Regional Geology and Mineral Deposits In and Near the Central Part of the Lemhi Range, Lemhi County, Idaho

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1480

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By EDWARD T. RUPPEL *and* DAVID A. LOPEZ

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A descriptive summary of the rocks and structure in part of east-central Idaho

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REGIONAL GEOLOGY AND MINERAL DEPOSITS IN AND NEAR THE CENTRAL PART OF THE LEMHI RANGE, LEMHI COUNTY, IDAHO

By EDWARD T. RUPPEL and DAVID A. LOPEZ

ABSTRACT

The Lemhi Range, in east-central Idaho, is one of the longest mountain ranges in the state. Its central part, a region of about 1,700 square kilometers, is an ice-sculptured mass of precipitous peaks that expose the underlying geology in abundant detail.

The oldest rocks in this region are included in the Middle Proterozoic Yellowjacket Formation, a thick sequence dominantly of feldspathic quartzite and siltite that is more widely exposed to the north and west. The base of the Yellowjacket Formation is not exposed; its top is the sole zone of the Medicine Lodge thrust system everywhere in east-central Idaho. The Yellowjacket Formation is the only autochthonous sequence of Proterozoic or Paleozoic sedimentary rocks in this region; all younger Proterozoic and Paleozoic sedimentary rocks overlie the Medicine Lodge thrust and are allochthonous, having been transported into east-central Idaho by thrust faulting in Late Cretaceous time from their original site of deposition in central or western Idaho.

The Medicine Lodge thrust sheet consists mainly of Middle Proterozoic feldspathic quartzite in the central Lemhi Range, but also includes younger rocks, mostly of early and middle Paleozoic age. The Middle Proterozoic rocks, more than 10,000 meters thick, are included in the Lemhi Group and Swauger Formation, which are more completely and widely exposed here than elsewhere in east-central Idaho. These rocks are almost entirely fine- to medium-grained quartzite, and all except the Swauger Formation are feldspathic. They were deformed and partly eroded in late Middle Proterozoic and Late Proterozoic time when the Lemhi arch was uplifted and are unconformably overlain by the lowest Paleozoic rocks of the Early Ordovician Summerhouse Formation or, more commonly, by the Middle Ordovician Kinnikinic Quartzite. Above the Kinnikinic, succeeding Paleozoic rocks are dominantly marine carbonate rocks, principally dolomite, of Ordovician, Silurian, and Devonian ages and limestone of Mississippian age. Younger Paleozoic and Mesozoic rocks are not preserved in the central Lemhi Range, but are present elsewhere in the range and in the adjacent Beaverhead Mountains to the east.

Strata within the Lemhi Range were deformed and crumpled in Late Cretaceous time when the Medicine Lodge thrust moved eastward, and were intruded in Eocene time by small stocks and sheets mostly of monzogranite, granodiorite, quartz monzodiorite, and quartz monzonite. The stocks were rapidly unroofed by erosion, and in later Eocene time were partly covered by the basaltic and andesitic flows, breccias, and tuffs of the Challis Volcanics. Later in the Tertiary, probably starting in early Oligocene time, thick tuffs, tuffaceous sands and gravels, and some freshwater limestone accumulated in the newly forming Lemhi and Pahsimeroi Valleys. In Miocene and Pliocene time, continuing to the present, major block uplifts formed the present northwest-trending mountain ranges.

Regional arching, parallel to the Snake River Plain and forming the Gilmore Summit and Donkey Hills divide, began in the late Tertiary, probably in late Pliocene time, simultaneously with development of the Snake River Plain and renewed faulting. As a result of this arching, the drainages of the Lemhi and Pahsimeroi Rivers had been reversed by early Pleistocene time, when the earliest glaciers accumulated in the Lemhi Range. Later, major episodes of glaciation carved the Lemhi Range into its present jagged form and left extensive deposits of till in mountain valleys and in terminal moraines at the valley mouths and deposits of glacial outwash gravels on the main valley floors. Late Pleistocene cirque glaciers have since resharpened the peaks and left small fresh moraines high in some valleys. Glacially steepened valley walls provide an ideal setting for snow avalanches; rock debris carried down with the avalanches chokes many of the glacial valleys.

Strike-slip and other faults along the range fronts are still active. Marked by abundant young fault scarps, they cut glacial deposits and the most recent alluvial fan deposits and disrupt drainages. Movement on these faults appears to be a major cause of landsliding, and young landslides, probably earthquake induced, are among the most conspicuous and widespread surficial deposits in and near the central Lemhi Range.

Mineral deposits in and near the central Lemhi Range include base- and precious-metal veins and replacement deposits in carbonate rocks, tungsten-copper-silver quartz veins in quartzite and siltite, and disseminated deposits of copper and molybdenum in granitic stocks. Smaller deposits of secondary copper minerals, of lead disseminated in granite, of copper-bearing magnetite in skarn, and small gash veins of hematite and quartz also occur. The most valuable deposits have been the rich base- and precious-metal veins and replacement deposits in the Pittsburgh-Idaho and Latest Out mines at Gilmore and the tungsten-copper-silver quartz veins in the Ima mine at Patterson. Most of the mineral deposits are closely associated with Eocene granitic intrusive rocks, which seem to have been the source of mineralizing solutions. The principal granitic stocks, and all the known major metallic deposits in east-central Idaho, are on the monoclinal limbs of block uplifts, above inferred steep reverse basement faults along which granitic magma was intruded to the base of the Medicine Lodge thrust plate where it spread laterally as sheets in imbricate thrust faults. The principal mineral deposits are clustered around the necklike stocks which fed the sheets, and only small deposits, few of which have yielded any ore, are related to the sheets that spread outward from the stocks. All the known mineral deposits are in the lower part of the Medicine

Lodge thrust plate. The primary structural controls so evident in the localization of granitic stocks and associated hydrothermal mineral deposits in the central Lemhi Range suggest that new mineral deposits should be sought, therefore, in the lower part of the Medicine Lodge thrust plate above steep faults along the edges of the linear mountain ranges of this region. Substantial orebodies have not been found and are not likely to have been deposited in the structurally flat, central parts of the block uplifts, where deep structural controls are not present.

INTRODUCTION

The Lemhi Range, in east-central Idaho, is one of the longest and loftiest mountain ranges in the State. Its central part, embraced by the Leadore, Patterson, and Gilmore quadrangles, lies between lat 44°15′ N. and lat 44°45′ N. and between long 113°30′ W. and long 114°00′ W. (fig. 1), and includes some of the highest and least accessible parts of the range. Leadore, the only community near the central part of the range, is on Idaho Highway 28, which links Idaho Falls, Idaho, and the Snake River Plain with Salmon, Idaho, and continues northward to Missoula, Mont. (fig. 1). Idaho Highway 29, only partly surfaced, extends eastward from Leadore through Bannock Pass into southwest Montana, joining U.S. Interstate Highway 19 south of Dillon, Mont. Idaho 29 closely follows the old track bed of the Gilmore and Pittsburgh Railroad, which served the area from 1910 until 1939, when dwindling shipments of ore and other freight from Gilmore and the Lemhi Valley led to its end. The railroad bed, in turn, partly followed the route of early miners and settlers along an ancient Indian trail through Bannock Pass and down Railroad Canyon to the pioneer community of Junction—a settlement, bypassed by the Gilmore and Pittsburgh Railroad, that was moved to form the nucleus of Leadore. The name of this early community is given in mining reports before 1910, and is preserved in the name of the nearby Junction mining district, but the town disappeared in 1910 with the establishment of Leadore. Leadore in turn, was named for the principal source of revenue and principal justification of the Gilmore and Pittsburgh Railroad—the lead and silver ores from the Gilmore mining district. The former community of Gilmore was founded about 1903 at the time of the first significant ore production from the newly opened mines. It was a major mining center in Lemhi County until about 1925 when ore production started to decline. When the railroad discontinued regular train service in 1935, the town was largely abandoned although the last original resident, G. Grover Tucker, agent for the Gilmore Mercantile Company, did not move away until 1965 following a robbery. Today, Gilmore is a ghost town, disappearing board by board to souvenir collectors.

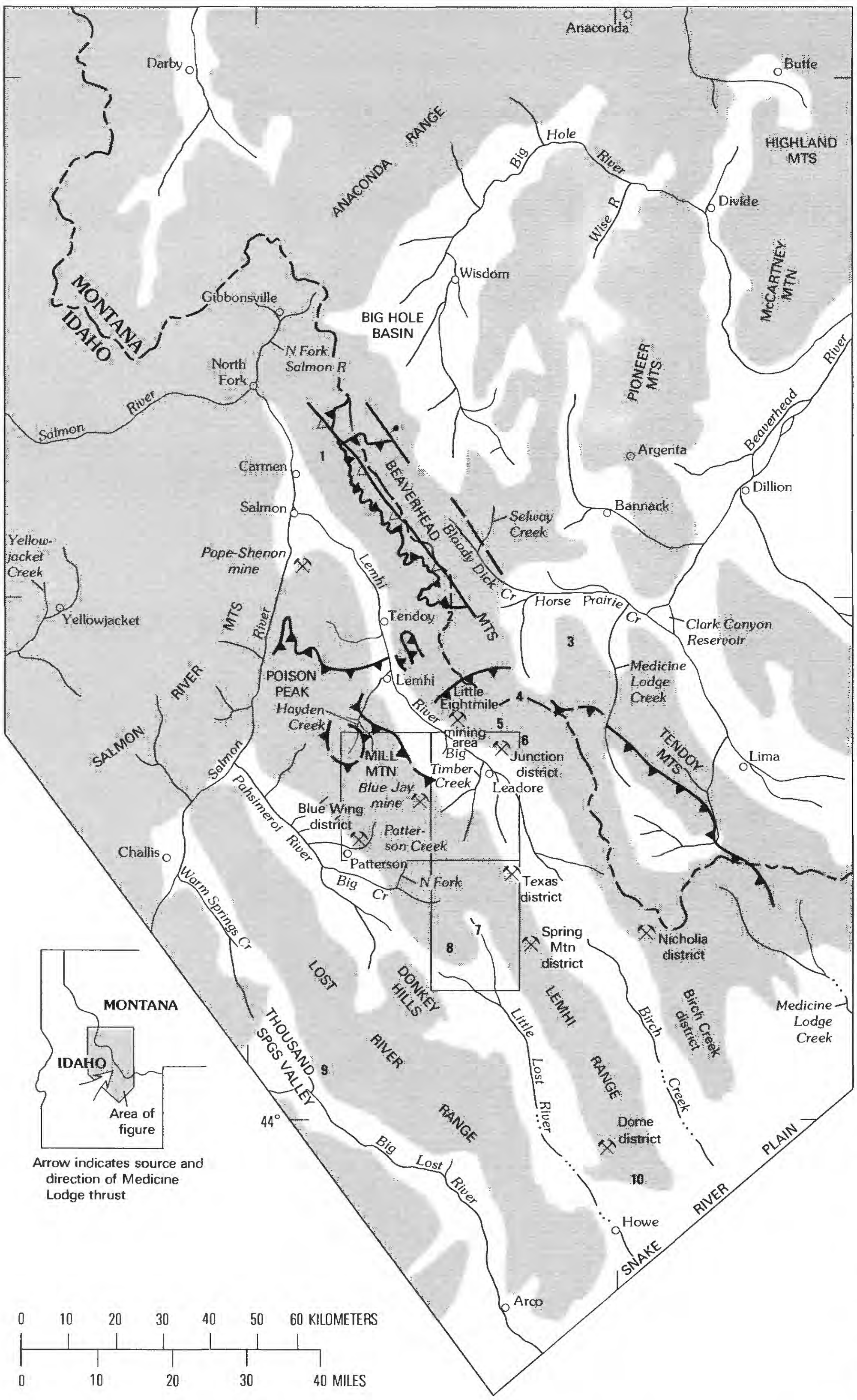
Patterson, the only other sometime community, in the southwest part of the Patterson quadrangle, was also founded in the early 1900's with the discovery of silver and later tungsten ores at the Ima mine. The town remained an active center until 1958, when tungsten production ceased because of withdrawal of price supports.

The Lemhi Range is crossed by only one road, in Spring Mountain Canyon and the valley of Squaw Creek, which connects the Birch Creek and Little Lost River Valleys in the south part of the Gilmore quadrangle. Other roads reach short distances into many of the mountain valleys, but travel in most of the Lemhi Range is by foot or horse on U.S. Forest Service trails. The rolling upper part of the Beaverhead Mountains, north of Leadore, is crossed by a road that reaches westward from Bannock Pass, across Grizzly Hill to connect with Idaho Highway 28 north of Leadore.

TOPOGRAPHY AND DRAINAGE

The central part of Lemhi Range is a mass of jagged, ice-carved peaks that stand at altitudes of more than 3,100 m and tower high above adjacent U-shaped valleys. Nearly all the peaks have supported major glaciers repeatedly, and the distribution of ice-carved features suggests that the mountain glaciers coalesced in at least the latest main glaciation to form an ice field in the highest part of the range. As a result, nearly all the peaks have been reduced to horns and a radiating series of sharp-crested ridges by the cumulative effects of glacial erosion (fig. 2). A few areas in the core of the range that remained above the ice persist today as nearly flat, rolling alpine meadows flanked by cirques or ice-carved valleys. Total relief of the central part of the range is about 1,800 m.

The precipitous walls of the range descend into the Lemhi and Birch Creek valleys on the east and the Pahsimeroi and Little Lost River valleys on the west—great, northwest-trending, structural trenches. The drainage in each trench is now divided, that of the Lemhi-Birch Creek valley at Gilmore Summit, and that of the Pahsimeroi-Little Lost River valley at the Donkey Hills. The Lemhi and Pahsimeroi Rivers flow northwest into the Salmon River, and Birch Creek and the Little Lost River flow to the southeast to the Snake River, joining it underground after disappearing into sinks in the lavas of the Snake River Plain. The valleys, which resemble one another except for opposite drainage directions, are broad and open, with gently sloping sagebrush-covered surfaces cut in places by steep-walled dry gulches. Perhaps the most prominent feature in the Lemhi Valley is Middle Ridge, which rises 100–200 m above the valley floor east of Gilmore



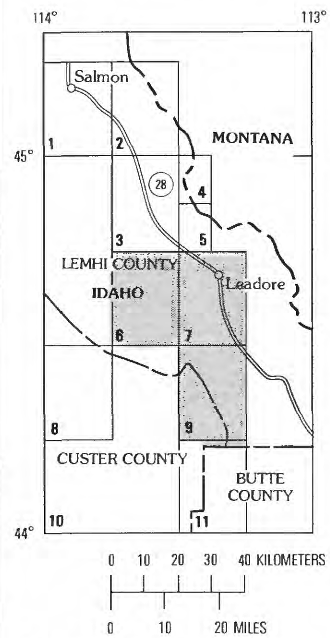
EXPLANATION

Medicine Lodge thrust system—Approximately located. Sawteeth on upper plate

Beaverhead Divide fault zone—Sawteeth on downthrown side

Locality index

1. **Carmen Creek**
2. **Lemhi Pass**
3. **Maiden Peak**
4. **Bannock Pass**
5. **Grizzly Hill**
6. **Railroad Canyon**
7. **Sawmill Canyon**
8. **Summerhouse Canyon**
9. **Elkhorn Ranch–Sawmill Gulch**
10. **Black Canyon**



Index map of east-central Idaho and southwestern Montana showing location of Patterson, Leadore, and Gilmore quadrangles and of adjacent areas described in recent reports

1. Salmon; Anderson, 1956
2. Goldstone Mountain; Anderson, 1957
3. Lemhi; Anderson, 1961
4. Lemhi Pass; Staatz, 1972
5. Goat Mountain; Staatz, 1973
6. Patterson; Ruppel, 1980
7. Leadore; Ruppel, 1968
8. Doublespring; Mapel and others, 1965
9. Gilmore; Ruppel and Lopez, 1981
10. Borah Peak; Ross, 1947
11. Hawley Mountain; Mapel and Shropshire, 1973

FIGURE 1.—Index map of east-central Idaho and southwest Montana.

FIGURE 2.—Photographs showing geologic units exposed in the ice-carved mountains of central Lemhi Range.
A, View looking south into cirque on north side of Mogg Mountain in northern part of Patterson quadrangle. Rocks are Swauger Formation, in nearly vertical beds cut by west-dipping thrust faults and shearing surfaces.

Summit. The Lemhi Valley abruptly narrows north of the Patterson quadrangle, to a north-trending trench only a few kilometers wide, but it resumes its northwest trend and broad, open character before it joins the valley of the Salmon River at Salmon, Idaho.

Although the mountain valleys contain small streams tributary to the main rivers, most of these streams do not carry enough water to maintain a flow across the gravels and other sediments that fill the main valleys. Most tributary streams disappear abruptly at the mountain front, and their water reappears in springs and swamps in the main river lowlands, like those at the head of the Lemhi River near Leadore. A few major tributary streams do have sufficient flow throughout the year to withstand the losses to subsurface flow and irrigation; chief among these are Big Timber Creek, Big Eightmile Creek, and Canyon Creek in Railroad Canyon, all of which empty into the Lemhi River; Patterson Creek, which empties into the Pahsimeroi River; and Canyon Creek in Sawmill Canyon, which forms the head of the Little Lost River. Hayden Creek, the main tributary to the Lemhi River, partly heads in the northern part of the Patterson quadrangle, and flows northeast to enter the Lemhi River near Lemhi, Idaho. It differs from most other streams in the central part of the Lemhi Range in having a valley that is mostly in or near bedrock, so that its volume is little diminished by losses to subsurface flow.

PRESENT WORK AND ACKNOWLEDGMENTS

This report is a general and largely descriptive summary of the geology of the central part of the Lemhi Range and of the relation of that geology to the geologic framework of east-central Idaho and southwest

B, View, looking southeast and south, of cirques and glacial valleys in upper part of Falls Creek, western part of Patterson quadrangle. Rocks are Gunsight and Apple Creek Formations, cut by multiple west-dipping imbricate thrust faults. 1969.

C, Junction Peak and Yellow Peak, glaciated peaks in east part of Patterson quadrangle. Cirque wall on north face of Yellow Peak is more than 350 m high, composed of Gunsight Formation overlain with angular unconformity by Kinnikinic Quartzite. Junction Peak is underlain by Kinnikinic Quartzite. Looking south. Shorty in foreground. 1962.

D, Glaciated valley of North Fork of Little Timber Creek, Gunsight Peak underlain by Kinnikinic Quartzite left-center. Type locality of Gunsight Formation. Looking east. Lemhi Valley and Beaverhead Mountains in background.

E, Compound cirques and glaciated valleys of Big Creek. Type locality of Big Creek Formation in Patterson quadrangle. Looking west from Park Fork of Big Creek. Cliffs in foreground are Big Creek Formation overlain by slabby talus of Apple Creek Formation.

Montana. The detailed geologic maps that are the basis for this report have been published separately (Ruppel, 1968; Ruppel, 1980; Ruppel and Lopez, 1981) (pl. 1), as have reports concerned with some regional stratigraphic and structural topics, including: Precambrian and Lower Ordovician sedimentary rocks (Ruppel, 1975; Ruppel and others, 1975); the Medicine Lodge thrust system (Ruppel, 1978); regional relations of thrust systems (Ruppel and Lopez, 1984); steep faults and their regional relations (Ruppel, 1964, 1982); late Cenozoic drainage changes and regional warping (Ruppel, 1967); glaciation (Ruppel and Hait, 1961); and the geologic, geochemical, and geophysical setting of the Texas (Gilmore) mining district (Ruppel and others, 1970).

