravel. The pay streak is approximately 6 inches above bedrock; 8-10 inches thick, and 150 feet wide. On the bends the gold was always on the inside.* (Choate, 1962)

Much of the gold in the Yankee Fork presumably came from Estes Mountain, down Jordan Creek which enters the Yankee Fork from the north between Bonanza and Custer. Drilling records indicate that the Yankee Fork above Jordan Creek contains only $0.16 gold value per cubic yard (at $35.00 per ounce) whereas Jordan Creek contains $1.00 to $3.00 per cubic yard (Choate, 1962).

In addition to these two major creeks that have been worked by dredges, Adair Creek, a tributary to the Yankee Fork near Custer, and Rankin Creek, a tributary near the southern end of the dredge workings have been either hand-placered or worked by bulldozer during the past century. They are too shallow to work by dredges.

### Twentieth Century Placer Mining and Dredging on the Yankee Fork

During the 1930s, State Senator R. E. Whitten assembled options on a group of inactive placer claims along the Yankee Fork. These claims covered a strip of river bottom approximately five miles long. In 1932, the Yankee Fork Placer Mining Company acquired these options, brought in a small dredge and began mining. The operations ceased before any significant production because of unexpectedly large boulders and the tightly cemented gravels in the Yankee Fork riverbed.

A few years later, in 1939, the Silas Mason Company began a program of systematic rotary drilling on 100-foot centers to test the dredging potential of the Yankee Fork. Successful drilling results led to the formation of the Snake River Mining Company. An amusing account of the naming of this company is given in Packard (1983):

*Showing little knowledge of the geography of the State of Idaho, and knowing only that the Snake River drained a large part of the state, the eastern-bound financiers casually adopted the name of the largest river for their company.*

Early in 1940, a contract was signed with the Bucyrus-Erie Company to construct the Yankee Fork dredge. Assembly of the dredge took place in the Yankee Fork valley near the present-day Polecamp Flat Camp-ground (see Fig. 2). The pontoons and frame of the dredge were produced by the Olsen City Manufacturing Company in Boise and trucked over Galena Summit to the assembly site. The heavy machinery manufactured by Bucyrus-Erie was shipped by train from Wisconsin to the railhead at Mackay and thence by truck to the Yankee Fork. Perhaps the single most impressive truck load was the 17.5-ton, 55-foot-long, steel spud. The assembly of the dredge was completed in just under four months during the summer of 1940.

Dredge operations began in late August 1940 and continued until October 12, 1942, when production was halted by the War Production Board Act that regulated nonessential mining activities during World War II. Dredging of the Yankee Fork by the Snake River Mining Company began again in March 1946 and continued until the fall of 1947. During its pre- and post-war operations, the Yankee Fork dredge worked the stream gravels from Polecamp Flat northward to a point just above West Fork Creek, where significantly lower gold values caused the cessation of dredging.

Subsequently, the Warren Mining Company, owned by Fred Baumhoff and J. R. Simplot, bought both the Yankee Fork dredge and the placer claims owned by the Snake River Mining Company. Dredging began again in 1950, and the dredge slowly worked its way northward until it encountered a prominent bedrock bar across the Yankee Fork about a quarter of a mile below Bonanza. A temporary earthen dam was built to the south of the dredge, and the dredge was successfully floated across the bedrock obstacle. High gold values were again found in the gravels on the north side of this bar, and dredging operations continued northward, past Bonanza, to the northern end of the property at the mouth of Jordan Creek. The dredge ceased operation for the last time in 1952, although it ran once more in 1953 when it was returned to its present site on the claims of the Warren Mining Company.

The abandoned and vandalized dredge was donated by J. R. Simplot to the U.S. Forest Service in 1966 for use as a museum. Through a cooperative agreement between the U.S. Forest Service and the Yankee Fork Gold Dredging Association, the dredge is slowly being restored and is open for public tours. The Yankee Fork dredge is the last Bucyrus-Erie gold dredge built for Idaho and is, presumably, the last remaining dredge of the sixty that once operated in Idaho. Table 1 shows production figures for the dredge operations from 1940 to 1952.

### OPERATION OF THE YANKEE FORK DREDGE

The hull of the Yankee Fork dredge (Fig. 3) is 112.5 feet long and 54 feet wide; it floats on 25 (10 x 10 x 27 foot) pontoons. The highest point on the dredge is the 64-foot-high stern gantry which supports the spud and the 105-foot-long stacker (Packard, 1983). The 17.5-ton...
Table 1. Production figures for the Yankee Fork dredge from 1940 to 1952. (from Choate, 1962).

<table>
<thead>
<tr>
<th>Year</th>
<th>Gold</th>
<th>Silver</th>
<th>Cubic Yards</th>
<th>Months Worked</th>
<th>Recovery per Yard</th>
</tr>
</thead>
<tbody>
<tr>
<td>1940</td>
<td>$31,745</td>
<td>$416</td>
<td>214,000</td>
<td>4</td>
<td>$0.15</td>
</tr>
<tr>
<td>1941</td>
<td>262,500</td>
<td>2,877</td>
<td>1,344,000</td>
<td>12</td>
<td>0.21</td>
</tr>
<tr>
<td>1942</td>
<td>169,927</td>
<td>2,195</td>
<td>1,428,000</td>
<td>10</td>
<td>0.12</td>
</tr>
<tr>
<td>1946</td>
<td>183,120</td>
<td>2,512</td>
<td>1,999,000</td>
<td>8</td>
<td>0.17</td>
</tr>
<tr>
<td>1947</td>
<td>90,140</td>
<td>1,359</td>
<td>715,000</td>
<td>8</td>
<td>0.13</td>
</tr>
<tr>
<td>1950</td>
<td>(423,150)</td>
<td>(1,064)</td>
<td>(631,000)</td>
<td>7</td>
<td>0.15</td>
</tr>
<tr>
<td>1951</td>
<td>178,515</td>
<td>3,010</td>
<td>804,000</td>
<td>8</td>
<td>0.23</td>
</tr>
<tr>
<td>1952</td>
<td>64,120</td>
<td>1,065</td>
<td>195,000</td>
<td>4</td>
<td>0.33</td>
</tr>
<tr>
<td>Totals</td>
<td>$1,023,025</td>
<td>$14,298</td>
<td>6,330,000</td>
<td>61</td>
<td>$0.16</td>
</tr>
</tbody>
</table>

Total gold & silver dredged by the Yankee Fork dredge ($1,037,323.00)
Total cost of digging 6,330,000 cu. yds. per $0.17 ($1,076,100.00)
Average gold and silver per cubic yard ($0.16)
Estimated cost of digging a cubic yard ($0.17)
Figures included in parentheses are estimates.

The spud is attached vertically to the stern of the dredge, and during dredging operations its point rested on the river bottom, providing a pivot point for the dredge as it swung and loaded from side to side in the channel.

The bucket line on the bow of the dredge contains 71, eight-cubic-foot capacity, buckets weighing over a ton each (Packard, 1983). The buckets would load and dump at a rate of 26 per minute and could reach to a maximum depth of 37 feet below the water surface. The gold-bearing gravel was then sized by screening and washed through sluice boxes where the ore was caught by a system of riffles and mercury traps. The coarser, non-gold-bearing gravels and cobbles were passed through the dredge to the stacker on the stern, which built the crescent-shaped dredge-tailings piles commonly seen at most places along the Yankee Fork. A schematic illustration of the Yankee Fork dredge in operation is shown in Figure 4.

REFERENCE


SURFACE FAULTING AT DOUBLESPRINGS PASS ROAD
(STOP 10)

Location

Refer to Figure 1 of Evenson and others (1988, this volume) for the location of Stop 10. The following description is modified from Crone (1987). The site can be reached by traveling 23 miles (37 km) northwest from Mackay, Idaho, or 30 miles (48 km) southeast from Challis, Idaho, on U.S. Highway 93 to the Doublesprings Pass Road turnoff. The Doublesprings Pass Road heads northeast from Highway 93. The intersection of the Doublesprings Pass Road with Highway 93 is identified by a sign indicating the direction to the towns of May and Patterson. The turnoff is also identified at the intersection by a historical marker commemorating William E. Borah, after whom Borah Peak was named. The Doublesprings Pass road crosses the fault scarps 2.5 miles (4 km) northeast of its intersection with U.S. 93 (Fig. 1).

Discussion and Description

The Doublesprings Pass road site is an excellent location at which to examine the surface faulting and ground breakage that accompanied the 1983 Borah Peak earthquake. The Ms 7.3 Borah Peak earthquake on October 28, 1983, was the first earthquake in the intermontane west to produce surface faulting since the Hebgen Lake, Montana, earthquake of August 17, 1959. An impressive, Y-shaped zone of fault scarps and surface ruptures, approximately 22 miles (36 km) long, formed during the Borah Peak earthquake (Fig. 1). The scarps and ground ruptures occur primarily along the Lost River fault that separates the Thousand Springs and Warm Spring valleys from the Lost River Range to the northeast. The largest scarps and most complex patterns of ground rupture occur along the fault at the northeast margin of Thousand Springs Valley, between Elkhorn Creek on the southeast and Arentson Gulch on the northwest. At the Doublesprings Pass Road site, the ground breakage is typical of the surface faulting that accompanied the earthquake.

1 U. S. Geological Survey, Denver, CO 80225

Figure 1. Location of Stops 10 and 11 and generalized map of fault scarps and ground rupures associated with the Borah Peak earthquake. Heavy solid lines indicate prominent scarps, bar and ball on the downthrown side; dashed lines indicate poorly defined scarps or cracks. Stipple pattern shows valley bottoms; hachure pattern shows mounbtaneous part of Lost River Range and hills near Willow Creek Summit.

Trenching studies of the fault scarps conducted at this site are unique. In 1976 a trench, located about 200 feet (60 m) northwest of the road, was excavated across a Holocene fault scarp (Hait and Scott, 1978). During the 1983 earthquake, the backfilled trench was faulted, and a cross section of it was exposed in the newly formed fault scarps. Because of these rare circumstances, the 1976 trench was re-excavated and remapped in 1984 to study the structures associated with the new faulting, and the stratigraphy of the colluvium associated with the new scarp (Schwartz and Crone, 1985).

General Geologic Setting

The Borah Peak earthquake occurred in a region of typical basin and range topography in east-central Idaho. The Lost River Range and adjacent ranges to the northeast are composed of Paleozoic and Precambrian sedimentary rocks that were complexly folded and thrust faulted during the Mesozoic (Skipp and Hait, 1977). Cenozoic normal faults bound one or both flanks of the ranges (Skipp and Hait, 1977). Much of the present topography probably results from late Pliocene and Pleistocene displacements on the normal faults (M.H. Hait, Jr., 1984, written communication). The net Cenozoic vertical displacement on the Lost River fault in the vicinity of the 1983 earthquake is at least 1.6 miles (2.5 km), on the basis of 1.2 miles (1.9 km) of topographic relief between the summit of Borah Peak, at 12,662 feet (3,859 m) the highest point in Idaho, and the floor of Thousand Springs Valley, plus an estimated 0.4 to 0.6 mile (0.6-0.9 km) of Cenozoic fill in the valley.

The Doublesprings Pass road site is located on the Willow Creek alluvial fan that slopes gently toward the valley. This broad, smooth fan, composed of upper Pleistocene (Pinedale) glacial outwash, was probably active until about 15,000 years ago (Pierce and Scott, 1982). At least several thousand years after outwash deposition ceased, a surface faulting event displaced the fan surface 4.9 to 6.6 feet (1.5-2 m) (Hait and Scott, 1978). The vegetated, rounded slope directly above the free-face of the main 1983 scarp is the remnant of the scarp from this older earthquake.

Characteristics of the Surface Faulting

The zone of surface faulting at the Doublesprings Pass road site is 115 feet (35 m) wide and contains numerous en echelon scarps produced by displacements on both synthetic and antithetic faults. As a result, the ground in the fault zone is a broad depression internally broken into horsts and grabens (Fig. 2). The violent shaking during the earthquake locally shattered the ground surface in the fault zone into randomly tilted blocks up to several feet across. The height of individual scarps varies from a few inches to more than 6 feet (2 m) for the main scarp that bounds the upthrown block at the northeast edge of the fault zone. The height of the main scarp commonly
exceeds the true tectonic displacement because, in most places, a broad graben has formed in the adjacent downthrown block. Along strike, individual scarps typically diminish in height until the ground surface is warped; eventually the warping decreases to zero. As displacement on one scarp decreases, a corresponding increase usually occurs on an adjacent scarp.

The near-surface fault movement during the earthquake was dominantly normal slip with a subordinate amount of left-lateral slip (Crone and others, 1987). The best geologic measurements of these slip components are from the area between Elkhorn Creek and Arentson Gulch (Fig. 1). Here, at the base of large scarps, clasts dragged along the fault plane formed corrugations and grooves. The rakes of these grooves show about 7 inches (17 cm) of left-lateral slip per 40 inches (100 cm) of dip slip. At the Doublesprings Pass road site, the left-lateral slip component can be observed by matching the graded drainage channel along the side of the old roadbed across the fault zone. Also at this site, as well as at many other locations, en echelon scarps and cracks typically form a pattern of right-stepping offsets in plan view that is characteristic of left-lateral strike-slip displacements (Fig. 2).

The focal mechanism, based on data from seismograph records, indicates that the displacement at depth had a larger component of lateral slip than that measured at the surface. This suggests that, as the rupture propagated upward from the focus, a smaller proportion of the lateral-slip component propagated to the surface than did the dip-slip component.

The throw (vertical component of slip across the entire fault zone) varies considerably along the fault zone and is difficult to measure accurately at places such as Doublesprings Pass Road where there are multiple fault strands, local warping and backrotation, and extensive ground breakage. Measurements of the throw at this site, compensated for these complicating factors, are 5.2 feet (1.6 m) from geodetic data (Stein and Barrientos, 1985) and 7.9 feet (2.4 m) from displaced fluvial terraces (Vincent, 1985). The maximum throw along the entire fault zone, 8.2 to 8.9 feet (2.5-2.7 m) was measured near the base of Borah Peak about 2.2 miles (3.5 km) southeast of Doublesprings Pass road.

The highest single scarp that formed in 1983, nearly 16 feet (5 m) high, is located about 1,180 feet (360 m) northwest of the Doublesprings Pass road (Fig. 2). The height of the scarp exceeds the actual throw on the fault zone because of a small graben and backrotation of the ground surface on the downthrown block.

Figure 2. Vertical aerial photograph showing scarps and ground breakage near Doublesprings Pass Road (Stop 10). Labeled features are discussed in the text. Photograph taken October 29, 1983 (print compliments of Robert Whitney, Mackay School of Mines, University of Nevada, Reno).
geomorphically older appearance than the nearby simple scarp of the same age.

Most of the 1983 surface faulting followed pre-existing fault scarps, and in many locations the new ground breakage has mimicked the older scarps in amazing detail. This is especially apparent near the trench site just northwest of the road. Both the 1976 trench and pre-1983 aerial photographs show a well-developed horst within the graben adjacent to the main scarp. During the Borah Peak earthquake, the fault at the main scarp and the smaller faults bounding the horst were all reactivated, renewing the topographic expression of the horst (Fig. 3).

Significance of Trenching Studies

Trenching studies of prehistoric fault scarps have become increasingly valuable sources of paleoseismic data for earthquake-hazard assessments. However, often the stratigraphic relationships observed in trenches permit several plausible interpretations of the faulting history. Displacement of the 1976 trench in 1983 provided an unprecedented opportunity to re-excavate a previously mapped trench and to record, in detail, the effects of multiple surface-faulting events. The re-excavated trench clearly documents the stratigraphy of scarp-derived colluvial deposits that are used to interpret the history of faulting (Schwartz and Crone, 1985).

Detailed interpretations of the relationships in the re-excavated trench are still in progress, but two important conclusions about the behavior of the Lost River fault at this site are already clear. First, many small 1983 scarps, some only a few tens of inches high, overlie pre-1983 shear zones, indicating that even faults with small displacements were reactivated in 1983 (Fig. 3). This is additional evidence that the pattern of 1983 ground breakage was virtually identical to that associated with the prehistoric earthquake. Second, stratigraphic relationships in the trench show that the amounts of 1983 dip-slip displacement on most faults were similar to those

![Figure 3. Schematic diagram of the stratigraphy in the re-excavated trench at the Doublesprings Pass Road (Stop 10). Map shows south wall of trench. Scale in meters; no vertical exaggeration. Offset of distinctive silty gravel shows net movement from 1983 and prehistoric earthquakes; offset of pre-1983 earthquake ground surface shows displacement from Borah Peak earthquake (Stop 10).]
associated with the earlier earthquake. The similarities in pattern of ground breakage and in amounts of displacement suggest that the prehistoric earthquake probably had a magnitude similar to the Borah Peak earthquake. These similarities support the concept of characteristic earthquakes in which a specific fault or fault segment tends to generate similar magnitude large earthquakes and the surface faulting displacements associated with these large earthquakes are similar. In addition, the trenching studies at the Doublesprings Pass road site suggest that the pattern of ground breakage can be remarkably similar for successive large earthquakes. Thus, the trenching studies have provided some new and valuable insight into the behavior of a seismogenic normal fault in an extensional tectonic environment.

CHILLY BUTTES GROUNDWATER ERUPTIONS AND SAND BOILS (STOP 11)

The following discussion is taken from Crone and Pierce (1985), as modified from Waag (1985), and Youd and others (1985).

Location

This stop is located one mile west of the abandoned town of Chilly. To reach the site, turn west onto Trail Creek Road from Highway 93. Proceed west on Trail Creek Road for 2.2 miles and turn north onto the dirt road leading to the Chilly townsite. Proceed 0.7 mile north and turn west on the dirt road leading to Chilly Buttes Road. Turn north on Chilly Buttes Road and proceed 0.1 mile. Craters are exposed in field to the east.

Discussion and Description

Immediately after the earthquake, violent eruptions of groundwater from the limestone bedrock on the flank of Chilly Buttes and from the adjacent alluvial plain created spectacular sand boils and huge temporary springs. The vigorous flow erupted from the alluvial plain, reportedly spewed water 4-6 meters into the air, and left numerous craters, some more than 6 meters in diameter and more than 2 meters deep (Fig. 4). The flowing water transported limestone clasts with maximum dimensions of 15-20 cm. The stratigraphy exposed in the wall of a 3-by-5-meter-wide crater shows that the sediment deposited during the eruption was 25-35 cm thick. Following the earthquake there were at least 47 craters larger than 0.6 meter and 19 larger than 6 meters. The largest crater was 22 meters in diameter and about 5 meters deep. Some of the craters existed prior to the earthquake and were reactivated during the event.

During the seismic shaking and for at least 48 hours thereafter, water at high artesian pressures eroded from solution-widened fractures in the limestone on the flank of the butte. The washout and fracture at this site can be traced upslope for at least 175 meters. The highest eruption is about 30 meters above the valley. No measurements of the flow of water from the fractures are available but elutriation of all but the largest clasts and the 4 meter depth of the washout trench show that the velocity and volume of water were substantial.

REFERENCES


Field Guides to the Quaternary Geology of Central Idaho:
Part E.
History of Quaternary Faulting and Scarp Degradation Studies, Southern Lost River Valley

Kenneth L. Pierce

TRENCH SITE OF MALDE (1971):
DISPLACEMENT HISTORY OF THE ARCO SEGMENT OF THE LOST RIVER FAULT (STOP 12)

Descriptions, discussion, figures and roadlog for Stop 12 are modified from Crone and Pierce (1985), Pierce (1985), and Pierce and Colman (1986).

Location

Refer to Figure 1 of Evenson and others (1988, this volume) for the location of Stop 12. From Arco proceed north on Highway 93 towards Mackay. At 2.5 miles, turn east onto gravel road. In 0.1 mile this road curves to the north. Continue north for 3.8 miles and turn on two-track east up flat-bottomed valley and take two-track to the south up onto the fan south of the valley.

Description and Discussion

The fault scarp here is 16 to 23 meters high and has maximum slope angles of about 23 degrees to 26 degrees. The scarp is formed on old alluvial fan gravels (unit Qf01) and is partly buried by a younger "old" fan gravel (unit Qf02; Fig. 1) at the power poles west of the scarp.

Estimates of the rate of carbonate-coat accumulation and isochron dating at Stops 12 and 13 provide a means of estimating the ages of the alluvial gravels along the Arco segment (Table 1). At this stop, the soil on unit Qf01 has stones with carbonate coats about 10 mm thick (Pierce, 1985) and the soil on unit Qf02 has stones with carbonate coats that average 5 mm thick. At Stop 13, carbonate coats on Qfm1 clasts average 2.0 mm thick, those on Qfm2 clasts average 1.6 mm thick, and those on Qfy clasts average 1 mm thick (Pierce, 1985). Uranium-series dating (230Th/234U isochron method) by John Rosolt (written communication, 1980) of the inner, middle, and outer layers in the coats on clasts from unit Qf01 indicate that this unit has an age of about 160±30 Ka, and similar data suggest that Qfm1 has an age of 30-40 Ka (Pierce, 1985). An average carbonate-
coat accumulation rate of 0.6 mm/10 Ka can be calculated by assuming that the carbonate accumulates at a constant rate, and by using the uranium-series ages and the inferred late glacial age of 15 Ka for unit Qfy. Using this rate, the ages of units Qfo2 and Qfm2 are estimated to be 80±30 Ka and 25-30 Ka, respectively (Table 1).

Table 1. Carbonate coat thicknesses and estimated ages of alluvial fan gravels at Stops 12 and 13 along the Arco segment of the Lost River fault.

<table>
<thead>
<tr>
<th>NAME OF UNIT</th>
<th>THICKNESS OF CaCO₃ COAT (mm)</th>
<th>ESTIMATED AGE (Kilo-annum, Ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Young gravel (Qfy)</td>
<td>1.6</td>
<td>15±17</td>
</tr>
<tr>
<td>“Younger” middle gravel (Qfm2)</td>
<td>2.0</td>
<td>30-40</td>
</tr>
<tr>
<td>“Older” middle gravel (Qfm1)</td>
<td>3.0</td>
<td>30-35</td>
</tr>
<tr>
<td>“Younger” old gravel (Qfo2)</td>
<td>5.2</td>
<td>80±30</td>
</tr>
<tr>
<td>“Older” old gravel (Qfo1), Stop 12</td>
<td>10</td>
<td>160±30</td>
</tr>
</tbody>
</table>

1 based on inferred late glacial age (Pierce and Scott, 1982)
2 based on 39Ar/40Ar dating
3 based on carbonate-coat accumulation rate of 0.6 mm/10 Ka

The Qfo2 fan gravels were deposited on the downthrown side of the fault in response to the early faulting of unit Qfo1 (C. in Fig. 1). The Qfo2 gravels were later downfaulted against unit Qfo1 (D. in Fig. 1) as seen from exposures in the trench. Near the power poles, note that the surface of unit Qfo3 is clearly back-tilted toward the fault. The flat-bottomed drainage cuts through units Qfo1 and Qfo2 as well as through a small terrace remnant of unit Qfm2, a “younger” middle fan gravel (Pierce, 1985). West of the main gravel road, this drainage valley debouches onto the surface of unit Qfy, a young fan gravel of late Pleistocene age. Deposition of unit Qfy is considered to have ceased about 15 Ka (Pierce and Scott, 1982).

Figure 1 shows the history of faulting inferred from geologic relations in and near the trench, and at Stop 13 at King Canyon (Pierce, 1985). The total amount of displacement is largely dependent on the depth to the top of unit Qfo1 on the downthrown side of the fault. Fifty meters south of the power poles, a combined backhoe pit and auger hole penetrated colluvium, then thick colluvial sands, and finally a gravel with carbonate coats. If this gravel marks the top of unit Qfo1, the total vertical displacement is 18 meters. Two units of fault-scarp derived colluvium are exposed in the bottom of Malde’s (1971) trench. Each unit has a calcic soil that contains clasts with carbonate coats 1 mm thick. The geometry of the colluvial units indicates that they are resting on the surface of a horizontal gravel (inferred to be unit Qfo1) just below the bottom of the trench. These relationships suggest a vertical offset of 19-20 meters of unit Qfo1, which is 1-2 meters more than that determined at the backhoe pit-auger hole site.

The colluvial units are overlain by loess containing a volcanic ash dated as 76±34 Ka (N. D. Naeser, written commun., 1981). The chemistry of the ash indicates a source from the Yellowstone area of Wyoming (G. A. Isett, oral commun., 1982) where the youngest eruption occurred about 70 Ka. Unit Qfo1 had been offset by 4-7 meters by the time the ash was deposited (D. in Fig. 1) as indicated by the dip and thickness of the colluvial units and the ash-bearing loess.

The youngest surface faulting along the Arco segment occurred about 30 Ka based on relationships to be discussed at Stop 13. Unit Qfy (about 15 Ka) is found in numerous places across the Arco fault segment, but is not faulted.

Figure 2 shows the inferred history of faulting for the Arco segment of the Lost River fault. The estimated overall slip-rate for the past 160 Ka is 0.12 meters/Ka, about an order of magnitude less than that on the most active basin and range faults. Two contrasting patterns of displacement history can be interpreted using the seven control “points” on the plot of age versus offset (Fig. 2): (1) a pattern in which the timing of individual faulting events is relatively evenly spaced, or (2) a pattern in which events cluster in time (Pierce, 1985). Predictions of the next faulting event vary greatly, depending upon which pattern is considered valid.

One pattern of offset versus age assumes a constant strain accumulation (Fig. 2, solid line). Faulting occurs at time intervals directly related to the amount of accumulated strain, resulting in a stair-step pattern that has a slope similar to the overall long-term slip rate. If constant strain accumulation is assumed, about 3.6 meters of strain has accumulated since the fault last moved about 30 Ka. Based on typical offsets of about 1 meter during the Borah Peak earthquake (Crone and others, 1987), this amount of strain is nearly four times that expected to be released in a single faulting event. Even if offsets of 1-2 meters are assumed, additional surface faulting should have occurred about 10-20 Ka.

A second pattern of offset history assumes that multiple faulting events cluster in time and are separated by intervals of quiescence. Wallace (1984) suggests that this pattern of displacements may be characteristic of some faults in the Great Basin. A proposed history of such temporal clustering of activity is shown by the dashed line in Figure 2. This pattern seems more compatible with the most probable age of the volcanic ash (70 Ka) and with the past 30 Ka of quiescence on the Arco segment. If temporal grouping of events is characteristic of the Arco segment, then predictions of future activity are uncertain. The quiescence of the past 30 Ka may continue for many thousands of years, or may soon be interrupted by new faulting.

Back track to the main gravel road and turn north.

**Mileage**

1.1 STOP vehicles but REMAIN IN vehicles. The fresh gravel at the head of the steep fan above the fault scarp at 4:00 was deposited in the past few years. The gravel was deposited during debris avalanches that were generated on steep slopes higher
Figure 1. Schematic cross sections showing progressive displacement (A-D) on the Arco segment of the Lost River fault during the past 160 Ka. Sections are based on the trench map of Malde (1971) combined with additional data from Pierce (1985). Dot patterns show alluvial units discussed in the text. Buried soils shown by vertically lined pattern, volcanic ash shown by "xxx" symbol.
in the Lost River Range. The avalanches appear to have been generated in areas that were burned by fires in the early 1980s. At 2:00, about 1.6 km away, the head of the King Canyon fan is locally offset by a fault scarp that can be seen in early morning light. Pinedale moraines containing closed depressions extend to just below timberline upslope on King Mountain.

1.1 TURN RIGHT (east) on dirt road marked by the sign "Beaverland Pass, 3 miles". Leave the wire gate in the fence as you found it. The road here is on a fan of unit Qfy, deposited during the late Pleistocene episode of gravel deposition estimated to have ceased about 15 Ka (Pierce and Scott, 1982). The calcic soil on unit Qfy is 25 cm thick, has stage I carbonate development, and has 1 mm-

---

**Figure 2.** Seven "control points" for the offset history of the Arco segment, and two contrasting patterns of offset history (taken from Pierce, 1985). Control points are solid dots; coarse stipple shows error limits of geologic data. Solid line is model of episodic offset and constant strain accumulation; dashed line is model of episodic offset and temporal clustering. For simplicity, individual offsets are arbitrarily assumed to be 2 meters.
thick carbonate coatings on clasts.

0.2 Road ascends obliquely onto somewhat older fan surface (Qfm1). The soil on unit Qfm1 has 2 mm-thick carbonate coats on clasts and has an estimated age of 30-40 Ka (Table 1).

0.5 Park vehicles for Stop 13.

YOUNGEST FAULTING ON THE ARCO SEGMENT AND SCARP DEGRADATION STUDIES (STOP 13)

Descriptions, discussion, figures and roadlog for Stop 13 are modified from Crone and Pierce (1985), Pierce (1985), and Pierce and Colman (1986). Refer to Figure 1 of Evenson and others, this volume, for the location of Stop 13.

A fault scarp on unit Qfm1 has maximum slope angles of 11 to 15 degrees, and a vertical offset of 2-3 meters. Units Qfm2 and Qfy are not faulted. The estimated age of 30 Ka for the last surface faulting event on the Arco segment is based on two observations. First, the faulting resulted in the stabilization of the surface of unit Qfm1, thus allowing carbonate coats to start accumulating. Second, the faulting predated the surface of unit Qfm2 which has an estimated age of 25-30 Ka (Table 1). The fault scarp near the northern edge of unit Qfm1 has been buried by a small, steep fan. This fan was deposited immediately after unit Qfm1 was faulted, but before the drainage incised unit Qfm1 and deposited unit Qfm2. The soil on unit Qfm1 has stage I-II carbonate morphology that extends to a depth of about 1 meter.

About 300-400 meters to the southeast, the fault scarps are considerably higher where it is formed in a Qfo unit. However, the scarp was partially buried by unit Qfm1, reducing the height of the scarp. East of the fault, clasts in unit Qfo have carbonate coats that average 6-7 mm thick.

Influence of Slope Aspect on Terrace Scarp Degradation Rates

Near the modern drainage channel, unit Qfy was deposited in a flat-bottomed valley that was incised into older fan gravels. The incision and lateral erosion formed terrace scarps. The evolution of the terrace scarps, schematically shown in Figure 3, is inferred to be quite similar to the evolution of fault scarps. The incision initially undercut the older gravels to slopes steeper than the angle of repose. The unstable slopes quickly raveled to an angle of repose, here considered to be about 33.5 degrees. This value is based on high, late Pleistocene, north-facing scarps in King Canyon that have slopes within 0.5 degrees of this angle. After the scarp ravelled to the angle of repose, scarp degradation is dominated by colluvial and slope wash processes.

Pierce and Colman (1986) examine the inter-relationships between the height, the maximum slope angle, and the orientation of scarps. Using data from approximately 100 profiles of terrace scarps of probable late Pleistocene age, they show that south-facing scarps have degraded to lower angles than west-facing scarps, which in turn, have degraded more than north-facing ones (Fig. 4). About one-third of the profiles used in the study are from King Canyon and the remainder are from Ramshorn Canyon about 4 miles to the north. The maximum slope angle of south-facing scarps is about two-thirds the angle of north-facing scarps (Fig. 4). For example, 10 meters-high, south-facing scarps have a

Figure 3. Stages in the evolution of a 5 meters-high terrace scarp. Stages are: (1) lateral undercutting by incised stream forms terrace scarp steeper than the angle of repose, (2) oversteepened slope rapidly raveling to rectilinear scarp at angle of repose (starting at 33.5 degrees), (3-5) scarp profiles and maximum slope angles predicted by diffusion-equation model for rate coefficient (c) of $12 \times 10^{-4} m^2/yr$ at labeled time intervals.
Figure 4. Relation between maximum scarp angle (θ) and log of scarp height (h) for scarps that face north (N), south (S), and west (W). See text for further explanation. Symbols for regression analyses: n, number of scarps in a group; SD, standard deviation of y about the regression line; r², coefficient of determination adjusted for degrees of freedom. Two highest south-facing scarps (circled by dashed lines) are excluded from regression analysis because they may be younger than other scarps in the same data set.
maximum angle of about 20 degrees, whereas equally high north-facing scarps have an angle of 30 degrees.

In the field there are obvious contrasts in the amount and type of vegetation on scarps with different orientations. High south-facing scarps have a desert shrub vegetation that leaves about 70 percent of the ground exposed. High north-facing scarps have a prairie grassland vegetation and a tough, root-bound turf that leaves about 10 percent of the ground exposed.

The difference in vegetation is controlled by soil moisture, which in turn is related to differences in solar radiation. The fast degradation rates on south-facing scarps probably result primarily from their poor vegetation cover and their large number of freeze-thaw cycles.

The diffusion equation has been applied to model the degradation of fault scarps (Colman and Watson, 1983; Hanks and others, 1984). Pierce and Colman (1986) show that not all the processes involved in the erosion of high scarps are modelled by the linear diffusion equation (Fig. 5). Instead scarp orientation, which controls the types and rates of slope processes through the microclimatic effects of freeze-thaw cycles, soil moisture and vegetation, has profound effects on the degradation of fault scarps (Crone and Pierce, 1985).

REFERENCES


Field Guides to the Quaternary Geology of Central Idaho:
Part F.
Holocene Volcanism at Craters of the Moon National Monument

George C. Stephens¹

LOCATION

Craters of the Moon National Monument is about 20 miles southwest of Arco on Idaho Highway 20-26-93. Refer to Figure 1 of Evenson and others (1988, this volume) for the locations of Stops 14 through 18.

INTRODUCTION

The rocks and landforms of Craters of the Moon National Monument (Fig. 1) are the result of Holocene volcanism on the Snake River Plain. For further discussion of Snake River Plain Quaternary volcanism see Hackett and Morgan (1988, this volume). The Craters of the Moon are located on the north end of the Great Rift (Fig. 2; Kuntz and others, 1982, 1986). The Craters of the Moon lava field contains more than 40 separate flows which range in age from 15 to 2 Ka. The lava flows cover approximately 1,650 square kilometers, making the Craters of the Moon lava field the largest Holocene basalt field in the conterminous United States.

Volcanism at Craters of the Moon occurred both as effusive lava flows and as pyroclastic eruptions from cinder cones. The numerous cinder cones at Craters of the Moon are asymmetric in shape because winds from the south-southwest caused the accumulation of volcanic ejecta on the downwind sides of the growing cones.

¹ Department of Geology, George Washington University, Washington, D.C. 20052

Figure 1. Location map of the Snake River Plain showing major geologic provinces, cities, and Craters of the Moon National Monument.
LOCALITY DESCRIPTIONS
(STOPS 14-18)

This guide briefly describes noteworthy features along the Craters of the Moon loop road. An excellent self-guided tour is contained in a booklet available at the monument headquarters (Henderson, 1986). From Idaho Highway 20-26-93, turn southeast at the Craters of the Moon National Monument entrance and follow signs to the visitor center. Figure 3 shows the locations of the field trip stops and major volcanic landforms in the monument.

Visitor Center (Stop 14)

Several interesting self-guided exhibits explain the geology and ecology of Craters of the Moon National Monument. A short film describing volcanic processes and the resulting volcanic rocks and landscapes is shown in the theater at the visitor center. Topographic maps of the monument as well as a variety of general interest scientific publications and field guides are available for purchase. Restrooms, water fountains and soft-drink machines are also located at the visitor center.

Continue southeast from the visitor center, through the entrance gate and past the campground on the right, and follow signs for the loop road.

North Crater Flow (Stop 15)

A paved 0.4 km-long foot-trail loops across the North Crater lava flow at this locality. Photogenic exposures of both pahoehoe and aa lavas are seen on this short walk. Pahoehoe lavas flow at temperatures of about 1,000 degrees C and can travel at velocities greater than 50 km/hr (Henderson, 1986). The ropy, folded surface of pahoehoe lavas indicate the direction of lava movement. The folds are convex in the down-stream direction.

Aa flows result from slightly lower lava temperatures (approximately 923°C) and more viscous fluid behavior (Henderson, 1986). Aa flows typically travel at speeds of 5-16 km/hr. As an aa lava flow moves forward, chunks of solidified lava from the top and the leading edge of the flow tumble down the front and are incorporated in the viscous material. Large blocks of lava are present near the far end of the foot-trail loop. These blocks, several meters in size, were carried by the viscous aa flow from the wall of North Crater to their present location. Other large monoliths from North Crater were carried farther eastward by the Devil’s Orchard and Serrate flows of approximately the same age (Kuntz and others, 1982).

The North Crater flow of Henderson (1986) is one of the youngest lava flows of the Great Rift. It belongs to Eruptive Period A of Kuntz and others (1982) and has a probable age of 2.2-2.3 Ka. The “Triple Twist Tree”, which grew on the surface of the North Crater flow adjacent to the trail at Stop 15, has been studied by dendrochronology and contains 1,350 annual growth rings. The relatively small size of this tree belies its ancient age. Note also the general lack of vegetation on the North Crater flow as compared to older lava flows exposed elsewhere in the monument.

Return to the parking area and continue southeast.
toward the loop road, past North Crater (on the right), source of the North Crater flow seen at Stop 15. Continue past the turn-off on left for Devil's Orchard Nature Trail and enter the one-way loop road. Turn left and park at the parking area for Inferno Cone.

**Inferno Cone (Stop 16)**

Inferno Cone is a large, undissected cinder cone. A short, steep walk leads from the parking lot to the top of the cone. Visible from the top are cinder cones extending along the Great Rift toward the southeast. Nearby cones include Broken Top Cone, Big Cinder Butte, Half Cone and Crescent Butte (Fig. 3). The view toward the northwest includes Paisley Cone, North Crater and Sunset Cone. In the distance, the Pioneer Mountains and Lost River Range are visible beyond the northwestern edge of the Snake River Plain.

Notice the asymmetric shape of Inferno Cone, with its steep profile on its western to southwestern flank and the...
gentler more extensive wind-blown cinder accumulations on the east to northeast flank.

Return to the parking area and continue a short distance on the loop road to the parking area for Big Craters and the Spatter Cones.

**Spatter Cone Area (Stop 17)**

The group of cinder cones north of the parking lot is known as Big Craters. These are typical fissure-related cones, similar to Inferno Cone. Several small spatter cones are located a short distance south of the parking area. The cinder cones formed early in an eruptive cycle from volatile-rich, gas-charged lavas. The spatter cones formed later from volatile-poor, pasty and viscous lavas (Henderson, 1986). Small masses of pasty lava adhered to one another as they accumulated near the base of their vents, building these characteristically steep-sided spatter cones. The absence of direct sunlight and the insulating properties of the surrounding volcanic rocks allow ice and snow accumulations within the deeper spatter cone vents to be preserved throughout the summer.

Return to the vehicles and continue around the loop drive to the parking lot for the lava tube area.

**Blue Dragon Flows (Stop 18)**

A fissure slightly south of Big Craters and the spatter cones of Stop 17 was the source of the Blue Dragon flows—pahoehoe lava flows with a characteristic iridescent blue surface that erupted about 2.1 Ka (Kuntz and others, 1982). The Blue Dragon flows have a surface area of approximately 320 square kilometers and are as much as 15 meters thick (Henderson, 1986). One lobe of the Blue Dragon lavas flowed southwestward from the Big Craters area, and another, larger group of flows moved east and southeast from the source (Kuntz and others, 1982).

The largest volume of the Blue Dragon flows is located to the east and southeast of the source area. Much of the lava flowed in large tubes (Kuntz and others, 1982), such as the ones seen at this locality. A short (0.8 km) walk leads to Dewdrop, Boy Scout, Beauty, and Surprise Caves and Indian Tunnel. During this walk, notice the characteristic color of the Blue Dragon flows and also the well-developed vertical fractures which are radial to the ropy "flow folds". These are extensional fractures formed by lateral flowage as the ropy surface advanced.

We will visit Indian Tunnel to examine the interior of a characteristic lava tube. Note the lava "stalactites" on the ceiling and the evidence of lava movement on the tube walls. Indian Tunnel is the only lava tube accessible without a flashlight.

Return to the parking area and continue around the loop road to its junction with the road to the visitor center. Retrace your route northwestward, to Highway 20-26-93. Turn right for Arco or left to return to Carey and Sun Valley.

**REFERENCES**


Chapter Five
Geology of the Snake River Plain

The Holocene Kings Bowl lava field, looking south along the Great Rift on the eastern Snake River Plain. Aerial photograph by W. R. Hackett.
INTRODUCTION

This three-day field trip is designed primarily to review the products of the extensive silicic volcanism that accompanied the evolution of the central and western portions of the Snake River Plain. The silicic volcanic rock localities that will be visited include nonwelded and welded ash-flow tuff units and rhyolite lava flows and domes. These classic Miocene and Pliocene volcanic sequences are exposed along the margins of the Snake River Plain and in canyons incised within its interior. As the trip progresses, visits to, and comments regarding, the basaltic volcanism, the structural development, the distribution of sediments, and the geomorphic features of the Snake River Plain volcanic province will also be made. The general location and route of the entire trip is indicated in Figure 1.

Day One

The first day of the trip, led by William P. Leeman, will emphasize the silicic volcanic rocks and the associated ferrolatites and basalts of the Magic Reservoir eruptive center. This area of volcanism lies along the northern margin of the Snake River Plain, in the eastern part of the Camas Prairie, a complex east-west graben, and in the Bennett Hills, an east-west horst situated between the Snake River Plain and the Camas Prairie. During the latter part of the first day, the trip will cross the central Snake River Plain from north to south, offering a view of various basalt features and providing an opportunity to visit the large rhyolite flow exposed at Shoshone Falls on the Snake River.

Day Two

The second day of the trip, led by Bill Bonnichsen, will concentrate on the silicic volcanic rocks and structural relations along the southern margin of the central Snake River Plain, both in the north-south graben between Rogerson and Jackpot, and farther west, within the Bruneau-Jarbidge eruptive center. This part of the trip will emphasize the general appearances of the welded-tuff units and rhyolite lava flows, and the distinctions between them. The variations in the types of internal structures within welded-tuff sheets that are dependent upon emplacement temperatures will also be stressed. During the latter part of the second day we will...
go back into the interior of the plain and view several basalt shield volcanoes on the way to Castleford. We will visit the rhyolite flows at Balanced Rock near Castleford and then proceed down Snake River canyon, where Pliocene and Pleistocene basalt flows, sedimentary units, and accompanying geomorphic features can be viewed.

Day Three

The third day of the trip, led by Bill Bonnichsen, will concentrate on the Browns Creek area, along the southwestern margin of the western Snake River Plain graben. Here, a rhyolite lava flow, a series of unwelded, pumice-rich, ash-flow tuff units, and a rheomorphically deformed, densely welded, ash-flow tuff sheet occur in close proximity to one another. By comparing these units, participants will gain an appreciation of the problems involved in distinguishing rhyolite lava flows from rheomorphically deformed welded-tuff units. During the latter part of this day participants will be introduced to the problem of identifying "water-affected" basalt, which is common in the western Snake River Plain, and will be shown views of several basalt volcanoes, some of which erupted into Lake Idaho, a large lake that existed in this part of Idaho between the late Miocene and the early Pleistocene.

DAY 1: SUN VALLEY TO GOODING VIA MAGIC RESERVOIR ERUPTIVE CENTER

Introduction

During this portion of the field trip we will examine the lithologies, stratigraphic relations and volcanological features of volcanic deposits of the Magic Reservoir area, which lies along the north-central margin of the Snake River Plain. This area is a structurally complex zone where the western and eastern limbs of the Snake River Plain converge. Here, northwesterly, east-west, and northeasterly trending fault systems are well developed. The silicic volcanism in the area is anomalously young compared to the general age distribution of rhyolites in the Snake River Plain-Yellowstone Province.

The Neogene volcanic rocks in the Magic Reservoir area overlie a basement of Paleozoic sedimentary rocks, Cretaceous and Tertiary granitic plutons, and erosional remnants of Eocene Challis volcanic rocks and associated sediments. The Neogene deposits include an early sequence (about 9 to 10 million years old) of rhyolitic ash-flow tuffs (Idavada Volcanics) erupted from sources to the south in the Snake River Plain. These rocks are exposed in upfaulted highlands along the margins of, and generally dip toward, the Snake River Plain. Ash-flow tuffs similar to the Idavada Volcanics progress in age from about 14 million years old in southwestern Idaho to 0.6 million years old at Yellowstone, and they record the earliest phase of the time-transgressive Snake River Plain volcanic activity across the province (Armstrong and others, 1975). Rhyolitic lavas and tuffs, basaltic lavas, and hybrid lavas younger than 6 million years in age were erupted in the Magic Reservoir area. This volcanic rock suite is essentially bimodal, as the intermediate-composition magmas were formed by the mixing of silicic and mafic magmas. Because of their anomalous age-setting and the constraints they provide on the magmatic and tectonic development of the Snake River Plain, the volcanic rocks from this eruptive center have been studied in detail (Honjo and others, 1986; Honjo and Leeman, 1987; Leeman and Honjo, unpublished data).

The volcanic stratigraphy for the Magic Reservoir eruptive center has been defined, and the distribution of the units there has been mapped in varied detail, by Schmidt (1961), Malde and others (1963), Smith (1966), Struhsacker and others (1982), and Leeman (1982). The generalized stratigraphic relations among the map units of some of these authors (slightly modified in this guide) are listed in Table 1, which also summarizes the available K-Ar age determinations (cf. Honjo and others, 1986). The volcanic section in the Magic Reservoir area consists of the following: (A) the Idavada Volcanics that consist of welded-tuff units and minor intercalated basaltic lavas (10 to 9 million years old); (B) the quartz latite lavas, hybrid ferrolatite lavas, and high-silica rhyolite tuffs and young domes (6 to 3 million years old) associated with the development of the Magic Reservoir eruptive center; and (C) the young basaltic lavas that are similar in composition to other basalts in the Snake River Plain. Figure 2 shows the areal distribution of these units.
The heavy line on Figure 2 traces the proposed elliptical boundary of the Magic Reservoir eruptive center. This boundary has been considered as a possible caldera margin (Leeman, 1982), but it is unlikely that this eruptive center is a typical caldera. This boundary coincides with the locations of several young rhyolite domes, several large basaltic shields, and the large hot spring at Hot Springs Landing. The Magic Reservoir eruptive center is bounded on the northeast by a major fault with subsidence to the southwest. Cretaceous granites are exposed in the upthrown block to the northeast, and correlative ferrolatite lavas (Tsf) are offset by at least 500 feet across this fault (Fig. 3). The inferred western edge of the eruptive center coincides with the easternmost exposures of Idavada Volcanics and older rocks. Elsewhere, the boundary is not well defined due to the cover of younger deposits. The interior of the Magic Reservoir eruptive center subsided and was filled by the voluminous rhyolite of Magic Reservoir and younger deposits (Fig. 4). The central portion of the eruptive center (in the easternmost Mount Bennett Hills) was subsequently uplifted along a major east-to-west-trending fault with downward displacement to the north. This fault, which offsets the ferrolatite lavas by several hundred feet, forms the southern boundary of the Camas Prairie. Pliocene ash-flow tuffs and related pyroclastic deposits were erupted from vents within the Magic Reservoir eruptive center (Fig. 5), but the volume of

<table>
<thead>
<tr>
<th>Table 1. Volcanic stratigraphy in the Magic Reservoir area.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>BELLEVUE AREA</strong></td>
</tr>
<tr>
<td>-----------------------------------------------------------</td>
</tr>
<tr>
<td>Recent</td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td>Pleistocene</td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td>Pliocene</td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td>Miocene</td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td>Eocene</td>
</tr>
<tr>
<td>Cretaceous</td>
</tr>
</tbody>
</table>

(+) indicates normal magnetic polarity, (-) indicates reversed magnetic polarity
these materials is small in comparison with the rhyolites at Yellowstone or at eruptive centers within the Snake River Plain.

The compositions of the major igneous units listed in Table 2 are averages of multiple samples. As a function of decreasing age, the rhyolites associated with the Magic Reservoir eruptive center become more siliceous and their phenocryst assemblages more hydrated; their calculated mineral equilibrium temperatures decrease from about 1000 degrees C (Tmr) to 700 degrees C (Tyd). These trends are consistent with progressive fractionation of the magmas. The Square Mountain Ferrolatite and some of the rhyolite of Magic Reservoir samples exhibit petrographic and compositional features that are consistent with magma mixing. Figure 6 illustrates some of the disequilibrium textures, including resorbed feldspars and reacted quartz xenocrysts. The ferrolatites contain abundant crustal xenoliths, some of which are described in Leeman and others (1985). The xenoliths are Archean in age (about 2.8 billion years old), based on Pb-Pb and Sm-Nd studies. Honjo and Leeman's (1987) detailed study of the ferrolatites suggests that they formed by the mixing of Magic Reservoir-type rhyolite and typical Snake River Plain basaltic magmas shortly before the "young tuff" unit was erupted. Violent stirring may account for the homogeneity of the ferrolatites and their entrainment of crustal xenoliths.
Road Log: Sun Valley To Gooding

The first day’s trip will allow participants to examine representative examples of most of the main stratigraphic units. Figure 7 shows the locations of the stops planned in the Magic Reservoir eruptive center.

Drive south from Sun Valley on Idaho Highway 75 to the junction with U. S. Highway 20, a distance of about 26 miles. This route passes through the Big Wood River Valley and the towns of Hailey and Bellevue. Other field trips in this volume that discuss the southern Big Wood River valley include Foley and Street (1988), Johnson and others (1988), and Link and others (1988). The southern end of this valley is filled to a great depth with alluvial and glacial outwash deposits which are intercalated with numerous thin basalt flows (Schmidt.
The mountains on both sides of the road consist mainly of Paleozoic shelf sediments and deep-water shales which host the disseminated and vein base metal and Au-Ag deposits of the Hailey Gold Belt. Many abandoned and reactivated mine workings can be seen in this vicinity.

Approaching the Highway 20 junction you will note a bank of low rolling hills due south. These are the Timmerman Hills, and they consist of poorly exposed volcanic rocks of the Challis Volcanics and deeply weathered granitic rocks of the Idaho batholith overlain by remnants of the Picabo Tuff of Schmidt (1961). Several miles south of the junction a graded road leads west from Highway 75 to the resort of Magic City (known locally as Baja Magic). This road leads to the fault scarp which bounds the east side of Magic Reservoir and provides a good overview of the eruptive center. A few miles south of the Magic City turnoff, and on either side of Highway 75, are several rhyolite domes (about 3 to 4 million years old) that are distinguished by their craggy profiles. A small quarry at the southern base of the largest of these domes (Wedge Butte), a short distance east of Highway 75, provides easy access to examine the rhyolite. Our travel route, however, will take us west on Highway 20 to where similar rhyolites will be seen.