These maps and reports are products of a study by the U.S. Geological Survey of the geology and mineral resources of the central part of the Lemhi Range. This study was begun in 1959 and continued intermittently until 1975. In the course of the work, we have had the benefit of advice and consultation with many colleagues—Professor Robert Scholten of The Pennsylvania State University, Ralph Nichols of the University of Montana, Professor David A. Bostwick of Oregon State University, Ora H. Rostad of American Metals-Climax, Inc., and J.E. Harrison, W.H. Hays, S.W. Hobbs, the late M.R. Klepper, G. Edward Lewis, R.J. Ross, Jr., the late C.P. Ross, W.J. Sando, M.H. Staatz, C.A. Wallace, and I.J. Witkind, all of the U.S.

Geological Survey. We also have had the help of many assistants—M.H. Hait, Jr., J.G. Smith, M.P. Mifflin, H.B. McFadden, R.G. Tysdal, C.K. Scharnberger, M.F. Gregorich, G.M. Fairer, D.A. Schleicher, H.R. Covington, and D.K. Fridrich.

Many residents of the Lemhi Valley have shown an interest in our studies, and have provided help and information: Clem Zook, Worth Hawley, the late Orion Lindskog, the late Roland P. Davidson, the late George Howell, and Dare Anderson, all of Leadore, Idaho, the late G. Grover Tucker of Gilmore, Idaho, and Sandy Sims, the late Howard J. Sims, and Marjorie Sims of Salmon, Idaho, provided helpful information on mines and mineral deposits and mining history. Lloyd and Beva Clark of Leadore provided horses, mules, and pack string that made accessible the most remote parts of the Lemhi Range, and Ralph and Ruth Sims of Tendoy, Idaho, provided housing and storage facilities. Orlo Johnson and C.P. Guillette of the U.S. Forest Service in Leadore provided helpful advice on access to different parts of the Lemhi Range, and we consulted with them frequently on many problems of mutual interest. The help and cooperation of these and many other residents of the Lemhi Valley contributed greatly to our work, and we are indebted to them.

SUMMARY OF EARLIER GEOLOGIC STUDIES

The earliest geologic study in the central part of the Lemhi Range was that of Umpleby (1913), whose report continues to be a standard reference on the region. Later, more detailed studies have been made in only a few areas: the Little Eightmile mining area (Thune, 1941); the southern part of the Leadore quadrangle (Knowles, 1961); the Gilmore quadrangle (Hait, 1965); and the Patterson area or Blue Wing mining district (Callaghan and Lemmon, 1941; Hobbs, 1945; Anderson, A.L., 1948). Glaciation in the Gilmore area and farther south in the Lemhi Range has been described by Dort (1962, 1965), and some of the glacial deposits near Gilmore have been discussed in detail by Funk (1976), Knoll (1977), and Butler (1982). A few other reports, referred to later, discuss special aspects of mineral deposits, mining districts, and ground water.

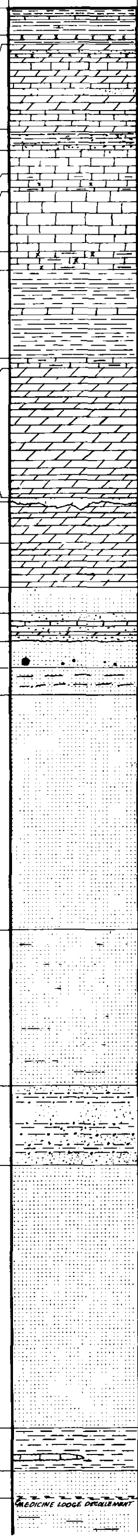
Reports on adjacent or nearby areas (pl. 1 and fig. 1 and 3) and topical reports on various aspects of regional geology are more numerous, and have provided much help in understanding the geology of the central Lemhi Range. Kirkham (1927, 1931) made a geologic reconnaissance of a large region that included the southern part of the Lemhi Range. C.P. Ross' discussion of the geology of the Borah Peak quadrangle (1947) has been particularly useful, as have his other reports on the rocks of surrounding regions (1934a,b, 1937, 1961a,b, 1962a,b, 1963). A.L. Anderson (1956, 1957, 1959, 1961) mapped and described the northern part of the Lemhi Range and adjacent areas farther north and east. Beutner (1968, 1972) discussed the geology of the southern part of the Lemhi Range. Scholten and others (1955) mapped and described a large part of the Beaverhead Mountains and southwest Montana, and Scholten (1957, 1960, 1967, 1973) has published other thought-provoking reports on the region. Smith (1961), Ramspott (1962), M'Gonigle (1965), and Lucchitta (1966) described the geology of part of the Beaverhead Mountains east of the Leadore quadrangle. Sloss (1950, 1954) and Sloss and Moritz (1951) reviewed the stratigraphy of Paleozoic rocks in southwest Montana, as did McMannis (1965) and Klepper (1950). Staatz (1972, 1973, 1979) recently described the geology of the Beaverhead Mountains north of the Leadore quadrangle. Hobbs and others (1968) redefined the Kinnikinic Quartzite in its type locality west of the Lemhi Range. Mapel (Mapel and others, 1965; Mapel and Shropshire, 1973) mapped part of the Lost River Range, west of the Lemhi Range.

The stratigraphy of Paleozoic rocks in central Idaho has been summarized by Ross (1962a,b). Upper Paleozoic rocks in this region have been described more recently by Huh (1967), Mamet and others (1971), Skipp and Hait (1977), Skipp, Sando, and Hall (1979), and Skipp, Hoggan and others (1979).

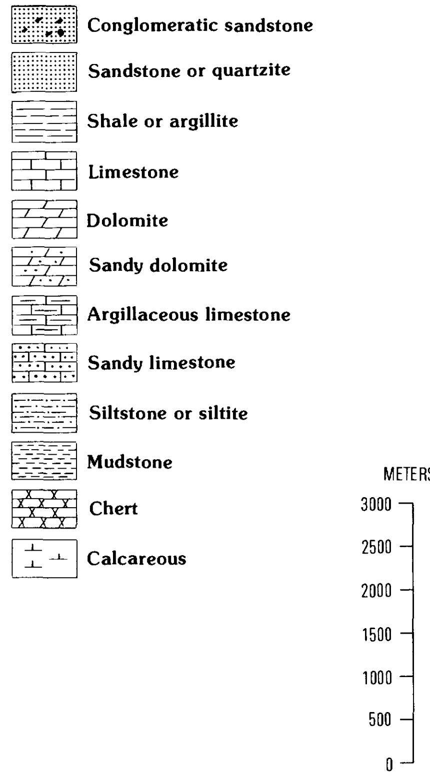
GEOLOGY

The ice-sculptured central part of the Lemhi Range exposes the underlying geologic framework in abundant and sometimes confusing detail and makes clear the need to consider the stratigraphic and structural framework in both local and regional contexts (fig. 3). The single most important element in understanding the geology of the region is recognition of the pervasive effects of thrust faulting. The Medicine Lodge thrust system underlies almost the entire region, and most of the Precambrian and all of the Paleozoic sedimentary rocks have been transported far to the east of their depositional areas by thrust faulting, to overlap rocks of similar age deposited in different sedimentary environments (Ruppel, 1978). Although the Medicine Lodge thrust fault was recognized more than 50 years ago by Kirkham (1927, p. 26-27), Sloss and Moritz (1951, p. 2160) were the first to understand its regional importance, for they pointed out that upper Paleozoic rocks above the thrust in the southern part of the

AGE	FORMATION NAME	THICKNESS DISTRIBUTION (METERS)	REFERENCE
TRIASSIC	Dinwoody Formation	36 Southern Beaverhead Mtns, 300 Central Beaverhead Mtns	4,9
PERMIAN	Phosphoria Formation and older rocks	55–100	2,4,14,26,27
PENNSYLVANIAN	Snaky Canyon Formation	400–1300 Thickens southward	1,2,4,9,15 25,26,27
	Bluebird Mountain Fm	0–130	27
	Railroad Canyon Formation	260 Beaverhead Mountains	15
MISSISSIPPIAN	South Creek Fm	90–200	7,10,26
	Scott Peak Formation	600–700	7,10,15,26
	Middle Canyon Fm	150–200	7,13,15,26
	McGowan Creek Formation	40–120 Beaverhead Mountains 80–150 Southern Lemhi Range 120–1000 Lost River Range	7,10,15,21,26
DEVONIAN	Three Forks Fm	15–90 Thickens westward	14,15,18,23
	Jefferson Dolomite	60–300 Beaverhead Mountains 90–820 Lemhi Range 600–900 Lost River Range Includes up to 400 Grand View Dolomite in Lost River Range	5,9,12,13,14, 15,18,20,22,24
	channel sandstone	0–50 Locally Lemhi and Lost River Ranges	3,17,22
SILURIAN	Laketown Dolomite	0–60 Lemhi Range Thins to east and north. 400 Lost River Range	15,18 10,11
ORDOVICIAN	Saturday Mountain Fm	0–370 Thins to east and south	11,12,14,47
	Kinnikinic Quartzite	0–300 Thins to east	4,8,17,22,23
	Summerhouse Formation	0–300	8,13,17,19
CAMBRIAN AND LATE PROTEROZOIC	Wilbert Formation	0–300	19
	Lawson Creek Formation	At least 1200 in Lost River Range and Salmon River Mtns	6
MIDDLE PROTEROZOIC	Swauger Formation	Maximum thickness 3100	16
	Lemhi Group: Gunsight Formation	1800	16
	Lemhi Group: Apple Creek Formation	760–900	16
	Lemhi Group: Big Creek Formation	3100	16
	Lemhi Group: West Fork Formation	460	16
	Lemhi Group: Inyo Creek Fm	215	16
	Yellowjacket Formation	8000+	This report



EXPLANATION



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8. James and Oaks (1977)
9. Lucchitta (1966)
10. Mamet and others (1971)
11. Mapel and others (1965)
12. Mapel and Shropshire (1973)
13. M'Gonigle (1965)
14. Ross (1961a)
15. Ruppel (1968)
16. Ruppel (1975)
17. Ruppel (1978)
18. Ruppel and Lopez (1981)
19. Ruppel and others (1975)
20. Ruppel and others (1970)
21. Sandberg (1975)
22. Scholten and Hait (1962)
23. Scholten and others (1955)
24. Scholten and Ramspott (1968)
25. Shannon (1961)
26. Skipp and Hait (1977)
27. Skipp and others (1979)

FIGURE 3.—Generalized section of sedimentary rocks in and near the central Lemhi Range, Idaho. Modified from Ruppel and Lopez (1984, p. 24).

Beaverhead Mountains are radically different from rocks of similar age below the thrust in the Tendoy Mountains only about 10 km farther east.

The abrupt changes in rock units across the edge of the thrust system led to a concept of changes in sedimentation across a zone of transition, or hinge line, between shelf and geosyncline (Sloss and Moritz, 1951, p. 2138; Scholten and others, 1955, p. 352; Scholten, 1957, p. 152). The changes are real, and do in fact reflect differences between shelf—or seaway, and miogeoclinal sedimentation. They do not occur across a zone of transition, however, but abruptly across the Medicine Lodge thrust system. The hinge line concept was useful in separating rocks deposited in different environments, but it obscured the real nature of the abrupt changes and delayed recognition of the significance of thrust faulting.

The thrust plate also now conceals remnants of the former Lemhi arch, a major landmass that separated the miogeocline on the west from an intermittent embayment or seaway on the east in Late Proterozoic and early Paleozoic time. The arch, which influenced late Paleozoic and perhaps early Mesozoic marine sedimentation and shed continental debris eastward later in the Mesozoic, was overridden by the thrust in Late Cretaceous time (Sloss, 1954; Ruppel, 1978) (fig. 4). The arch is evident in shoaling patterns in lower Paleozoic rocks, and in the many arches and uplifts described previously (Sloss, 1954; Scholten, 1957, p. 167; Armstrong, 1975, p. 452) to explain anomalous early Paleozoic marine sedimentation patterns. What had seemed to be a system of islands bobbing up and down now seems more likely to represent parts of a single major landmass, the Lemhi arch. Rocks from the western, miogeoclinal side of the arch have been thrust eastward across the arch, and conceal it; they overlap rocks deposited on the eastern, seaway side of the arch. As a result rocks originally deposited on opposite shores of the arch are now placed nearly together.

The Lemhi arch was first proposed as a major middle Paleozoic landmass in east-central Idaho by Sloss (1954), and its definition was revised by Ruppel (1978, p. 12; 1983; 1986) to include its earlier history in the Late Proterozoic and early Paleozoic. Scholten (1957, p. 166–167) proposed other names, the Skull Canyon uplift, the Tendoy dome, and the Beaverhead arch, for various parts of the Lemhi arch; and Armstrong (1975, p. 452–455) suggested a new name, the Salmon River arch, for a Precambrian uplift partly identical to the Lemhi arch. The original name, Lemhi arch, is used in this report.

The effects of regional thrust faulting are pervasive throughout the central and southern parts of the Lemhi Range. The most extreme deformation is in the lower

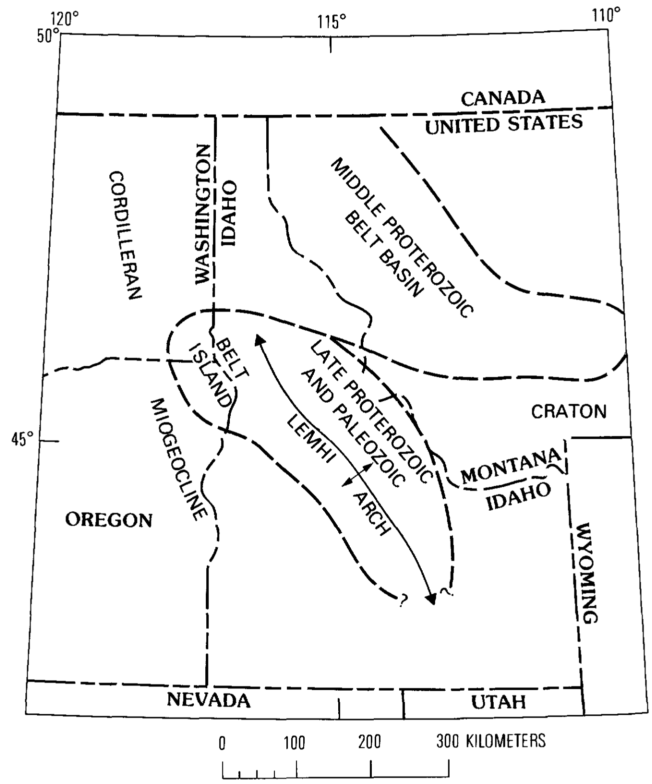


FIGURE 4.—Map showing approximate location of the Lemhi arch and flanking regions of Late Proterozoic and early Paleozoic marine sedimentation (from Ruppel, 1986).

part of the thrust-faulted plate, where the rocks are partly mylonitized and extensively sheared and brecciated. Higher in the plate, the rocks are in tight, nearly isoclinal overturned folds, asymmetric to the east, and broken by closely spaced, west-dipping imbricate thrusts that shuffle upright and overturned limbs. About half the sedimentary rocks in the central Lemhi Range are upside down.

The thrust system also influenced the emplacement of post-thrusting intrusive igneous rocks in the central Lemhi Range, where both stocks and related deposits of metallic minerals are found only in the lower part of the allochthonous block.

All Proterozoic and Paleozoic sedimentary rocks of the Medicine Lodge thrust plate in east-central Idaho and southwest Montana, and all sedimentary rocks in the central Lemhi Range except the Yellowjacket Formation, therefore have been transported far to the east of their depositional area, the Cordilleran miogeocline in central or western Idaho. These far-traveled rock units differ from those of similar age east of and beneath the thrust plate. Correlation with rocks of somewhat similar ages in the Bayhorse area west of the Lost River Range is made difficult by the presence of other major thrust plates there.

Sedimentary rocks in the central Lemhi Range are mostly of Middle Proterozoic and early to middle Paleozoic age, but Mississippian rocks crop out near Gilmore, and in the northern part of the Leadore quadrangle and a few kilometers farther north near the mouth of Little Eightmile Creek (Staatz, 1973) (fig. 3). Mississippian and younger Paleozoic rocks dominate the southern part of the Lemhi Range—where Proterozoic and lower Paleozoic rocks occur only on the western side of the range. Paleozoic rocks are the most widespread rocks in the southern part of the Beaverhead Mountains, east of the Lemhi Range, and in the Lost River Range farther west. Devonian and Mississippian rocks also crop out along the lower reaches of Hayden Creek north of the Patterson quadrangle, their northernmost exposures in the Lemhi Range. Mesozoic sedimentary rocks are absent in the Lemhi Range, but do occur in the Beaverhead Mountains east of Leadore in imbricate thrust slices. The sequence of sedimentary rocks in and near the central Lemhi Range is summarized in figure 3.

The Middle Proterozoic sedimentary rocks of the central Lemhi Range are slightly metamorphosed, fine- to medium-grained quartzite, quartzitic sandstone, siltite, and argillite; the sequence is more than 10,000 m thick, and is overwhelmingly sand—about 95 percent. These rocks were deformed and eroded in the Late Proterozoic, when the Lemhi arch formed, so that succeeding Late Proterozoic(?) and lower Paleozoic rocks, the Wilbert Formation, the formation at Tyler Peak of McCandless (1982), and the Lower Ordovician Summerhouse Formation, overlie them unconformably in the southern part of the Lemhi Range. These Late Proterozoic and lower Paleozoic formations wedge out northward, and the Middle Proterozoic quartzites are overlain directly, with strong angularity, by the Middle Ordovician Kinnikinic Quartzite. Above the Kinnikinic, the rocks are dominantly marine carbonates, principally dolomite of Ordovician, Silurian, and Devonian age, and limestone of Mississippian age.

After Medicine Lodge thrusting and related deformation in Late Cretaceous time, the now intensely crumpled and faulted rocks of the Lemhi Range were intruded in Eocene time by small stocks and sheets, mostly of monzogranite, granodiorite, quartz monzodiorite, and quartz monzonite. The stocks were rapidly unroofed by erosion, and in later Eocene time were partly covered by basaltic and andesitic flows, breccias, and tuffs of the Challis Volcanics. Later in the Tertiary, thick tuffs, tuffaceous sands and gravels, and some freshwater limestone accumulated in the newly forming Lemhi and Pahsimeroi Valleys. In middle and late Tertiary time, continuing to the present, major block uplifts formed the present northwest-trending mountain ranges.

In the late Tertiary, regional arching, parallel to the Snake River Plain, formed the Gilmore Summit and Donkey Hills divide simultaneously with development of the Snake River Plain and renewed steep faulting. As a result of this arching, the drainages of the Lemhi and Pahsimeroi Rivers were reversed before early Pleistocene time, when the earliest glaciers accumulated in the Lemhi Range. Later, major episodes of glaciation carved the Lemhi Range into its present jagged form and left extensive deposits of till in mountain valleys and in terminal moraines at the valley mouths, and widespread deposits of glacial outwash gravels on the main valley floors. Late cirque glaciers have since resharpened the peaks, and left small, fresh moraines in the heads of some valleys. The glacially oversteepened valley walls provide an ideal setting for snow avalanches, and rock debris carried down by the avalanches chokes many of the glacial valleys.