Figure 5. Panoramic overview looking south from the top of the Moonstone Mountain rhyolite dome complex (Tyd). The low, flat area at the right side of the view is the eastern end of the Camas Prairie. The bounding fault along the northeastern margin of the Magic Reservoir eruptive center is visible in the center of this view. The craggy exposures in the foreground are rhyolites of the dome complex.
Mileage

0.0 Road log mileages are measured from the junction of Highways 20 and 75.
Proceed west from the junction on Highway 20 and cross the Big Wood River in about one mile. In this area are two intracanyon basalt flows, the older is the Upper Wind Ridge Basalt and the younger, which forms part of the valley floor, is the Lower Wind Ridge Basalt. The younger basalt is better exposed to the west. Most of the basaltic lavas associated with the Magic Reservoir eruptive center are olivine tholeiites of roughly similar lithology and composition; they are discussed at Stop 6.

1.9 Here we pass by road cuts in biotite granodiorite (Kg) that probably is Cretaceous in age. This rock is compositionally similar to the analyzed samples from other Idaho batholith localities (Table 2), although it has not been dated. At Stop 4 we will see similar granitic rocks.

4.4 At the intersection with the graded road to Hot Springs Landing, on the north arm of Magic Reservoir, turn south and proceed 1/2 mile to the parking area near the boat landing, which is the site of a large hot spring. This will be Stop 1. A relatively detailed geologic map of this area is available in Struhsacker and others (1982) and in Foley and Street (1988, this volume).

For some years the 79-m-deep well at Hot Springs Landing has remained uncapped. It produces some 57 liters per minute of the hottest (71
degrees C) water in the area (Mitchell and others, 1980). This water is distinguished by an unusually high boron content and, based on its chemical composition, seems to have equilibrated with rhyolitic reservoir rocks (L. Street, personal communication, 1987).

The low hills to the west-northwest consist primarily of various members of the "young tuff," including several ash-flow-tuff cooling units and an overlying plinian pumice deposit (cf. Struhsacker and others, 1982). At low reservoir levels, one of the ash-flow units is exposed just west of the landing. The pumice deposit and two ash-flow cooling units will be seen at the next stop.

Stratigraphic relations in the Hot Springs Landing area establish the relative ages of almost all of the eruptive units associated with the Magic Reservoir eruptive center. The "young tuff" section locally overlies the rhyolite of Magic Reservoir, which has been dated as 5.81 ± 0.69 million years old. The lower member of the "young tuff" and a rhyolite dome which crosscuts the "young tuff" section have been dated as 5.64 ± 0.23 and 4.77 ± 0.29 million years old, respectively (Struhsacker and others, 1982). This area straddles the northwest-trending fault zone which bounds the eastern margin of the reservoir. Because of the placement of the rhyolite domes, the occurrence of near-vent eruptive deposits, and the presence of

Table 2. Average compositions of geologic units in the Magic Reservoir area.

<table>
<thead>
<tr>
<th>Analyses</th>
<th>Idaho Batholith (Kg) (7)</th>
<th>Idaho Adava Volcanics (Tiv) (7)</th>
<th>Rhyolite of Magic Reservoir (Tmr) (6)</th>
<th>Square Mtn Ferrolatite (Tsmf) (12)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>66.32 (4.00)</td>
<td>71.31 (2.48)</td>
<td>68.06 (0.43)</td>
<td>59.67 (0.66)</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.58 (0.17)</td>
<td>0.55 (0.13)</td>
<td>0.70 (0.07)</td>
<td>1.80 (0.05)</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>15.86 (1.27)</td>
<td>12.51 (0.35)</td>
<td>13.37 (0.19)</td>
<td>14.57 (0.14)</td>
</tr>
<tr>
<td>Fe₂O₃*</td>
<td>3.98 (1.05)</td>
<td>3.64 (1.08)</td>
<td>5.13 (0.49)</td>
<td>9.21 (0.48)</td>
</tr>
<tr>
<td>MnO</td>
<td>0.07 (0.02)</td>
<td>0.06 (0.01)</td>
<td>0.08 (0.01)</td>
<td>0.13 (0.01)</td>
</tr>
<tr>
<td>MgO</td>
<td>1.69 (0.68)</td>
<td>0.40 (0.21)</td>
<td>0.46 (0.05)</td>
<td>1.91 (0.15)</td>
</tr>
<tr>
<td>CaO</td>
<td>3.71 (1.25)</td>
<td>1.51 (0.53)</td>
<td>2.14 (0.19)</td>
<td>5.06 (0.25)</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.66 (0.40)</td>
<td>2.88 (0.20)</td>
<td>3.30 (0.16)</td>
<td>3.29 (0.11)</td>
</tr>
<tr>
<td>K₂O</td>
<td>3.20 (0.79)</td>
<td>5.06 (0.32)</td>
<td>5.00 (0.22)</td>
<td>3.14 (0.12)</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.18 (0.06)</td>
<td>0.09 (0.04)</td>
<td>0.19 (0.02)</td>
<td>0.40 (0.03)</td>
</tr>
<tr>
<td>LOI</td>
<td>0.31 (0.10)</td>
<td>1.78 (0.61)</td>
<td>1.30 (0.22)</td>
<td>0.52 (0.05)</td>
</tr>
<tr>
<td>Total</td>
<td>99.43</td>
<td>99.47</td>
<td>99.31</td>
<td>99.27</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Analyses</th>
<th>Young tuff (Tyt) (4)</th>
<th>Young rhyolite domes (Tyd) (5)</th>
<th>McHan Basalt (2)</th>
<th>Olivine Tholeiites (19)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>74.61 (0.27)</td>
<td>75.27 (0.95)</td>
<td>50.52 (0.61)</td>
<td>46.97 (0.94)</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.24 (0.01)</td>
<td>0.13 (0.05)</td>
<td>1.75 (0.04)</td>
<td>2.59 (0.63)</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>12.57 (0.09)</td>
<td>12.47 (0.32)</td>
<td>16.08 (0.12)</td>
<td>15.07 (0.67)</td>
</tr>
<tr>
<td>Fe₂O₃*</td>
<td>2.19 (0.07)</td>
<td>2.02 (0.40)</td>
<td>13.45 (0.20)</td>
<td>14.38 (1.50)</td>
</tr>
<tr>
<td>MnO</td>
<td>0.03 (0.01)</td>
<td>0.03 (0.01)</td>
<td>0.19 (0.01)</td>
<td>0.20 (0.01)</td>
</tr>
<tr>
<td>MgO</td>
<td>0.11 (0.03)</td>
<td>0.03 (0.03)</td>
<td>4.91 (0.14)</td>
<td>7.37 (0.90)</td>
</tr>
<tr>
<td>CaO</td>
<td>0.64 (0.10)</td>
<td>0.56 (0.17)</td>
<td>7.45 (0.24)</td>
<td>10.12 (0.59)</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.24 (0.03)</td>
<td>3.60 (0.35)</td>
<td>3.82 (0.04)</td>
<td>2.57 (0.17)</td>
</tr>
<tr>
<td>K₂O</td>
<td>5.55 (0.07)</td>
<td>5.00 (0.16)</td>
<td>1.69 (0.09)</td>
<td>0.55 (0.16)</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.03 (0.00)</td>
<td>0.05 (0.03)</td>
<td>0.44 (0.01)</td>
<td>0.55 (0.27)</td>
</tr>
<tr>
<td>LOI</td>
<td>(n.a.)</td>
<td>0.48 (0.09)</td>
<td>(n.a.)</td>
<td>(n.a.)</td>
</tr>
<tr>
<td>Total</td>
<td>99.20</td>
<td>99.34</td>
<td>100.28</td>
<td>100.38</td>
</tr>
</tbody>
</table>

All analyses were determined by XRF (Edinburgh University) except for Mg which was analyzed by ICP (Rice University). Numbers in parentheses indicate standard deviation; (n.a.) means not analyzed.
Figure 6. Photomicrographs of representative reaction and disequilibrium textures observed in samples from the rhyolite of Magic Reservoir and the Square Mountain Ferroliteite. A: Partially resorbed anorthoclase phenocryst in the rhyolite of Magic Reservoir showing tartan twinning and a plagioclase core. B: Sieve-texture in a plagioclase phenocryst in the Square Mountain Ferroliteite. C: Quartz xenocryst in the Square Mountain Ferroliteite surrounded by aggregates of Mg-rich clinopyroxene. D: Resorbed and sieve-textured anorthoclase xenocryst in the Square Mountain Ferroliteite (core is unaltered). Each view is 4 mm wide.
geothermal activity, this fault zone has been considered analogous to a caldera-margin, ring-fracture structure (Leeman, 1982).

Return to the highway and proceed west.

6.0 Stop 2 is located at the pumice quarry on the south side of the highway. Here we will examine the pumice deposits in the quarry and then walk a short distance to the road cuts along Highway 20, where multiple cooling units within the "young tuff" are exposed. These outcrops are good examples of the near-vent facies of ash flows and of related pyroclastic deposits. The features of note are the zonation in the degree of welding, or lack thereof, and the variations in the size and lithology of the clasts. Similar deposits are exposed sporadically in the foothills of the eastern Mount Bennett Hills (to the south-southwest), in the hills at the southeast end of Magic Reservoir, and in the low hills to the north. The coarse, unsorted, and angular nature of the capping pumice deposit suggests that we are near the vent. The ash flows contain xenoliths of the rhyolite of Magic Reservoir, of the Square Mountain Ferrolatite, and of basement rocks (granitoid rocks and rare gneisses). The volcanic units at this stop are relatively crystal-poor, high-silica rhyolites (Table 2) containing phenocrysts of sanidine, plagioclase, quartz, orthopyroxene, minor opaques, and traces of apatite and zircon in a glassy, eutaxitic groundmass.

Return to the vehicles and proceed west on Highway 20.

6.7 The road crosses Poison Creek here. Along this creek, south of the highway, are good exposures of the lower part of the "young tuff" section. Proceed west past the Moonstone Ranch. The prominent craggy hill north of the highway is Moonstone Mountain, a dome-flow complex. Two samples from the north and southwest sides of the dome have been dated as 3.88 +/- 0.06 and 3.02 +/- 0.05 million years old, respectively (Honjo and others, 1986).

8.7 Turn right onto the graded ranch road and proceed 0.7 mile north toward Moonstone Mountain.

9.4 Stop 3 is at the intersection with the ranch road at the base of the rhyolite lava flow on the southeast side of Moonstone Mountain. At least seven young domes are distributed in an arcuate pattern along the northern and eastern margins of the Magic Reservoir eruptive center and in a more diffuse cluster to the southeast. Dates from these domes suggest that they were emplaced between 4 and 3 million years ago (Honjo and others, 1986). Except for variations in the degree of devitrification and vapor-phase alteration, most of the domes are lithologically uniform. However, the Moonstone Mountain complex consists of several distinct emplacement units which do vary slightly in composition, although they have not been mapped in detail. The high-silica rhyolites (Table 2) of the young domes typically are extremely porphyritic, containing abundant large phenocrysts of smoky quartz and sanidine, small amounts of plagioclase, amphibole, biotite, and opaques, and traces of apatite, zircon, allanite, and fayalite. Chevkinite, an unidentified Nb oxide (up to 20% Nb₂O₃), and at least two other unidentified accessory phases have been observed in the young dome samples. The young domes are usually extensively devitrified. At this locality we will examine part of a lava flow in which an upper vitrophyre selvage is locally preserved.

Return to Highway 20 and proceed west across the surface of the loess-covered Macon Basalt, which consists of several successive flows. Paleomagnetic measurements indicate normal polarity for these basalts, and one sample has been dated as 0.9 million years old (Armstrong and others, 1975).

11.2 The road crosses Camp Creek here. The Macon Basalt is exposed along this creek.

14.1 Camas County-Blaine County line. Turn right here on the graded road to the north.
16.0 Turn right (north) at the intersection with the Camp Creek road. We will drive through exposures of the Square Mountain Ferrolatite lavas. The thicknesses and distribution of these lavas were apparently controlled by pre-existing topography. They probably flowed from vents in this area, but actual vent locations are not known with certainty. On the east side of Square Mountain and north of Moonstone Mountain, the local concentrations within the lavas of large xenoliths of crustal materials are thought to be proximal to vents because it is unlikely that the xenoliths could have been transported far. Similar lavas occur in several areas: in the hills east of Magic Reservoir, in the foothills on the north and northeast sides of the eastern Mount Bennett Hills (where small crustal xenoliths have also been found), along the southwestern margin of the Magic Reservoir eruptive center, and in a large area north of Moonstone Mountain. All of the ferrolatite lavas are remarkably similar in spite of their wide distribution, and evidence for ferrolatite vents has not been recognized in any part of the Magic Reservoir eruptive center, except near Square Mountain.

Stratigraphic and field relations near Stops 1 and 2 and in the eastern Mount Bennett Hills indicate that the ferrolatite lavas are approximately 5.7 million years old, because they postdate the rhyolite of Magic Reservoir and predate the “young tuff.” The ferrolatites (or “tholeiitic andesites”) (see Table 2 for composition) are rich in phenocrysts, including plagioclase, augite, pigeonite, minor hypersthene and magnetite, and sparse ilmenite. They also contain markedly embayed xenocrysts, up to 3 mm in diameter, of anorthoclase (commonly sieve textured) and quartz, and relics of Nb-rich ilmenite. The hyalopilitic groundmass contains sodic plagioclase, pyroxenes, and Fe-Ti oxides. Detailed microprobe analyses of these minerals and the bulk compositions of the ferrolatite lavas suggest that the magmas were mixtures of subequal proportions of the rhyolite of Magic Reservoir (Tmr) and olivine tholeiite basalt (Honjo and Leeman, 1987).

18.0 Where we cross the first of two small bridges over Camp Creek, the thick flows of the Square Mountain Ferrolatite contain a “cylindrical” jointing pattern.

18.9 Stop 4 is at the roadcut in foliated biotite granodiorite. We will cross the creek at this point, climb 500 feet to the top of Square Mountain, and walk to its southeast corner where large crustal xenoliths are abundant in the Square Mountain Ferrolatite. The total walking time will be about one hour. This excursion will take us through a poorly exposed section of granitic rocks to the unconformable ferrolatite cap on the west side of Square Mountain. We will walk to one of the better xenolith localities where gneissic blocks in excess of 1 m in diameter can be observed; locally such blocks constitute more than 20 percent of the outcrop. Many xenoliths were extensively fused and injected with veins of mafic magma. Representative samples from this location that were analyzed isotopically (Leeman and others, 1985) were found to have Sm-Nd and Pb-Pb model ages of about 2.8 billion years. Thus, these xenoliths suggest a Precambrian basement beneath this portion of the Snake River Plain province.

Return to the vehicles and drive back to Highway 20 along the same route.

23.7 At Highway, 20, turn right (west) and continue across Macon Flat. The next few miles will take us over the sediment-veneered Pliocene-Pleistocene basaltic lavas that floor the valley.

26.7 Pass the junction with the Princess Mine road.

32.6 At the junction with Idaho Highway 46, turn south toward Gooding.

37.0 The road crosses a small-displacement fault scarp in the Macon Basalt.

39.6 Stop 5 (lunch) will be at the scenic overlook near the top of Johnson Hill. The view from this locality gives a good perspective of the structural setting of the Camas Prairie, which appears to be a large graben. However, the origin and age of the structure is certainly debatable. Cluer and Cluer (1986) proposed that the graben may be a pull-apart structure that formed in response to the downwarping, to the south, of the Snake River Plain. They cite the presence in the Mount Bennett Hills of post-Challis (younger than Eocene) alluvial gravels containing clasts derived from granitic, sedimentary, and Challis volcanic terranes, and they propose that these deposits were part of an extensive outwash fan derived from sources north of the Camas Prairie. If these deposits originally were continuous across the valley, then they provide a maximum time constraint for the faulting. It can also be demonstrated, however, that several of the Idavada ashflow tuffs thinned against preexisting topography during their northward transport (ca. 10 to 8 million years ago; Honjo and others, 1986) from eruptive centers in the Snake River Plain; hence, the Mount Bennett Hills must have had positive topographic relief at that time. In addition, because only one or two thin remnants of the Idavada ashflow tuffs can be found north of the Camas Prairie, the Mount Bennett Hills must have been a
significant barrier to the northward distribution of these ash-flow tuffs. The occurrence of significant structural offsets in the ferrolatite lavas indicates that many of the northwest- and east-to-west-trending faults bounding the graben have been active since approximately 5.7 million years ago. In fact, much of the relief on the bounding fault on the north side of the eastern Mount Bennett Hills must have formed since that time. Furthermore, the small scarp in the Macon Basalt indicate that at least minor faulting has occurred since 0.9 million years ago. The evidence thus suggests that formation of the Camas Prairie is not simply related to the subsidence associated with initial volcanism in the Snake River Plain, but that it developed over a considerable interval of time.

40.1 At the summit of the Johnson Hill grade, a small quarry east of Highway 46 has fresh exposures of the McHan Basalt, a sample of which was dated as 9.44 +/- 0.11 million years old (Honjo and others, 1986). Because its composition is relatively "evolved" (Table 2), this basalt is distinctive compared with younger basalts of the area. Similar basaltic lavas cap the numerous horsts and grabens in the northern part of the eastern Mount Bennett Hills. At least one related flow is interbeded with Idavada ash-flow tuffs on the southern flank of the eastern Mount Bennett Hills (see Stop 8). Thus, it is unlikely that the McHan Basalt is genetically related to the younger Pliocene-Pleistocene olivine tholeiites.

42.7 Pass the junction with a ranch road that leads west to a diatomite quarry. Excavations at this quarry expose a sequence of tuffaceous sediments that are overlain by an ash-flow-tuff unit (the Fir Grove Tuff of Smith, 1966). The basal contact is exceptionally well exposed and consists of a thin, air-fall unit beneath a densely welded vitrophyre.

44.0 Pass the Camas-Goolding County line just south of the rest stop/picnic area on the left side of Highway 46.

44.9 Stop 6 is at the junction with the graded road to Thorn Creek Reservoir. Turn east here and park along the road. From this point we can walk north a short distance to examine the lower tuff of the Gwin Springs Formation, one of the older ash-flow-tuff units in the region, and the Schooler Basalt, a relatively young basalt that is typical of many of the Snake River Plain basalt units.

The upper portions of the Schooler Basalt (Smith, 1966) are exposed in road cuts along Highway 46. This unit vented nearby and flowed southward onto the Snake River Plain, forming a hummocky-surfaced apron. Like most of the Pliocene-Pleistocene Snake River Plain basalts, it is a porphyritic olivine basalt (Table 2) containing phenocrysts of olivine (with rare traces of Cr-spinel) and labradoritic plagioclase; clinopyroxene occurs only in the groundmass as subophitic grains.

The prominent ash-flow tuff that caps the hill (the lower tuff of the Gwin Springs Formation; Smith, 1966) is virtually identical to the Fir Grove Tuff (near Mile 42.7) in its zoning characteristics, lithology, and composition. Both probably are part of the same flow unit. This ash flow consists of a porphyritic basalt vitrophyre that is overlain, in sharp unconformity, by a crystal-poor, devitrified, upper zone. Thus, it is a compound cooling unit. The vitrophyre is a low-silica rhyolite (Table 2) containing phenocrysts of plagioclase and mica, plagioclase, and ophitic andesine. The glassy groundmass also contains traces of apatite and zircon. Beneath the lower Gwin Springs tuff unit and an intervening slope break is a partially exposed, contorted and flowbanded, porphyritic rhyolite. This unit resembles the vitrophyre of the overlying ash-flow tuff in its phenocryst assemblage and bulk composition, but many features that characterize pyroclastic flows (shards, eutaxitic structure, etc.) are not present. In exposures along Thorn Creek, to the southeast, this unit is very thick and dense, and has a well-developed flow-top breccia. There is sure to be some debate as to whether this unit is a rhyolite lava flow or a densely welded, ash-flow tuff.

Return to the vehicles and proceed east toward Thorn Creek Reservoir.

44.9 At this junction take the east fork. The road intermittently crosses a Square Mountain Ferrolatite flow which, to the south, overlaps the rhyolite of Magic Reservoir along Thorn Creek.

44.9 At this junction take the south fork, continuing across the ferrolatite.

48.9 Stop 7 is at the spillway of Thorn Creek Reservoir. This spillway is constructed on the rhyolite of Magic Reservoir, which is well exposed over a broad area in the central portion of the Magic Reservoir eruptive center. Only two K-Ar dates are available for this unit: 5.8 million years for a sample collected west of Hot Springs Landing and 4.15 ± 0.06 million years (Honjo and others, 1986) for a sample from the southeast border of the eruptive center. However, because the rhyolite of Magic Reservoir in the central portion of the eruptive center is overlain in many places by ferrolatite, the older age is probably more representative. At this stop, the rhyolite of Magic
Reservoir is distinctively coarsely porphyritic, with more than 30 percent phenocrysts in a glassy black groundmass. More typically, however, the groundmass is devitrified completely, although locally the upper vitrophyre carapace of the flow is preserved. Even though the groundmass glasses are low-silica rhyolites, the rhyolite of Magic Reservoir is quartz latite in bulk composition (Table 2) because of its high crystal content.

Representative rhyolite of Magic Reservoir samples from throughout the eruptive center are very similar in composition. The phenocrysts include potassium feldspar, quartz, andesine, ferroaugite, pigeonite, magnetite, ilmenite, and accessory apatite and zircon. Hydrous phases are notably absent. The potassium feldspar phenocrysts commonly exceed 5 mm in diameter and are very inhomogeneous in composition, ranging from anorthoclase to sanidine. Cores of plagioclase also commonly occur within anorthoclase, but discrete sanidine phenocrysts occur only rarely. Most of the quartz and anorthoclase grains are markedly resorbed. Glomeroporphyritic clots of anorthoclase, ferroaugite, pigeonite, magnetite and ilmenite are common. Relatively calcic plagioclase and magnesium orthopyroxene (hypersthene-bronzite) are associated in xenocrystic clots, which apparently were derived from more mafic magma. These observations indicate that the rhyolite of Magic Reservoir magmas may have been mixtures of rhyolitic and basaltic magmas. In addition, xenoliths of granitic rocks, gneiss, and quartzite have been found in this unit.

Return along the same route to Highway 46.

58.3 Stop 8 is near the abandoned farm buildings. Here, we will climb through the section and see the McHan Basalt overlain by the City of Rocks Tuff, which has a well-exposed basal vitrophyre. The basalt is not particularly remarkable except that it is less porphyritic than typical olivine tholeiites of this region. The tuff is a quartz latite (Table 2) that is distinctively crystal-rich, with phenocrysts of plagioclase, minor quartz, clinopyroxene, orthopyroxene, and opaques, and traces of zircon and apatite in a glassy to devitrified groundmass. This unit thins and ultimately feathers out a short distance to the north.

Return to the vehicles and follow the same route back to Highway 46.

59.9 At the junction with Highway 46, turn south toward Gooding and proceed across the surface of the Schooler Basalt.

61.5 Pass the junction with the ranch road that goes west to the Little City of Rocks. A short distance west of the highway is an area which contains hoodoos similar to those at the City of Rocks and at the Little City of Rocks.

64.4 The road crosses the margin of the Schooler Basalt flow onto an older, more vegetated and loess-mantled, flow surface. Occasional pressure ridges are discernable.

65.1 Pass the junction with the road that goes east to Turkey Lake. From here, a low, but conspicuous, basaltic vent can be seen a short distance east of the highway.

68.3 The road crosses another contact between basalt flows, onto an even more subdued flow surface.

72.8 Cross the Big Wood River at the north side of Gooding.

GOODING TO JACKPOT VIA SHOSHONE FALLS

Follow Highway 46 through Gooding to Wendell. Turn east on Interstate Highway 84 and follow it to U.S. Highway 93. Turn south and cross the Perrine Bridge over the Snake River.
Figure 8. Shoshone Falls, from the north rim of Snake River canyon (Stop 9). The waterfall plunges over the massive interior of a rhyolite lava flow. Basalt flows form the upper canyon walls.

Between Gooding and the Perrine Bridge, the trip crosses the basalt-covered interior of the central Snake River Plain. Features to be seen include the surfaces of several basalt flows and some typical basalt shield volcanoes. The flows are relatively young (mainly of Pleistocene age) and are only partially mantled by soil. Pressure ridges are evident on the tops of some. One of the youngest flows, which we will cross between Gooding and Wendell, erupted from Notch Butte.

Perrine Bridge Area

Typical Snake River Plain basalt flows can be seen in the walls of Snake River canyon from this bridge. A rhyolite lava flow is exposed below the basalt. The flow lobes at the margin of this flow can be seen downstream (to the west) near the golf course. At Stop 9 (Fig. 1), a few miles upstream, we will visit this same rhyolite flow at Shoshone Falls. The stone monument at the south end of the bridge was erected by the Twin Falls Chamber of Commerce to commemorate daredevil Evel Knievel's attempted rocket-propelled motorcycle jump across the Snake River canyon 2 miles east of this point in 1974.

Continue south past the Perrine Bridge for about 1.5 miles. Turn left onto Falls Avenue and go east 3 miles to the Shoshone Falls turnout. Follow this road north and park in the main parking lot.