Strike-slip faults and many range front faults are considered to be still active, for they cut glacial deposits and the most recent alluvial fan deposits, disrupt drainages, and are marked by widespread, fresh fault scarps. Movement on these faults appears to be a major cause of landsliding, and young landslides, probably earthquake induced, are among the most conspicuous and widespread surficial deposits in and near the central Lemhi Range.

MIDDLE PROTEROZOIC TO LOWER CAMBRIAN ROCKS

Middle Proterozoic rocks in the central Lemhi Range are part of a belt of similar rocks that extends from the southwest side of the Lemhi Range near its south end at the Snake River Plain northward for about 250 km to the Idaho-Montana line at Lost Trail Pass; westward to the vicinity of Challis, Idaho; and eastward into the Beaverhead Mountains north and east of Leadore, Idaho. The rocks were recognized as Precambrian and correlated with the Belt Series (now Supergroup) of Montana by Umpleby (1913, p. 30–32), and were partly described by Ross (1934a, p. 16–18; 1947, p. 1097–1102; 1961a, p. 195; 1962a, p. 14), and Anderson (1959, p. 16–18; 1961, p. 16–21). The formations named by Ross and Anderson subsequently were redefined and described more completely by Ruppel (1975), who also removed these rocks from the Belt Supergroup. The Yellowjacket Formation has been studied most recently by Lopez (1981), who divided it for the first time. The Proterozoic sequence (table 1) retained in modified form (Ruppel, 1975) the original names given by Ross ("Swauger quartzite, Lemhi quartzite, Yellowjacket formation") (1934a, p. 16; 1947, p. 1096–1099) and

Anderson ("Apple Creek phyllite") (1961, p. 19–21), but changed the Swauger Quartzite to Swauger Formation; elevated the Lemhi from a formation to a group consisting of five formations; placed the Apple Creek Phyllite in the Lemhi Group as the Apple Creek Formation and corrected its position in the sequence; and extended the name Yellowjacket Formation as suggested by Ross (1962a, p. 14). Because these rocks have been redescribed so recently, only summary descriptions are given in this report. The Middle Proterozoic Lawson Creek Formation (Hobbs, 1980), which overlies the Swauger Formation in the north part of the Lost River Range, is not present in the Lemhi Range, probably because of Late Proterozoic or early Paleozoic erosion. The Wilbert Formation of Late Proterozoic(?) and Cambrian age, described in the south part of the Lemhi Range (Ruppel and others, 1975; Derstler and McCandless, 1981; McCandless, 1982), is absent in the central Lemhi Range and farther north, either because of later Proterozoic or early Paleozoic erosion or nondeposition.

TABLE 1.—*Revised sequence of Middle Proterozoic sedimentary rocks, Lemhi Range, Idaho*

[Modified from Ruppel (1975)]

	Swauger Formation Pale-purple to grayish-green, medium-grained hematitic quartzite; maximum thickness about 3,100 meters
Lemhi Group	Gunsight Formation Light-brownish-gray to grayish-red-purple, fine-grained feldspathic quartzite; minimum thickness 1,800 meters
	Apple Creek Formation Grayish-green siltite with abundant lenses of fine-grained sandstone cemented by ferrodolomite; thickness about 760–900 meters
	Big Creek Formation Greenish-gray to light-gray, fine-grained feldspathic quartzite; thickness about 3,100 meters
	West Fork Formation Medium-gray to greenish-gray siltite that contains lenticular algal limestone; thickness 460 meters
	Inyo Creek Formation Medium-gray to light-gray, fine-grained to very fine grained, thin-bedded feldspathic quartzite; thickness unknown, greatest measured thickness more than 215 meters; base not exposed
	Yellowjacket Formation Medium-gray to medium-dark-gray, fine-grained, feldspathic finely bioitic quartzite, and interbedded siltite; thickness 6,500–8,000 meters; base not exposed

YELLOWJACKET FORMATION

The Yellowjacket Formation consists of the oldest Proterozoic sedimentary rocks known in east-central Idaho. In the central part of the Lemhi Range, it is exposed only in the northern part of the Patterson quadrangle and in the canyons of Squaw Creek and Lake Creek in the Gilmore quadrangle. These rocks are the predominant ones of the northern parts of the Lemhi Range, Beaverhead Mountains, and Salmon River Mountains and from there westward to the area near the Yellowjacket Mine (Ross, 1934a, p. 16) (fig. 1). They are cut off in the Beaverhead Mountains on the east, by the Miner Lake-Beaverhead Divide fault zone, and are absent in southwest Montana (Lopez, 1981, p. 93–107; Ruppel and Lopez, 1984, p. 23–26).

The formation appears to be as much as 8,000 m thick (Lopez, 1981, p. 137), although estimates of thickness by Ross (1934a, p. 16) were about 2,750 m near the Yellowjacket mine, and by Anderson (1956, p. 18; 1959, p. 17; 1961, p. 23) were at least 1,500 m and possibly as much as 3,100 m in the northern part of the Lemhi Range. Tucker (1975, p. 58) estimated a possible thickness of almost 5,500 m in the Beaverhead Mountains. Lopez (1981) subdivided the formation in the Salmon River Mountains into five members, and concluded that their composite thickness is 6,500–8,000 m. The base of the formation has not been found, however, and the top in the Lemhi Range, Beaverhead Mountains, and Salmon River Mountains is the Medicine Lodge thrust zone. The total thickness and stratigraphic relations to the other Precambrian rocks are not known. Yellowjacket rocks are present only beneath the Medicine Lodge thrust zone and are the only pre-Tertiary rocks interpreted to be autochthonous in and near the central Lemhi Range (Ruppel, 1978, p. 8; Ruppel and Lopez, 1984, p. 23–26).

The formation consists of medium-gray to medium-dark-gray fine-grained to very fine grained feldspathic, finely biotitic quartzite in beds mainly 6–60 cm thick but locally as much as 2 m thick, and of dark-greenish-gray to medium-dark-gray argillite and siltite in thin to medium beds. Interbeds and interbedded units of light-gray to medium-light-gray or greenish-gray quartzite are common throughout the formation and generally are somewhat coarser grained than the dark-colored rocks. The formation is dominantly quartzite in the Lemhi Range (Anderson, 1956, p. 17) but includes thick units of siltite and argillite in the Beaverhead Mountains (Tucker, 1975, p. 32–63). In isolated exposures recognition of the five members separated by Lopez (1981) is made difficult by compositional similarities but differences in grain size, range in grain sizes, thickness of graded beds, sedimentary structures, paleocurrent directions, and proportions of quartzite and argillite can be used to distinguish the members. All the contacts between members are transitional through thicknesses of 100–200 m.

Member A, the lowest exposed member, consists of graded beds of the coarsest grained quartzites and argillaceous quartzites in the formation. Coarse- to medium-grained quartzite, which is compositionally subarkose and quartz arenite, grades upward into fine-grained or very fine grained argillaceous quartzite or arkosic wacke, which in turn grades into sandy argillite. These graded beds are 10 cm to 1 m thick. The upper part of member A is finer grained, consisting of very fine-grained and fine-grained argillaceous quartzite and sandy argillite. Colors in member A range from light gray to dark gray, darkening both upsection and upward within individual graded beds. Member A is about 700 m thick.

Member B is dominantly very fine grained, very argillaceous quartzite and siltite. Graded beds are 5–50 cm thick. Colors are dominantly medium gray and greenish gray. Beds of slightly calcareous, argillaceous quartzites and lenticular bodies of banded and mottled, highly calcareous quartzite (20–35 percent calcite) are characteristic of this member. Member B is about 1,600 m thick.

Member C is characterized by couplets of light-colored argillaceous quartzite and (or) siltite, and dark-colored sandy and silty argillite that occur together in graded genetic units. The quartzite and siltite are arkosic wacke in composition. The graded intervals thicken upsection from couplets a few mm thick at the base to beds 5–15 cm thick near the top. The sand to shale ratio is about 1 near the base of the member and increases to 2 or more at the top. Convolute laminations and pseudomudcracks are very common in this member. Member C is about 1,300 m thick.

Member D is mainly very fine grained argillaceous quartzite in graded beds 10 cm to about 1 m thick. Grading is exhibited mainly by an increase in the content of clay-size matrix upward, although the maximum grain size rarely decreases upward. Colors darken upward within graded beds, generally from medium gray to dark gray. Member D is at least 2,500 m and may be as much as 3,500 m thick.

Member E contains less argillaceous quartzite than other members of the formation. Quartzite, or subarkose and quartz arenite, grades upward into argillaceous quartzite, or arkosic wacke. Grading is exhibited by the increase of argillaceous matrix content. The graded beds are, in general, thicker than in the other members and range from 30 cm to 2 m thick. Grain sizes are dominantly very fine and fine sand. Colors are light gray to medium dark gray, typically darkening upward within graded beds. Member E probably is at least 1,000 m thick.

A minimum age of 1.4 b.y. for the Yellowjacket Formation is suggested by isotopic dates from several porphyritic quartz monzonite plutons that intrude it (Lopez, 1981, p. 105). No data are available for its maximum age, except that by comparison with the probably correlative Prichard Formation of the Belt Supergroup, it is probably younger than about 1.7 b.y. (Lopez, 1981, p. 48–49).

The Yellowjacket Formation is interpreted to be a deep-marine turbidite on the basis of features characteristic of turbidites and their internal organization as set forth in the Bouma model (Lopez, 1981, p. 84–97). In addition, no shallow-water features were observed within the Yellowjacket. The formation's great thickness and sequence of alternating argillaceous quartzite and sandy argillite are consistent with this interpretation. The original extent of the Yellowjacket depositional basin must have been much larger than present rock distribution suggests, because strata preserved in east-central Idaho all indicate deposition in the distal portions of subsea-fan systems.

REASONS FOR INTERPRETING THE YELLOWJACKET FORMATION TO BE AUTOCHTHONOUS

The Medicine Lodge thrust plate is underlain by the Yellowjacket Formation everywhere in east-central Idaho that the sole zone of the thrust plate is exposed. Although only a few of these exposures are in the central Lemhi Range, mapping elsewhere in the region shows that deformation in the Yellowjacket Formation is very different from that in the complexly folded and faulted structurally and stratigraphically higher strata of the Medicine Lodge thrust plate, and shows that much of the deformation in the Yellowjacket Formation is Precambrian in age, and not related to thrusting. These relations suggest that the Yellowjacket Formation is autochthonous (Ruppel, 1978, p. 8–9; Lopez, 1981, p. 103–118; Evans, 1981, p. 125–134; Ruppel and Lopez, 1984, p. 23–26).

The Yellowjacket Formation west of Salmon, Idaho, is folded into very large, tight to isoclinal folds that commonly are overturned to the southwest, in contrast to the small tight folds overturned to the east and cut by multiple imbricate thrust faults in the Medicine Lodge plate. The Yellowjacket folds are cut by granitic rocks that are 1.3–1.4 b.y. old (Lopez, 1981, p. 105; Evans, 1981, p. 32–53, 64–94), a relation that shows the folds to be older than the granitic rocks.

Precambrian folding was accompanied by movement on the Miner Lake-Beaverhead Divide fault zone in the Beaverhead Mountains and along a major zone of northeast-trending faults west of Salmon, Idaho, all of which have moved recurrently in Cenozoic time. The northwest-trending Miner Lake-Beaverhead Divide fault zone, a wide zone that dips 70° or more southwest in the Beaverhead Mountains east of Salmon, cut off the Yellowjacket Formation on the east in early Middle Proterozoic time. Reverse movement along it folded and broke the Medicine Lodge thrust plate in mid-Cenozoic time (Ruppel, 1982, p. 11; Ruppel and Lopez, 1984, p. 23–26). Northeast-trending steep faults break the Yellowjacket Formation in the Panther Creek and

Yellowjacket areas, west of Salmon (Lopez, 1981, p. 108), and partly controlled the development of the Yellowjacket depositional basin (Hahn and Hughes, 1984). Some of these faults also controlled the emplacement of the 1.3–1.4 b.y. old granitic intrusive rocks that cut folds in the Yellowjacket Formation; later movement along this zone has broken the granitic rocks and continued recurrently to the present time (Lopez, 1981, p. 108–109; Evans, 1981; Evans and Zartman, 1981). These faults are part of a broad zone of steep, northeast-trending faults that reaches from central Idaho far to the northeast in Montana (Hyndman and others, 1977; O'Neill and Lopez, 1985; Ruppel, 1982, p. 17–19).

The recurrent movement on Precambrian faults in Cenozoic time suggests that these faults must still extend to great depths and that they cannot have been cut off and moved eastward on some still-hidden major thrust fault. This indicates that the Yellowjacket is still where it was deposited and not displaced by any thrusts comparable to, or related to the Medicine Lodge thrust. The differences in structural style of the Yellowjacket rocks, compared to those of the Medicine Lodge plate, and their different ages, also suggest that the Yellowjacket Formation is autochthonous.

LEMHI GROUP

The Lemhi Group (Ruppel, 1975, p. 6–12) includes the rocks originally included in the Lemhi Quartzite by Ross (1947, p. 1096) and is divided into five formations, in ascending order: the Inyo Creek, West Fork, Big Creek, Apple Creek, and Gunsight Formations. The Apple Creek Formation was originally named the Apple Creek Phyllite (Anderson, 1961, p. 19). The Lemhi Group is more than 6,100 m thick in the central Lemhi Range and is mostly greenish-gray or brownish-gray, fine-grained feldspathic quartzite and lesser amounts of greenish-gray to purplish-gray siltite and argillite. The lowest two formations of the group, the Inyo Creek and West Fork Formations, are known only in the southern part of the Patterson quadrangle and for a short distance farther south in the headwater area of the north fork of Big Creek. The other formations are widespread throughout the central and southern Lemhi Range, in the Beaverhead Mountains north and east of Leadore, and much farther north, along the crest of the Beaverhead Mountains from near Lemhi Pass to Lost Trail Pass, and westward to the vicinity of Challis, Idaho.

INYO CREEK FORMATION

The lowest formation of the Lemhi Group is the Inyo Creek Formation, which is known to crop out only in its type locality in the southern part of the Patterson quadrangle and for a short distance farther south in the central Lemhi Range. The base of the formation is cut by a thrust fault and concealed by surficial deposits; the top is gradational through a thickness of about 30 m into the overlying siltite of the West Fork Formation. The greatest measured thickness of the formation is about 215 m.

The exposed part of the Inyo Creek is mainly light-greenish-gray to medium-dark-gray, very fine grained to fine-grained feldspathic (5–25 percent) quartzite in beds 0.2–1.5 m thick. Some beds are finely cross laminated, some are ripple marked, and many beds in the lower part of the formation are separated by silty partings and a few contain mudstone chips. Thin beds of medium-gray to greenish-gray siltite are present throughout the formation, and become increasingly abundant in the upper part where it grades into the overlying siltite of the West Fork Formation. Some beds of quartzite are silicified and extremely hard, and some are pyritic. The rocks of the formation are grouped into alternate sequences 3–10 m thick of light- and dark-colored beds, reflecting greater amounts of feldspar and sericite in light-colored beds, and of heavy minerals and pyrite in dark-colored beds.

WEST FORK FORMATION

The West Fork Formation, like the Inyo Creek, is exposed only in the southern part of the Patterson quadrangle and for a short distance farther south. The formation is about 460 m thick. The lower contact is gradational from the underlying Inyo Creek Formation and the upper contact is gradational into the quartzite of the overlying Big Creek Formation through a zone of interbedded siltite and quartzite 7–12 m thick. The formation is mostly thick bedded to massive, medium-gray to dark-greenish-gray siltite, some beds of which are laminated or cross laminated and ripple marked. Interbeds of brownish- to greenish-gray, very fine grained to fine-grained feldspathic quartzite occur throughout the formation and are as much as 15 m thick in the lower part of the formation. Lenticular interbeds of yellowish-gray to greenish-gray, partly sandy, algal limestone are common in the upper part of the formation, and are present, but are less common in the lower part; these are the only stromatolites known in Proterozoic rocks in east-central Idaho.

BIG CREEK FORMATION

The Big Creek Formation is one of the thickest and most widespread formations in the Lemhi Group, and is the lowest formation of the group in most of the region, where the Inyo Creek and West Fork Formations are not known. It is widely exposed in the west-central part of the Lemhi Range, on Grizzly Hill in the Beaverhead Mountains north of Leadore, and southward from there at least locally along the Continental Divide. It is present also in much of the region along the the Continental Divide in the northern part of the Beaverhead Mountains (Tucker, 1975, p. 64–86; Ruppel and others, 1983). The formation is about 3,100 m thick in the Lemhi Range and probably about the same thickness in the Beaverhead Mountains (Tucker, 1975, p. 79). The base of the formation is gradational from the siltite of the underlying West Fork Formation; the top is gradational into the siltite of the overlying Apple Creek Formation through a thickness of about 45 m.

Despite its great thickness, the Big Creek is a strikingly homogeneous sequence of light-gray, pale-greenish-gray or light-brownish-gray, fine-grained feldspathic (10–35 percent) quartzite. The rock is typically speckled with grains of iron oxide which also are concentrated along laminae. Beds are 0.1–1 m thick or massive; the upper half of the formation tends to be thicker bedded than the lower half, but thick bedded or massive rocks are common throughout the formation. The uppermost 150 m is thin to medium bedded. These rocks contain scattered chips of greenish-gray argillite and siltite and interbeds of medium-dark-gray siltite as much as 3 m thick in the gradational zone with the Apple Creek, and many beds are laminated or cross laminated. A few layers of quartzite as much as 12 m thick in the lower half of the formation contain abundant irregular, small lenses of ferrodolomite-cemented, fine-grained quartz sandstone that weathers dark yellowish brown and gives the quartzite a mottled look. These beds also are strongly cross laminated and contrast sharply with the other rocks of the formation.

In the northern part of the Beaverhead Mountains, Tucker (1975, p. 64–86) included medium-grained to very coarse grained feldspathic quartzite and some conglomerate in the Big Creek Formation, rocks that MacKenzie (1949, p. 13–15) had earlier described in the Carmen Creek area as part of a "gray quartzite" unit. Tucker attributed the coarser grain size to deposition in a shallow basin close to the source area, and contrasted the rocks of the northern Beaverhead Mountains with the miogeoclinal, fine-grained Big Creek rocks of the Lemhi Range. Regional geologic mapping in the Beaverhead Mountains (Ruppel and others, 1983) has shown that although the Big Creek Formation becomes coarser grained northward and includes beds of medium-grained feldspathic quartzite, the coarser grained quartzite and conglomerate discussed by Tucker (1975, p. 64–86) are part of the Mount Shields Formation of the Missoula Group, in the upper part of the Belt Supergroup. These Missoula Group rocks are in fault-contact, across the Miner Lake-Beaverhead Divide fault zone, with rocks of the Yellowjacket Formation and Lemhi Group, including the Big Creek Formation (Ruppel, 1982, p. 11; Ruppel and Lopez, 1984).

APPLE CREEK FORMATION

The Apple Creek Formation is one of the most distinctive and useful stratigraphic keys in the Proterozoic sequence in east-central Idaho. The formation is exposed in a northwest-trending band across the central Lemhi Range, in a few places in the type area north of the Patterson quadrangle, and in the Beaverhead Mountains on the flanks of Grizzly Hill north of Leadore. Part of the formation is well exposed along U.S. Highway 93, at Ellis, Idaho, where the Pahsimeroi River joins the Salmon River, and in more extensive but scattered outcrops within a 6 km-wide area west of the highway (McIntyre and Hobbs, 1978). It is also present in the Beaverhead Mountains as far north as Big Hole Pass; in and east of Goldstone Pass, where it is widely exposed; and much farther south along the Continental Divide east of Leadore. It has not been mapped separately in most of the Beaverhead Mountains, however. It is present locally in the Salmon River Mountains, west and northwest of Salmon, Idaho.

The Apple Creek is 760–900 m thick in the Lemhi Range, and is composed mainly of grayish-green to medium-dark-gray or grayish-red-purple siltite. Its lower and upper contacts are both gradational, the lower one from the Big Creek Formation as previously described, and the upper one into the overlying quartzite of the Gunsight Formation through a thickness of about 60–90 m in which interbeds of Gunsight-like quartzite are increasingly common in the siltite. The siltite is in beds 6–15 cm thick. Much of it contains distinctive irregular streaks and lenses typically 1–2 cm thick and as much as 0.6 m long of light-gray, pale-brown-weathering, fine-grained sandstone cemented by ferrodolomite. A few beds of medium-dark-gray argillite, about 0.3 m thick, are present in the lower part of the formation. Greenish-gray to pale-red, very fine grained feldspathic, partly cross laminated quartzite, in beds 0.3–0.6 m thick, is interbedded with the siltite in the upper and lower parts of the formation but is rare in the middle part. Ripple marks and mud cracks are common throughout the formation. Many beds are laminated or cross laminated. Many bedding surfaces are coated with coarse detrital muscovite, and some beds contain sparse pyrite cubes. Many small deposits of secondary copper minerals are either in or near outcrops of the Apple Creek Formation, an association so consistent that it suggests that the Apple Creek Formation is the source of the copper.

The type area of the Apple Creek Formation is in the vicinity of Apple Creek, north of the Patterson quadrangle (Anderson, 1961, p. 19–21; Ruppel, 1975, p. 10). This area is now known to be even more complex, structurally, than was recognized when the formation was named and redefined, and the Yellowjacket Formation in this area includes siltites and fine-grained quartzitic rocks that in places resemble the rocks of the Apple Creek Formation. Because the type area is so confused, both stratigraphically and structurally, the most useful section for study is the principal reference section described when the formation was redefined (Ruppel, 1975, p. 9–10), which is near Golden Trout Lake in the east part of the Patterson quadrangle (Ruppel, 1980).

GUNSIGHT FORMATION

The uppermost formation of the Lemhi Group, the Gunsight Formation, is widely exposed in the central Lemhi Range and farther south along the west flank of the range. The formation is present in the Beaverhead Mountains east of Leadore and Salmon, Idaho, although it has not been mapped separately there, and near North Fork and Gibbonsville, Idaho (Lopez, 1981, pl. 1). The minimum thickness of the formation is about 1,800 m, but no complete, unfaulted sections are known. The lower contact is gradational from the Apple Creek Formation. The upper contact into the succeeding Swauger Formation is gradational through a thickness of about 45 m in which the quartzite of the Gunsight becomes medium grained and contains rounded, glassy quartz grains like those of the Swauger. In many places, however, the Swauger has been eroded away and the Gunsight is overlain with strong angularity by the Wilbert or Summerhouse Formations or by the Kinnikinic Quartzite.

The Gunsight Formation is mostly light brownish gray to grayish-red-purple, fine-grained feldspathic (5–40 percent, typically 10–20 percent) quartzite that contains abundant magnetite in rounded detrital grains scattered through the rock and concentrated in laminae. The weathered magnetite gives the rock a speckled appearance. The rock is in beds 0.3–1.2 m thick, many of which are separated by silty or muddy partings, are prominently laminated and cross laminated, and are deformed by soft-sediment slumping. About half of the rocks in the lower 60–90 m of the formation are grayish-red-purple to medium-gray siltite and argillite interbeds, as much as 0.3 m thick. Similar interbeds decrease in proportion upward in the formation, and are rare in rocks more than 300 m above the base. The quartzites in the lower 300 m mostly are silty or argillitic, but a few massive units as much as 6 m thick, of quartzite like that higher in the formation also are present in the lower 300 m.

The rocks of the Gunsight and Big Creek Formations are similar in many respects, and the two cannot always be distinguished with any confidence. In general, the Gunsight is darker, dirtier, and more limonite speckled than the Big Creek. The Gunsight also contains more siltite and argillite beds and partings and more sedimentary deformation structures such as soft-sediment slumping. The Big Creek tends to be lighter colored and clean appearing. But the distinction between the two is not always clear.

SWAUGER FORMATION

The uppermost Middle Proterozoic sedimentary rocks in the Lemhi Range are the quartzites of the Swauger Formation, which is widely exposed in the central Lemhi Range and the adjacent Donkey Hills, and crops out in what may be nearly complete sections where cut by the Salmon River northeast and southwest of Ellis, and by Morgan Creek near the west border of the Challis quadrangle (McIntyre and Hobbs, 1978). It has not been separately mapped in the Beaverhead Mountains, although some of the rocks along the Continental Divide east of Leadore are Swauger. The greatest known thickness of the formation is about 3,100 m but no unfaulted sections have been found, and in most places part, or all, of the formation was eroded before deposition of the overlying rocks. The lower contact of the formation with the underlying Gunsight Formation is gradational through a zone about 45 m thick in which rocks characteristic of both formations are interbedded. The top of the formation is not preserved in the central Lemhi Range because the rocks in the thickest known section, in the northwest part of the Patterson quadrangle, are overturned and cut off by a thrust fault. In the Lemhi Range the Swauger Formation is overlain with strong angularity by the Wilbert Formation in the southern part of the range; by the Summerhouse Formation in the west-central part of the range; and by the Kinnikinic Quartzite in the east-central part of the range. Farther west, in the northern part of the Lost River Range, the Swauger is overlain gradationally by the Lawson Creek Formation (Hobbs, 1980).

The Swauger Formation is composed of distinctive quartzites that make it fairly easy to identify, and that differ substantially from the underlying rocks. Most of the quartzite is pale purple to pale red purple, or less commonly light brown to grayish green, medium to coarse grained (typically about 0.5 mm), hematitic, and only slightly feldspathic (<10 percent) if at all. Beds are generally about 1–2 m thick and commonly are prominently and coarsely cross laminated; some are ripple marked. Quartz grains are glassy, well rounded, and well sorted; commonly some grains are amethyst. Hematite grains 0.1–1 mm in diameter weather to stain matrix and sand grains, giving the rock a distinctive blotchy appearance. The Swauger quartzites thus differ from the older quartzites in having a purplish color; coarser, rounded, glassy grains; almost no feldspar or black heavy minerals; and purplish hematitic blotching and speckling.

LAWSON CREEK AND WILBERT FORMATIONS

The Lawson Creek and Wilbert Formations are not present above the Swauger Formation in the central or northern parts of the Lemhi Range, but they are preserved in adjacent areas.

The Lawson Creek Formation (Hobbs, 1980) gradationally and conformably overlies the Swauger Formation in the northern part of the Lost River Range (McIntyre and Hobbs, 1978), west of the Lemhi Range, but has not yet been recognized elsewhere. The formation is a heterogeneous sequence of reddish-purple to purplish-gray and maroon quartzite, impure quartzite, siltstone, and argillite, and is mostly thin bedded although it locally includes thick beds of quartzite much like that of the Swauger Formation. The greatest exposed thickness of the formation is more than 1,200 m, but the top is concealed by Challis Volcanics, and the total thickness is unknown. The formation is the uppermost Middle Proterozoic formation known in east-central Idaho.

The Wilbert Formation (Ruppel and others, 1975, p. 27–29) is exposed in the southern parts of the Lemhi Range and Beaverhead Mountains, and probably also is present on the west-central flank of the Lost River Range (Baldwin, 1951, p. 887–889; Ross, 1947, p. 1099–1102; 1961a, p. 198) and in the northern part of the Lost River Range, where its relation to the Lawson Creek Formation is obscured by younger volcanic rocks (McIntyre and Hobbs, 1978; Hobbs, 1980, p. 3). The Wilbert Formation overlies the Swauger and Gunsight Formations with angular unconformity. Originally the Wilbert was tentatively assigned a Late Proterozoic age (Ruppel and others, 1975, p. 28). Derstler and McCandless (1981) and McCandless (1982, p. 30) later reported fossils that indicate that at least the uppermost 35 m of the formation in the southernmost part of the Lemhi Range is of middle Early Cambrian age. The age of the lower part of the formation, in which no fossils have been found, remains uncertain but is tentatively considered to be Late Proterozoic(?) and perhaps partly Cambrian.

The rocks in the Wilbert Formation, and those that overlie the Wilbert and underlie the Kinnikinic Formation of Middle Ordovician age, have been the subject of controversy and confusion since they were first described by Umpleby (1917, p. 23; Ruppel and others, 1975, p. 25–27). Recently, McCandless (1982) redescribed the sequence, and suggested that the Wilbert is overlain by a previously unrecognized group of rocks at Tyler Peak. The formation of Tyler Peak, of late Early Cambrian age, in turn is overlain by the Summerhouse Formation of Early Ordovician age (Ruppel and others, 1975, p. 29–31). The Wilbert Formation, as revised by McCandless (1982), overlies Middle Proterozoic rocks, the Gunsight and Swauger Formations, with angular unconformity, and is conformably, and in places gradationally, overlain by the formation of Tyler Peak (McCandless, 1982).

The Wilbert Formation is about 120 m thick in the south part of the Lemhi Range, thinning northward to pinch out about 30–35 km south of the Gilmore quadrangle, although the north edge of the formation is only approximately known and has not been mapped. The northward thinning and disappearance is a result of both onlap onto the Late Proterozoic erosion surface, the Lemhi arch, and early Paleozoic erosion. The formation consists mainly of light-gray to brownish-gray and pale-red, fine- to coarse-grained quartzitic sandstone, grit, and conglomerate. McCandless (1982, p. 27) stated that about half of the strata included in the formation are fine to very fine grained, well sorted, and unimodal, or medium to coarse grained, moderately well sorted to well sorted, and bimodal, and about half of the strata are medium grained to conglomeratic, and moderately to poorly sorted. Many beds are laminated or cross laminated. The hematitic, glauconitic, and partly calcareous beds originally included in the Wilbert (Ruppel and others, 1975, p. 27) were apparently assigned to the overlying formation of Tyler Peak by McCandless (1982, p. 69–84), perhaps partly explaining the differences in reported maximum thickness of the Wilbert—about 120 m (McCandless, 1982, p. 26), as contrasted to about 300 m (Ruppel and others, 1975, p. 27).

In other areas in east-central Idaho and southwest Montana, some strata previously considered equivalent to the Wilbert Formation are now known to be older, and part of the Missoula Group of the Belt Supergroup, most of them as part of the Mount Shields Formation. These include conglomeratic rocks in the north part of the Beaverhead Mountains (MacKenzie, 1949, p. 13; Tucker, 1975, p. 64–86; Ruppel, 1975, p. 19; Ruppel and others, 1983), and conglomeratic rocks in depositional contact with Archean crystalline metamorphic rocks near Maiden Peak in the Beaverhead Mountains in southwest Montana (M'Gonigle, 1965, p. 18–19; McCandless, 1982, p. 99–100).

REGIONAL RELATIONS AND CORRELATION

Proterozoic rocks of east-central Idaho are mostly different from those of the Belt Supergroup in west-central Montana, and were deposited in the Belt miogeocline rather than in the Belt basin (Ruppel, 1975, p. 5, 18; Ruppel and Lopez, 1984). Because of the differences in lithology and depositional setting, direct correlations with formations included in the Belt Supergroup are not possible. Nonetheless, the Proterozoic rocks of the two regions are thought to be more or less equivalent in age.

The Middle Proterozoic rocks of east-central Idaho are all in or beneath the Medicine Lodge thrust plate, and extend only a short distance, if at all, east of the Idaho-Montana State line. The Yellowjacket Formation, which is widely exposed in central and eastern Idaho (Lopez, 1981) is considered autochthonous—the only autochthonous Proterozoic formation of the region. The Yellowjacket is cut off on the east by Proterozoic faults (Ruppel and Lopez, 1984, p. 23-26) and has not been found in southwest Montana. The Lemhi Group, Swauger Formation, Lawson Creek Formation, and Wilbert Formation are all confined to the Medicine Lodge thrust plate. These units are not known east of the leading edge of the thrust plate, where Proterozoic rocks in the underlying Grasshopper thrust plate are clearly part of the Missoula Group of the Belt Supergroup (Winston, 1978; Ruppel and Lopez, 1984, p. 17-20).

South of the Grasshopper thrust plate, Archean crystalline metamorphic rocks directly underlie Paleozoic and Mesozoic rocks. Proterozoic rocks, if they ever overlapped this region of cratonic crystalline rocks in southwest Montana, appear to have been removed almost completely by erosion before the Paleozoic rocks were deposited. The only known remnants of Proterozoic rocks in depositional contact with crystalline metamorphic rocks south of the Horse Prairie fault zone, which formed the south edge of the Belt embayment (Scholten, 1981; Ruppel and Lopez, 1984, p. 13), are arkosic conglomerate and coarse-grained sandstone in a small area between Medicine Lodge Creek and Maiden Peak (M'Gonigle, 1965, p. 15-19). The conglomerate and sandstone seem to have been deposited in an estuary or river valley tributary to the Belt embayment about 20 km to the north, and to be a nearshore facies of the Missoula Group immature arkosic and quartzitic sandstones along the south edge of the Belt embayment.

The Proterozoic rocks of east-central Idaho cannot be correlated directly with rocks of presumably similar age in western Montana, but comparison of events that have affected both groups of rocks suggested an earlier tentative correlation that still seems likely (Ruppel, 1975, p. 18; Hobbs, 1980, p. 10-12) (table 2).

TABLE 2.—*Tentative regional correlations of Middle Proterozoic sedimentary rocks*

[Modified from Ruppel (1975) and Hobbs (1980)]

	East-central Idaho	Western Montana-northern Idaho
	Middle Proterozoic Miogeocline	Belt Basin
		Belt Supergroup
	Lawson Creek Formation (Missing in Lemhi Range) Swauger Formation	Missoula Group
Lemhi Group	Gunsight Formation Apple Creek Formation	Helena and Wallace Formations
Lemhi Group	Big Creek Formation West Fork Formation Inyo Creek Formation	Ravalli Group
	Medicine Lodge Thrust	
	Yellowjacket Formation	Prichard Formation

Paleomagnetic studies by Elston and Bressler (1980) of the lower part of the Swauger Formation exposed in Falls Creek, north of Patterson (Ruppel, 1980); of the middle(?) Swauger exposed along the Salmon River north of Ellis, Idaho; and of the uppermost Swauger Formation and lower Lawson Creek Formation in the northern part of the Lost River Range (Hobbs, 1980, p. 8-9) suggest that these rocks exhibit both normal and reversed polarities like those of the Missoula Group of the Belt Supergroup, and so suggest correlation of the Swauger and Lawson Creek Formations with the Missoula Group. On the basis of the paleomagnetic data of Elston and Bressler and of stratigraphic comparisons, Hobbs (1980, p. 12) suggested that the upper Swauger and the Lawson Creek correlate most closely with the upper Mount Shields, Bonner, and lower McNamara Formations of the Missoula Group, and that they may reflect deposition in proximate or interconnected basins in latest Middle Proterozoic time.

The Wilbert Formation had been correlated tentatively with the Windermere System of Canada, and with part of the Mutual Formation of southeast Idaho and the Brigham Group of northeast Utah (Ruppel and others, 1975, p. 28–29). At least the upper part of the formation is now known to be Early Cambrian, and perhaps at least that part of the formation is more directly related to the Early or Middle Cambrian Cash Creek Quartzite of central Idaho (Hobbs and others, 1968, p. J18–J19).

PALEOZOIC ROCKS

Most of the central part of the Lemhi Range is underlain by Proterozoic rocks, but Paleozoic rocks are exposed in much of the Leadore quadrangle, particularly in the Beaverhead Mountains north of Leadore, in a few places near the eastern margin of the Patterson quadrangle, and in the eastern and central parts of the Gilmore quadrangle (pl. 1). Paleozoic rocks are much more widespread in the southern parts of the Lemhi Range and in the adjacent Beaverhead Mountains and Lost River Range. In the central Lemhi Range, Paleozoic rocks are mostly of Ordovician to Devonian age, in contrast to adjacent areas where mainly younger Paleozoic rocks are exposed.

Paleozoic rocks in the central part of the Lemhi Range have a total thickness of about 1,500–1,600 m, and are divided into 14 formations (fig. 3). Although most of these rocks are limestone or dolomite, the most conspicuous formation is the nearly white, vitreous Kinnikinic Quartzite. The oldest strata, of Early Ordovician age, are overlain by formations that represent every system of the Paleozoic but significant hiatuses are present, and no system is completely represented.

ORDOVICIAN AND ORDOVICIAN-SILURIAN

Ordovician and Ordovician-Silurian rocks in the central part of the Lemhi Range are divided into three formations, the Summerhouse Formation, Kinnikinic Quartzite, and Saturday Mountain Formation. The Summerhouse Formation is a recently defined unit that is largely, perhaps entirely, restricted to the Lemhi and Lost River Ranges, and which includes carbonate rocks that contain a distinctive fauna of Early Ordovician age (Ruppel and others, 1975, p. 30–33).

The Kinnikinic Quartzite and Saturday Mountain Formation were named and described first in central Idaho (Ross, 1934b, p. 947; 1937, p. 15–17); the formations later were extended into east-central Idaho by Ross (1947, 1961a) and Anderson (1961, p. 25–26). The Kinnikinic had been recognized earlier, but not named, in the Lost River, Lemhi, and Beaverhead Ranges (Umpleby, 1913, p. 32; 1917, p. 24; Shenon, 1928, p. 5; Kirkham, 1927, p. 18), although in some of these areas it had been assigned tentatively to the Cambrian(?). Ross (1934b, p. 950) dated the quartzite as Middle Ordovician on the basis of its position between the Lower Ordovician Ramshorn Slate and the Upper Ordovician Saturday Mountain Formation. Ross (1947, p. 1104) later reported that fossils of early Late Ordovician age had been collected from the upper part of the Kinnikinic Quartzite in the Lemhi Range at Meadow Lake in the northeast part of the Gilmore quadrangle, and (1961a, p. 203) that fossils of Middle or Late Ordovician age occurred in the formation at the southern end of the range. When the Gilmore quadrangle was mapped, the Meadow Lake fossil locality was revisited, and the fossils reported by Ross were found to have come from faulted, sandy dolomite in the lower part of the overlying Saturday Mountain Formation. The stratigraphic position of the fossils reported from the Kinnikinic in the southern part of the Lemhi Range is also in doubt. As a result, the formation in the Lemhi Range is not known to contain any undoubted fossils other than *Scolithus* (James and Oaks, 1977, p. 1492). Its age clearly is Middle Ordovician as Ross concluded (1934b, p. 950), however, because it overlies the Early Ordovician Summerhouse Formation with slight angular unconformity and is conformably overlain by the Middle and Late Ordovician and Early Silurian Saturday Mountain Formation. In addition, fossils of Middle Ordovician age have been found in the formation elsewhere in central Idaho (Hobbs and others, 1968, p. F11; Oaks and James, 1980, p. 6).