Stop 9: Shoshone Falls Park

Notable features: massive interior of rhyolite lava flow, Shoshone Falls, Bonneville flood erosion

The Shoshone Falls rhyolite lava flow is exposed by Shoshone Falls (212 feet high) and in the adjacent canyon walls (Fig. 8). Note its massive interior, the sheeted and folded rocks at the top of the massive section, and the large lenses of breccia at the top of the
Objectives

During today's trip (Fig. 9) we will examine the following geologic features: (1) the distal facies of medium- and high-temperature welded-tuff sheets; (2) the interior and marginal facies of rhyolite lava flows; (3) the southern boundary of both the Bruneau-Jarbidge eruptive center and the central Snake River Plain; and (4) aspects of regional basaltic volcanism.

Drive north from Jackpot on Highway 93. Turn right (east) at Rabbit Springs rest stop (0.5 mile north of milepost 4) and park vehicles. Walk about 0.1 mile east to a south-facing escarpment.

Stop 10: High-Temperature Welded Tuff

Notable features: basal vitrophyre, devitrified zone, flow marks, axiolites

This exposure contains both the lower vitrophyre section and an overlying devitrified zone of a welded-tuff unit. Approximately one foot of fused, air-fall ash is exposed below the basal vitrophyre. Within the lower vitrophyre section are numerous elongate spherulites or axiolites (Fig. 10, A and B) with an average elongation of N.65°W., and within the devitrified zone are flow marks with a similar orientation. Note that the axiolites higher in the exposure have a greater degree of elongation. This welded-tuff sheet was mapped by Earl Cook (Alicef, 1962) as part of his Tvo unit (older silicic volcanics).
Stop 11: Medium-Temperature Welded Tuff

Notable features: vitrophyre, fused top of air-fall ash, pumice layers

This unit consists of vitrophyre with some spherulites and overlies several feet of layered air-fall ash. The top of the basal ash layer was fused by the heat of the overlying ignimbrite. The ash contains pumice-rich layers and lies on tan, structureless silt. Note that this ash flow does not contain the type of flow marks that occur at Stop 10. This welded-tuff unit was mapped by Earl Cook (Alief, 1962) as unit Tv2 (ignimbrite 2).

Medium- vs. high-temperature welded tuff.

A lack of flow marks similar to those at Stop 10 suggests that this unit was emplaced at too low a temperature for vesicles to form. High-temperature welded-tuff units, as defined for this guide (Table 3), refer to those units which became molten silicate liquid after eruption and emplacement, permitting the development, deformation, and preservation of gas bubbles. Medium-temperature welded-tuff sheets, on the other hand, lack liquid state flowage features, such as elongate vesicles or flow marks, because their emplacement temperatures were too low for the ash to coalesce to a layer of silicate liquid. Although precise temperature ranges have not been investigated thoroughly, the transition between the medium- and high-temperature types probably is in the 750-850 degree C range.

Description of Drive from Jackpot to Rogerson

Return to vehicles and proceed north on Highway 93 toward Rogerson. As you travel north on Highway 93, many welded-tuff units can be seen on either side of the north-to-south-oriented graben in which the highway is located. The escarpment to the east, 2 to 3 miles away, marks the margin of the Cassia Mountains (Fig. 9). These mountains are an uplifted block cored by Paleozoic rocks with an overlying veneer of welded-tuff sheets which were erupted from sources to the west, perhaps including from the graben in which we are driving, from the north in the Twin Falls area, and from the Oakley area to the east.

Browns Bench escarpment

The Browns Bench escarpment, which was uplifted along a major fault, is visible 6 to 8 miles to the west and contains many welded-tuff sheets. These units were erupted from the Snake River Plain region and probably include some tuff units that were erupted from the Bruneau-Jarbidge eruptive center several miles to the northwest. The detailed mapping needed to trace the
welded-tuff units across the graben has yet to be conducted.

**Description of Drive From Rogerson to Salmon Dam**

At the turnoff at the gasoline station at the south edge of Rogerson, turn left (west) here and proceed through town. In town, turn left (west) on the hard-surface road that heads west to Murphy Hot Springs and Jarbidge.

Note your mileage as you turn toward Murphy Hot Springs because the distances below are measured from this intersection.

**Salmon Dam area**

Between Rogerson and Salmon Dam several basalt shield volcanoes can be seen to the north. While driving the first few miles west from Rogerson, the hill located straight ahead is Salmon Butte, the youngest basalt volcano in this part of the Snake River Plain. It is situated where the large fault along the base of the Browns Bench escarpment intersects the southern margin of the Snake River Plain. At Salmon Dam you can see that the basalt flows from Salmon Butte partially filled a canyon incised in older basalt and rhyolite units. Also visible at Salmon Dam (7.7-7.8 miles from Rogerson) are scattered rhyolite exposures, where the pre-Salmon-Butte basalt canyon wall has been exhumed.

**Description of Drive from Salmon Dam to Stop 12**

At 13.6 miles from Rogerson the road crosses a small basalt volcano. In the road cut inward-dipping basalt dikes separate unlike sections of sediments and basaltic tuff. These dikes extend to the southeast about 0.2 mile. Overall, these features seem to form a small crater that was partially filled by basaltic ash and sediments during and after the final stage of the eruption. It may be a diatreme with later dikes injected along the interface between the footwall and crater-fill zones.

Ahead, on the right (north of the road) as you leave the small crater, is Cedar Creek Reservoir. Northwest of it is Cedar Butte, one of the taller basalt shield volcanoes in this part of the Snake River Plain. The southwest side of Cedar Butte has been cut off by one of the many northwest-trending normal faults that occur in the south-central part of the Snake River Plain.

Along this stretch of the drive, a topographic break is...
Table 3. Ash-flow tuff types based on emplacement temperatures.

<table>
<thead>
<tr>
<th>Low-temperature (Type L)</th>
<th>Medium-temperature (Type M)</th>
<th>High-temperature (Type H)</th>
<th>Very high-temperature (Type V)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Up to 650°C.</td>
<td>550-850°C.</td>
<td>750-1050°C.</td>
<td>950-1200°C+</td>
</tr>
<tr>
<td>Ash-flow tuff sheets are not welded and show only modest amounts of compaction.</td>
<td>Welded and densely compacted but lacking liquid-state flowage features.</td>
<td>Thoroughly welded and compacted with restricted-size liquid-state flowage features.</td>
<td>The entire ash-flow tuff sheet may have undergone en masse liquid-state flowage.</td>
</tr>
</tbody>
</table>

After emplacement the ash-flow tuff does not revert to a silicate liquid. After emplacement the ash-flow tuff reverts to a mass of silicate liquid that can flow.

visible 1 to 2 miles to the south. Beyond this break, where the land dips more steeply to the north, are welded-tuff units that were erupted from the Snake River Plain. On this side of the break the rock units are more nearly horizontal rhyolite and basalt lava flows that fill the subsided zone from which the ash flows erupted. Exposures of a rhyolite lava flow are visible near Cedar Creek at 17.0 miles.

Between 17.4-18.4 miles, along the grade west of Cedar Creek, are well-exposed basalt flows and fine-grained sediments. Farther west, to about 21 miles from Rogerson, the road travels on the Bud Lewis Hill basalt flow. This flow was erupted from a volcano (Hill 5991) that has less than 50 feet of relief, and which is located to the south (best seen from near Stop 12).

Stop at 22.1 miles at the turnout on the left (southeast) side of the road (opposite the large pit excavated in sediments) and walk northeast 0.1-0.2 miles to a quarry.

Stop 12: Post-Emplacement Flowage in Welded Tuff

Notable features: stretched vesicles, post-emplacement flowage structures, boundary between vitrophyre and devitrified rhyolite

The welded-tuff sheet exposed in the small quarry southeast of the road contains flow marks, relatively large stretched vesicles, various small-scale, post-emplacement flowage features, and the boundary between the basal vitrophyre and the overlying devitrified zone. The flow marks and stretched vesicles have an average azimuth of N 5°W.

Basalt Dike-Vent. Visible from this quarry, about 0.3 mile to the west, is a small basalt volcano (Hill 5822) from which an unnamed basalt flow erupted from a nearly vertical northwest-trending dike that is a few feet thick. This feeder dike has been traced for about a mile and is clearly exposed in a roadside silt bank just south of the volcano, where it consists of two parts. The western part of the dike is 6 feet thick and the eastern part is 2 feet thick. The two parts are separated by about 15 feet of sedimentary material. As you drive away from this area, you will pass this dike on the north (right) side of the road about 0.2 mile west of Stop 12. The Hill 5822 vent has only about 50 feet of relief, but the basalt flow from it extends at least half a mile northward. This dike-vent is one of several similar tiny volcanoes located in this area. They are within the boundary zone between the down-dropped Snake River Plain to the north and the highland area to the south, and may be associated with major buried plain-margin faults.

Description of Drive From Stop 12 to Jarbidge Canyon

Return to vehicles and continue driving west. At 26.6 miles (0.2 mile west of Devil Creek crossing), another basalt dike cuts sediments and silicic ash beds on the north (right) side of the road. This dike fed a small volcano about 0.1 mile to the north, which erupted a small basalt flow that ran to the north. At 29.3 miles white silicic tuff is exposed within the sediments on the east (left) side of the road.

Between 30.7 and 31.4 miles, a section of sediments is exposed along the grade down to Mud Flat Draw. Here, gravel caps a section of sandy and silty material which, in turn, lies on tuffaceous beds. This gravel, which extends northward for many miles, probably was eroded during the uplift of Elk Mountain, 7-10 miles to the south. Similar, generally postrhyolite and prebasalt, sequences of sediments occur at other places along the southern margin of the Snake River Plain.

Between 32.0 and 34.0 miles are several exposures of
Bedded, white silicic ash. At 34.2 miles, the Three Creek Rhyolite is exposed. This will be Stop 16 when we return along this road. At 41.4 miles the road goes over a rise on the north flank of Horse Hill. Horse Hill, whose top, complete with a loess-filled crater, is located about half a mile to the south, is a basalt shield volcano from which part of what Bonnichsen and Jenks (in press) have called the basalt of Big Flat was erupted and flowed northward. We will see this basalt unit again at Stop 13, where it forms the eastern rim of Jarbidge Canyon. At about 45.5 miles, approximately half a mile southeast of the road, is another low hill. This is Hill 5981 and it was another source of the basalt of Big Flat.

**Jarbidge Mountains**

The Jarbidge Mountains appear ahead and slightly to the left along the stretch of road from Horse Hill to Jarbidge Canyon. The Matterhorn, at an elevation of 10,839 feet, is the highest peak. These mountains are composed primarily of the Jarbidge Rhyolite (Schlauer, 1923; Coats, 1964, 1987). This unit is 15 to 17 million years old and hosts quartz-adularia veins that contain gold which has been mined episodically in the Jarbidge area since the early part of this century. Although the Jarbidge Rhyolite has not been studied in much detail, it is apparent that it is part of a large field of pre-Snake River Plain rhyolitic rocks exposed along the south side of the Snake River Plain to as far east as the Nevada-Utah border (Stewart and Carlson, 1978; Coats, 1987). The Jarbidge Mountains were uplifted principally along a set of northeast-trending faults which merge into the large fault system that bounds the east side of Browns Bench escarpment, which we viewed previously.

Near the airstrip, just before you reach the rim of Jarbidge Canyon at 46.1 miles from Rogerson, take the dirt road that turns to the northwest (right) at the cattle guard. Follow this road for approximately 5.5 miles to reach the Diamond A Trail (not marked). Turn west, drive 0.1 mile to the rim of Jarbidge Canyon, and park.

**Stop 13: Dorsey Creek Rhyolite**

**Notable features: interior and marginal rhyolite-flow facies, canyon-rim basalt**

From this point on the rim the view to the northwest, at the bottom of the canyon, shows the interior of the Dorsey Creek Rhyolite (Fig. 13). To the southeast the marginal portion of the flow is exposed at the bottom of the canyon (Fig. 14). The rim basalt on the east side of the canyon is the basalt of Big Flat, which was erupted from Horse Hill, Hill 5981, and perhaps other sources to the east. The rim basalt on the west side of the canyon is the basalt of the Diamond A Desert (Bonnichsen and Jenks, in press) and was erupted from two or more shield volcanoes located in northern Nevada, 7 to 8 miles to the southwest. Locally exposed beneath the Dorsey Creek Rhyolite in this part of Jarbidge Canyon is the Columbet Creek Basalt (Bonnichsen and Jenks, in press).

**Interior and marginal facies of rhyolite lava flows.**

The rhyolite lava flows in the Snake River Plain show variations in thicknesses and types of internal structures that vary from the interiors to the margins of the flows. A generalized longitudinal section from the vent area to the margin of an idealized flow is shown in Figure 15.

Rhyolite lava-flow interiors generally have three zones: a thick central zone of massive devitrified rhyolite, which overlies a complex basal zone that is capped by a structurally complicated upper zone. The basal zones consist of either massive vitrophyre layers or breccia layers, or a combination of both materials. The upper zones generally contain both glassy and devitrified portions and consist of massive, sheeted, folded, flow-layered, and brecciated rhyolite. All of the zones, but especially the upper ones, contain gas cavities of widely varying dimensions and abundance.

Flow margins typically consist of bulbous lobes of massive or sheeted rhyolite overlying basal breccia layers and separated by chaotic-appearing zones of steeply jointed or flow-layered rhyolite. The thicknesses of the interior portions of the flows typically are 50 m or more and vary up to 250 m. The flows are thinner near their margins, but only in a few places thin to less than 25 m (Bonnichsen, 1982b; Bonnichsen and Kauffman, 1987).

**Dorsey Creek Rhyolite.**

The Dorsey Creek Rhyolite is well exposed for 25 miles in Jarbidge Canyon, from Murphy Hot Springs to near the confluence of the Bruneau and Jarbidge Rivers. This flow contains a minimum volume of 75 cubic km of rhyolite and is one of the largest in southwestern Idaho (Bonnichsen, 1982b). It exceeds 200 m in maximum thickness. The Dorsey Creek is the youngest rhyolite lava flow exposed in Jarbidge and Bruneau Canyons, with an age of between 8.0 and 8.22 million years (Hart and Aronson, 1983).

The compositions of two Dorsey Creek Rhyolite samples are included in Table 4. These analyses suggest that the northern part of the unit is somewhat richer in silica and poorer in femic constituents than the southern part. Generally, the lava flow contains phenocrysts of quartz, plagioclase, augite, pigeonite, and opaque oxides. Its initial Sr isotope ratio suggests that the unit was derived by the fusion of crustal materials (Bonnichsen and Citron, 1982).

**Description of Drive From Jarbidge Canyon Rim to Murphy Hot Springs**

Backtrack to the Rogerson-Murphy Hot Springs road.
Turn right (southwest) and follow the road over the rim and down the grade into the East Fork of Jarbidge Canyon toward Murphy Hot Springs.

Panoramic View from Jarbidge Canyon

From the top of the Murphy Hot Springs grade at the rim of Jarbidge Canyon, some interesting topographic features are visible. In a clockwise order, beginning to the south-southwest, are the following features: a glacial cirque (called the crater by some of the locals) on the north end of Jarbidge Mountain (south-southwest); Copper Mountain and Bear Paw Mountain (southwest); Wilkins Island, the table land between the east and west forks of Jarbidge Canyon (west); the Black Rock hill basalt shield volcano (west, in the distance); the Diamond A Desert which is the table land lying east of Jarbidge Canyon (north); several other basalt shield volcanoes (north and northeast); Hill 5981, one of the basalt of Big Flat sources (east-northeast); Elk Mountain (east-southeast); and Cougar Point on the East Fork of Jarbidge Canyon (south).

Bruneau-Jarbidge Eruptive Center

The Bruneau-Jarbidge eruptive center is a 95-by-55-kilometer structural basin within the southwestern part of the Snake River Plain (Figs. 16 and 17). During the late Miocene, a sequence of eleven or more welded, ash-flow-tuff, cooling units (the Cougar Point Tuff), twelve or more large rhyolite lava flows, and a series of olivine tholeiite basalt flows from forty or more shield volcanoes were erupted from, and filled in, this basin. The Bruneau-Jarbidge eruptive center does not now have the physiographic form of a caldera, but it may contain one or more buried calderas. Most of the subsidence of its interior occurred between 11 and 10 million years ago,
during or just after the eruption of the voluminous Cougar Point Tuff (Bonnichsen, 1982a).

Along its southern margin, the boundary zone of the eruptive center is outlined by a combination of down-to-the-north faults of small displacement, a moat zone filled with sediments and basalt, and lateral transitions from older welded ash-flows on the south to younger rhyolite lava flows on the north. These relations will be seen at Stop 14. The eastern and northern margins of the eruptive center are buried beneath younger geologic units, primarily basalt flows. The extent of the eruptive center has been partially defined on the basis of aeromagnetic anomalies (Bonnichsen, 1982a).

**View From Murphy Hot Springs**

As you start down the grade into Jarbidge Canyon, the prominent rock ledges that dip gently northward are some of the Cougar Point Tuff units. The thick sequence of sediments between the Cougar Point Tuff and the rim basalt was deposited within the moat zone along the southern margin of the Bruneau-Jarbidge eruptive center.

**Cougar Point Tuff**

The Cougar Point Tuff is a sequence of densely welded, rhyolitic, ash-flow tuff cooling units that erupted from the Bruneau-Jarbidge eruptive center during late Miocene time. The Cougar Point Tuff is best exposed in the canyons of the Bruneau River and the East and West Forks of the Jarbidge River near the Idaho-Nevada border. At the Black Rock escarpment reference section (Bonnichsen and Citron, 1982) in Bruneau Canyon, eight units are exposed and have an aggregate thickness of 400-475 meters. At Cougar Point in the East Fork of Jarbidge Canyon, about 5 miles south of Stop 14, six of
the units are exposed with a thickness of about 250 meters. By the end of Cougar Point Tuff volcanism, the eruptive center had developed into a large structural and physiographic basin, which was filled by later volcanic and sedimentary deposits. The exposed part of the Cougar Point Tuff consists of outflow-facies rocks that extend beyond the margins of the eruptive center. The emplacement of each unit was a distinct volcanic event, and the units are separated by sedimentary layers. Four or more reversals of the Earth’s magnetic field occurred during the deposition of the Cougar Point Tuff (Bonnichsen, 1982a).

In general, each of the Cougar Point Tuff units is similar to the others in appearance and welding characteristics. Commonly, the units were formed by multiple ash emplacements, but are simple cooling units. In a vertical section through a unit, the typical zones from base to top consist of the following: (1) a basal layer of thinly bedded, air-fall ash, (2) a vitrophyre layer at the base of the ash-flow sheet, (3) a massive, relatively thick, devitrified central zone of the ash-flow sheet in which most structures have been obliterated by matrix crystallization, and (4) an upper zone of the ash-flow sheet with abundant flow marks and folds.

The various Cougar Point Tuff units are similar in composition and in their temperature and mode of eruption. Quartz, sanidine, plagioclase, augite, pigeonite, fayalite, and opaque oxides are the principal phenocryst minerals. Hornblende and biotite are essentially absent, suggesting that the magmas were relatively hot and dry. Magma-temperature estimates based on the compositions of coexisting pyroxenes and feldspars suggest eruption temperatures for the Cougar Point Tuff units in the 850-950 degrees C range (Honjo and others, 1987). Chemically analyzed samples for each unit from locations many kilometers apart reveal that each has a relatively narrow compositional range. Representative analyses for two units (XIII and XV) are shown in Table 4. Overall, the Cougar Point Tuff becomes increasingly femic upwards, but local reversals in the trend exist. A K-Ar age of 11.3 million years was reported for one of the lower Cougar Point Tuff units. Initial strontium isotope ratios indicate a predominantly, if not exclusively, crustal origin for the Cougar Point Tuff (Bonnichsen and Citron, 1982). This is similar to many other rhyolitic units in the Snake River Plain volcanic province.

Stop at about 47.5 miles, get out of the vehicles, and walk the rest of the way down the grade. This will be stop 14. Rejoin the vehicles at the base of the grade near the Murphy Hot Springs store at mileage 48.2.

Stop 14: Upper Part of the Cougar Point Tuff and the Margin of the Bruneau-Jarbidge Eruptive Center

Notable features: top of Cougar Point Tuff unit XV, small spherulites, end of Dorsey Creek Rhyolite

The exposures along the road are the thin distal facies of Cougar Point Tuff unit XV. Here, the unit is almost entirely glass, with small spherulites as the only sign of devitrification, indicating that the unit cooled rapidly after it was emplaced. At the very top of unit XV, an upwards transition from strongly welded ash that is quite hard, to weakly fused ash that easily can be crushed by hand is exposed over a vertical distance of 1 to 2 feet. Near the bottom of the canyon the next welded-tuff unit down (unit XIII) is also visible. It is much thicker and is almost entirely devitrified.

At this location the Cougar Point Tuff is overridden by the end of the Dorsey Creek Rhyolite. This relationship, along with the significant thickness of
moat-zone sediments and the hot spring, are all related to the margin of the Bruneau-Jarbidge eruptive center (Fig. 18). Bonnichsen (1982a) proposed that the hot spring marks the location of a covered, post-Cougar Point Tuff and pre-Dorsey Creek Rhyolite fault within the boundary zone of the eruptive center. Evidence for this fault includes the occurrence of the unconsolidated volcanic ash and sediment, which contains friable cobbles of sediment, that is exposed below the end of the Dorsey Creek Rhyolite at the bottom of the grade (Fig. 19). This exposure is postulated to be near the buried scarp.

**Cougar Point Tuff Unit XV**

The uppermost Cougar Point Tuff unit, XV, occurs prominently along the rim of Bruneau Canyon in the Black Rock escarpment area, 15 miles west of Stop 14, where it ranges from 30 to nearly 100 m in thickness. That thick part of unit XV is considered to be proximal to the buried source. Unit XV is discontinuously exposed as far east as here at Stop 14. In its distal exposures here in Jarbidge Canyon, unit XV is generally less than 5 meters thick and consists mainly of vitrophyre. In the thickest part of the unit at Black Rock escarpment, the basal zone contains abundant, greatly flattened, pumice fragments, but such fragments have not been observed in the more distal areas.

The phenocryst minerals in unit XV are plagioclase, sanidine, quartz, augite, pigeonite, fayalite, and magnetite. Plagioclase is considerably more abundant than either quartz or sanidine. Cumulophyric aggregates of plagioclase, pyroxenes, and magnetite are common throughout the unit. Phenocryst minerals are much more abundant in the proximal portion of this unit than they are in the distal part. Here at Stop 14, which is at a location far removed from the source of the ash, fayalite takes the place of pigeonite, which is common closer to the unit’s source. An analysis of one sample of unit XV is included in Table 4. Unit XV is one of the most fenic of the Cougar Point Tuff units.

**Stop 15: Interior of Dorsey Creek Rhyolite**

**Notable features:** basal breccia, sheeting joints, flow lobes, internal structures

Well-developed sheeting joints and other features in the central zone of the flow are well exposed near where the East and West Forks of the Jarbidge River join, at about 2.3 miles past the Murphy Hot Springs store, and park in the large lot.

---

### Table 4. Chemical analyses of representative Snake River Plain rhyolite samples from southwestern Idaho.

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Cougar Point Tuff, XIII</td>
<td>Cougar Point Tuff, XV</td>
<td>Dorsey Creek Rhyolite (south)</td>
<td>Dorsey Creek Rhyolite (north)</td>
<td>Three Creek Rhyolite</td>
<td>Balanced Rock Area, Upper Lava Flow</td>
</tr>
<tr>
<td>Sample</td>
<td>X-37</td>
<td>I-459</td>
<td>I-529</td>
<td>I-1001</td>
<td>I-356</td>
<td>I-15</td>
</tr>
<tr>
<td>SiO₂</td>
<td>75.27</td>
<td>73.65</td>
<td>72.14</td>
<td>73.64</td>
<td>73.73</td>
<td>69.50</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>12.30</td>
<td>12.83</td>
<td>12.77</td>
<td>12.93</td>
<td>12.56</td>
<td>13.59</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.28</td>
<td>0.42</td>
<td>0.52</td>
<td>0.45</td>
<td>0.46</td>
<td>0.77</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>2.40</td>
<td>3.06</td>
<td>4.24</td>
<td>3.23</td>
<td>3.46</td>
<td>4.78</td>
</tr>
<tr>
<td>MnO</td>
<td>0.04</td>
<td>0.05</td>
<td>0.01</td>
<td>0.03</td>
<td>0.04</td>
<td>0.07</td>
</tr>
<tr>
<td>CaO</td>
<td>0.81</td>
<td>1.40</td>
<td>1.64</td>
<td>1.31</td>
<td>1.14</td>
<td>2.28</td>
</tr>
<tr>
<td>MgO</td>
<td>0.23</td>
<td>0.50</td>
<td>0.39</td>
<td>0.34</td>
<td>0.23</td>
<td>0.83</td>
</tr>
<tr>
<td>K₂O</td>
<td>6.10</td>
<td>5.17</td>
<td>5.37</td>
<td>4.86</td>
<td>5.66</td>
<td>4.95</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.52</td>
<td>2.89</td>
<td>2.77</td>
<td>2.15</td>
<td>2.55</td>
<td>3.05</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.04</td>
<td>0.06</td>
<td>0.09</td>
<td>0.06</td>
<td>0.06</td>
<td>0.16</td>
</tr>
</tbody>
</table>

**Sample Locations:**
2. Sec.9, T.16 S., R.7 E., Owyhee Co., Idaho (basal vitrophyre).
5. Sec.7, T.16 S., R.11 E., Owyhee Co., Idaho (vesicular vitrophyre).