Hobbs and others (1968, p. J11) redefined and subdivided the Kinnikinic in its type locality in central Idaho, and concluded that its age falls between the lower and upper parts of the Middle Ordovician. They pointed out that the redefined Kinnikinic of the type locality closely resembles, both in lithology and age, the quartzites that had been called Kinnikinic in other areas, including the Lemhi Range. As a result, the Kinnikinic of east-central Idaho now conforms closely to the redefined and restricted Kinnikinic of the type locality. Regional stratigraphic relations, lithology, and petrology of the Kinnikinic Quartzite have been discussed recently by James and Oaks (1977) and Oaks and James (1980).

The gray dolomite of the Saturday Mountain Formation in east-central Idaho differs from the type Saturday Mountain in Custer County, Idaho, which is dominantly shaly, argillaceous and carbonaceous dolomite and magnesian limestone (Ross, 1937, p. 18–22; 1962a, p. 31–34). Because of the lithologic differences, Sloss (1954, p. 365) extended the name Fish Haven Dolomite from southeastern Idaho, to include the dominantly dolomitic rocks of Ordovician age in east-central Idaho. Churkin (1962, p. 579) suggested restricting the Saturday Mountain Formation to its type locality and extending the Fish Haven into east-central Idaho to emphasize the different lithologies of Ordovician rocks in the two regions. Churkin (1962, p. 576) also named the sandy and argillaceous rocks at the base of his Fish Haven Dolomite the Lost River Member of the formation, and noted (p. 579) the possible time equivalency of the lower Fish Haven of east-central Idaho and the shaly lower Saturday Mountain Formation of central Idaho. The name Fish Haven has been used subsequently in the southern parts of the Lemhi and Lost River Ranges (Beutner, 1968; Skipp and Hait, 1977), and in the central Lemhi Range (Hait, 1965), even though C.P. Ross (1961a, p. 208) pointed out that the lithologic changes in Ordovician carbonate rocks between the type locality of the Saturday Mountain Formation and the southern part of the Lemhi Range are gradual, and largely a variation in proportions of components, and that the two sequences are closely related. Ross (1962a, p. 34) also suggested that a local name would be more meaningful, at least until regional stratigraphic relations are better known, and so he retained the name Saturday Mountain in the Lemhi Range. R.J. Ross, Jr. (1959) continued the use of the name Saturday Mountain Formation in the southern part of the Lemhi Range, in a definitive study of the brachiopod fauna in the formation that also discussed regional stratigraphic relations.

Later studies in the central parts of the Lost River and Lemhi Ranges have consistently used Saturday Mountain Formation in preference to Fish Haven Formation (Mapel and others, 1965; Mapel and Shropshire, 1973; Ruppel, 1968; Ruppel, 1978; Ruppel, 1980; Ruppel and Lopez, 1981). We consider the Ordovician and Silurian dolomitic rocks between the Kinnikinic Quartzite and the Silurian Laketown Dolomite in east-central Idaho to be an eastern, carbonate facies of the Saturday Mountain Formation. Our reasons are the same as those of C.P. Ross (1961a; 1962a): these rocks are most closely related to those of the Saturday Mountain Formation in the Bayhorse quadrangle in central Idaho and occupy the same stratigraphic position; the lithologic changes suggest gradually changing proportions of components although regional stratigraphic relations remain uncertain because of major thrust faulting; and the relation of the Ordovician and Silurian dolomites of central Idaho to the Fish Haven Dolomite of southeast Idaho is unknown, other than that the rocks are partly temporal equivalents. Use of a local name, the Saturday Mountain Formation, seems less likely to introduce another element of confusion. The name Saturday Mountain Formation has priority, because Ross (1934b, p. 956) extended it into the Lemhi Range, on what seem to be reasonable stratigraphic grounds, long before the name Fish Haven was introduced in the region. We adopt the Middle Ordovician Lost River Member, described by Churkin (1962, p. 576), but remove it from the Fish Haven and assign it to the Saturday Mountain Formation as its basal member.

Ross (1934b, p. 952-956; 1947, p. 1104-1105) originally concluded that the Saturday Mountain Formation is of Late Ordovician age. Later studies have shown, however, that the lowest part of the formation in the type locality is of late Middle Ordovician age (Hobbs and others, 1968, p. J11-J12) and that the upper part of the formation in some areas includes Silurian fossils (C.P. Ross, 1962a, p. 33-34; R.J. Ross, Jr., 1959, p. 444). Mapel and Shropshire (1973) also reported the presence of *Halysites* sp., indicating a Silurian age, in the upper 30 m of the formation at Hawley Mountain in the Lost River Range to the west.

Fossils from the Saturday Mountain Formation in the central Lemhi Range are mainly of Late Ordovician age; the brachiopod fauna resembles that of the southern part of the Lemhi Range (R.J. Ross, Jr., 1959, p. 444-445), and has more Late than Middle Ordovician affinities; the coral fauna is clearly Late Ordovician. The upper 140-150 m of the formation, however, includes *Halysites* sp., and is of Early Silurian age. The Saturday Mountain Formation in the Lemhi and Lost River Ranges thus is mainly of Late Ordovician age, but includes late Middle Ordovician strata in its lower part and 140-150 m of Early Silurian strata at its top in the central and northern parts of these ranges.

SUMMERHOUSE FORMATION

The Lower Ordovician Summerhouse Formation unconformably overlies the Middle Proterozoic Swauger and Gunsight Formations in the south part of the Gilmore quadrangle (Ruppel and Lopez, 1981), but is not present farther north where the basal Paleozoic rocks are those of the Kinnikinic Quartzite. The type section of the formation is in Summerhouse Canyon in the southwest part of the Gilmore quadrangle (Ruppel and others, 1975, p. 29-32). The formation extends southward along the west flank of the Lemhi Range to the Snake River Plain, and is exposed farther west in the Donkey Hills and in the northern part of the Lost River Range (McIntyre and Hobbs, 1978). Strata described on the west side of Hawley Mountain by Mapel and Shropshire (1973), and at Elkhorn Ranch by Ross (1947, p. 1099-1102) and Baldwin (1951, p. 887-889) probably are equivalent to the Summerhouse Formation.

The Summerhouse Formation differs greatly in thickness from place to place. The formation is about 210 m thick in the southernmost part of the Lemhi Range; at the type section it is about 305 m thick. Farther east, in the southeast corner of the Gilmore quadrangle, it is 50–60 m thick, but thickens northward to about 120 m near Warm Creek. North of Warm Creek, it thins to 0 m near Squaw Creek, and only a few beds, perhaps equivalent to the uppermost part of the formation, are known farther north. The thickness changes are due partly to depositional onlap onto an irregular erosion surface, the Lemhi arch, and partly to erosion before deposition of the overlying Kinnikinic Quartzite (Ross, 1961a, p. 200; Ruppel and others, 1975, p. 29–31; James and Oaks, 1977, p. 1496–1498).

The type Summerhouse includes about 20 m of light-gray to yellowish-gray, fine to very fine grained, vitreous quartzite at its base, overlain successively by about 25 m of light-olive-gray to medium-gray, sandy, locally pebbly fossiliferous limestone and dolomite; by about 150 m of irregularly interbedded, pale-red to nearly white, fine- to medium-grained, partly calcareous sandstone and quartzite that contains interbedded dolomite in its lower part; by about 50 m of yellowish-brown to light-gray, fine-grained, slightly calcareous, partly dolomitic, partly glauconitic, *Scolithus*-bearing sandstone and quartzite; and by about 60 m of pale red to light gray *Scolithus*-bearing sandstone. Because the formation was deposited in a near-shore environment, its composition differs from place to place. In the southeast corner of the Gilmore quadrangle, the formation consists of a basal yellowish-gray to grayish-red, medium-grained, feldspathic quartzite, about 5–15 m thick, that is composed of poorly sorted and poorly rounded sand grains; about 3–5 m of yellowish-brown to brownish-gray dolomite and dolomitic sandstone; 5–15 m of light-gray to pale-pinkish-gray, medium-grained quartzite composed of moderately well rounded and sorted quartz grains, which resembles the Kinnikinic Quartzite, but is not vitreous; 20–30 m of yellowish-brown to brownish-gray, medium-grained dolomitic, commonly glauconitic sandstone and sandy dolomite that is partly *Scolithus*-bearing; and an uppermost thin and discontinuous unit of pale-red to light-gray, *Scolithus*-bearing sandstone. Despite the lateral differences, the formation is characterized by a basal, poorly sorted sandstone overlain by limestone or dolomite, pinkish-gray and white quartzite, and an upper, *Scolithus*-bearing sandstone. Most of the sandstone above the basal zone is composed of moderately to well rounded and sorted quartz grains, and is in beds 0.2–0.8 m thick. The light-colored sandstone at the top of the formation reaches farther north than the rest of the formation, to Meadow Lake in the northeast part of the Gilmore quadrangle. James and Oaks (1977, p. 1494–1498) suggested that these beds could represent either onlap of the Summerhouse Formation onto an older surface or a basal transgressive phase of the overlying Kinnikinic.

KINNIKINIC QUARTZITE

The Middle Ordovician Kinnikinic Quartzite, a distinctive unit of white or light-gray quartzite, is widely and prominently exposed in the east-central Lemhi Range; in most places, it is the basal Paleozoic formation and overlies Proterozoic rocks with angular unconformity. It extends south along the west side of the Lemhi Range to the Snake River Plain, but is preserved in only a few places in the northern part of the range, mainly along and west of the range crest. Its northernmost known outcrops are near Rattlesnake Creek, west of the north end of the Lemhi Range, about 30 km south of Salmon, Idaho (Landreth, 1964, p. 16). It is preserved in a few thrust slices along the west flank of the Beaverhead Mountains, north and east of Leadore, Idaho (Ruppel, 1968; Lucchitta, 1966), but is absent a short distance farther south. It reappears in the southern part of the Beaverhead Mountains where it thins southward, again to disappear, against the Lemhi arch (Sloss, 1954; Scholten, 1957, p. 159; Scholten and Ramspott, 1968, pl. 1; Ruppel, 1978, p. 9, 12; 1985). The formation is more widely exposed in the Lost River Range west of the Lemhi Range, than in the Beaverheads, in discontinuous fault blocks in the Donkey Hills, at Hawley Mountain, and elsewhere in the eastern and central parts of the range (Mapel and Shropshire, 1973; Mapel and others, 1965; Ross, 1947; McIntyre and Hobbs, 1978).

The Kinnikinic Quartzite is 300–400 m thick in the central Lemhi Range, but thins southward to about 150–200 m in the southern Lemhi Range (Ross, 1961a, p. 201–203; James and Oaks, 1977, p. 1497), and eastward to 0 m in the Beaverhead Mountains, reflecting shoaling against the west flank of the Lemhi arch. It thickens westward in part of the Lost River Range, to about 600 m (Mapel and Shropshire, 1973; Mapel and others, 1965), but thins farther west and south within that range. Regional thickness patterns of the formation have been discussed by James and Oaks (1977), who suggested that the unit was deposited on a shallow, open-marine shelf undergoing differential subsidence, which, combined with irregularities on the pre-Kinnikinic erosion surface, led to the lateral differences in thickness.

The Kinnikinic Quartzite is a light-gray to white, fine- to medium-grained, vitreous quartzite, commonly in beds 0.3–1.5 m thick or massive, although fracturing and shearing make bedding features impossible to recognize in many exposures. Discontinuous interbeds of white quartzite mottled with irregular lenses of reddish-brown to reddish-orange sandstone cemented with ferroan dolomite are locally present in the formation. The finely cross laminated lenses of sandstone commonly are from a few millimeters to a few centimeters long and as much as a centimeter thick. Locally the lenses may be as much as a meter long and 15 cm thick; north of Sheephorn Peak, in the Leadore quadrangle, one exceptional lens is about 10 m long and 2 m thick. Typically though, the Kinnikinic is nearly white vitreous quartzite, composed of subangular to subrounded, well-sorted, fine to medium grains of quartz sand cemented by silica. The formation is without known fossils other than sparse fucoidal markings and *Scolithus* (James and Oaks, 1977, p. 1492) in most of east central Idaho, but has yielded a fauna of Middle Ordovician age in the Bayhorse area in central Idaho (Oaks and James, 1980, p. 6).

The contacts of the Kinnikinic with underlying and overlying formations are sharp and distinct. The basal part of the Kinnikinic commonly contains sand derived from the underlying rocks in a zone from a few centimeters to a few meters thick, and in a few places includes a thin zone of sparse pebbles at the base. The upper contact is at the abrupt change from nearly white, vitreous quartzite to dark shale and quartzite or dolomite of the lower part of the Saturday Mountain Formation.

SATURDAY MOUNTAIN FORMATION

The Saturday Mountain Formation, which conformably overlies the Kinnikinic Quartzite, is widely exposed in the central and southern parts of the Lemhi Range, where it commonly forms dark-colored, stepped cliffs above the white cliffs of the Kinnikinic. The formation also is present in the Lost River Range to the west, in exposures that parallel those of the Kinnikinic. In the Beaverhead Mountains, only two small areas of Saturday Mountain Formation are known, both of them part of an imbricate thrust slice at the mouth of Railroad Canyon, east of Leadore (Ruppel, 1968). Elsewhere in this range, the formation is absent as a result of onlap and depositional thinning against the west flank of the Lemhi arch (Scholten, 1957, p. 159; Scholten and Ramspott, 1968).

The Saturday Mountain Formation is about 370 m thick on the east face of Rocky Peak in the southwest part of the Leadore quadrangle (measured section 1), and seems to maintain this thickness throughout the central parts of the Lemhi and the Lost River Ranges (Ruppel, 1968; Mapel and Shropshire, 1973; Mapel and others, 1965). The formation thins southward, and is about 85 m thick at the south end of the Lemhi Range (Ross, 1961a, p. 204; 1962a, p. 31–34), where it includes sandstone interbeds, lenticular channel fills, and bedding partings that are not present in the formation farther north. It thins eastward to 0 m in the Beaverhead Mountains. The sandstones in the formation in the southern Lemhi Range suggest nearness to a shoreline, and the regional thinning relations suggest eastward shoaling against the western shore of the Lemhi arch.

The basal Lost River Member of the Saturday Mountain Formation is a clastic unit about 17 m thick in the measured section at Rocky Peak, but thins to the east and southeast. It is only 4–5 m thick at Meadow Lake, about 16 km southeast of Rocky Peak, and is absent 0.5 km farther east, and also absent on Leadore Hill, about 10 km east of Rocky Peak. At Rocky Peak, the member consists of interbedded sandstone, quartzite, shale and mudstone. The sandstone is pale brown to medium dark gray, medium grained, partly argillaceous or silty, partly dolomitic, and is composed of well-rounded and well-sorted grains of quartz sand. Some beds contain moderately abundant chips of shale and mudstone, angular to rounded quartzite pebbles as much as 3 cm in diameter, and a small percentage of detrital magnetite grains altered to limonite. Most of the sandstone is in beds a few centimeters to 15 cm thick, or in irregular lenses a few centimeters thick; some beds and lenses are laminated or cross laminated in the upper part of the member. Quartzite is subordinate to the other rocks in the member, but conspicuous because of its resistance to weathering. Typically it is medium gray with a distinct bluish tint, fine to medium grained, composed of well-sorted and well-rounded quartz grains, in beds as much as 1 m thick and in lenses as much as 10 cm thick and about a meter long, and partly thinly cross laminated. Shale and mudstone form most of the lower one-third of the member, as well as interbeds and bedding partings in the sandstone above. These argillaceous rocks are pale red to greenish gray and dark gray, fissile and papery to blocky fracturing, and mostly in beds a few centimeters to 0.8 m thick, separated by thin beds of sandstone and quartzite. At Meadow Lake, the shale and mudstone are absent, and the member consists of a basal single bed, about 2 m thick, of bluish-tinted medium-light-gray, fine-grained, clean quartzite, overlain by 2–3 m of light-olive-gray, fine-grained dolomitic sandstone. The sand grains in both the quartzite and sandstone are well-sorted and well-rounded, frosted, clear quartz; similar sand grains are common in the dolomite that immediately overlies the sandstone.

The Lost River Member is overlain gradationally by dolomite that forms the main part of the Saturday Mountain Formation. The basal part of the dolomite is a distinctive thin-bedded sequence of yellowish-gray to medium-gray, partly fossiliferous, finely crystalline dolomite that is characterized by irregular, interlaced networks of hairlike wisps, stringers, and veinlets of white dolomite 0.5–2 mm thick, spaced 5–15 mm apart, and mostly about parallel to bedding. The upper and lower parts of this unit contain abundant black chert in irregular nodules 1-2 cm thick and 1-8 cm long. The thickness of the wispy dolomite differs from place to place, from 15 to 50 m; most commonly it is about 20-30 m thick and forms a definitive key unit in the basal Saturday Mountain Formation above the clastic rocks of the Lost River Member.

The wispy dolomite grades upward into medium-dark-gray to medium-gray, finely crystalline dolomite, about 35 m thick, that forms a prominent dark unit in the lighter colored rocks that make up most of the formation. This dark dolomite is faintly to distinctly mottled lighter shades of gray on weathered surfaces, and is in beds 0.3-1.5 m thick. Much of the dolomite contains black chert in irregular nodules 2-4 cm thick and as much as 10 cm long. Many beds contain abundant fossils and fossil fragments. The upper part of the unit includes beds and lenses of dolomite sandstone, typically about 0.3 m thick. The dark dolomite so closely resembles the Devonian Jefferson Formation that in isolated outcrops they cannot always be distinguished with certainty.

Above the dark dolomite unit, the rest of the formation is medium-gray to medium-light-gray dolomite, which is finely crystalline, mainly thick bedded to massive but with some beds 10-60 cm thick, and mottled lighter shades of gray on weathered surfaces. Many beds contain abundant fossils and fossil fragments. Some beds contain black chert nodules although chert is much less abundant than in the underlying dark-colored dolomites and is absent or very rare in the upper half of the formation. The uppermost 140-150 m of the formation, which contains corals of Silurian age, is lithologically similar to underlying dolomites that contain Ordovician fossils; the rocks of different ages are not separable. In the southeast part of the Gilmore quadrangle, the upper part of the formation includes sparse thin interbeds of yellowish-gray to yellowish-brown, fine-grained dolomitic sandstone that contains moderately abundant euhedral crystals of pyrite as much as 5 mm in diameter.

The top surface of the Saturday Mountain Formation, beneath the rocks of the overlying Laketown Dolomite of Middle Silurian age, is a deeply weathered, irregular erosional unconformity. The unconformity is well exposed in outcrops near the mouth of Nez Perce Creek in the southern part of the Leadore quadrangle, where the upper 3-4 m of the Saturday Mountain is a weathered and bleached light-gray to pinkish- or yellowish-gray solution breccia made up of subrounded fragments of dolomite 3-7 cm in diameter in a sparse matrix of dolomitic sand and mud, with abundant solution cavities. The gently rounded mounds and hills on the erosion surface are from 2 or 3 m to as much as 30 m high. This Early Silurian episode of uplift and erosion may have occurred at about the same time as the emplacement of granitic magmas in central Idaho, from 430 to 500 million years ago (Ruppel, 1968; Scholten and Ramspott, 1968, p. 21; Ruppel, 1978, p. 18; Evans and Zartman, 1981).