These rocks were analyzed by a combination of methods in several laboratories, and have been normalized to sums of 100 percent. Total iron is expressed as Fe₂O₃.
Figure 16. Reference map of eastern Owyhee County, Idaho, and vicinity. The Bruneau-Jarbidge eruptive center is indicated by the shaded area. The asterisk symbols are basalt volcanoes (from Bonnichsen, 1982a).
Figure 17. Generalized geologic map of eastern Owyhee County, Idaho, and vicinity (from Bonnichsen, 1982a).
Figure 18. View from the east rim of Jarbidge Canyon, near Stop 14, of the Murphy Hot Springs area. The southern end of the Dorsey Creek Rhyolite (DC) is the prominent rock outcrop on the right. Small exposures of the upper Cougar Point Tuff unit (XV) occur upstream. An inferred, deep-seated fault is thought to be covered by the end of the Dorsey Creek Rhyolite, where a hot spring is located (below arrow).

base of the flow is exposed along the road about 0.4 mile southeast of the river confluence. Note the absence of air-fall ash at the base of this lava flow in contrast to the situation beneath the ash-flow units. Detailed descriptions of the internal structural features of rhyolite lava flows in southwestern Idaho are discussed in Bonnichsen (1982b) and in Bonnichsen and Kauffman (1987).

Turn around at this point and backtrack along the Rogerson-Murphy Hot Springs road toward Rogerson. Drive 16.3 miles to Stop 16 to examine the Three Creek Rhyolite. Park alongside the road, about 1.5 miles past (east of) the Three Creek School. This is between the road that goes south (right) to the Elk Mountain area and the top of the grade at the rim overlooking Three Creek.

Stop 16: Three Creek Rhyolite at Three Creek

Notable features: flow-layering, spherulites, flow-base breccia, basal silicic air-fall ash

The Three Creek Rhyolite lava flow forms the rim on the west side of Three Creek. The flow at this exposure is about as thin as any rhyolite lava flow within the Snake River Plain volcanic province. At this locality the flow is markedly flow-layered, contains numerous spherulites, has breccia at its base, and lies on about a foot of air-fall ash, which in turn was deposited on many feet of silt. The air-fall ash beneath this rhyolite lava flow is an unusual but not unprecedented amount of basal air-fall ash, although most rhyolite lava-flow exposures have no underlying air-fall ash.

Three Creek Rhyolite

In the southeastern part of the Bruneau-Jarbidge eruptive center, an area measuring 17 km from southeast to northwest is underlain by the Three Creek Rhyolite lava flow (Citron, 1976). This flow has an estimated minimum volume of 10 cubic km (Bonnichsen, 1982b). It is less femic than many of the other rhyolite lava flows (Table 4) and contains phenocrysts of plagioclase,
pigeonite, augite, and opaque oxides, and locally, of quartz.

**Description of Drive from Three Creek School to Lily Grade**

Continue east for 9.7 miles. Turn north (left) at the unpaved House Creek road to Roseworth (16 miles) and Castleford (31 miles). **Note your mileage at this intersection, as the mileages below are measured from this point.** Pavement will begin again after about 18.4 miles.

An excellent view of many of the basalt shield volcanoes in the southern part of central Snake River Plain is available from the House Creek road, between the Rogerson-Murphy Hot Springs road and Salmon Falls Creek canyon. This drive will take us through the middle of a region in which the morphologies of many basalt shield volcanoes were studied by Jenks (1984). At 2 miles north of our start on the House Creek road is Signal Butte, one of these basalt shields.

**Basalt Shield Volcanoes**

Jenks (1984) found that there were six morphological classes of basalt volcanoes in the south-central part of the Snake River Plain: multiple-vent volcanoes, large round volcanoes, large elongate volcanoes, large steep-sided volcanoes, small volcanoes, and phreatic volcanoes. She suggested that these different classes of volcanoes formed during three successive periods of volcanism: (1) volcanoes of the large, round class erupted first and exhibit smooth, rounded, unfaulted and cratered edifices and appear unrelated to the northwest-trending structural grain of the area; (2) volcanoes associated with regional, northwest-trending faults and graben formed second and are of three classes: large elongate, multiple-vent, and small-sized. The shields are elongate, have irregular surfaces, and in the multiple-vent volcanoes, were formed by the intergrowth of several smaller volcanoes into one large edifice. (3) Volcanoes of the large steep-sided class formed last and consist of a small number of large volcanoes with steeper flank slopes than the older volcanoes.

**Lily Grade Area**

At about 23.7 miles the road starts its descent into Salmon Falls Creek canyon on Lily Grade, where several basalt flows are well-exposed. The upper group of flows on the southwest side, as well as all of the flows exposed on the northeast side of the canyon, are subaerial flows. However, along the lower part of the grade on the southwest side are excellent exposures of pillow basalt, clearly the result of the basalt being deposited in water. A small amount of rhyolite is exposed beneath the basalt near the bridge across Salmon Falls Creek. As the road ascends the northeast side of the canyon, excellent exposures of sedimentary interbeds between the basalt flows are visible.

**Description of Drive From Lily Grade to Balanced Rock**

Proceed across the canyon on Lily Grade, and then to the intersection at 30.6 miles at the east edge of Castleford. Turn west (right), travel through Castleford, and proceed about 5.8 miles to the Balanced Rock area, following the directional signs along the road. Proceed across Salmon Falls Creek Canyon to the Balanced Rock viewpoint on the west side. Park at the turnout and walk down the grade to the bottom of the canyon, examining the rhyolite exposures along the way.

**Stop 17: Balanced Rock Area**

Notable features: Balanced Rock, interior features of rhyolite lava flows

Two rhyolite lava flows are exposed in the Balanced Rock area (Fig. 1). Balanced Rock (Fig. 20) is within the upper flow, and the cliffs along the canyon bottom are within the lower unit. The grade on the west side of Salmon Falls Creek affords an excellent cross section through the internal zones of these flows. Phenocryst minerals observed in these flows include plagioclase, augite, hypersthene, and opaque oxides. An analysis of the upper flow is included in Table 4. This flow is about as femic as any rhyolite lava flow in the Snake River Plain volcanic province.
Figure 20. Balanced Rock at Stop 17. This amazing feature has been eroded from the central zone of a rhyolite lava flow.

Description of Drive From Balanced Rock to Mountain Home

Return to Castleford and continue east to Buhl, following the highway signs. Between Castleford and Buhl we will be traveling across a basalt-covered portion of the Snake River Plain. At Buhl, join U. S. Highway 30 and head west toward Bliss.

A few miles west and north of Buhl the highway descends into the Snake River canyon. Interesting items to be seen in the canyon include the features formed during the Bonneville flood, the springs flowing from the east wall of the canyon in the Thousand Springs area, the basalt flows that ran into standing water, and the remnants of basalt vents that were erupted into Lake Idaho, a large lake of Pliocene and early Pleistocene age. North of Hagerman, as the highway climbs out of the canyon, the world-famous Hagerman fossil beds are present within the thick section of Glenns Ferry Formation sediments that is exposed on the other side of the river. Exposed in road cuts and at the surface north of the canyon is the McKinney Basalt of Leeman and Vitaliano (1976).

A few miles south of Bliss, after climbing out of the canyon, the road traverses another area covered by basalt lava flows. Here, numerous “pressure ridges,” the result of flowage and crust deformation of the lava prior to its final consolidation, are present on the flow top.

At Bliss, take I-84 west toward Mountain Home. Between Bliss and Glenns Ferry the road again descends into the Snake River canyon. Abundant exposures of lake sediments of the Idaho Group that were deposited in Lake Idaho are present in this area. These sediments typically are capped by basalt flows. The sediments in this area have been mapped and described by Malde and others (1963), Malde and Powers (1972), and Malde (1972).

A few miles west of Glenns Ferry, the highway again leaves the Snake River and climbs onto another basalt-covered plain. As you approach Mountain Home several basalt shield volcanoes can be seen. Also visible, to the northeast (right) at a distance of a few miles, are the
Mount Bennett Hills. These are composed of rhyolite lava flows and welded-tuff units that are approximately the same age and very similar in character to those observed yesterday in the Magic Reservoir area and visited today in the south-central portion of the Snake River Plain. A preliminary map and description of these silicic volcanic units has been prepared by Wood and Gardner (1984).

Leave I-84 at exit 95 in Mountain Home and stop for the night.

DAY 3: MOUNTAIN HOME TO BOISE AIRPORT VIA OREANA AND KUNA

Objectives

During today’s trip (Fig. 21) we will examine the following features: (1) the southwestern margin of the western Snake River Plain graben, (2) the results of en masse rheomorphic flowage in a welded-tuff unit, (3) a rhyolite lava flow, (4) non-welded, pumice-rich ash flows, (5) water-affected basalt, (6) the Melon Gravel deposited by the Bonneville flood, and (6) shields and other basalt volcanoes.

Description of Drive from Mountain Home to Browns Creek Area

Drive through Mountain Home on Idaho Highway 68, then take Idaho Highway 67 southwest toward Grand View. This road passes the Mountain Home Air Force Base and, just beyond where it descends into Snake River canyon, a large feed lot.

Between Mountain Home and the Snake River canyon we will drive across a basalt plain, mostly of Pleistocene age. A few miles to the north is Cinder Cone Butte, the only basaltic cinder cone of any size in the western Snake River Plain.

Cross the Snake River and proceed through Grand View, turning west (right) on Highway 78 toward Murphy. Proceed west on Highway 78 to the Oreana turnoff. After you descend into the Snake River canyon, you will mainly be driving through lacustrine sediments of late Pliocene age and younger lacustrine and fluvial sediments, including those deposited during the Bonneville flood. As you cross Castle Creek, 11.5 miles west of the junction in Grand View (0.8 mile west of milepost 49), the view to the north includes Castle Butte about 3 miles away. Castle Butte is one of several basalt volcanoes that developed beneath the surface of a large lake, Lake Idaho (Jenks and Bonnichsen, 1987), which occupied the western part of the Snake River Plain during the Pliocene and early Pleistocene.

Go through Oreana and proceed south for 2.0 miles, to where the road turns east (left). At 0.2 mile past this turn, take the gravel road to the south (right). Follow this road for approximately 2 miles to just past the top of the grade. Stop and walk to the top of the slope to observe the view to the west and south.

Stop 18: Silver City Range and Margin of Snake River Plain

Notable features: Silver City Range, silicic volcanics of Browns Creek, Idaho Group sediments

From the top of this slope, looking west, we have an excellent view of the Silver City Range about 12 miles away, with War Eagle Mountain on the right, Hayden Peak in the middle (highest point), and Quicksilver Mountain on the left. In the foreground, one to two miles west and southwest, are the silicic volcanic rocks of the Browns Creek area. Northwest-trending faults occur near the base of the escarpment, downdropping the rhyolite units to the northeast. These faults that bound the southwestern side of the western Snake River Plain graben are almost totally buried by younger Idaho Group sediments.
Silver City Range

Hayden Peak and Quicksilver Mountain are rhyolite domes that are part of the Silver City Rhyolite and which are older (at 15 to 16 million years) than the silicic volcanism that we examined the past two days, and which accompanied the formation of the western part of the Snake River Plain. War Eagle Mountain is predominantly granitic rocks and was the location, starting in 1863, of the early gold and silver mining activity in the Silver City area. More recently, silver and gold mining has concentrated in the De Lamar area, farther to the west.

Silicic volcanics of Browns Creek

The Browns Creek rhyolite (Ekren and others, 1981, 1982, 1984) consist of both rheomorphically deformed welded-tuff sheets and rhyolite lava flows, as well as nonwelded ash-flow units. These are part of the silicic volcanism that accompanied formation of the western graben of the Snake River Plain. Similar units at other localities have been dated at about 11 million years old.

Lake Idaho Sediments

Lying stratigraphically above the Browns Creek rhyolite, as well as overlapping the rhyolite and extending to the north, are Idaho Group lake sediments. These sediments were deposited within a lake, or series of lakes (Lake Idaho, Jenks and Bonnichsen, 1987) that filled the western part of Snake River Plain between late Miocene and early Pleistocene time. Visible near the top of the Idaho Group sediment section is an angular unconformity over lain by oolitic limestone beds (Swirydczuk and others, 1979, 1980).

Stop 19: Overview of Browns Creek Basin

Notable features: rhyolite lava flow, nonwelded ash flows, rheomorphically deformed welded tuff, postrhyolite lacustrine sediments

In the basin along Browns Creek (Fig. 22) is an older rhyolite unit to the west (left) that is overlain by pumice-rich, nonwelded, ash-flow sheets (straight ahead to the northwest). Both are overlain by a rheomorphically deformed, welded ash-flow unit to the north (right). We will visit these three subdivisions of the Browns Creek rhyolite at Stops 20, 21, and 22. Remnants of the Idaho Group sediments overlie these volcanic units. One of these remnants caps the hill with the road climbing up its south side in the middle of the basin, northwest of this viewpoint. At the west rim of the basin another welded-tuff sheet is exposed. Evidently it is the distal edge of a rheomorphically deformed unit that thickens westward and includes the small peaks visible about 1.5 miles northwest of this viewpoint. This welded-tuff sheet is very similar to the welded tuff exposed at the north side of the Browns Creek basin, but the two exposures are not connected. Therefore it is not yet clear if they were separate parts of the same eruption, or if they were deposited by different eruptions.

From here we will hike to Stops 20, 21, and 22. This hike probably will take 3 to 4 hours, so please take sufficient water and your lunch. Follow the trail to the base of the slope, then walk southward for about half a mile. Proceed up to slope to the west (right) to a location in the eastern part of sec. 26 (Fig. 22) overlooking the Browns Creek gorge.

Stop 20: Rhyolite Lava Flow (?) at Browns Creek

Notable features: intermixed glassy and devitrified rhyolite, steeply dipping layering

Walk northwest for about half a mile along the canyon rim and the adjacent hill in order to examine the upper part of the rhyolite lava flow. As you proceed, note the intermixing of glassy and devitrified rhyolite, the occurrence of spherulites, the generally steeply dipping development of flow layering, and the local occurrences of vuggy zones (Fig. 23). These phenomena

Figure 22. Geologic sketch map of the Browns Creek area, located about 7 miles south of Oreana, Idaho. The general locations of Stops 19, 20, 21, and 22 are indicated.
Stop 20, to the cliffs of welded tuff on the north side of the basin (Fig. 22). The route for this part of the walk will depend on the water level in the streams and the amount of interest in scaling the steep slopes.

Stop 22. Rheomorphically Deformed Welded-Tuff Unit

Notable features: ramping and other rheomorphic structures, structural changes dependent on unit thickness, basal vitrophyre and breccia

The half-mile-long canyon cut by Browns Creek exposes a section through the welded-tuff unit. Where this unit is thin, it generally has a horizontal attitude. Where the canyon exposes the thickest part of the unit, however, one can see the pronounced development of rheomorphic features, including the rotation (ramping) of originally subhorizontal emplacement boundaries and lithophysal zones to a near vertical attitude (Fig. 26).

Much of this rheomorphic adjustment took place as a result of a combination of plastic deformation of devitrified rhyolite and movement along fractures that formed as the rhyolite cooled. The vitrophyre and breccia at the base of this unit are exposed in the bed of the tributary to Browns Creek in Sec. 24 (Fig. 22).

Description of Drive from Browns Creek Area to Sinker Creek

Return to the vehicles and head north towards Oreana. Go through Oreana to Highway 78, and turn northwest (left) towards Murphy. About 5 miles from Oreana we...
Murphy, the county seat of Owyhee County, to the junction of Highway 78 with Idaho Highway 45. Turn toward the north (right) and cross the Snake River on Highway 45. Just after you cross the Snake River, Bonneville flood gravels are well exposed on the northeast side of the river. Proceed about 0.4 mile past the river to Ferry Road that turns east (right) at the top of the grade. Turn here and follow Ferry Road eastward for about 2 miles to its intersection with Hill Road.

**View from Vehicle**

**North of Snake River**

At this intersection, the large hill to the northeast is Walters Butte. It consists of a thick section of palagonite tuff that overlies lacustrine sediments of the Idaho Group. Walters Butte is one of several basalt volcanoes in this part of the Snake River Plain that erupted within Lake Idaho during the Pliocene or early Pleistocene. Most of the material that was erupted came from a north-south trending dike system along the east side of the volcano. White Butte, a smaller subaqueous basalt volcano also composed mainly of palagonite tuff, is visible about a mile south of this intersection.

Turn right and drive south on Hill Road about a mile toward White Butte, then follow the road generally eastward for another mile. At the junction near the large feed lot, turn south (right) and drive about 2.5 miles on the gravel road. Stop at the old railroad bridge across the Snake River.

---

**Stop 23. Water-Affected Basalt at Sinker Creek**

Notable features: distribution of altered basalt, lack of pillows

Various features in these basalt flows suggest that they were affected by water as they cooled. Pillows seem to be absent from the exposure, but chilled zones along joints and zones that became altered as they cooled more slowly are prominent. These basalt flows evidently were erupted into a body of standing water, probably Lake Idaho, from a small basalt volcano (Hill 3337) located about a mile northeast of this stop.

**Description of Drive from Sinker Creek to Stop 24**

Return to vehicles and continue west, passing through Murphy, the county seat of Owyhee County, to the junction of Highway 78 with Idaho Highway 45. Turn toward the north (right) and cross the Snake River on Highway 45. Just after you cross the Snake River, Bonneville flood gravels are well exposed on the northeast side of the river. Proceed about 0.4 mile past the river to Ferry Road that turns east (right) at the top of the grade. Turn here and follow Ferry Road eastward for about 2 miles to its intersection with Hill Road.

**View from Vehicle**

**North of Snake River**

At this intersection, the large hill to the northeast is Walters Butte. It consists of a thick section of palagonite tuff that overlies lacustrine sediments of the Idaho Group. Walters Butte is one of several basalt volcanoes in this part of the Snake River Plain that erupted within Lake Idaho during the Pliocene or early Pleistocene. Most of the material that was erupted came from a north-south trending dike system along the east side of the volcano. White Butte, a smaller subaqueous basalt volcano also composed mainly of palagonite tuff, is visible about a mile south of this intersection.

Turn right and drive south on Hill Road about a mile toward White Butte, then follow the road generally eastward for another mile. At the junction near the large feed lot, turn south (right) and drive about 2.5 miles on the gravel road. Stop at the old railroad bridge across the Snake River.

---

**Figure 26. View, about 1.2 miles northwest of Stop 22, of the steeply dipping layering and lithophysal zones in the thick, rheomorphically deformed part of the welded-tuff unit.**
Stop 24: Melon Gravel and Guffey Butte Volcanic Complex

Notable features: basalt flows, dikes, and cinders across the river, flood-deposited boulders

Here we will examine the Melon "Gravel," which is composed of basalt boulders plucked from various flows and deposited by the Bonneville flood as a large bar. The Guffey Butte volcanic complex is located across the river. In it you can see dikes and cinder-rich zones that developed in this subaqueous basalt volcano, and the capping basalt flows.

Description of Drive From Stop 24 to Boise

Backtrack to Highway 45 along the same route. Turn north (right) and follow Highway 45 about 3 miles to its intersection with the Melba Road. Along Highway 45, about a mile south of its intersection with the Melba Road, the route passes through the eastern part of the crater area in the Grrouch Drain volcano. From the highway the floor and western and northern walls of this small volcano are clearly visible.

Turn east (right) onto Melba Road and follow it for 3 miles to its junction with the Can-Ada road. The hill to the north (left) of Melba Road at about a mile past (east of) the turn off from Highway 45 is McElroy Butte, a small shield volcano.

At Can-Ada Road turn north (left) and drive north for about 1.5 miles to Dickman Road. While driving north on Can-Ada Road, the large broad hill straight ahead is Powers Butte, a basaltic shield volcano which is about 3 miles across and about 200 feet in height.

Turn east (right) onto Dickman Road and follow it eastward for a mile to Robinson Boulevard. Turn north (left) onto Robinson Boulevard and follow it for about 5.5 miles to Kuna Way. While driving north on Robinson Boulevard Kuna Butte will be visible to the east (right). Kuna Butte is another broad shield volcano, with base dimensions of about 2 by 3 miles, and a height of about 400 feet. The upper part of this volcano is cut by a series of west-northwest-trending faults to form a shallow graben across the top of the shield.

At Kuna Way turn east (right) and proceed to the town of Kuna. From Kuna take Idaho Highway 69 to Meridian. At the south edge of Meridian turn to the east (right) and follow I-84 to the Boise airport (Vista Avenue exit).

ACKNOWLEDGMENTS

This Magic Reservoir eruptive center portion of the field guide was prepared with support of National Science Foundation Grants EAR 83-20358 and EAR 80-18580. We especially wish to thank Bill Hackett for his assistance in preparing the panorama photographs. Also, we thank Don Fiesinger for his review.

REFERENCES

Alief, M. H., 1962, Variation in SiO₂, Al₂O₃, Fe₂O₃, CaO, MgO, Na₂O, K₂O, and H₂O in some ignimbrites in southern Twin Falls County, Idaho: Masters thesis, University of Idaho, Moscow, 60 p.


resources: Idaho Department of Water Resources

Schmidt, D. L., 1961, Quaternary geology of the
Bellevue area in Blaine and Camas Counties, Idaho:
Ph. D. dissertation, University of Washington,
Seattle, 125 p.

Schrader, F. C., 1923, The Jarbidge mining district,
Nevada, with a note on the Charleston district: U. S.

Smith, C. L., 1966, Geology of eastern Mount Bennett
Hills, Camas, Gooding, and Lincoln Counties, Idaho:
Ph. D. dissertation, University of Idaho, Moscow,
129 p.

Stewart, J. H., and Carlson, J. E., 1978, Geologic map

Struhsacker, D. W., Jewell, P. W., Zeisloft, Jon, and
Evans, S. H., Jr., 1982, The geology and geothermal
setting of the Magic Reservoir area, Blaine and
Camas Counties, Idaho, in Bonnichsen, Bill, and
Breckenridge, R. M., editors, Cenozoic geology of
Idaho: Idaho Bureau of Mines and Geology Bulletin
26, p. 377-393.

Swirydczuk, Krystyna, Wilkinson, B. H., and Smith,
G. R., 1979, The Pliocene Glenns Ferry oolite: lake
margin carbonate deposition in the southwestern
Snake River Plain: Journal of Sedimentary Petrology,
v. 49, p. 995-1004.

_____, 1980, The Pliocene Glenns Ferry oolite:
sedimentology of oolitic lacustrine terrace deposits:
Journal of Sedimentary Petrology, v. 50, p. 1237-
1248.

rocks of the Miocene Idavada Group, Bennett
Mountain, southwestern Idaho: final contract report to
the Los Alamos National Laboratory from Boise
State University, 55 p.
INTRODUCTION

In this guide, we discuss Neogene rhyolite and Quaternary basalt on the eastern Snake River Plain and along its margins. The physical processes and deposits of explosive volcanism are emphasized, including rhyolite ignimbrites (ash-flow tuffs) and basaltic hydrovolcanic features that formed during interaction between basic magma and external water. Figure 1 is an index map for the field trip, which has nine stops and is intended to take two days. The first day covers rhyolitic and basaltic deposits along the southern margin of the eastern Snake River Plain from Idaho Falls to Heise and along the South Fork of the Snake River (Stops 1-5). The second day includes basaltic tuffs of the Menan volcanic complex near Rexburg and rhyolitic tuffs along the northern margin of the eastern Snake River Plain from Dubois to Howc Point (Stops 6-9).

The Snake River Plain is the dominant geomorphic feature of southern Idaho and covers nearly a quarter of the state. Volcanic and sedimentary rocks form a 40- to 60-mile-wide, arcuate belt that extends about 400 miles from the Idaho-Oregon border to Yellowstone National Park. A remarkable feature of the Snake River Plain is that the rhyolitic volcanism generally becomes younger from southwest to northeast (Armstrong and others, 1975) (Fig. 2). This has led to speculation that the eastern Snake River Plain is the trace of a mantle plume beneath western North America or the tip of an eastward-propagating lithospheric fracture (Christiansen and McKee, 1978). In either case, Tertiary rhyolites of the Snake River Plain are now largely covered by Quaternary basalts and the focus of Quaternary rhyolite volcanism has been the Yellowstone Plateau-Island Park region, where comparatively small volumes of basaltic lava have erupted.

Neogene rhyolitic ash-flow tuffs and lava flows are the most voluminous and widespread volcanic deposits of the Snake River Plain. Recent work on these rhyolites includes Kellogg and Embree (1986), Kellogg and Marvin (1988) and Bonnichsen and others (1988, this volume). In this guide, we discuss late Miocene to early Pliocene rhyolites that are present around the eastern Snake River Plain, from Idaho Falls to the Teton Range on the southern margin of the plain and from the Centennial Mountains to the Pioneer Mountains along its northern margin.

NEOGENE RHYOLITIC VOLCANISM OF THE EASTERN SNAKE RIVER PLAIN

A stratigraphic sequence and regional correlations have been established for three voluminous and extensive ignimbrites of the eastern Snake River Plain, ranging in age from 4.3 to 6.5 Ma. This stratigraphic package of densely welded ignimbrite sheets, together with associated volcanic rocks and sediments, forms the Heise volcanic field (Morgan and others, 1984) which is analogous in its development to the adjacent Quaternary Yellowstone Plateau volcanic field to the northeast (Christiansen and Blank, 1972, Christiansen, 1984).
Thirty years of local studies on Tertiary rhyolites around the eastern Snake River Plain have produced the complex “previous nomenclature” that is outlined in Table 1. Regional correlation of major ignimbrites from the Heise volcanic field has recently been accomplished with the aid of paleomagnetic remanent directions, radiometric ages, geochemistry, stratigraphic position, the compositions, sizes and relative abundances of phenocrysts, and macroscopic lithology. These correlations (e.g., Morgan and others, 1984; Morgan, 1988a) lead to the simpler terminology that is given under “present nomenclature” in Table 1. Formal stratigraphic names will soon be introduced (Morgan, 1988b and unpublished data), but we use the established terminology of Morgan and others (1984) throughout this guide.

Deposits of the Heise volcanic field cover more than 35,000 square kilometers in southeastern Idaho. Exposures are mostly limited to the mountains along the margins of the eastern Snake River Plain, but correlative rhyolitic units have also been identified in subsurface boreholes drilled through Quaternary basalt on the plain (Doherty and others, 1979). The three widespread ignimbrites of the Heise volcanic field are the 6.5 Ma tuff of Blacktail, the 6.0 Ma tuff of Blue Creek, and the 4.3 Ma tuff of Kilgore. In many places, these major ignimbrite sheets are found together as a stratigraphic package with local intercalated rhyolite lava flows, smaller ignimbrites, pyroclastic fall deposits, basaltic lava flows and tephra, and sediments. Although facies variations occur, all three of the major ignimbrites are composed of densely welded, crystal-poor, high-silica peraluminous rhyolite.

The three widespread ignimbrites of the Heise volcanic field are inferred to have erupted from major calderas that are now largely buried beneath Quaternary basalts of the eastern Snake River Plain (Fig. 3). The earliest major eruption of the Heise volcanic field produced the immense, plainwide Blacktail caldera, which was the source of the tuff of Blacktail and within
which the later calderas were either nested or largely overlapped. The second major event, perhaps a phase of the initial event, produced the smaller, nested Blue Creek caldera from which the tuff of Blue Creek erupted. A third major eruption produced the immense plainwide Kilgore caldera, the source of the tuff of Kilgore (Morgan, 1988a). Important evidence of these inferred source calderas is found along both margins of the eastern Snake River Plain, in the Poplar-Heise area on the southern margin (Stops 1-3), and in the Howe Point and Lidy Hot Springs areas on the northern margin (Stops 8 and 9).

The dimensions of Neogene calderas beneath the eastern Snake River Plain are not accurately known because of burial by Quaternary basalts and because most rhyolite exposures occur along the margins of the plain. However, careful study of the lateral variations in ignimbrite geometry and lithology allow inference of the directions and approximate distances to the source calderas. Facies changes within each ignimbrite as well as analyses of flow directions, volcanic and tectonic structures, and geophysical anomalies have all been used to infer the locations of the source calderas that are shown in Figure 3. The estimated volumes of the

Table 1. Stratigraphic nomenclature for the three major ignimbrites of the Heise volcanic field. References: 1, Scholten and others (1955); 2, Stearns and Isotoff (1956); 3, Carr and Trimble (1963); 4, Albee and others (1975); 5, Christiansen and Love (1978); 6, Prostka and Embree (1978); 7, Skipp and others (1979); 8, McBroome (1981); 9, McBroome and others (1981); 10, Morgan and others (1984); 11, Morgan (1988a). Previous correlation of the tuff of Elkhorn Spring with the tuff of Blue Creek (McBroome and others, 1981; Morgan and others, 1984) is now known to be erroneous, and the true correlation is between the tuff of Elkhorn Spring and the tuff of Kilgore (Morgan, 1988a). See Figure 3 for locations of inferred source calderas.

<table>
<thead>
<tr>
<th>PREVIOUS NOMENCLATURE</th>
<th>CURRENT NOMENCLATURE</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Northern Margin</strong></td>
<td><strong>Southern Margin</strong></td>
</tr>
<tr>
<td>tuff of Spencer 7</td>
<td>tuff of Heise 4,6,9</td>
</tr>
<tr>
<td></td>
<td>tuff of Heise C 4</td>
</tr>
<tr>
<td></td>
<td>tuff of Elkhorn Spring 11</td>
</tr>
<tr>
<td></td>
<td>Conant Creek Tuff 5,8</td>
</tr>
<tr>
<td>tuff of Blue Creek 7,8,9</td>
<td>tuff of Elkhorn Spring 9,10</td>
</tr>
<tr>
<td></td>
<td>Walcott Tuff 2,3</td>
</tr>
<tr>
<td>tuff of Edie School 7</td>
<td>tuff of Spring Creek 6,9</td>
</tr>
<tr>
<td>tuff of Edie Ranch 8,9</td>
<td></td>
</tr>
<tr>
<td>Edie School Rhyolite 1</td>
<td></td>
</tr>
</tbody>
</table>
Neogene Heise ignimbrites are comparable with those of the Quaternary Yellowstone ignimbrites (Fig. 4). We therefore suggest that the buried Heise calderas have dimensions comparable to those of the better-exposed Yellowstone Plateau volcanic field.

QUATERNARY BASALTIC VOLCANISM

Basalts and sediments of the eastern Snake River Plain are part of the Snake River Group, a stratigraphic unit composed largely of tholeiitic basalt lava flows that erupted during the past 2 million years (Fig. 2). Typical landforms of Quaternary basaltic volcanism on the eastern Snake River Plain are illustrated in Figure 5. Basaltic volcanism has generally been effusive, with lava flows erupted from broad, low, central-vent shield volcanoes and from fissures; many flows were fed by major lava tubes. Small shield volcanoes are typical of the eastern Snake River Plain, each having slope angles less than 0.5 degree, covering tens to hundreds of square kilometers, and composed almost entirely of fluid basalt lava flows. Cinder cones are relatively uncommon on the eastern Snake River Plain and most are confined to volcanic rift zones where viscous, differentiated magma erupted, such as at the Craters of the Moon lava field (Kuntz and others, 1988; Stephens, 1988, this volume).

Quaternary basalt vents of the eastern Snake River Plain commonly form linear arrays of fissure flows, small shields, spatter and cinder cones, pit craters and open cracks. These features define north-northwest-trending volcanic rift zones that are parallel to the fault-bounded margins of the surrounding basin and range mountains. Most recently active was the Great Rift (Kuntz and others, 1986; 1988), where Holocene eruptions occurred at the Craters of the Moon and several other lava fields. Holocene lava flows also erupted locally along several other rift zones of the eastern Snake River Plain, but most of the volcanism is Pleistocene in age (Kuntz, 1978a,b; Kuntz and Dalrymple, 1979; Kuntz and others, 1979).

An important geological feature of the eastern Snake River Plain is the voluminous Snake River Plain aquifer (Whitehead, 1986; Wood and Low, 1986). The aquifer is largely hosted by Quaternary basalt lava flows and intercalated sediments. It underlies 28,000 square kilometers of the eastern Snake River Plain, and its recharge area in the surrounding basin and range mountains covers 65,000 square kilometers. The annual recharge and discharge are currently about 8 million acre-feet, and the storage capacity is estimated to be about equal to the present volume of Lake Erie (U. S. Geological Survey
Hackett and Morgan—Basaltic and Rhyolitic Volcanism of the Eastern Snake River Plain, Idaho

287

Figure 4. Estimated volumes of Quaternary Yellowstone Group ignimbrites, Neogene Heise volcanics, with selected Holocene and historical eruption volumes shown for comparison. Adapted from Christiansen, 1984.

In prehistoric times, the aquifer was fed mostly by underflow through the surrounding alluvium-filled valleys, and by infiltration of surface runoff. Since the early 1900s, an additional source of recharge has been the percolation of irrigation water diverted from the Snake River and its tributary streams.

Hydrovolcanic constructs have formed on and around the eastern Snake River Plain as a result of interaction between basaltic magma and shallow ground water (King, 1982; Womer and others, 1982; Hackett and others, 1987). Tuff cones, tuff rings and phreatic explosion craters range from late Miocene (e.g., Massacre volcanic complex; Luessen, 1987a;b) through Pleistocene (e.g., Menan volcanic complex; Ferdock and Hackett, 1986; Ferdock, 1987; Creighton, 1987; and Conant Valley volcanics; Jobin and Schroeder, 1964; Roberts, 1981; this paper) and Holocene in age (e.g., Kings Bowl; King, 1982).

Depth and availability of ground water are major factors controlling the locations, lithologies and volumes of hydrovolcanic constructs on the eastern Snake River Plain. In addition to the depth of the water table, the lithology of the aquifer (which affects permeability) is also an important factor.

Large Pleistocene basaltic hydrovolcanic complexes such as Massacre, Menan, and Conant Valley occur along the southern margin of the Snake River Plain, near the Snake River and where the aquifer is shallow and voluminous. Large tuff cones and rings formed most commonly on permeable, weakly cemented alluvial deposits. In contrast, hydrovolcanic features rarely occur and are small in volume on the northern part of the eastern Snake River Plain, where the water table is deep (in many places greater than 300 m; Whitehead, 1986) and the aquifer is hosted by fracture-permeable basalt lava flows.

Tuffs, tuffaceous sediments and pillow lavas of the Conant Valley volcanics (Fig. 1; Stops 4 and 5), and tuff cones and rings of the Menan volcanic complex (Stops 6 and 7) illustrate the variety of hydrovolcanic features that resulted from explosive interaction between basaltic magma and external water.

**RHYOLITE AND BASALT ALONG THE SOUTHERN MARGIN OF THE EASTERN SNAKE RIVER PLAIN**

**Travel from Idaho Falls to Blacktail Recreation Area**

The trip begins in Idaho Falls (Fig. 1). From the intersection of Broadway (Highway 20) and Yellowstone Avenue (Highway 191), proceed 1.8 miles north on
Figure 6. Sketch of volcanic units exposed at Stop 1, Blacktail recreation area. The view is toward the east, with the Ririe Reservoir at the lower right.

Yellowstone to the intersection of Yellowstone and Lincoln Road. Turn right (east) on Lincoln Road and proceed about 12 miles to the Blacktail recreation area on Ririe Reservoir.

Stop 1. Rhyolitic Tuffs at the Blacktail Recreation Area

Stop 1 includes two localities within continuous exposures of rhyolitic tuffs at the Blacktail recreation area. The Stop 1 localities are near the inferred structural margin of the Blacktail caldera, source of the 6.5 Ma tuff of Blacktail. The first locality is a road cut exposing gray and tan rhyolitic tuffs along the left (north) side of the paved road, where the road begins its steep descent to the reservoir. The second locality is the tuff of Blacktail exposed at the boat ramp, which is reached by walking or driving down the paved road to Ririe Reservoir. The Stop 1 road cut is an excellent vantage point for an overview of the Heise volcanic units in this region. Exposed in the valley to the east (Fig. 6) is a stratigraphic section along the rim of the Blacktail caldera (Morgan, 1988a). At the boat ramp an important exposure of the tuff of Blacktail, the 6.5 Ma ignimbrite of the Heise volcanic field (Morgan and others, 1984; see below for more detailed discussion).

At the Stop 1 road cut, two members of the 2.1 Ma (Christiansen and Blank, 1972; Christiansen, 1984) Huckleberry Ridge Tuff form the top of the exposure and cap the hill. The lower Member A is nonwelded and poorly sorted, and it grades from tan at the base to pink at the top. Member A is separated from overlying Member B by a fines-depleted, lithic- and pumice-rich layer that ranges from 1 cm to 12 cm in thickness. Member B contains a pumice concentration zone about 18 cm from its base, and the upper part is pink, platy and crystal-poor. The Huckleberry Ridge Tuff rests conformably on a sequence of intercalated fine, planar-bedded lacustrine ash units, and pumiceous surge deposits with complex bed forms (Fig. 7). These deposits in turn rest unconformably on the tuff of Wolverine Creek, a local, nonwelded, obsidian-rich ignimbrite.

At this locality the tuff of Wolverine Creek contains crystal- and obsidian-rich, fines-depleted elutriation pipes (Fig. 8), together with pods of laminated, obsidian-rich ignimbrite that have abrupt terminations and soft-sediment deformation features (Fig. 9). The laminated pods may be xenoliths of an earlier pyroclastic flow that remained coherent when picked up and incorporated in later emplacement units of the tuff of Wolverine Creek.

Drive or walk downhill to the boat ramp, where an excellent section of the tuff of Blacktail is exposed (Fig. 6). At the boat ramp the tuff is a relatively thin (about 8-11 meters) and undeformed outflow facies ignimbrite composed of a basal vitrophyre (exposed only when water level is low), a devitrified, densely welded interior, a massive vapor-phase zone, a lithophysal zone, and a cap of black, densely welded vitrophyre. A 50 centimeter thick, red ash deposit overlies the black vitrophyre. A few hundred meters northeast of the boat ramp the tuff of Blacktail becomes greatly thickened (>150 m), is silicified and crystal-rich and has tight flowage folds. This thick tuff represents the ponded intracaldera facies (Fig. 6). Lower in the valley is a mesa of canyon-filling Quaternary basalts. The ridge behind the mesa is capped by the Huckleberry Ridge Tuff, underlain successively by the 4.3 Ma tuff of Kilgore, the tuff of Wolverine Creek, and the tuff of Blacktail.

Travel from Blacktail Recreation Area to Meadow Creek Dugway

Leave the Blacktail recreation area by heading west along Lincoln Road. At the first intersection, turn right and proceed north on Johnson Road. Isolated outcrops of the 4.3 Ma tuff of Kilgore are exposed along the road. Cross the railroad tracks, and at the stop sign turn right and proceed east on Highway 26 for about 1.5 miles. Turn right (south) on Meadow Creek Road. About 3 miles past the Ririe Dam picnic area, bear right at a
junction in the road, continue 100 feet and park near the top of the cliff.

Stop 2. Meadow Creek Dugway

Walk northwest on the surface of the Huckleberry Ridge Tuff along the rim of the cliff for about 1/3 mile. Proceed down a small ravine through an excellent section of the 4.3 Ma tuff of Kilgore. Here the tuff of Kilgore is an outflow facies ignimbrite that is about 11 meters thick (Fig. 10); it contains at least three cooling breaks that are marked by internal variations in the degree of welding. Subtle zonations of mineralogy and trace element concentrations occur within the tuff and are further evidence that it was emplaced as several pyroclastic flows in rapid succession. The tuff of Kilgore rests on a thin (10 cm) well-sorted ash, in turn resting on a thin (50 cm) orange, lithic-rich, nonwelded ignimbrite. The ignimbrite is separated from the underlying tuff of Wolverine Creek by a thin (1 cm) paleosol.

Walk back toward the vehicles but continue down the road into Meadow Creek valley. The 2.1 Ma Huckleberry Ridge Tuff forms the rim of the cliff (Fig. 11) and is separated from the underlying 4.3 Ma tuff of Kilgore by a 40-centimeter-thick paleosol. Continuous exposures of the tuff of Kilgore show great thickness variation from less than 3 meters to more than 13 meters over a distance of 2 kilometers, as a result of emplacement over irregular topography. The tuff of Kilgore contains multiple pyroclastic flow units with locally derived lithics, lithic concentration zones, and fines-depleted, crystal- and lithic-rich ground layer deposits. Beneath the tuff of Kilgore is a coarse fluvial gravel, then a sequence of plinian and lacustrine ash deposits, and the tuff of Wolverine Creek. Farther down the road across Mud Spring Creek is an outflow facies exposure of the tuff of Blacktail.

Travel from Meadow Creek Dugway to View of Heise Cliffs

Return to the cars and backtrack (north) along Meadow Creek Road to Highway 26. Turn right (east) on Highway 26 and proceed 1.5 miles to a sign on the left, “Heise Hot Springs, 4 miles.” Turn left, head north about 1 mile and then pull off onto the gravel shoulder for a view of the Heise cliffs.
Stop 3. View of the Heise Cliffs

About 2 miles to the northeast, the Heise cliffs expose a thick section of the Heise volcanics in the uplifted block of the Grand Valley fault. The Heise cliffs are one of the main areas in which the stratigraphy of the Heise volcanic field has been established. The great thickness and coarseness of the tuff of Kilgore and the tuff of Blacktail, together with the large number of intercalated rhyolite lava flows, suggest that the Heise cliffs section is located near the overlapping margins of the Kilgore and Blacktail calderas (Fig. 3). It was originally thought that all three major ignimbrites of the Heise volcanic field were present in the Heise cliffs (Morgan and others, 1984), but more recent paleomagnetic data (Morgan 1988a) show that the tuff of Blue Creek is not present.

The cliffs are capped by a thick section of 4.3 Ma tuff of Kilgore. Three lines of evidence suggest that the tuff of Kilgore was erupted from a source near Heise: (1) the average thickness of the tuff of Kilgore on the Heise cliffs is about 30 meters, (2) the tuff contains coarse crystals and large lithic and pumice clasts and (3) paleomagnetic data (Morgan 1988a) suggest a source in the Poplar-Heise area. Below the tuff of Kilgore is a sequence of rhyolitic ignimbrites (sheetlike in cross

Figure 10. Schematic section of the tuff of Kilgore exposed in a small ravine at Stop 2, Meadow Creek dugway.

Figure 11. Photograph of 2.1 Ma Huckleberry Ridge Tuff at Stop 2, Meadow Creek dugway. Conspicuous parting and color change above meter stick is the contact between members A and B.
section) with intercalated rhyolite lava flows (lenticular cross sections). Ignimbrites include the local tuffs of Hawley Gulch, Wolverine Creek, Dora Springs, and Newby Ranch (8.6 ± 0.5 Ma) and the local rhyolites of Kelly Canyon (5.7 ± 0.1 Ma) and Hawley Spring (7.2 ± 0.1 Ma). Near the base of the section is the 6.5 Ma tuff of Blacktail.

Travel from Heise Cliffs to Conant Valley Volcanics

Mileage

0.0 Reset odometers at intersection of Heise Hot Springs road and Highway 26. Proceed east on Highway 26.

3.9 Rhyolite outcrops on opposite (northeast) side of the South Fork of the Snake River are unconformably overlain by Quaternary basalt lava flows.

5.3 Rest area with toilets and Snake River overlook on the left. On the near (southwest) side of the river, about 12 subaerial basalt lava flows are exposed. These sheetlike flows probably erupted from Snake River Plain sources to the west.

17.2 Beginning of extensive road cut and descent into Conant Valley. Uppermost volcanic deposits in Conant Valley are subaerial lava flows up to about 3 meters thick, locally overlying or interbedded with palagonite tuffs. Interbedded subaerial lava flows and palagonite tuffs suggest that several vents were concurrently active, some "dry" and effusive and others "wet" and explosive. Conspicuous vertical fractures in the tuffs contain palagonite and calcite.

17.9 On the sweeping curve to the right, carefully turn off to the right, through the opening in the guard rail and onto the flat graded surface beyond. This is Stop 4 (optional). Continue east on Highway 26 to reach Stop 5.

18.5 Turn off to the right, onto the narrow shoulder just beyond the end of the guard rail. Walk about 0.1 mile northwest, to the base of the cliff. This is Stop 5.

Stop 4 (optional). Conant Valley Tuff Cone

Basalt deposits in Conant Valley (which we informally refer to as the "Conant Valley volcanics") are a complex assemblage of Pliocene-Pleistocene lava flows, tuffs and fluviatile sediments (Jobin and Schroeder, 1964; Roberts, 1981). Lava flows from the eastern Snake River Plain and from local sources partially filled the valley, disrupting the drainage of the Snake River. Most eruptions involved the subaerial effusion of lava flows, but local pillow lavas in the valley fill indicate that shallow, impounded bodies of water were present. Primary and redeposited palagonite tuffs also indicate that local explosive eruptions occurred when basaltic magma interacted with shallow ground water or surface water. (We use the term "palagonite" to mean hydrated basaltic glass; it is a yellow-brown mineraloid that often coexists with other alteration products such as calcite and zeolites. Palagonite forms most efficiently in the presence of high-temperature steam during hydrovolcanism, but protracted hydration at lower temperatures also produces interstitial palagonite cement.)

Extensive road cuts expose massive and bedded palagonite tuffs of the Conant Valley tuff cone; the exposure at Stop 4 is about 40 meters high. The deposits are planar- and cross-bedded, moderately sorted palagonite tuffs and lapilli tuffs. The juvenile component is lapilli-sized fragments of basaltic scoria with outer glassy rinds of sideromelane that were quenched during interaction with external water; the smaller, more vesiculated juvenile fragments are entirely palagonitized. Accidental clasts are pebbles and cobbles of (1) quartz arenite (rounded cobbles of Pennsylvanian-Permian Wells Formation, derived from underlying alluvium), (2) cobble-sized, angular, dense tachylitic basalt with large vesicles (derived from older basalt lava flows) and (3) subangular, dense, devitrified rhyolite pebbles (derived from older ash-flow tuffs).

The deposits generally lack the good sorting, planar bedding and sag structures beneath ballistic blocks that are typical of near-vent airfall material. Rather, the bedding structures suggest transport and deposition as sheet wash or grain flows on the steep outer flanks of a tuff cone that was formed when vesiculating basaltic magma erupted through water-saturated alluvial deposits.

Stop 5. Tuffaceous Sediments and Lava Flows of Conant Valley

About 80 meters of tuffaceous sediments and lava flows are exposed along the cliff face. The deposits are divided into three lithologic units for descriptive purposes (Fig. 12).

Basal deposits (unit 1) are massive, heterolithologic, matrix-supported tuff breccias with palagonitic matrix and clasts of scoria, pillow fragments and accidental rock fragments as described under Stop 4. Subtle variations in grain size and clast proportions suggest numerous lobate depositional units, each a debris flow or mudflow that probably originated by failure and mass flow of water-saturated palagonite tuffs onto the valley floor. Several thin, anastomosing dikes of vesicular basalt intrude the tuff breccias and were probably feeders for the lava flows at the top of the section.

Unit 2 is composed of fine-grained, well-cemented, yellow-brown palagonite tuffs and lapilli tuffs (tuffaceous sand and gravel) containing planar beds and low-angle cross beds. Internal unconformities are present and many beds have lenticular cross sections, suggesting fluvial deposition. Planar-bedded, faintly laminated fine tuffs at the top of unit 2 are interpreted as lacustrine
deposits that formed in shallow, standing water behind a natural dam of lava and tuff.

Unit 3 consists entirely of basalt lava flows, emplaced both subaerially and into water. Along the cliff face, sheet-like subaerial pahoehoe flows are intercalated with pod-like pillow lavas. Individual flows commonly grade upward from basal pillow lava into subaerially emplaced pahoehoe, suggesting flowage into water that was only a few meters deep. Pillow cross sections are well displayed, with glassy rims, multiple palagonite rinds and interstitial pods of hyaloclastite. Loess deposits cap the volcanic sequence.

Pahoehoe lava flows, pillow lavas and tuffaceous sediments are also the dominant lithologies in road cuts along the next 6 miles of Highway 26 east of here. These exposures suggest that the South Fork drainage was disturbed numerous times by Pleistocene lava flows and tuff constructs, involving erosion or the deposition of coarse fluvial sediments downstream of the natural dams, and the deposition of lake sediments and pillow lavas into the impounded water upstream of the dams. Similar volcanic facies relationships have been reported for other Pleistocene deposits exposed along the canyon of the Snake River on the central Snake River Plain (Malde, 1982) and along the Boise River canyon (Howard * and others, 1982).

Travel from Conant Valley to Menan Volcanic Complex

Reverse direction and travel west on Highway 26, back toward Idaho Falls. Watch for turnoff to Heise Hot Springs, 18.5 miles west.

Mileage

0.0 Reset odometers at intersection of Highway 26 and Heise Hot Springs road, but do not turn there. Continue west on Highway 26.
1.5 Road to Meadow Creek recreation area on left (south); continue west on Highway 26.
3.1 Turn right (north) on Highway 48 toward Ririe.
4.1 Town of Ririe. Turn left (west) on Highway 48 at stop sign.
4.3 Continue straight on Highway 48 toward Rigby.
Between Ririe and Menan the road travels over Pleistocene “younger mainstream gravel” of the Snake River (Pierce and Scott, 1982).
14.2 Turn right on business route 20 at stop sign in Rigby.
14.4 Turn right (continue on business route 20) at traffic light.
15.5 Turn left on Highway 20 north toward Rexburg.
19.2 Intersection with State Highway 80; sign indicates “Menan 7.” This is the end of the first day.

MENAN VOLCANIC COMPLEX AND NORTHERN MARGIN OF THE SNAKE RIVER PLAIN

Introduction to the Menan Volcanic Complex

The Menan Buttes are known to many geologists because they have been used in topographic map exercises of popular introductory laboratory manuals; they are outstanding morphological examples of tuff cones. Stearns and others (1938) first noted the Menan Buttes as conspicuous features on the eastern Snake River Plain, and Hamilton and Myers (1963) did a reconnaissance geomorphic and petrographic study of the buttes. Recent work includes Creighton (1982, 1987),
Ferdock and Hackett (1986), Hackett and others (1987), and Ferdock (1987). The following discussion is largely adapted from the latter three references.

Six late Pleistocene hydrovolcanic constructs make up the Menan volcanic complex, an assemblage of basalt vents that occurs along a 5-kilometer-long, northwest-trending lineament about 8 miles west of Rexburg, Idaho (Fig. 13). Tuff cones and tuff rings were formed during the injection of basaltic magma into near-surface, water-saturated alluvial sediments and basalt lava flows of the Snake River Plain aquifer, producing voluminous deposits of palagonite tuff (Fig. 14).