The Saturday Mountain Formation in the central Lemhi Range contains a rich fauna that has been studied by R.J. Ross, Jr. and W.A. Oliver, Jr. (written commun., 1961, 1962, 1963). Many of the fossil collections described below by Ross and Oliver were collected by Ross from Rocky Peak (measured section 1) in the west part of the Leadore quadrangle; these collections are keyed by number to the measured section. Other collections are from near Meadow Lake in the Gilmore quadrangle and from the southern part of the Lemhi Range.

USGS fossil locality D-1041-CO about 4.3 m above base of wispy dolomite, and above upper contact of Lost River Member of Saturday Mountain Formation, unit 23 of measured section 1 (Rocky Peak). Identified by R.J. Ross, Jr.

Lordorthis variabilis Ross
Sowerbyella (large species)
Zygospira cf. *A. circularis* Cooper
rhynchonellid fragments (indeterminate)
bryozoans (indeterminate), two genera

USGS fossil locality D-1042-CO about 18.7 m above base of wispy dolomite and above upper contact of Lost River Member, unit 23 of measured section 1 (Rocky Peak). Identified by R.J. Ross, Jr.

Platystrophia sp.
Lordorthys variabilis Ross
Plaesiomys subquadratus var. *idahoensis* Ross
Austinella two spp.
Therodonta sp.
Paucicrura? sp.
Diceromyonia sp.
Zygospira recurvirostris Hall
Z. cf. *Z. sequvalvia* Twenhofel
Z. cf. *Z. lebanonensis* Cooper
Calliops sp.
trilobite fragments, incomplete

Ross noted (written commun., 1963) that the brachiopod fauna of the two collections above is almost the same as that from the Saturday Mountain Formation in the southern part of the Lemhi Range (R.J. Ross, Jr., 1959, p. 445), which he stated had more Late than Middle Ordovician affinities.

USGS fossil locality D-1043-CO, in the middle part of the Saturday Mountain Formation, unit 41 of measured section 1 (Rocky Peak). Identified by W.A. Oliver, Jr.

Catenipora sp.
favositid coral
streptalasmatoid horn corals, including *Bighornia* sp.
Age, Late Ordovician

USGS fossil localities D-1044-CO and D-1045-CO, about 145-150 m below the top of the Saturday Mountain Formation, unit 45 of measured section 1 (Rocky Peak). Collection D-1044-CO is from lower part of unit, and collection D-1045-CO is from the upper part. Identified by W.A. Oliver, Jr.

D-1044-CO
Halysites sp.
D-1045-CO
favositid coral
Halysites sp.
syringoporoid coral
phaceloid rugose coral, possibly *Palaeophyllum* sp.
Age, Silurian, both collections

USGS fossil locality D-964, 30-45 m above base of Saturday Mountain Formation on ridgecrest southeast of Meadow Lake, Gilmore quadrangle. Identified by W.A. Oliver, Jr.

labechiid stromatoporoid?
Catenipora sp. cf. *C. rubra*
Favistella? sp.
Grewinkia? sp.
Streptelasma sp. cf. *S. angulatum*
Lobocorallum sp. cf. *L. trilobatum*
phacelloid rugose coral

Oliver noted (written commun., 1961) that this fauna is of Late Ordovician age and virtually the same as the fauna of the Bighorn Dolomite. Other fossils from higher in the Saturday Mountain Formation at this same locality, also identified by Oliver, included:

Catenipora sp.
Catenipora sp. cf. *C. rubra* type
Tollina? sp.
small streptelasmoid corals
including *Streptelasma* sp.
cf. *S. prolongatum*
Aulacera sp. (several species)
Age, Late Ordovician

The genus *Aulacera*, noted in the fauna from the Saturday Mountain Formation near Meadow Lake, is found in the middle part of the formation throughout the central Lemhi Range. Oliver (written commun., 1961) described another specimen of the genus, collected on Rocky Peak, as follows:

> The concavely hexagonal, prismatic object is very poorly preserved and its core is completely destroyed, if, indeed, it had one. The remaining structure is apparently stromatoporoidal, but can be fitted to known forms only on the assumption of a core, now destroyed.
>
> The specimen probably represents a new species of some labechiid genus, most likely *Aulacera*. In gross character, the nearest thing to the specimen at hand is *A. radiata* which is well illustrated by Galloway and St. Jean (Bull. Amer. Pal., no. 194, pl. 12, fig. 4) and by Shimer and Shrock (pl. 19, figs. 16, 19 and 20), as *A. undulata*. If, in these illustrations, the coarse and fine vesicular tissue were destroyed, and the number of corners reduced to six, the result would approximate your specimen.
>
> I suspect then, that you are dealing with a species of *Aulacera*, which may be of some zonal significance. The genus, and all stromatoporoids of this general type, are limited to the Ordovician.

Fossils of long, orthoceraconic nautiloid cephalopods, including some shell fragments more than a meter long, are locally present in the lower part of the Saturday Mountain Formation. The siphuncle of one of these, collected in the wispy dolomite about 3 m above the contact with the Lost River Member on the ridgecrest southeast of Meadow Lake in the Gilmore quadrangle, is described by Rousseau Flower (written commun., 1961) as follows:

> This siphuncle has the structure common to two genera, *Williamsoceras*, known only from Whiterock beds, and *Perkinsoceras* of the Chazy, which differs in that the apical part of the siphuncle is swollen as in "Nanno". As the apex of the siphuncle is not here, certain generic assignment is not possible, but the pattern is more like that of known *Williamsoceras* than like any *Perkinsoceras* known. The siphuncle attains twice the diameter of any previously known species of either of these genera.
>
> Either this is a first occurrence of this type of endoceroid above the Whiterock-Chazy interval, or else it indicates that in the Saturday Mountain dolomite there are some beds of Whiterock or, at the earliest, Chazy age. Chazy age is unconvincing; no typical Chazy is known in the western U.S., but the inclusion of some Whiterock beds with materially younger beds (Red River or Richmond) is perfectly possible. This form is being described as *Williamsoceras*? *gilmorense*.

In the course of this study, new collections of fossiliferous rocks from the Saturday Mountain Formation were made near the mouth of Black Canyon at the south end of the Lemhi Range, an area where C.P. Ross (1961a, p. 205-206) had earlier described channel sandstone units interbedded with dolomite in the upper part of the formation, and where R.J. Ross, Jr. (1959) described the brachiopod fauna from the lower part of the formation. The new collections were made by R.J. Ross, Jr., and L.A. Wilson from dolomite between two quartzitic channel sandstone beds in the upper part of the formation, and yielded the following conodonts (J.W. Huddle, written commun., 1968; revised by A.G. Harris, 1987):

USGS fossil locality 6276-CO
17.7 m above lower brown quartzitic sandstone in upper Saturday Mountain Formation, Black Canyon, southern Lemhi Range, Idaho.
1 *Acontiodus* sp. s.f. element
9 rastrate elements of *Belodina* sp.

5 *Drepanoistodus suberectus* (Branson and Mehl) elements
1 Pb element of *Oulodus* sp.
14 *Panderodus gracilis* (Branson and Mehl) elements
Plectodina florida Sweet
2 Pb, 1 M, 3 Sa, and 4 Sc elements

The presence of representatives of *Plectodina florida* Sweet restricts the age of this collection to the late Edenian through Richmondian.

USGS fossil locality 6277-CO
23.8 m above lower brown quartzitic sandstone at above locality.
2 rastrate elements of *Belodina* sp.
1 *Drepanoistodus suberectus* (Branson and Mehl) element
28 *Panderodus gracilis* (Branson and Mehl) elements
1 *Panderodus panderi* (Stauffer) element
Plectodina florida Sweet
1 Pa, 3 Pb, 1 M, 1 Sa, 1 Sb, and 1 Sc elements

Age the same as USGS fossil locality 6276-CO, Late Edenian through Richmondian.

These collections show that the upper part of the Saturday Mountain Formation at the south end of the Lemhi Range is of Late Ordovician age, and not Silurian as in the central part of the range. Thin remnants of Silurian rocks are present in the southern part of the range (Ross, 1961a, p. 206-209), but these all are north of the Black Canyon area. The formation thickens northward in the range, from about 85 m at Black Canyon to 370 m in the central part of the range, at least partly by addition of the Silurian dolomites. The erosional remnants of Silurian rocks suggest that they originally extended farther south, but were removed by pre-Late Devonian erosion in the southern region (Ross, 1961a, p. 207).

SILURIAN

LAKETOWN DOLOMITE

The Silurian upper part of the Saturday Mountain Formation is overlain with erosional unconformity by light-colored dolomite of the Laketown Dolomite, which is of Middle and Late Silurian age (Ross, 1947, p. 1106). The name was extended from southeast Idaho into east-central Idaho by Ross (1934b, 957-960; 1937, p. 23-25), for Silurian rocks in the Lost River Range and on Lone Pine Peak just west of the northern part of the Lost River Range. The formation has been recognized subsequently in most of the central and northern parts of the Lost River Range, (Ross, 1947, p. 1105-1107; Mapel and others, 1965; Mapel and Shropshire, 1973), and in the central part of the Lemhi Range. Its westernmost exposures in the Lone Pine Peak region are the thickest—more than 400 m (Hays and others, 1978; Hays and others, 1980). The formation thins to the east and south and is not present in the southern part of the Lemhi Range or in the Beaverhead Mountains. The regional thinning is a result both of shoaling against the western shore of the Lemhi arch and of pre-Devonian erosion.

The Laketown Dolomite is exposed in many places along the east flank of the central Lemhi Range and also near Summerhouse Canyon on the west flank of the range. The formation thickens and thins abruptly as a result of the relief on the underlying surface; its thickness ranges from 0 m to about 60 m. The formation typically consists of light-gray-weathering, light-olive-gray to medium-light-gray and light-gray, finely to medium-crystalline dolomite in beds 0.6-3 m thick. The dolomite is characterized by its light color, by its sparkling or "live" appearance on freshly broken surfaces as contrasted to the dull luster of most of the dolomite of the Saturday Mountain Formation, and by a generally porous appearance with moderately abundant vugs 1-2 mm in diameter. In most places the formation is without recognizable fossils, but Ross (1962a, p. 37) reported a coral fauna of Silurian age from exposures near Meadow Lake, west of Gilmore.

DEVONIAN

Devonian rocks in the central Lemhi Range are divided into the Jefferson and Three Forks Formations. Both formations include rocks that closely resemble those in the area near Three Forks, Mont. (Peale, 1893, p. 27-32), as Ross (1934a, p. 25-27; 1947, p. 1110) noted when he extended these names into central Idaho. Earlier reports described these rocks as being much like the Devonian rocks of western Montana but did not apply formal names (Kirkham, 1927, p. 19; Shenon, 1928, p. 7). The age and some of the regional relations of Devonian rocks in east-central Idaho have been discussed by Scholten and Hait (1962, p. 13-22), Scholten (1957, p. 159-162), Churkin (1962, p. 582-584), and Ross (1961a, p. 209-216; 1962a, p. 37-39).

The Jefferson (Middle and Upper Devonian) and Three Forks (Upper Devonian) Formations are widely exposed in east-central Idaho, but the Jefferson differs from place to place in thickness and lithology. Devonian rocks below the Three Forks Formation in the Lost River Range include the Jefferson Formation and Grand View Dolomite, and are more than 900 m thick. The Grand View has not been recognized east of the Lost River Range, and time-equivalent rocks are included in the Jefferson Formation. In the Lemhi Range, the thickest section of the Jefferson is near Gilmore, where it is about 825 m thick. The formation thins southward to about 90 m at the south end of the range (Ross, 1961a, p. 210). In its northernmost exposures, near Hayden Creek, a short distance north of the Patterson quadrangle, the thickness of the Jefferson is uncertain because it is tectonically brecciated; it may be about 300 m thick (Anderson, 1961, p. 29). In the Beaverhead Mountains, the Jefferson is about 30-90 m thick (Rainspott, 1962, p. 33-41; Smith, 1961, p. 18-21) and overlies the Kinnikinic Quartzite or Proterozoic rocks.

As Sloss (1954) and Scholten and Hait (1962, p. 21-22) pointed out, the lithologic changes and eastward and southeastward thinning of the Jefferson Formation indicate that it laps onto an erosion surface on an older high area, the Lemhi arch of Sloss or Tendoy dome of Scholten (1957, p. 165-167). The Jefferson overlapped the Lemhi arch in Late Devonian time; the uppermost part of the formation appears to have been continuous with the formation farther east in the cratonic region. The Jefferson Formation in the Lemhi Range and Beaverhead Mountains thus shares the general eastward-thinning pattern of all the older Paleozoic rocks, reflecting onlap onto the western shore of the Lemhi arch. The arch was finally submerged in Late Devonian time, however, and marine sediments were deposited in a continuous blanket from the miogeocline far onto the craton, probably for the first time since deposition of the Yellowjacket Formation early in Middle Proterozoic time.

The Three Forks Formation is fairly widely exposed in east-central Idaho from the Lost River Range eastward, but is not present west of Lost River Range (Ross, 1947, p. 1110) except a short distance to the west in the Lone Pine Peak area (Hays and others, 1978). It is lithologically fairly uniform across this region, and resembles the Three Forks near Three Forks, Mont. The formation is thickest, about 90 m, in the central Lemhi Range and the Lost River Range, and thins eastward from there.

CHANNEL SANDSTONE AND JEFFERSON FORMATION

In the Lemhi Range the Jefferson Formation is most completely exposed in the vicinity of the Gilmore mining district and from there southward along the front of the range to Spring Mountain Canyon, where its dark dolomites form most of the canyon walls (measured sections 2 and 3). In other parts of the central Lemhi Range, only the lower part of the formation is preserved in faulted blocks—east of Summerhouse Canyon in the southwest part of the Gilmore quadrangle; on Leadore Hill, and north of Rocky Peak in the southwest part of the Leadore quadrangle; in a thrust slice south of Stroud Creek in the east-central part of the Patterson quadrangle; and in the Beaverhead Mountains in a thrust plate at the mouth of Railroad Canyon, east of Leadore. At the head of Spring Mountain Canyon, a channel sandstone deposit of Middle Devonian age is present beneath the sandstone and dolomite that elsewhere form the basal part of the formation (Hait, 1965, p. 33-35; Scholten and Hait, 1962, p. 19; Hoggan, 1981; Ruppel and Lopez, 1981).

The channel sandstone at the head of Spring Mountain Canyon fills a northeast-trending channel eroded through the Laketown Dolomite and deeply into the underlying Saturday Mountain Formation, and is a lenticular deposit about 170 m thick in the deepest part of the channel. The sandstone is light to moderate brown, fine to medium grained, and dolomitic. It is interbedded with subordinate, similarly colored dolomitic siltstone and shale, and minor, thin beds of dolomite pebble conglomerate derived from the adjacent Laketown and Saturday Mountain dolomites. The sandstone locally contains abundant fossil fragments of fish of Middle Devonian age, which include both fresh and brackish water forms and suggest deposition under estuarine conditions (Hait, 1965, p. 33-35; Denison, 1968). Similar channel deposits have been found farther south in the Lemhi Range at Badger Creek (Churkin, 1962, p. 583; Beutner, 1968, p. 15), and at Hawley Mountain in the Lost River Range (Mapel and Shropshire, 1973). The rocks at Hawley Mountain were interpreted to be a sinkhole deposit equivalent to the Lower Devonian Beartooth Butte Formation of Wyoming, but most probably they are of Middle Devonian age like the similar rocks at Spring Mountain Canyon.

In other places in the central Lemhi Range, the basal rocks of the Jefferson Formation most commonly are light-olive-gray and yellowish-gray, fine- to medium-grained quartzitic sandstone, dolomitic sandstone and sandy dolomite that fill shallow channels eroded earlier into the underlying surface. These basal sandstone units typically are from less than 1 m to about 3 m thick, but locally they are as much as 10 m thick; they occur in lenses from a few meters to several tens of meters long.

Above the basal lenticular sandstones the Jefferson Formation is mainly medium-dark-gray to dark-gray, finely crystalline, sugary dolomite. Most of these dark-colored rocks have the fetid odor characteristic of the Jefferson in other areas. The formation also includes thick units of lighter colored rocks, and of limestone and limestone sedimentary breccia. It has been separated into six informal members (Hait, 1965, p. 32-33; Scholten and Hait, 1962, p. 19; Ruppel and others, 1970, p. 39-54) that have an aggregate thickness of 700-900 m and form the thickest sequence of Devonian rocks in the Lemhi Range.

The basal member (member 1), which in earlier reports included the basal sandstones that were just described separately, is dark- to light-gray and yellowish-gray, thinly laminated, very fine grained, sugary, partly sandy fetid dolomite in thin to medium beds, and interbedded light- to medium-gray and yellowish-gray, fine- to medium-grained dolomitic sandstone that is most common in the lower part of the member. The member is 60–100 m thick, and generally consists of repeated, cyclic sequences of sandstone or sandy dolomite, medium-dark-gray dolomite, and medium-gray dolomite. The cycles are from 0.2 to 1.5 m thick; in each of them the dark-gray dolomite dominates to give the entire unit a dark color. The dolomites typically are fetid, finely to medium crystalline, and thinly laminated; they commonly contain rounded grains of quartz sand 0.25–0.5 mm in diameter that are relatively abundant low in a cycle and sparse near its top.

Member 1 is overlain by the dominantly light colored dolomite of member 2, which is 60–100 m of light- and medium-gray to yellowish-brown, partly sandy, mostly finely crystalline dolomite in beds 0.3–1 m thick. Member 2 also includes subordinate but conspicuous interbeds 0.5–1 m thick and interbedded units as much as 12 m thick of dark-gray, finely to medium-crystalline, fetid, commonly fossiliferous, partly laminated dolomite.

Member 3, 150–180 m thick, is made up mainly of medium- and dark-gray, finely to medium-crystalline, fetid dolomite in beds 0.5 to more than a meter thick. The member includes many algal beds, including *Amphipora* beds like those in the Jefferson in other areas.

Member 4 is dominantly light colored limestone, with interbeds of sandstone and minor dark-gray dolomite, in contrast to the predominantly dolomitic character of the lower three members. Scholten and Hait (1962, p. 19) suggested that this member forms the basal part of the Jefferson farther east in the southern part of the Beaverhead Mountains. The member is 100–120 m thick, and consists of light- to medium-gray, thin- to medium-bedded limestone and sandy limestone, and interbedded light-gray or pale-red to moderate-red, mostly fine to medium-grained, but partly coarse grained calcareous sandstone, and minor dark-gray, fetid dolomite.

Member 5, which is 150–240 m thick, consists of dark- to medium-gray, finely crystalline, partly laminated, partly sandy, medium- to thick-bedded fetid dolomite, and interbedded dark-gray, dolomite sedimentary breccia, light- to dark-gray, mostly thick bedded limestone, and limestone sedimentary breccia (M'Gonigle, 1982, p. 685). It includes a few thin units of yellowish-brown, fine- to medium-grained, thin-bedded calcareous sandstone.

The uppermost member, member 6, is about 150 m thick, and is light- to medium-gray, very finely crystalline limestone and limestone sedimentary breccia. The member occurs only in the central part of the Lemhi Range.

The thin thrust slice of Devonian rocks at the mouth of Railroad Canyon, east of Leadore in the Beaverhead Mountains, includes about 70 m of the Jefferson Formation, overlying the Saturday Mountain Formation. The top of the Jefferson is not present in the thrust slice and the total thickness is not known. In other thrust slices farther south in the Beaverhead Mountains, the Jefferson overlies the Kinnikinic Quartzite, where it is present, or Proterozoic rocks (Smith, 1961, p. 18–21; Lucchitta, 1966, p. 34–36); it is about 55 m thick in a section measured by Smith (1961, p. 20) about 30 km southeast of Railroad Canyon.

The formation at Railroad Canyon can be divided into two parts. The lower half consists of medium-gray to medium-dark-gray, finely crystalline, partly laminated, thick-bedded to massive dolomite and interbedded yellowish-gray, medium-grained dolomitic sandstone. The upper half consists of medium-gray to medium-light-gray, partly laminated, partly sandy, thin- to medium-bedded dolomite, interbedded yellowish-gray sandstone and bedding partings of black fissile shale. The Jefferson at Railroad Canyon most closely resembles the upper part of member 1 and the lower part of member 2 in the Gilmore area. The formation is in conformable sedimentary contact on the Saturday Mountain Formation; the lower part of member 1 does not appear to have been deposited here.

The age of the Jefferson Formation in east-central Idaho is well known from earlier studies (Scholten and Hait, 1962, p. 19–20; Mapel and Sandberg, 1968, p. D115–D125), and no significant new fossil collections were made in the central Lemhi Range. As Scholten and Hait (1962, p. 20) pointed out, the formation is of Middle and Late Devonian age in the Lemhi Range, but entirely of Late Devonian age in the Beaverhead Mountains.

Members 1 through 3 of the Jefferson Formation contained most of the ore mined in the Gilmore (Texas) mining district, and seem to be especially favorable host rocks for mineral deposits in this district.

THREE FORKS FORMATION

The Three Forks Formation is present in the central Lemhi Range only in the Gilmore area. It is present northeast of Gilmore in the Beaverhead Mountains near the mouth of Railroad Canyon. It is more widely exposed in the southern parts of both of these mountain ranges, where it has been described by Sloss (1954, p. 366), Ross (1961a, p. 209–216), Scholten and Hait

(1962, p. 20–21), Smith (1961, p. 24–25), Lucchitta (1966, p. 38–40), and Ramspott (1962, p. 41–45). The formation is about 90 m thick in the Gilmore area, and about the same thickness in the Lost River Range to the west (Mapel and others, 1965; Mapel and Shropshire, 1973). It thins to the east and south. In the southern Lemhi Range and in the Beaverhead Mountains, where its thickness ranges from a few meters to about 35 m it has not been mapped separately, but has been included either with the underlying Jefferson Formation or with overlying Mississippian rocks.

The formation consists of yellowish- and pinkish- to light-gray, thin-bedded, platy, very finely crystalline limestone, silty limestone, and calcareous siltstone. Much of the formation is fossiliferous. Hait (1965, p. 37–38) reported the fossils *Pugnoides* sp., *Cyrtospirifer* sp., *C.* cf. *whitneyi*, *Tenticosperifer* cf. *utahensis*, and *Camarotoechia* cf. *contracta*, which indicate a Late Devonian age. Additional collections from the formation in the Gilmore area yielded the following fossils, identified by J.T. Dutro, Jr. (written commun., 1967):

Trifidirostellum? *madisonensis* (Haynes)
Cyrtospirifer whitneyi (Hall)
Tenticospirifer? sp.
Camarotoechia sp.
Rhipidomella sp.
"*Cleiothyridina*" *devonica* (Raymond)
ambocoelid brachiopod, undet.
Cyrtospirifer cf. *C. animasensis* (Girty)
Cyrtiopsis? *C.*? *monticola* (Haynes)
Trifidirostellum dunbarense (Haynes)
Cyrtospirifer sp.

Dutro noted that the collection contains species characteristic of the Three Forks Formation, and is of mid-Famennian (Late Devonian) age.

MISSISSIPPIAN AND MISSISSIPPIAN-PENNSYLVANIAN

Mississippian rocks in and near the central Lemhi Range include the McGowan Creek Formation, at the base, conformably and gradationally overlain by the Middle Canyon Formation, and, above the Middle Canyon, the Scott Peak Formation, Railroad Canyon Formation, and the lower part of the Mississippian and Pennsylvanian Bluebird Mountain Formation. Because of structural complexity, the Mississippian limestones of the Middle Canyon and Scott Peak Formations were mapped as an undivided unit, mainly of Late Mississippian age.

The nomenclature and ages of Mississippian rocks in east-central Idaho have become clear only recently, principally as a result of subdivision of these rocks by Huh (1967, p. 31–62), and of subsequent further definition by Skipp and others (Skipp and Mamet, 1970; Mamet and others, 1971; Skipp, Sando, and Hall, 1979) of the formations named by Huh. In earlier reports, these rocks were assigned to the Milligen Formation together with either the Brazer Formation (Ross, 1947, p. 1112–1117; 1961a, p. 216–228; Sloss and Moritz, 1951, p. 2160–2162;) or, tentatively, the Madison Group (Lucchitta, 1966, p. 40–49; Ruppel, 1968; Staatz, 1973; 1979, p. A11). Scholten (1957, p. 162–164), however, had questioned the continued use of both Milligen and Brazer in east-central Idaho, and cited evidence to show that rocks of the Madison Group are not present there. Ross also questioned extension of the Milligen into the Lemhi Range (Scholten 1957, p. 162). Sandberg (1975) later pointed out that the Milligen Formation is wholly of Devonian age in west-central Idaho, and named and described the McGowan Creek Formation, of Early Mississippian age, to include the rocks in the basal part of the Mississippian sequence in east-central Idaho that earlier had been called Milligen Formation. The extension of the Madison Group into east-central Idaho is inappropriate, as Scholten (1957) suggested, because the rocks above the McGowan Creek Formation are largely or entirely of Late Mississippian age and clearly are related to the now better known Middle Canyon and Scott Peak Formations of east-central Idaho. The Brazer Formation now has been restricted to its type area (Sando and others, 1959), and is no longer used in central Idaho (Ross, 1962c, p. 385).

The Railroad Canyon Formation (Wardlaw and Pecora, 1985, p. B4) includes the strata originally mapped as Big Snowy Formation in the Leadore quadrangle (Ruppel, 1968) and farther south in the Beaverhead Mountains (Smith, 1961; Ramspott, 1962; Lucchitta, 1966; Embree and others, 1975). It is interpreted by Wardlaw and Pecora (1985, p. B4) to be a marginal or back-bank facies of the foredeep carbonate bank sequence of the South Creek, Surrett Canyon, and Arco Hills Formations that are present farther southwest (Skipp, Hoggan, Schleicher, and Douglas, 1979, p. 5–18), and to be the temporal equivalent to part of the Snowcrest Range Group of southwest Montana.

The rocks assigned in this report to the Bluebird Mountain Formation were included previously (Ruppel, 1968) in the lower third of the Pennsylvanian Quadrant Sandstone in the Leadore quadrangle, because they are lithologically similar to those in the Quadrant in southwest Montana, and because the effects of regional thrust faulting were not recognized until later (Ruppel, 1978, 1982; Ruppel and Lopez, 1984). Subsequent studies in the southern parts of the Beaverhead Mountains and the Lemhi Range and in the Lost River Range led to definition of the Bluebird Mountain Formation, of Late Mississippian and Early Pennsylvanian age (Skipp, Hoggan, Schleicher, and Douglas, 1979; Skipp and others, 1981). Thus, the rocks previously included in the Quadrant in the Leadore area (Ruppel, 1968) and elsewhere in the Beaverhead Mountains (Smith, 1961, p. 36–39; Lucchitta, 1966, p. 54–60; Scholten and Ramspott, 1968, p. 12–13) are now assigned to the formations proposed by Skipp, Hoggan, Schleicher, and Douglas (1979). Bluebird Mountain Formation now is the appropriate name for the lower one-third of the unit called Quadrant Sandstone in the Leadore quadrangle (Ruppel, 1968).

Rocks of Mississippian age underlie most of the southern parts of the Beaverhead Mountains and Lemhi Range, and almost all of the Lost River Range; the type sections of all of the formations are in these ranges. The McGowan Creek Formation is more than 1,100 m thick at its type section on the west side of the Lost River Range (Sandberg, 1975, p. E3), but thins eastward to 150–260 m in the eastern part of the Lost River Range (Mapel and Shropshire, 1973), and to perhaps 60-150 m in the Beaverhead Mountains, where it has been mapped with the Three Forks Formation (Smith, 1961, p. 24–27; Skipp and Hait, 1977, p. 504). The Middle Canyon Formation is about 300 m thick at its type section at the south end of the Lemhi Range (Huh, 1967, p. 37), but its thickness seems to range from 150 to about 400 m in different places across east-central Idaho, perhaps partly reflecting uncertainty in the location of its gradational contact with the McGowan Creek Formation. The Scott Peak Formation is about 685 m thick at its type section in East Canyon in the southern Lemhi Range (Huh, 1967, p. 39–42) and maintains a thickness of 600 m or more throughout east-central Idaho (Skipp and Hait, 1977, p. 504). The Railroad Canyon Formation is known only in the Beaverhead Mountains, where its thickness ranges from about 200 m in the south part of the range (Embree and others, 1975) to 260 m at its northernmost exposures near Leadore. The Bluebird Mountain Formation is 105 m thick at its type section in the southern Beaverhead Mountains, and about 40–50 m thick in the southern Lemhi and Lost River Ranges (Skipp, Hoggan, Schleicher, and Douglas, 1979, p. 18–19). It is about 130 m thick at its northernmost exposure near Leadore.

These thick and widespread formations are rarely preserved in the central part of the Lemhi Range. The lower part of the sequence is present south of Gilmore, and is exposed again much farther north in and near the lower reaches of Hayden Creek, where it was mapped by Anderson (1961, p. 29) as Grand View Dolomite. In the Beaverhead Mountains north of Leadore, the Mississippian formations are more widely preserved than in the central Lemhi Range, but the only complete sequences in this area of closely spaced imbricate thrust faults are of the Railroad Canyon and Bluebird Mountain Formations. The Mississippian rocks extend only a short distance north of the Leadore quadrangle (Ruppel, 1968), to the flat, upper part of Grizzly Hill, where they are thrust across Proterozoic rocks of the Lemhi Group. The northernmost exposures in the Beaverhead Mountains, mainly of Scott Peak limestones, are about 15 km northwest of Leadore, on the west flank of Goat Mountain (Staatz, 1973; 1979, p. A11), in an isolated thrust slice on Lemhi Group rocks.

McGOWAN CREEK FORMATION

The McGowan Creek Formation is present just south of Gilmore, where it overlies the Three Forks Formation with apparent conformity, and is gradationally overlain by the Middle Canyon Formation. This poorly exposed formation weathers to form a shale-littered slope. The Mississippian rocks are partly hidden beneath the brecciated lower part of a thrust plate and are partly metamorphosed by the Gilmore stock. The formation also is present in Railroad Canyon, in the Beaverhead Mountains east of Leadore. It is about 30–60 m thick in the Gilmore area and about 40–45 m thick in Railroad Canyon. In both areas, the formation consists of medium-gray to dark-gray, chippy to fissile, partly carbonaceous siltstone, mudstone, and shale, with thin interbeds of dark-gray, finely crystalline limestone. The limestone interbeds are sparse in the lower part of the formation, but increasingly abundant in the upper part as it grades into the overlying thin-bedded, cherty limestones of the Middle Canyon Formation.

No fossils have been found in the McGowan Creek Formation in the central Lemhi Range, but lithologically similar rocks in an erosionally isolated remnant of a thrust plate about 8 km east of Railroad Canyon have yielded the following Lower Mississippian fossils, identified by W.J. Sando, J.T. Dutro, Jr., and E.L. Yochelson (written commun., 1960):

USGS fossil collection 19139-PC, collected from center of sec. 13, T. 16 N., R. 27 E., by E.R. Cressman.

Lophophyllum?
Vesiculophyllum sp.
Sychnoelasma sp.
Amplexizaphrentis sp.
Unispirifer sp.
Punctospirifer sp.

Sando, Dutro, and Yochelson considered the assemblage to be of Early Mississippian age, and characteristic of coral zone II of Sando and Bamber (1985), equivalent in age to the middle and upper Lodgepole Limestone, the correlative in southwest Montana of the McGowan Creek Formation (Sandberg, 1975, p. E9–E10). In this area east of Railroad Canyon, the lower part of the Mississippian sequence has been mapped as Lodgepole Limestone, including a unit of dark-gray, thin-bedded, shaly limestone, silty limestone, siltstone, and siliceous mudstone and a unit of dark-gray cherty, laminated limestone (Lucchitta, 1966, p. 41–42). Although the stratigraphic relation of the two units is confused by faulting, the descriptions given by Lucchitta suggest that the shaly limestone, siltstone, and mudstone unit is the McGowan Creek Formation, and that the cherty limestone is part of the overlying Middle Canyon Formation. The assignment of these rocks to the Lodgepole was based on their thin bedding, texture, and composition. The more recent definition of the McGowan Creek Formation and younger Mississippian formations indicates that the names McGowan Creek and Middle Canyon are more appropriate for these rocks (Huh, 1967; Sandberg, 1975).

MIDDLE CANYON AND SCOTT PEAK FORMATIONS

Only faulted and incomplete sections of the Middle Canyon and Scott Peak Formations are preserved in the central Lemhi Range and on Grizzly Hill in the adjacent Beaverhead Mountains north of Leadore. These rocks were mapped as undivided Mississippian limestones near Gilmore and in the southwest corner of the Gilmore quadrangle, where only a little of the Middle Canyon Formation is exposed (Ruppel and Lopez, 1981), and incorrectly mapped as Madison Limestone on Grizzly Hill (Ruppel, 1968). The sequence of Mississippian rocks is thickest on Grizzly Hill, but it has been so complexly cut by thrust faults that no complete section remains, and the limestone in any one thrust slice generally cannot be related with much confidence to that in any other slice.

Only the basal 45 m of the Middle Canyon Formation and the uppermost 240–300 m of the Scott Peak Formation are in well-defined stratigraphic positions on Grizzly Hill; not many of the rocks in these sequences can be recognized in thrust slices that make up the main exposures. As a result, the original thicknesses of the formations are unknown, but they must exceed greatly those of the abbreviated well-defined sections on Grizzly Hill. In the type sections in the southern Lemhi Range, the Middle Canyon Formation is about 300 m thick, and the Scott Peak Formation is almost 700 m thick (Huh, 1967, p. 38, 40). In the southern part of the Beaverhead Mountains, the Middle Canyon Formation is 150–400 m thick, and the Scott Peak Formation is 580–600 m thick (Skipp and Hait, 1977, p. 504). The original, unfaulted thicknesses of the formations on Grizzly Hill could have been as great as these.

The basal part of the Middle Canyon Formation is well exposed, although intensely contorted, in cliffs in Railroad Canyon near its mouth and north of Salt Creek on the canyon wall. The rocks overlie, apparently gradationally, the carbonaceous black shale of the McGowan Creek Formation. The rocks above the basal 45 m have not been found in normal stratigraphic position. The basal rocks are thin-bedded, medium-gray, finely crystalline limestone that weathers to medium-light-gray platy fragments. Very abundant medium-dark-gray, yellowish-weathering chert is interbedded with the limestone in 2–10-cm-thick irregular beds and lenses spaced 2–30 cm apart.

The uppermost rocks of the Scott Peak Formation are exposed best on Grizzly Hill, just north of the Leadore quadrangle, but also crop out in some of the thrust slices south of the quadrangle boundary. They underlie the shaly rocks of the Railroad Canyon Formation with apparent conformity. The rocks are mainly thick bedded to massive, medium-dark- to medium-light-gray, fine- to coarse-grained, partly fetid limestone that weathers medium light gray. Most of the rocks are characterized especially by a bluish tint that seems to be unique to this formation. Many beds are very fossiliferous, containing abundant horn corals as well as other marine fossils; many beds are bioclastic. Much of the limestone in the upper 150 m is thin to medium bedded, even though thick-bedded to massive limestones still dominate. These rocks are more siliceous than the more massive rocks below, with abundant siliceous weathering crusts and dark-gray chert in nodules and thin beds. About 45 m below the top of the formation, a solution breccia zone with associated mudstone, about 12 m thick, is made especially conspicuous by its red and yellowish-orange colors.

The limestones that underlie much of the Beaverhead Mountain front north of Leadore are clearly part of a sequence of rocks that includes the Middle Canyon and Scott Peak Formations, because they share many of the lithologic characteristics and the fauna of these formations. But the complex thrust faults that cut the rocks have shuffled them into layers that cannot be restored to their original position. In addition, thrusting has sheared some of the rocks so strongly that they now have a pronounced shaly to flaggy parting. Many of the rocks are somewhat bleached, although the typical bluish tint of the Scott Peak Formation is rarely destroyed. In some places, the intense shearing that accompanied the thrusting seems to have concentrated the chert into thrust zones, perhaps partly by secondary silicification of the zone and partly by mobilization of the limestone. The chert remained behind to form layers from about a meter thick to more than 30 m thick in the exceptional layer in Jakes Canyon. These layers are made up of moderately to strongly brecciated to locally mylonitized dark-gray chert. They are the source of the widespread blocks of chert that litter most of the slopes in this part of the Beaverhead Mountains.

The Middle Canyon Formation in Railroad Canyon has not yielded any fossils, but its Late Mississippian (Meramecian, coral zones III C and D) age is well established in the southern Lemhi Range and in the Lost River Range (Huh, 1967, p. 37; Mamet and others, 1971, p. 24–25; Sando, 1976, p. 3–4).

The upper part of the Scott Peak Formation on Grizzly Hill has yielded a distinctive fauna that clearly indicates its Late Mississippian age. J.T. Dutro, Jr., W.J. Sando, and Helen Duncan have examined several collections of fossils from the uppermost 75 m of these beds, and they stated (written communs., 1962, 1964) that all of the collections contain elements of the *Faberophyllum* assemblage (coral zone IV of Sando and Bamber, 1985) that is common in Upper Mississippian rocks of eastern Idaho. The assemblage is younger than any of those found in the Madison Group in southwest Montana (Dutro and Sando, 1963, p. 1983). These collections contain the following fossils, identified by Sando, Dutro, and Duncan:

USGS fossil collections 21920–PC, 21921–PC, 21922–PC, 21923–PC, and 21924–PC, all from the same locality along the head of the south branch of Hood Gulch, about 2–4 km northwest of the northeast corner of the Leadore quadrangle.

- *Pleurosiphonella virginica* (Butts)
- *Ekvasophyllum* sp.
- *Ekvasophyllum*? sp.
- *Faberophyllum*? sp.
- *Dorlodotia*? sp.
- *Schoenophyllum* cf. *S. aggregatum* Simpson
- *Striatifera* sp. aff. *S. brazeriana* (Girty)

USGS fossil collections 20217–PC, 20218–PC, and 20219–PC, same locality as above.

- *Faberophyllum* sp.
- syringoporoid and lithostrotionoid corals, undet.
- productoid brachiopod, indet.
- bellerophontacean, loxonematid, and straparollid gastropods, indet.