Mileage

0.0 Intersection of State Highways 20 and 80, about 8 miles south of Rexburg. Travel west on Highway 80 toward Annis and Menan.
2.5 Town of Annis.
3.6 Bear right onto narrow paved road.
3.9 Turn left onto road 400 N.
4.1 Turn left into gravel quarry. This is Stop 6.

Stop 6. North Little Butte Tuff Ring

A few miles south of the Menan Buttes tuff cones are the remnants of two tuff rings: the North and South Little Buttes (Fig. 13). In contrast to the tuff cones, the tuff rings are much smaller and have lower slope angles; the deposits are strongly cyclic and contain abundant accidental quartzite pebbles derived from the underlying permeable alluvium. The tuff rings are dominated by two types of bed sets (Figs. 15 and 16). Black, planar-bedded, coarse, scoriaceous lapilli tuff bed sets (mostly fall, with minor surge deposits) were formed during relatively "dry vent" violent strombolian eruptions that were driven partly by magmatic gas expansion (vesiculation). Intercalated bed sets of tan, cross-stratified, fine palagonite tuffs (fall and surge deposits) were formed during relatively "wet vent" surtseyan eruptions, driven largely by the flashing of external water to steam (hydrovolcanic explosions).

Since groundwater was probably always available, the alternating strombolian and surtseyan bed sets suggest a fluctuating magma supply, with repeated dike injection into water-saturated alluvium. When actively vesiculating magma was present at the surface, scoriaceous ejecta (strombolian deposits) predominated. When explosions occurred within the alluvial deposits over-lying shallow dikes, fine palagonitic ejecta (surtseyan deposits) predominated. Accidental rounded quartzite pebbles are abundant in both varieties of tuff but are more thoroughly fragmented in the strombolian deposits as a result of thermal shock.
STROMBOLIAN BED SETS: black, scoriaceous, massive lapilli tuff and bomb beds; mostly airfall with some surge deposits

SURTSEYAN BED SETS: tan, massive to thin-bedded, fine palagonite tuff; surge and airfall deposits

Figure 15. Stratigraphic column of North Little Butte tuff ring, illustrating the cyclic nature of the deposits. Modified from Ferdock, 1987.

Travel from North Little Butte to North Menan Butte

Mileage

0.0 Reset odometers at North Little Butte quarry. From the quarry turn left and head west on road 400N.

1.9 At junction, turn left (north) on road 400W. Road traverses quartzite-dominated Quaternary alluvium of the Snake River.

3.7 Bridge over Snake River. Exposures along the riverbank are planar-bedded, distal airfall tuffs from South Menan Butte. About one mile upstream from the bridge, tightly folded tuffs are exposed at the southeastern base of South Menan Butte (Fig. 17). The folds appear to have formed by the crumpling of wet, cohesive tuff at the bottoms of slide sheets that detached and moved downhill along bedding surfaces.

4.5 The subdued crater of Center Menan Butte is visible on the right (east) between North and South Menan Buttes.

5.1 Turn right (east) onto the unmarked, improved...
gravel road that heads northeast, between North and South Menan Buttes.

6.0 Turn left onto unmarked gravel road that heads up the south flank of North Menan Butte.

6.2 Turn around and park along the wide part of the road. This is Stop 7. Walk a short distance west until reaching a gully. Walk up the gully to see excellent tuff exposures and eventually reach the southern crater rim where densely palagonitized tuffs and large ballistic blocks are exposed.

Stop 7. North Menan Butte Tuff Cone

The Menan Buttes tuff cones (Fig. 18) are the largest features of the volcanic complex, with reconstructed volumes of 0.7 (North Menan Butte) and 0.3 (South Menan Butte) cubic kilometer (the volume of Center Menan Butte is unknown). Dense basalt volume equivalents are 0.4 and 0.2 cubic kilometer, respectively (Ferdock, 1987). The Menan Buttes are among the largest terrestrial tuff cones, with volumes comparable to those of Diamond Head, Oahu (0.6 cubic kilometer; Wentworth, 1926) and Surtsey, Iceland (0.6 cubic kilometer; Thorarinsson, 1967).

Deposits of the North and South Menan Buttes tuff cones are monotonous, massive to thin-bedded, tan, lithified palagonite lapilli tuffs with minor accidental clasts of dense basalt and rounded quartzite pebbles. Near-vent deposits (within about 0.5 km from the crater centers) are massive, poorly sorted, palagonite lapilli tuffs, suggesting mass emplacement as slurry flows of wet, cohesive tuff. Along the crater rim of North Menan Butte, ballistic blocks of accidental basalt up to 1 meter in size occur, and dense palagonite alteration obscures the bed forms in the tuff matrix (Fig. 19). Flank deposits are crumpled, planar- and cross-stratified, moderately sorted palagonite tuffs with local armored lapilli horizons, suggesting deposition by airfall, pyroclastic surge and sheet wash (Fig. 20). Distal deposits beyond the cone flanks are fine, planar-laminated palagonite airfall tuffs. Deformation of the wet, cohesive tuffs during or soon after their deposition is indicated by monoclines, complex folds and possible detached slide blocks that are found on the lower flanks of the cones.

The monotonous palagonite tuffs of the Menan Buttes tuff cones were formed during protracted, steady-state eruptions of wet basaltic tephra. Hydrovolcanic explosions were fueled by steam produced by dike emplacement into the underlying aquifer. Explosions occurred above the aquifer, within the tephra-slurry of the central craters. As a result, accidental lithic clasts are uncommon: on the Menan Buttes tuff cones they comprise less than 5 volume percent of the deposits, as compared to more than 50 percent of some deposits on the Little Buttes tuff rings.
Figure 19. Photograph of dense palagonite tuff exposed at Stop 7, along the southern crater rim of North Menan Butte. Bed forms are nearly obliterated by palagonitization. Subrounded clast near hammer is a 20 cm accidental fragment of basalt lava. Other, smaller clasts are also accidental basalt and quartzite derived from underlying alluvial sediments and lava flows.

Travel from Menan Buttes to Lidy Hot Springs

0.0 Reset odometers at the bottom of the steep gravel road. Turn left (east) and proceed around the east side of North Menan Butte on improved gravel road.

1.1 Eroded monoliths of palagonite tuff have been interpreted as detached slide blocks from North Menan Butte (Creighton, 1982). However, we can find little evidence to support this interpretation and we instead consider them to be in situ erosional remnants. The monoliths contain similar lithologies and structures as the surrounding lower flank, distal facies deposits, and do not appear to have slid from substantially higher elevations near the crater rim. Although some of the tuffs in the monoliths have been deformed into low-amplitude folds, we have not identified any basal detachment surfaces.

1.8 Nearly horizontal, thin-bedded to laminated palagonite tuffs are typical of distal airfall deposits from North and South Menan Buttes.

3.0 Continue straight at intersection, toward sanitary landfill.

3.1 Road climbs the margin of Little Grassy Ridge Basalt lava flow (Kuntz, 1979).

4.1 Stop sign and intersection with paved Highway 33. Turn left (west) on Highway 33.

4.8 Low hills to the north are the Juniper Buttes, where a sequence of Quaternary and Pliocene basalts and rhyolites are present (Kuntz, 1979) (Fig. 21). Here, the 2.1 Ma Huckleberry Ridge Tuff overlies the 4.3 Ma tuff of Kilgore. The Juniper Buttes complex is the inferred resurgent dome of the Kilgore caldera (Fig. 3), source of the tuff of Kilgore. This is suggested by the 3.5 Ma rhyolite of Juniper Buttes, the presence of 4.3 Ma tuff of Kilgore at Juniper Buttes, and the location of the complex relative to the position of the Kilgore caldera as determined from flow direction and facies analyses of the tuff of Kilgore (Morgan, 1988a). Furthermore, the structural patterns that are present on Juniper Buttes are similar to those described at other resurgent domes such as Yellowstone (Christiansen, 1984), Timber Mountain dome (Smith and Bailey, 1968; Byers and others, 1976), and the Redondo dome in the Valles caldera (Smith and Bailey, 1968; Nielson and Hulen, 1984).

14.0 (Approximate mileage) Intersection with Interstate Highway 15. Turn right and proceed north on Interstate 15 toward Dubois.

39.0 (Approximate mileage) Exit Interstate 15 at Dubois and turn left (west) on State Highway 22.

54.0 (Approximate mileage) Road climbs a small hill. Road cut near Lidy Hot Springs is Stop 8.

Figure 20. Thin-bedded, weakly convolute palagonite tuff exposed in stream gully at Stop 7, southern flank of North Menan Butte. These deposits are typical of the Menan Buttes tuff cones.
Stop 8. Lidy Hot Springs

At the road cut along Highway 22 near Lidy Hot Springs, the 4.3 Ma tuff of Kilgore is emplaced against and on top of the 5.4 Ma rhyolite of Lidy Hot Springs (Morgan and others, 1984). Here the rhyolite of Lidy Hot Springs has well-developed spherulites and flow banding. In places, a thin matrix-supported unit containing unsorted blocks of locally derived tuff of Blue Creek and rhyolite of Lidy Hot Springs occurs between the rhyolite of Lidy Hot Springs and the overlying tuff of Kilgore. The tuff of Kilgore here is a proximal facies ignimbrite (Morgan, 1988a; Morgan and Doherty, unpublished data) containing coarse and relatively abundant phenocrysts, multiple ground layer deposits and lithic concentration zones near its base. The ignimbrite grades upward from a tan nonwelded base into a densely welded interior. A well-developed lithophysal zone, characteristic of the tuff of Kilgore, caps the hill on the east side of the road.

Travel and Description from Lidy Hot Springs to Howe Point

From Lidy Hot Springs, proceed west on State Highway 22. About 9 miles from Lidy Hot Springs several large, conical hills occur on the right (west) side of the road. The hills consist of rhyolitic lava flows including the 6.2 Ma rhyolite of Reno Gulch (McBroome, 1981), which may represent precaldera effusive volcanism along the margin of the Blue Creek caldera (Fig. 3), source of the 6.0 Ma tuff of Blue Creek. Alternatively, they may be postcaldera lavas associated with the 6.5 Ma Blacktail caldera.

Continue south on Highway 22. At the intersection with Highway 28, continue on Highway 22 for an additional 14 miles until reaching a stop sign at the intersection with Highway 33. Turn right (west) and pause to look at the southern Lemhi Range to the west. The nearly horizontal units forming the southern portion of the Lemhi Range are the major ignimbrites of the Heise volcanic field. The view direction is parallel to the topographic rim of the Blue Creek caldera. A major fault runs parallel to Highway 33, and separates a slumped block of tightly folded, intracaldera facies tuff of Blue Creek on the south, from outflow facies tuff of Blue Creek with mostly planar flowage features on the north side of the highway.

Continue west on State Highway 33 approximately 1 mile and park in the turnout on the south side of the road. This is Stop 9.

Stop 9. Howe Point

The southern tip of the Lemhi Range at Howe Point (Fig. 1) contains stratigraphic and structural features that are important in reconstructing the nature of silicic volcanism and associated caldera collapse of the Heise volcanic field. Stop 9 involves short walks to two localities on the north and south sides of Highway 33.

On the north side of Highway 33, a stratigraphic section of Heise volcanic units is exposed at Howe Point (Fig. 22). The 6.0 Ma tuff of Blue Creek (Richard F. Marvin, written communication, 1986; Morgan, 1988a, b) forms the top of the section. At Howe Point a number of features indicate that the tuff of Blue Creek is a proximal deposit. These features include: (1) the tuff of Blue Creek ignimbrite is relatively crystal-rich; it contains phenocrysts as large as 5 millimeters and lithic rock fragments up to 6 centimeters in size; (2) a relatively thick (1 m) welded, pumice-rich, plinian fall deposit directly underlies the tuff of Blue Creek; (3) zones of autobrecciation, marked by poorly sorted, angular clasts of the tuff of Blue Creek within a white, fine-grained vapor-phase matrix of the same tuff, occur along faults that are interpreted as caldera ring fractures.

Below the tuff of Blue Creek at Howe Point are ash deposits underlain by the 6.5 Ma tuff of Blacktail (Fig. 22). The tuff of Blacktail also has lithologic features that suggest deposition proximal to source: the ash-flow tuff is coarser in comparison with other exposures farther from the margin of the Snake River Plain.

On the west side of the Lemhi Range below the tuff of Blacktail (Fig. 22) is a light brown to orange soil that is underlain by a sequence of basaltic lavas in turn underlain by a crystal-rich, quartz- and biotite-bearing, nonwelded ignimbrite. This ignimbrite may be correlative with the 7.7 Ma (Kellogg and Marvin, 1988) tuff of Arbon Valley, a similar-appearing and distinctive ignimbrite exposed along the southern margin of the eastern Snake River Plain near American Falls (Trimble and Carr, 1976). At the base of the exposure at Howe Point, the 12.4 Ma tuff of Lost River Sinks (McBroome, 1981) is exposed.

Cross Highway 33 and climb the small hill on the south side of the road. The hill is a slumped block of tuff.
of Blue Creek containing tight flowage folds, strong flow lineation and numerous subparallel tension fractures lined with vapor-phase minerals. Deformation of the tuff of Blue Creek in this block was probably a result of its collapse near a caldera ring fracture shortly after emplacement and while the tuff was still plastic. This is supported by paleomagnetic remanence directions that do not vary across secondary flowage features, indicating that plastic deformation occurred while the welded tuff was still above its Curie temperature (McBroome, 1981; Morgan and others, 1984).

Travel to Sun Valley; End of Trip

Return to the vehicles and continue west on State Highway 33. At the town of Howe, turn left and continue west on Highway 33/22. Drive about 16 miles to the intersection of Highway 33/22 with U. S. Highway 20/26 and turn right (west) on Highway 20/26. At the town of Arco, turn left (southeast) on Highway 20/26 and follow the road signs to Craters of the Moon and Sun Valley.

Stephens (1988, this volume) presents a guide to Craters of the Moon National Monument, which is about 20 miles southwest of Arco on Highway 20/26. Link and others (1988, this volume) provide a road log from Arco to Sun Valley.

ACKNOWLEDGMENTS

Hackett thanks his present and former graduate students for their hard work and valuable insight into Snake River Plain volcanism, especially Michael J. Luessen and Gregory C. Ferdock. Some of this research was supported by grants to Hackett from the Idaho State University Faculty Research Committee and grants to graduate students from the Idaho State University Graduate Research Committee. Morgan thanks David J. Doherty, Hector Z. Albedo and Jan Morzel for discussions and assistance in the field. This paper stems partly from Morgan's doctoral studies at the University of Hawaii with George P. L. Walker. We are grateful to Paul Karl Link and Falma J. Moye for constructive reviews of a preliminary manuscript. However, we take responsibility for our data and interpretations.

REFERENCES


Armstrong, R. L., Leeman, W. P., and Malde, H. E., 1975, K-Ar dating, Quaternary and Neogene rocks of


King, J. S., 1982, Selected volcanic features of the south-central Snake River Plain, Idaho, in Bonnichsen, Bill, and Breckenridge, R. M., editors,


, 1988b, Paleomagnetic correlations of major ignimbrites on the eastern Snake River Plain: Geological Society of America, Abstracts with Programs, v. 20, no. 5.


Chapter Six
Geology of Southwest Montana

Granite blades of the late Cretaceous Boulder Batholith, southwest Montana. Photograph by D.W. Ilyndman.
Southwest Montana Thrust Belt: Bannack to Melrose

INTRODUCTION

On this field trip we will examine changes in structural style in a segment of the southwestern Montana fold and thrust belt. This segment coincides with the southern boundary of a Late Cretaceous volcanic-plutonic province (Fig. 1), which formed coevally with the fold-thrust belt (Scholten, 1968). Heat associated with the widespread igneous activity induced low pressure/high temperature metamorphism and reduced the plastic strength of the rocks in the growing thrust belt. When a structurally significant percentage of the rock sequence lost its plastic strength, the thrust system failed in a relatively short episode of plastic deformation. This episode followed early injection of sills and eruption of volcanics, but preceded final crystallization of large plutons.

This field trip focuses on the deformation along the McCarthy Mountain salient, between Bannack and Melrose, Montana (Fig. 2), where south-to-north thermal and strain gradients can be studied.

SETTING

The McCarthy Mountain salient is an east-projecting

bulge of the Rocky Mountain fold and thrust belt in southwestern Montana (Brumbaugh and Hendrix, 1981). It is defined by curving traces of Late Cretaceous thrusts, axial surfaces and cleavage planes. In the south, folds plunge northeastward; in the north, they plunge to the southeast. The salient occupies a structural depression on the western edge of Archean-cored Rocky Mountain

Figure 1. Index map for the area of field trip. The area of the box is enlarged in Figure 2. Large Late Cretaceous plutons are shown by the dashed pattern. “D” indicates the location of Dillon.
foreland uplifts in the Ruby and Highland Mountains. A well penetrated basement rocks in the central part of the salient at a depth of 11,835 feet. On the west, the Pioneer batholith and Grasshopper thrust plate form the high peaks of the Pioneer Mountains.

The Grasshopper plate contains a 7 to 8-km-thick section of Middle Proterozoic quartzites and argillites (Ruppel and others, 1981; Ruppel and Lopez, 1984). It is the dominant thrust sheet in this sector of the fold-thrust belt and was emplaced along the Kelley thrust (Fig. 2a) before growth of the McCarthy Mountain salient. In the northern part of the salient, northeast of the Pioneer batholith, the plate is folded into a broad synform, in which it is preserved as a klippe (Fig. 2a). The Grasshopper plate overrode Paleozoic and Mesozoic strata as a strong structural beam (Brandon, 1984) before the region was thermally weakened.

The folded and thrust rocks of the salient (Table 1) are Middle Proterozoic quartzites and argillites, Cambrian through Triassic platform sediments, Jurassic through Late Cretaceous foreland basin sediments and Late Cretaceous volcanics and intrusives. Tertiary intermontane-basin sediments and volcanics unconformably overlie the fold-thrust structures.

We will examine the structural styles of different lithologies along the length of the salient, principally contrasting the behavior of pelites, micrites, sparites and quartzites within the Paleozoic and Mesozoic section. We will also examine the important timing constraints provided by the Upper Cretaceous Beaverhead Group conglomerates and associated volcanics and by Late Cretaceous sills and plutons.
STRUCTURAL STYLES

With increasing temperature, different rock types successively weaken and undergo plastic deformation at low differential stress (Suppe, 1985). The specific lithology determines the relative temperature and stress necessary for the initiation of plastic deformation. From south to north along the McCarthy Mountain salient, pelite, micrite and sparite successively show plastic failure, with increasing strain to the north. Quartzite, however, never surpassed its plastic strength in the area of our study. Clay mineralogy and conodont color alteration indices (CAI) show that the paleotemperature of the rocks increased from south to north across this region, from 50°C near Dillon to 250-350°C near Melrose, at stratigraphic depths <3 km (Geiger, 1986; Sharkey, 1986; Sweet and others, 1981). The geothermal gradient near Dillon was average for continental crust, but near Melrose was more characteristic of the Abukuma facies series of volcanic arcs (see Miyashiro, 1961).

Figure 2a shows five areas we will visit to contrast structural styles. These areas are shown in greater detail in Figures 3 through 7. In the Bannack-Badger Pass area (Area A of Fig. 2a, shown in detail in Fig. 3), no rocks exhibit penetrative strain. Mississippian Lodgepole Formation micrites are brecciated and thrust over Late Cretaceous Beaverhead Group conglomerates along the Ermont and associated faults. The conglomerates contain abundant clasts of Mississippian limestone and rest unconformably on the Mississippian Madison Group (Lowell, 1965). Numerous local angular unconformities within the Beaverhead Group conglomerates and associated volcanics show that this region was near the surface during deformation. East of Badger Pass (Stop 2), the Mississippian Mission Canyon Limestone forms a rigid thrust plate above Beaverhead Group conglomerates and volcanic rocks along the Ermont thrust. Sparites and quartzites are extensively brecciated. Upper Mississippian mudstones flowed between disharmonically folded micrite beds, but did not cleave.

Along Rattlesnake Creek and Argenta Road (Area B of Fig. 2a, shown in detail in Fig. 4) in the southern Pioneer Mountains, the Mississippian limestones behaved brittlely despite their position immediately beneath the Kelley thrust, which carries the massive Grasshopper plate in this area. The Grasshopper plate itself contains exclusively brittle deformation fabrics along its leading edge.

South of Birch Creek (Area C of Fig. 2a, shown in detail in Fig. 5), the Ermont thrust passes laterally into tight folds with minor thrusts. Pennsylvanian Quadrant Quartzite is the structural key-bed controlling the major folds and is extensively brecciated. Some units show minor penetrative strain. A micrite bed in the Lower Cretaceous Kootenai Formation is locally cleaved, as are some middle Cretaceous Colorado Group mudstones. Where the cleavage exists, it is a spaced cleavage gener-

### Table 1. Map units for figures 3 through 7.

<table>
<thead>
<tr>
<th>Quaternary</th>
<th>Tertiary/Quaternary</th>
</tr>
</thead>
<tbody>
<tr>
<td>Qs</td>
<td>Quaternary sediments</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Tertiary</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tq</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Late Cretaceous/Early Tertiary</th>
</tr>
</thead>
<tbody>
<tr>
<td>KTa</td>
</tr>
<tr>
<td>Andesitic agglomerates, Bannack Badger Pass area</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Late Cretaceous/Early Tertiary</th>
</tr>
</thead>
<tbody>
<tr>
<td>KTt</td>
</tr>
<tr>
<td>Platy tuffs, Bannack-Badger Pass area</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Late Cretaceous/Early Tertiary</th>
</tr>
</thead>
<tbody>
<tr>
<td>KTB</td>
</tr>
<tr>
<td>Beaverhead Group conglomerate, Bannack Badger Pass area</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Late Cretaceous/Early Tertiary</th>
</tr>
</thead>
<tbody>
<tr>
<td>KTI</td>
</tr>
<tr>
<td>Fine-grained andesite intrusives, Bannack area</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Late Cretaceous/Early Tertiary</th>
</tr>
</thead>
<tbody>
<tr>
<td>KTo</td>
</tr>
<tr>
<td>Granitic/granodioritic intrusives</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Cretaceous</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kk</td>
</tr>
<tr>
<td>Cretaceous Kootenai Formation: shales, sandstones, limestones, local basal conglomerate. Mapped with Jurassic Morrison Formation shales.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Permian and Triassic</th>
</tr>
</thead>
<tbody>
<tr>
<td>PTb</td>
</tr>
<tr>
<td>Triassic Dinwoody Formation: shales and platy limestones; Permian Phosphoria Formation: shales, bedded cherts, quartzites, limestones</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Pennsylvanian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pq</td>
</tr>
<tr>
<td>Quadrant Quartzite</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Mississippian/Pennsylvanian</th>
</tr>
</thead>
<tbody>
<tr>
<td>MPa</td>
</tr>
<tr>
<td>Amsden Formation: shales, limestones</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Mississippian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mu</td>
</tr>
<tr>
<td>Undivided</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Mississippian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mn</td>
</tr>
<tr>
<td>Madison Group, Mission Canyon Limestone: massive biosparite</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Mississippian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ml</td>
</tr>
<tr>
<td>Madison Group, Lodgepole Formation: pelite, micrite</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Mississippian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mb</td>
</tr>
<tr>
<td>Big Snowy Group of Hildreth (1981)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Devonian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Du</td>
</tr>
<tr>
<td>Undivided: dolomite, shale</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Cambrian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cu</td>
</tr>
<tr>
<td>Undivided: dolomite, shale</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Middle Proterozoic</th>
</tr>
</thead>
<tbody>
<tr>
<td>pC8</td>
</tr>
<tr>
<td>Belt Supergroup: quartzite, argillite</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Archean</th>
</tr>
</thead>
<tbody>
<tr>
<td>pCa</td>
</tr>
<tr>
<td>Undivided: gneiss, schist, marble, amphibolite</td>
</tr>
</tbody>
</table>
ally at a high angle to bedding. Tear faults are very common in sandstones and limestones along steep fold limbs.

Along the Burma Road (Area D of Fig. 2a, shown in detail in Fig. 6), Quadrant Quartzite and Permian Phosphoria Formation chert beds form rigid thrust sheets with steep ramps, while shaley Mesozoic beds are tightly folded. Sandstone beds are prominently jointed throughout the Mesozoic section, but mudstones and micrite of the Kootenai Formation are distinctly cleaved, with cleavage normal to bedding. The cleavage is folded with the bedding and is cut by thrust faults. Biosparite beds of the upper part of the Kootenai Formation are thrust and buckled dis harmonically above black shale layers.

Along Trapper Creek, west of Melrose (Area E of Fig. 2a, shown in detail in Fig. 7), Kootenai Formation pelites have slatey and phyllitic axial planar cleavage with stretched mineralogic spots; micrites have continuous axial planar cleavage; biosparite beds have spaced axial planar cleavage with strained gastropods; quartzite beds have quartz-veined tension gashes; and conglomerates exhibit pressure-solution pits. Quartzites and cherts remain brittle and supported thrusts. Along Camp Creek, east of Melrose, Lodgepole Formation micrites and pelites have continuous slatey cleavage oblique to bedding. Sparry dolomites have brittle fabrics and minor thrust faults.

TIMING

The oldest part of the Pioneer batholith, a hornblende gabbro body with a minimum 40Ar/39Ar age of 79.9 Ma, cuts the Grasshopper plate (Snee, cited in Pearson and Zen, 1985). Structures of the McCarthy Mountain salient fold the Grasshopper plate and are therefore younger than the plate. From Bannack to Rattlesnake Creek, the Ermont thrust and associated structures were active during and after deposition of Beaverhead Group conglomerates and volcanics, but before intrusion of several small satellite stocks of the Pioneer batholith (Johnson, 1986; Thomas, 1981). Volcanics of the southeastern Pioneer Mountains have K-Ar minimum ages of 71 to 67 Ma (Snee and Sutter, 1979). The Beaverhead Group appears to be of Campanian to Maastrichtian age, based on palynology studies southeast of Bannack (Nichols and others, 1985).

The Ermont thrust passes laterally into folds of the McCarthy Mountain salient which are cut by the Uphill Creek granodiorite of the Pioneer batholith, with an 40Ar/39Ar age of 74.9 Ma (Snee, 1982, in Pearson and Zen, 1985). Near the Burma Road, an andesite sill in the Colorado Group pre dates folding, but the McCarthy Mountain stock cuts the folds. Brumbaugh and Hendrix (1981) bracketed the formation of the salient in this area between 75 and 77 Ma.

TECTONOTHERMAL MODEL

The changes in structural style show that rocks within the McCarthy Mountain salient were increasingly weak from south to north during the deformation episode. The CA1 data show that the rocks along Trapper Creek were as much as 350°C at their peak temperature (Sweet and others, 1981). At that temperature, micrite and sparite have virtually no plastic strength, and quartzite has a very low plastic strength of a few 10's of megapascals (Suppe, 1985). Micrite and sparite did indeed deform plastically near Trapper Creek. Since the Quadrant Quartzite failed brittlely however, either the peak temperature was reached after all deformation ceased or the brittle strength of the Quadrant Quartzite was very low, and the Quadrant Quartzite failed brittlely under a very low differential stress. Since the Quadrant Quartzite is the strongest structural unit in the Paleozoic-Mesozoic stratigraphic sequence, the thrust system could not support any load in excess of what the Quadrant Quartzite could bear.

In the McCarthy Mountain salient, the rocks failed during a period of intense igneous activity which followed the emplacement of the Grasshopper thrust plate. The Grasshopper plate formed a mountainous upland that required the structural support of rigid footwall rocks. Increasing temperature reduced the plastic strength of various rock types, while water, driven off by high temperature/low pressure metamorphism, may have increased the pore water pressures of more competent units and decreased their brittle strength (see Suppe, 1985). The footwall rocks may have become too weak to bear the burden of the Grasshopper plate, and consequently "collapsed" toward the foreland in a short-lived deformational episode. The rising magma then cross-cut the structures, resulting in the patterns we see today.

ROAD LOG FOR MONTANA THRUST BELT: BANNACK TO MELROSE

Day 1. Dillon-Bannack-Rattlesnake Creek-Birch Creek

Mileage

0.0 Road log starts on Atlantic Street in Dillon, Montana, at Western Montana College of the University of Montana. Proceed south on Atlantic Street.

0.7 Turn left onto Interstate 15 South.

3.2 Take Exit 59 toward Jackson and Wisdom.

3.5 Turn right on Montana Highway 278 toward Jackson. Tertiary volcanic rocks form the low hills
at 2:00. The highway traverses the Quaternary outwash plain of Rattlesnake Creek for the next 5 miles.

7.4 The Pioneer Mountains lie ahead. The high peaks are composed of Late Cretaceous batholiths and Middle Proterozoic rocks of the Grasshopper thrust plate (Ruppel and Lopez, 1984).

10.3 The road on the right leads to Argenta. Continue west on Highway 278. The prominent ridge at 3:00 is Dutchman Mountain, an anticline cored by Quadrant Quartzite. Folded Mesozoic rocks underlie the lower ridges to the east.

11.0 The road on the left is the old stage road to Bannack, dating from 1860s. It climbs up to a pass below the tree-covered ridge of Quadrant Quartzite known as Robber's Roost. Continue west on Highway 278.

12.3 The road now follows a pediment surface of probable Pliocene-Pleistocene age which rises toward Badger Pass.

13.1 The sharp knobs to the left at 9:00 are in a Late Cretaceous (?) stock intruded into Late Cretaceous volcanic agglomerates, which occupy a major syncline in the low valley. The volcanics and the intrusive are overlain by poorly consolidated Tertiary gravels forming the rounded grassy hills.

15.4 The quarry on the left is in Late Cretaceous bedded tuffs. These may have been laid down subaqueously (Thomas, 1981; Johnson, 1986).

15.6 The highway crosses the Ermont thrust, which here places Mississippian rocks over Late Cretaceous volcanics and Beaverhead Group conglomerates.

15.8 The roadcut on the right is in Pennsylvanian Quadrant Quartzite in the hangingwall of the Ermont thrust.

16.2 The highway crosses float of Permian Phosphoria Formation in the axis of a syncline. Triassic Dinwoody Formation is present in the valley to the north (Lowell, 1965).

16.7 The highway crosses an anticline of Quadrant Quartzite. Both this fold and the syncline to the east are cut by Cretaceous intrusions 2 miles south of here (Thomas, 1981).

17.0 Badger Pass.

17.2 Grasshopper Prairie lies ahead. This Tertiary basin is superimposed on Late Cretaceous structures. Here, Tertiary sediments overlap the French Creek thrust. This fault places the Madison Group over the Dinwoody and Phosphoria Formations. It is cut by a Cretaceous intrusion 2 miles south of here. The Kelley thrust places Proterozoic rocks of the Grasshopper plate over the Madison Group along Taylor Creek, 1.5 miles northwest of here (Thomas, 1981).

18.3 The roadcut on right provides a rare exposure of the Amsden Formation beneath the Quadrant Quartzite.

19.2 Tertiary sediments of Bozeman Group form low outcrops.

20.7 Turn left on road to Bannack State Park.

21.2 The tree-covered hills to the right are underlain by Middle Proterozoic rocks of the Grasshopper thrust plate.

21.6 A prominent Pliocene-Pleistocene pediment lies on the right.

23.2 A cemetery is on the left.

23.5 Drive past ranch buildings.

23.7 The road to Grant is on the right; proceed to Bannack on the left fork.

24.4 Stop 1: Bannack State Park and Grasshopper Creek traverse (Fig. 3). This stop will establish critical timing relationships for the Ermont thrust and associated structures. It provides views of brittlely deformed rocks near the toes of thrust sheets.

Park in the lot on the left at the west edge of town and proceed on foot through town. We will traverse about 2 miles down the canyon of Grasshopper Creek and return. Gold was discovered here in 1862, and Bannack became Montana's first major gold camp and territorial capital. Vigilantes hanged the outlaw sheriff Henry Plummer here in 1864.

We will proceed through Bannack and downstream along Grasshopper Creek.

0.0 Log begins at the old hotel.

0.2 Hike down the dirt road on the northeast side of Grasshopper Creek. Beyond the edge of town, the canyon narrows in highly brecciated and silicified Madison Group sparry limestones. A thrust fault within the Madison Group crosses Grasshopper Creek here (Thomas, 1981; Johnson, 1986). It may link up with the French Creek thrust crossed by Highway 278 near Badger Pass. This fault is cut
Figure 3. Area A, Bannack to Badger Pass, stops 1 and 2 are circled. See Figure 2a for location. Rock unit symbols explained in Table 1. Geologic symbols given in inset box on lower right are also used in Figures 4-7. After Thomas, 1981; Johnson, 1986.
by a Late Cretaceous granodiorite stock and its contact aureole.

0.7 The rounded hillslope on the southwest side of the canyon is underlain by Late Cretaceous granodiorite. A number of mines are located on the altered host rock near the contact.

1.0 Remain on the old road along the northeast side of the valley bottom. The canyon crosses the Ermont thrust, which places Madison Group over Beaverhead Group conglomerate. The conglomerate contains abundant clasts of Madison Group limestone, and rests unconformably on the Madison Group. The gently tilted Ermont thrust cuts up section to the east across steeply tilted Beaverhead Group beds in the footwall, for 1.5 km along the north side of the canyon. The Beaverhead Group is tilted along the west limb of major east-verging syncline which was overridden by the Ermont thrust. Farther east, the fault climbs into tuffs and agglomerates which stratigraphically overlie the conglomerates. In the hanging wall, the Ermont fault cuts down through the Madison Group section toward the east and thus appears to truncate an older structure. Johnson (1986) concluded that the older structure was related to the Archean-basement-cored Armstead anticline, part of the Laramide Rocky Mountain foreland which lies just to the south of Grasshopper Creek.

1.3 The canyon crosses back into the hangingwall of the Ermont thrust, and narrows in Madison Group limestones.

1.9 Here Madison Group limestone is in contact with steeply dipping Beaverhead Group conglomerate. This contact may be a high-angle fault which has dropped the hangingwall of the Ermont thrust (on the west) against the footwall of the Ermont thrust (on the east). Alternatively, it may be a steeply tilted angular unconformity (Johnson, 1986), or a thrust with the west side up (Thomas, 1981). Beaverhead Group conglomerate is well exposed here. Johnson (1986) correlated it with the Campanian Lima Conglomerate of Nichols and others (1985) on the basis of similar clast composition and stratigraphic position in the lower Beaverhead Group. Thomas (1981) and Johnson (1986) correlated the overlying volcanic rocks with Maastrichtian age (67-71 Ma) volcanics of the southeastern Pioneer Mountains dated by K-Ar (Snee and Sutter, 1979). The Ermont thrust is younger than the conglomerates and volcanics but is intruded by granodiorite and andesite stocks 2.5 miles north of this locality. The stocks are compositionally similar to, but crosscut, the andesite agglomerates.

Return to Bannack along the route we followed down the canyon.

Return to Highway 278. Resume road mileage log.

28.1 Turn right on Highway 2/78, toward Dillon.

31.8 Badger Pass.

33.2 **Stop 2. Ermont thrust (Fig. 3).** To the north, the tree-covered ridge is underlain by Quadrant Quartzite in the hangingwall of the Ermont thrust. It dips gently to the west and overlies the slope-forming Amsden Formation. The light grey outcrops along the lower parts of the slope are Mississippian limestones which Hildreth (1981) correlated with the Big Snowy Group. The wide grey terrace to the north is underlain by massive upper Madison Group sparry limestone, in a broad anticline with very gentle dips. Down slope from the Madison Group limestone, the prospect pits are in vertical conglomerate beds of the Beaverhead Group. These beds lie along the west limb of the major syncline overridden by the Ermont thrust. The Ermont thrust follows the break in slope, placing Madison Group over Beaverhead Group. The overlap along the flat-lying fault is at least 3 km, based on the erosional trace of the thrust fault.

As the fault is traced toward the highway, it cuts up section in both hangingwall and footwall. Along the highway, the thrust places the upper Mississippian limestones against white and red bedded tuffs which overlie Beaverhead Group conglomerates. Further south, the fault continues to climb in the footwall, into thick volcanic agglomerates which overlie the tuffs. However, it proceeds down-section in the hangingwall into the Madison Group. South of the highway at the New Departure mine, a klippe of the thrust plate is cut by a stock. The New Departure mine is one of a series of silver mines localized along zones where the Ermont thrust carries limestone and is cut by stocks.

Continue east on Highway 278.

38.5 Turn left on the gravel road to Argenta, following the glacial outwash plain of Rattlesnake Creek.

41.5 The road passes through Argenta. This silver district is associated with the Argenta stock, which intrudes Madison Group limestone and Quadrant Quartzite. Enter canyon of Rattlesnake Creek. The Argenta stock crops out as rounded boulders. The stock intrudes northeast-trending folds which appear to represent the northeastward termination of the Ermont thrust. Less than 2 miles southwest of here, the Ermont thrust places the Madison Group over Late Cretaceous volcanics along
312 Guidebook to the Geology of Central and Southern Idaho

Figure 4. Area B, Argenta Road, stop 3. See Figure 2a for location. Symbols explained in Table 1. After Brandon, 1984.

Ermont Gulch.
Continue along dirt road up Rattlesnake Creek. For next 2 miles, the road traverses down through the Paleozoic section, on the east limb of the Argenta anticline.

43.5 The road crosses the Argenta thrust, which places Proterozoic rocks over the Cambrian Flathead sandstone (Brandon, 1984). The Argenta thrust continues north from here for 7 miles, where it is cut by the Pioneer batholith along Birch Creek. To the south, the Argenta thrust dies out into the Argenta anticline, which is carried on the Ermont thrust. The Ermont thrust appears to transfer its displacement to the Argenta thrust in this area.

45.1 The road passes through a narrow canyon in Proterozoic quartzite on the west limb of the Argenta anticline.

45.2 Stop 3. Kelley thrust/Grasshopper thrust plate (Fig. 4). Park at the fork in the road. A down-to-the-west normal fault complicates the structure here on the northeast side of the canyon, and places Mississippian and Devonian rocks against Proterozoic along the west limb of the Argenta anticline. We will traverse approximately 1 mile north-northwest up Rattlesnake Creek. At the location of the splendid Quaternary glacial moraine, the French Creek thrust places the Madison Group over the Amsden Formation. This fault extends for 8 miles to the south, where it is cut by a granodiorite stock (Thomas, 1981). It may be the fault zone we crossed at Bannack. Further up Rattlesnake Creek, the canyon narrows against a large bluff of Quadrant Quartzite which occupies the tightly overturned Red Butte syncline (Red Butte is the prominent knob west of Rattlesnake Creek). The Red Butte syncline is a footwall syncline beneath the Kelley thrust, which carries the Grasshopper plate of thick Proterozoic rocks. The high ridge 2 miles due west of here is underlain by Proterozoic rocks of the Grasshopper plate; the Kelley thrust follows the break in slope between Madison Group limestones below and Proterozoic Belt Supergroup quartzites above. Further north, the thrust puts Middle Proterozoic Belt rocks over the Kootenai Formation (Brandon, 1984). The Kelley thrust extends 8 miles north of here, where it is cut by the Pioneer batholith. The
Grasshopper plate is cut by a 79.9 Ma hornblende gabbro pluton 5 miles to the west (Pearson and Zen, 1985). All lithologies along the Rattlesnake Creek transect from Argenta to the Kelley thrust display exclusively brittle deformation fabrics.

Return to Interstate 15 North along State Highway 278. Restart mileage at junction of Highway 278 and Interstate 15.

Mileage

0.0  Turn left unto the Interstate 15 northbound entrance ramp.

2.0  The hills to the left with the letters “B” and “M” are underlain by Tertiary volcanics and gravels.

2.5  Cross the Beaverhead River.

2.8  A Madison Group limestone erosional inlier lies to the right.

3.6  The hills to left are made up of Tertiary gravel.

4.4  Overpass. The highway traverses a Pliocene/ Pleistocene pediment surface for the next 8 miles.

7.7  Overpass. The tree-covered hills to left are underlain by Quadrant Quartzite in the Dutchman Mountain anticline.

11.2  McCarthy Mountain is to the right at 2:00. It is cored by a Late Cretaceous stock. The Tobacco Root Mountains in the distance at 3:00 are an Archenoroved Laramide Rocky Mountain foreland uplift.

12.9  Take Apex exit.

13.2  Take road to Birch Creek, Western Montana College recreation camp.

16.2  Cross Birch Creek and park.

Stop 4. Birch Creek Folds (Fig. 5). We are parked near the axis of the Cave Gulch syncline, in the middle Cretaceous Colorado Group. This syncline lies between the Dutchman Mountain anticline to the east and the Birch Creek anticline to the west. These folds are asymmetric and overturned to the east, and they form the northern termination of the Ermont thrust. The Dutchman Mountain anticline is cut by small stocks; the Birch Creek anticline is cut by the 74.9 Ma Uphill Creek granodiorite pluton of the Pioneer batholith.

We will traverse the Colorado Group and Kootenai Formation to observe the deformation styles of different rock types. This region contains the southernmost penetrative cleavage in the McCarthy Mountain salient, in beds of the Colorado Group in the Cave Gulch syncline, and locally in the Kootenai Formation middle limestone micrite bed and calcareous mudstone.

Continue up Birch Creek after traverse.

17.5  The canyon of Birch Creek narrows as it crosses the Quadrant Quartzite.

17.6  Madison Group limestone forms the canyon walls.

19.1  The road crosses the contact zone of Pioneer batholith. This is the southernmost reach of the batholith along its eastern side. North of here the batholith follows a zone near the base of the Quadrant Quartzite for ten miles in a very concordant fashion (Brandon, 1984).

21.2  Aspen Campground. Turn around and return to Dillon.

Day 2. Dillon-Burma Road-Trapper Creek-Camp Creek

Mileage

0.0  Start road log mileage at Apex Exit, Interstate 15 North. Stay on I-15 North.

2.1  The hill on the right is the Apex anticline with Quadrant Quartzite at the surface. This structure was drilled in 1980 by American Quasar in a dry well located about 3 miles to the south. The well penetrated Precambrian basement at a depth of 11,835 feet.

4.8  Road cuts are in the Colorado Group, which occupies the structural depression in the center of the McCarthy Mountain salient.

6.7  Tertiary Renova Formation basin fill crops out to the left.

9.0  The McCarthy Mountain stock intrudes Colorado Group on the north end of the mountain to right. Note two separate pediment surfaces on the flank of the mountain.

10.2  Take Glen Exit 85.

10.5  Turn right toward Glen.

10.6  Cross tracks, stop, turn right toward Glen.

13.2  Glen Post Office is on the left.
Figure 5. Area C, Birch Creek Road, stop 4. See Figure 2a for location. Symbols explained in Table 1. After Brandon, 1984.
14.1 Turn left on Glen fishing access road (Burma Road).

14.9 Cross first branch of the Big Hole River.

15.0 Cross second branch of the Big Hole River.

18.9 In the gulch on the left, Six Mile Creek Formation gravels overlie Renova Formation mudstones with angular unconformity.

19.3 Colorado Group sandstone dips to west. The sandstones have conspicuous bc-joints throughout this area.

19.4 Stop 5. Andesite sill in Colorado Group (Fig. 6). This sill has bc-joints similar to the overlying sandstones, showing that it was intruded and crystalline before the rocks here were folded.

19.5 Upper Kootenai Formation limestone forms the ridge.

19.6 Stop 6. Cleaved calcareous mudstone of upper Kootenai Formation clastic unit. Nodular red mudstones exhibit spaced cleavage normal to bedding throughout this area. The cleavage wraps around the diagenetic calcareous nodules. Overlying biosparite beds in the upper Kootenai Formation limestone contain buckle folds and thrust faults and are uncleaved.

19.8 On the left are the Creasy Gulch anticline and Ziegler syncline. Upper Kootenai Formation limestone is tightly folded in the syncline on the high ridge at 10:00.

20.0 Stop 7. Thrust faults in middle Kootenai Formation limestone. Two west-dipping thrust faults cut the beds in this spectacular outcrop. Also prominently displayed is the typical cleavage of the micrite marker bed in the middle Kootenai Formation limestone. Throughout the area, this unit has a spaced, somewhat anastomosing pressure-solution cleavage essentially perpendicular to bedding. Where the bedding is vertical on the steep limbs of folds, the cleavage is horizontal. Thus, the cleavage represents a bedding-parallel shortening that affected the rocks before buckling. The buckling was related to the cleavage, however, because the intersection of bedding and cleavage parallels the macroscopic fold hinges throughout the area. The cleavage is clearly cut by small-scale thrusts in this outcrop. In this outcrop, only the micrite bed has any cleavage. Other units remained rigid.

20.2 Ridge of sandstone in lower Kootenai Formation clastic unit lies on the left.

20.3 Basal Kootenai Formation conglomerate, Jurassic Morrison Formation and upper part of Triassic Dinwoody Formation crop out on left.

20.5 Stop 8. Deformation styles in Dinwoody and Phosphoria Formations. Dinwoody Formation platy limestones are disharmonically folded above basal Dinwoody Formation shales, but no cleavage is present. The Dinwoody Formation is also broken by low-angle thrusts, visible to the north on the high ridge.

Phosphoria Formation cherts and quartzites form rigid thrust plates and tight kink-style folds. The Phosphoria Formation is thrust over the basal Kootenai Formation clastic unit along a spectacular footwall ramp along the Sandy Hollow fault.

20.9 Basal Kootenai Formation conglomerate, Morrison Formation shale and Dinwoody Formation limestone crop out on the left.

21.3 The Phosphoria Formation is in contact with the Quadrant Quartzite on the left. The Quadrant forms the narrow canyon of the Big Hole River in the core of the Hogback anticline. The Quadrant formed a brittle structural beam. To the south, the anticline is thrust over the Colorado Group along the Hogback thrust.

22.2 Sandy Hollow road is on the left. Turn left, drive up Sandy Hollow 1.2 miles, and park.

Stop 9. Sandy Hollow duplex and thrust ramp. This spectacular area has confused hundreds of field geology students and has been the focus of papers by Hendrix and Porter (1980), Brumbaugh (1973), Brumbaugh and Dresser (1976), and Brumbaugh and Hendrix (1981). The cliff of upper Kootenai Formation limestone to the northeast has many tight folds and thrusts which are disharmonic with respect to the underlying and overlying beds. These folds have been stripped off the basal bed of the limestone unit along a black shale layer, at the locus of a thrust ramp in the hangingwall and footwall of the major Sandy Hollow thrust.

Return to Interstate 15; continue mileage log.

34.2 Turn north on Interstate 15.

36.2 To left, the cliff exposes the Angler's thrust, which places upper Kootenai Formation limestone over Colorado Group (Brumbaugh, 1973). The limestone contains pyrite crystals here and farther north near Trapper Creek, but nowhere to the south. These may be the product of increasing metamorphism to the north.

39.5 Beal's Mountain, to left, is underlain by a klippe
Figure 6. Area D, Burma Road, stops 5 through 9. See Figure 2a for location. Symbols explained in Table 1. After Brandon, 1984.
of Proterozoic rocks of the Grasshopper plate resting on the Colorado Group in a large, southeast-plunging syncline. The structures in the Mesozoic rocks here thus postdate the emplacement of the Grasshopper plate.

41.3 Take Melrose exit.

41.5 Turn left toward Melrose. Cross tracks and proceed straight on road to Trapper Creek.

44.0 Fishing access road is on the right. Park here.

**Stop 10. Metamorphic effects in the Kootenai Formation (Fig. 7).** This stop contrasts vividly with the stops along the Burma Road in the same stratigraphic units. Here, in very tight southeast-plunging folds, Kootenai Formation mudstones are slates and phyllitic slates with stretched mineralogic spots on cleavage surfaces; micritie has continuous slaty cleavage in the axial plane of tight folds; quartzites and conglomerates are quartz veined; Dinwoody Formation shales have pencil cleavage, and limestones have cleavage. However, Quadrant Quartzite and Phosphoria Formation chert still acted brittlely.

Drive up fishing access road 1.3 miles and park near junction of road to Trapper Spring.

**Stop 11. Limestone of upper Kootenai Formation.** Walk 0.3 mile north up poor dirt road through the Colorado Group on the nose of the Canyon Creek anticline. At the ridge of upper Kootenai Formation limestone, turn right off the road and climb up to the ridge of limestone. Here, the upper Kootenai Formation biosparite is penetratively cleaved and contains deformed gastropods. The cleavage lies in the axial plane of the Canyon Creek anticline.

The folds in the Trapper Creek area are younger than the emplacement of the Grasshopper plate, because it is folded into a syncline in the Colorado Group. The Canyon Creek anticline is cross cut by a Late Cretaceous (76 Ma) pluton along the Big Hole River 10 miles northwest of here, which overprints the cleavage (Geiger, 1986). Thus, the folds at Trapper Creek fall within the same age bracket as the folds and thrusts further south.

Return to Melrose.

50.1 Cross under Interstate 15 and proceed northeast up the dirt road in Camp Creek.
52.5 Stop 12. Cleaved Lodgepole Formation limestone. The Lodgepole Formation limestone crops out continuously for 6 miles southeast of here along the flank of the Highland Mountains. At the south end of the outcrop belt, the Lodgepole Formation has only a local, nonpenetrative cleavage. The cleavage consistently increases in intensity northwestward to this point, where it dominates the outcrop morphology and is clearly continuous in thin section. Fine clay seams and flattened calcite grains define the fabric. Crinoid ossicles are only slightly deformed, commonly with pressure shadows. In the same interval in which the cleavage increases in strength, the conodont alteration index rises from 4.5 to 5.5, equivalent to a temperature increase from 250 to 350°C (Geiger, 1986).

CONCLUSION

The changes in structural style from south to north along the McCarthy Mountain salient correspond closely to increases in paleotemperature from about 50°C to 250-350°C. The increasing temperature led to the progressive softening and plastic failure of pelite, micrite and sparite. In the south, the older, Archean-cored Armstead anticline was broken and faulted by the deformation pulse. Near Melrose, the Paleozoic and Mesozoic rocks were plastically deformed against the Archean-cored Highland Mountains (Geiger, 1986).

ACKNOWLEDGMENTS

This paper is based on theses by W. C. Brandon, B.C. Geiger, L. M. Johnson, J. Sharkey and G. M. Thomas at the University of Montana, and on observations made during University of Montana field geology courses over a number of years. This manuscript was reviewed by W.S. Snyder and P. K. Link.

REFERENCES