RAILROAD CANYON FORMATION

Rocks of the Railroad Canyon Formation crop out in a number of thrust slices on the south face of Grizzly Hill in the Beaverhead Mountains, where they underlie smooth, grassy slopes. Like the other rocks in this area, they have been so cut by faults that no complete section remains, but a well-exposed complete section is present just north of the Leadore quadrangle boundary about 3 km north of the mouth of Waugh Gulch in the northeast corner of sec. 31, T. 17 N., R. 27 E., Beaverhead Mountains (measured section 4). The rocks included in the Railroad Canyon Formation are about 260 m thick in the section north of Waugh Gulch. The lower 70 m is mainly grayish red or grayish-red-purple to yellowish-brown, brownish-gray to brownish-black, shaly to flaggy, partly calcareous shale and mudstone that overlies the limestone of the Scott Peak Formation with apparent conformity, although the contact is nowhere exposed. The lower half of the shale and mudstone unit is red- or orange-weathering, and gives the soil above the massive limestones a distinct reddish cast. Most of the upper half is nearly black, papery shale that contrasts strongly with the less fissile and reddish basal rocks; it includes a few thin interbeds of yellowish-brown limestone. The shale and mudstone is overlain by a distinctive olive-gray limestone unit about 20–25 m thick that is characterized by abundant hairlike irregular veins of white calcite. The remaining 165–170 m of the formation is mainly gray and brownish-gray, partly silty, partly fossiliferous limestone with some similarly colored interbeds of siltstone. Some distinctive beds of limestone pebble conglomerate and olive- and brownish-gray paper shale and shaly mudstone are present about in the middle of the formation.

Many beds in the Railroad Canyon Formation section north of Waugh Gulch are fossiliferous, and they have yielded a fauna that establishes the Late Mississippian age of these rocks. The fossils were studied by J.T. Dutro, Jr., W.J. Sando, and Helen Duncan, who considered the assemblage to be much like that in the Snowcrest Range Group (Wardlaw and Pecora, 1985) of southwest Montana (formerly Big Snowy Formation). The collections from the formation, which are located by collection number in measured section 4, yielded the following fossils identified by Sando, Dutro, and Duncan (written communs., 1962, 1968).

USGS fossil collection 20221–PC (unit 40 in measured section 4).

- horn corals, undet.
- *Fistulipora* sp.
- *Dichotrypa* sp.
- *Inflatia* sp.
- *Diaphragmus* aff. *D. phillipsi* (Norwood and Pratten)
- *Flexaria*? sp.
- productid brachiopods, undet.
- *Anthracospirifer* aff. *A. leidyi* (Norwood and Pratten)
- *Anthracospirifer* aff. *A. increbescens* (Hall)
- *Composita* cf. *C. subquadrata* (Hall)
- *Cleiothyridina*? sp.
- *Dielasma*? sp.
- pelecypod? indet.

USGS fossil collections 20220–PC and 21930–PC (unit 38 in measured section 4)

Orbiculoidea sp. (abundant)
"Leiorhynchus" cf. *"L." carboniferum* (Girty)
brachiopod fragments, indet.

USGS fossil collection 21929–PC (unit 37 in measured section 4)

Orbiculoidea sp. (abundant)

USGS fossil collection 21928–PC (unit 31 in measured section 4)

orthotetid brachiopod, indet.
Inflatia? sp.
Composita sp.
pelecypod fragment, undet.

USGS fossil collection 21927–PC (unit 27 in measured section 4)

chonetid fragments, indet.
Diaphragmus sp.
Inflatia sp.
Flexaria? sp.

USGS fossil collection 21926–PC (unit 25 in measured section 4)

chonetid brachiopod, undet.
Inflatia sp.
Composita sp.
productid fragments, indet.
spiriferoid fragments, indet.

Fossil collection from unit 22 in measured section 4; no USGS collection number

Composita sp.
productid fragments (possibly *Inflatia*?)
burrows of probable organic origin

USGS fossil collection 21925–PC (unit 15 in measured section 4)

Diaphragmus sp.
Inflatia sp.
Antiquatonia? sp.
linoproductid brachiopod, indet.

Dutro (written commun., 1968) noted that all of these collections contain brachiopods of Late Mississipppian age, probably coral zone V of Sando and Bamber (1985) or possibly younger. Conodonts collected from this section, and described by B.R. Wardlaw and W.C. Pecora (written commun., 1985), belong to the long-ranging *Cavusgnathus* zone and *Rhachistognathus* zone, indicating a middle to latest Chesterian age.

BLUEBIRD MOUNTAIN FORMATION

The Bluebird Mountain Formation crops out on Grizzly Hill in the Beaverhead Mountains, north of Leadore. It is best exposed in cliffy outcrops in Waugh Gulch, north of Railroad Canyon, where it is about 130 m thick. It is composed predominantly of gray sandstone with some interbedded dolomite and limestone in the upper part of the formation.

The Bluebird Mountain Formation overlies the upper silty limestone and mudstone of the Railroad Canyon Formation with apparent conformity. The contact is sharply defined by the distinctive basal sandstones, about 15 m thick of the Bluebird Mountain, which are light-olive-gray, weathering to grayish-orange and yellowish-brown, fine-grained rocks that are in alternating calcareous and quartzitic interbeds. The sandstone beds typically are 5–15 cm thick, and rarely more than 30 cm thick. The quartzitic and calcareous beds weather differentially to form a honeycombed or fretted weathered surface that distinguishes them from other sandstone beds higher in the formation, or in other formations.

The basal sandstone is overlain by about 40–45 m of light-gray to medium-dark-gray, fine-grained, thick-bedded to massive quartzitic, partly calcareous sandstone, and, above these massive sandstones, by a sequence of somewhat similar but increasingly calcareous sandstones, and interbedded thick units of medium- to dark-gray, cherty limestone and dolomite. The contact with the carbonate rocks of the overlying Bloom Member of the Snaky Canyon Formation of Pennsylvanian age is gradational; the contact is placed at the top of the highest bed of sandstone like those that form most of the Bluebird Mountain Formation.

The formation has not yielded fossils to establish its age in this area, but is of Late Mississippian age in the type section (Skipp, Hoggan, Schleicher, and Douglas, 1979, p. 19–20; Skipp and others, 1981). The Bluebird Mountain Formation is a thin, western tongue of the lower part of the Quadrant Sandstone of southwest Montana (B.R. Wardlaw and W.C. Pecora, written commun., 1985).

PENNSYLVANIAN

Pennsylvanian rocks are not present in the central part of the Lemhi Range, but are preserved in thrust slices on Grizzly Hill in the Beaverhead Mountains north of Leadore. The exposures near Leadore are the northernmost occurrence of Pennyslvanian rocks in the Beaverhead Mountains. These rocks were mapped as the Quadrant Formation (Ruppel, 1968), but later studies by Skipp, Hoggan, Schleicher, and Douglas (1979) suggest that they are part of the Snaky Canyon Formation.

At the type locality in the southern part of the Beaverhead Mountains the Snaky Canyon Formation includes three members, the Bloom (basal), Gallagher Peak, and Juniper Gulch (top) Members, and is 1,125 m thick (Skipp, Hoggan, Schleicher, and Douglas, 1979). The Bloom Member, 550–650 m thick, is mainly thin to medium-bedded, medium- to medium-dark-gray limestone, and includes thin interbeds of medium-gray, very fine grained sandstone and siltstone. The member is of Early to Late Pennsylvanian age in this region. The Bloom is gradationally overlain by the Gallagher Peak Sandstone Member, about 56 m of thin-bedded, medium-gray to light-brownish-gray, very fine grained calcareous sandstone of Late Pennsylvanian age.

The middle carbonate unit of the Quadrant Formation as used by Ruppel (1968) near Leadore is lithologically similar to the Bloom Member of the Snaky Canyon Formation, although much thinner, and is of the same age. The upper sandstone unit of the Quadrant as used by Ruppel (1968) is similar to the Gallagher Peak Sandstone Member of the Snaky Canyon Formation. The Bloom and Gallagher Peak Members of the Snaky Canyon Formation are extended to include the Pennsylvanian rocks near Leadore that originally were assigned to the upper two units of the Quadrant (Ruppel, 1968; Lucchitta, 1966, p. 54–60; Smith, 1961, p. 36–39). The dominantly carbonate and calcareous sandstone Bloom and Gallagher Peak Members of the Snaky Canyon Formation are, in part, the western age-equivalents of the quartzite and sandstone beds included in the upper part of the Quadrant Sandstone in southwest Montana.

The uppermost member of the Snaky Canyon Formation, the Juniper Gulch, is about 520 m thick in the type section, and is interbedded light-gray-weathering, sandy and cherty limestone and dolomite of Late Pennyslvanian to Early Permian age. It is overlain by rocks assigned to the Phosphoria Formation of Early Permian age. The Juniper Gulch Member does not appear to be present in the Leadore area, unless it is represented by cherty carbonate rocks included here in the upper part of the Gallagher Peak Member, or by the carbonate rocks included here in the Permian Park City Formation. The Juniper Gulch is the age equivalent, in part, of beds included in the upper part of the Quadrant Sandstone in southwest Montana, which contain an Early Permian fauna (B.R. Wardlaw and W.C. Pecora, written commun., 1985).

SNAKY CANYON FORMATION

In exposures on Grizzly Hill and in Railroad Canyon, the Bloom Member of the Snaky Canyon Formation gradationally overlies the Bluebird Mountain Formation, and is gradationally overlain by the Gallagher Peak Member of the Snaky Canyon Formation. The member is about 120 m thick in exposures in Railroad Canyon, and consists of medium-gray to medium-light-gray, finely crystalline, thin- to medium-bedded (5–60 cm) limestone and dolomite that contains abundant dark-gray chert in thin interbeds and in nodules as much as 15 cm long. The carbonate rocks are separated in places by thin interbeds of black shale and mudstone. Many beds in the member are fossiliferous; they have yielded fusulinids that suggest the member is of older Pennsylvanian age than most of the Quadrant Sandstone in southwest Montana (Axelsen, 1973, p. 10–29). David A. Bostwick (written commun., 1968; Axelsen, 1973, p. 18–19) identified the fusulinids *Pseudostaffella*?, *Profusulinella*, *Nankinella*, *Fusulina*, and *Wedekindellina* from the Bloom Member, which suggest it is of early Atokan age at the base and Desmoinesian at the top (Axelsen, 1973, p. 22–23). Other collections from exposures of the member in and near Railroad Canyon yielded the following fossils, identified by J.T. Dutro, Jr., and W.J. Sando (written communs., 1962, 1968):

USGS fossil collection 21916–PC, along Railroad Canyon, NE corner, sec. 6, T. 16 N., R. 27 E.
- *Pseudozaphrentoides* sp.
- *Antiquatonia*? aff. *A. coloradoensis* (Girty)
- *Linoproductus* sp.
- *Anthracospirifer*? sp.
- *Composita* sp.
- *Phricodothyris*? sp.
- *Hustedia* sp.
- chonetid fragments, indet.
- pelecypod fragments, indet.

USGS fossil collection 21917–PC, same locality as above.
- *Pseudozaphrentoides* sp.
- *Chonetinella* sp.
- *Antiquatonia*? aff. *A. coloradoensis* (Girty)
- *Linoproductus* sp.
- productid fragments, indet.

USGS fossil collection 20222–PC, near center of sec. 30, T. 17 N., R. 27 E., south of Hood Gulch in Bannock Pass quadrangle.
- *Fenestella* sp.
- *Polypora* sp.
- *Linoproductus*? sp.
- *Spiriferella*? sp.
- *Squamularia*? sp.
- punctate spiriferoid, indet.
- gastropod fragments, indet.
- echinoderm debris, indet.

Dutro (written commun., 1968) considered the fauna to be of Middle or Late Pennsylvanian age. It closely resembles the fauna in the Bloom Member reported by Skipp, Hoggan, Schleicher, and Douglas (1979, p. 25–28), but the fusulinids reported by Bostwick suggest a somewhat narrower Pennsylvanian age for the member in this area than the latest Mississippian to Late Pennsylvanian range suggested by Skipp, Hoggan, Schleicher, and Douglas (1979, p. 25–28) for the member farther south in the Beaverhead Mountains and in the Lemhi and Lost River Ranges.

The Gallagher Peak Sandstone Member of the Snaky Canyon Formation gradationally overlies the Bloom Member, and is gradationally overlain by carbonate rocks thought to be part of the Grandeur Member of the Park City Formation of Permian age. The member is about 140 m thick in Railroad Canyon, and consists mainly of pale-yellowish-brown and light-brownish-gray to light-gray, fine-grained (0.2–0.3 mm), partly thinly laminated, partly cross laminated, clean, calcareous and quartzitic sandstone in beds from 8 cm thick to massive. The lower beds in the member are conspicuously cross laminated. The sandstone includes thin, inconspicuous interbeds of cherty, sandy dolomite. The upper 30 m of the member, gradational into the overlying formation, includes two 3–5-m-thick units of medium-light-gray to light-gray, fine-grained, thin-bedded dolomite and dolomitic limestone that in places contain abundant chert nodules as much as 10 cm in diameter, and sparse, thin sandstone interbeds. The top of the member is drawn at the highest, massive sandstone bed, beneath the cliff-forming, massive basal limestone of the Grandeur Member of the Park City Formation.

The sandstone of the Gallagher Peak Member has yielded the following fossils, identified by J.T. Dutro, Jr., and W.J. Sando (written communs., 1965, 1968):

USGS fossil collection 21918–PC, from upper part of Member, sec. 5, T. 16 N., R. 27 E.

Pseudozaphrentoides sp.

Antiquatonia? sp.

These genera are long ranging, of latest Mississippian to Late Pennsylvanian age. The age of the member in this area appears to be compatible with the Late Pennsylvanian age assigned in the type section (J.T. Dutro, Jr., written commun., 1968; Skipp, Hoggan, Schleicher, and Douglas, 1979, p. 28–29).

PERMIAN

Permian rocks are preserved only in a small area on the east wall of Railroad Canyon (Ruppel, 1968) in the Beaverhead Mountains, and are not present in the central part of the Lemhi Range. They are present in a much wider area in the southern parts of the Beaverhead Mountains and Lemhi Range, however, and in the Lost River Range farther west. In the Beaverhead Mountains, the Permian rocks most commonly have been mapped as the Park City and Phosphoria Formations (Cressman and Swanson, 1964, p. 296–300; Lucchitta, 1966, p. 60–66; Smith, 1961, p. 40).

The thin, incomplete sequence in Railroad Canyon is assigned to the Grandeur Member of the Park City Formation on the basis of lithologic similarity and stratigraphic position. The clean limestone and dolomite assigned to the Grandeur Member at this location are not much like the sandy and very cherty carbonate rocks of the Juniper Gulch Member of the Snaky Canyon Formation in the southernmost Beaverhead Mountains, Lemhi Range, and Lost River Range (Skipp, Hoggan, Schleicher, and Douglas, 1979, p. 29–34). Accordingly the name Grandeur is retained in this area for the carbonate rocks of Permian age above the Gallagher Peak Member of the Snaky Canyon Formation and beneath the Phosphoria Formation. As Cressman and Swanson pointed out (1964, p. 297), however, the assignment is an uncertain one and should be considered tentative until the regional relations of Permian rocks in east-central Idaho are better known.

GRANDEUR MEMBER OF PARK CITY FORMATION

The rocks tentatively assigned to the Lower Permian Grandeur Member of the Park City Formation in Railroad Canyon gradationally overlie the Gallagher Peak Member of the Snaky Canyon Formation. Only a thin and incomplete section of the member is present, about 75 m thick; the top of the member is cut off by a thrust fault that carries Devonian and Mississippian rocks across the Permian ones. In nearby areas, rocks equivalent to the Grandeur Member are 90–190 m thick, and are overlain by 70–170 m of chert, sandstone, dolomite, phosphatic sandstones and siltstones, and phosphate rock of the Phosphoria Formation (Cressman and Swanson, 1964; Lucchitta, 1966, p. 60–65). The total thickness of Permian rocks is about 250 m.

The Grandeur Member in Railroad Canyon is composed of very light gray to medium-light-gray, very fine to fine-grained dolomite, dolomitic limestone, and limestone. The basal 25 m of the member is massive, vuggy limestone that forms cliffy outcrops; the rest of the member is thin bedded and platy weathering. These rocks are not fossiliferous, but equivalent beds exposed near Hawley Creek, about 10 km southeast of Railroad Canyon, have yielded colonial rugose corals that suggest the beds involved are somewhat older than the Grandeur Member of the type section in the Wasatch Mountains of Utah (Duncan, 1961, p. B235–B236).

The outcrops of Grandeur carbonate rocks in Railroad Canyon are the northernmost exposures of Permian rocks in the Beaverhead Mountains. No Permian rocks are present in the central or northern parts of the Lemhi Range.

The Permian rocks are overlain by the Dinwoody Formation, of Early Triassic age, which is preserved at only two places on the west side of the Beaverhead Mountains (Lucchitta, 1966, p. 66–72; Skipp, Hoggan, Schleicher and Douglas, 1979, p. 34). Younger Mesozoic rocks are absent in east central Idaho as a result of uplift and erosion that began in post-Early Triassic to Jurassic time (Scholten, 1968, p. 115) and culminated in regional thrust faulting in Late Cretaceous time. As a consequence, the Precambrian and Paleozoic sedimentary rocks in the Lemhi Range are overlain with angular unconformity by the early Tertiary Challis Volcanics, and by later Tertiary tuffaceous sedimentary rocks that mainly are preserved in the valleys that flank the range.

CHALLIS VOLCANICS AND RELATED INTRUSIVE ROCKS

Early Tertiary volcanic rocks are widespread throughout much of east-central Idaho. All these rocks have been grouped under the broad term "Challis Volcanics," which includes all the dominantly volcanic strata of early Tertiary age of central Idaho from the Snake River Plain north to the westward-flowing segment of the Salmon River (Ross, 1961b). Ross (1961b, 1962a) included tuffaceous sedimentary rocks of Oligocene and younger ages in the Challis, but the term "Challis Volcanics" generally has been applied only to dominantly volcanic rocks, which are separated from the overlying tuffaceous sedimentary rocks by a distinct angular unconformity (Anderson, 1956, 1957, 1979, 1961; Ruppel, 1968, 1980; Ruppel and Lopez, 1981; Staatz, 1972, 1979).

The age of Challis Volcanics in the region surrounding the central Lemhi Range generally is considered to be Eocene; radiometric dates range from 39 to 54 m.y. (Armstrong, 1974, 1975; McIntyre and others, 1982; Staatz, 1979). No radiometric dates are available for the Challis in the Lemhi Range, but geologic relations indicate the volcanics are considerably younger than the granitic intrusive rocks. These 48–50-m.y.-old intrusive rocks were cooled, uplifted, and exposed to surface weathering before being buried under flows of Challis Volcanics. The volcanic rocks are overlain by sedimentary rocks that have been dated on the basis of plant remains as Oligocene or tentatively late Eocene in age. Therefore, the Challis Volcanics in the Lemhi Range are late Eocene in age, possibly 40–45 m.y. old. They probably are of about the same age as the Challis Volcanics in the Beaverhead Mountains that have an average age of 41.3 m.y. (Staatz, 1979, p. A22).

Attempts have been made to divide the Challis Volcanics into lithologic units that could be used as regionally correlative stratigraphic horizons (Ross, 1961b; Anderson, 1961). As has been discussed by Staatz (1979), however, these lithologic units typically are only of local extent, and correlating units of volcanic rocks of similar composition can lead to serious stratigraphic errors. In general, however, there is a consistent change in lithology with time. In the eastern half of the Challis 1° × 2° quadrangle (McIntyre and others, 1982), Challis volcanic rocks can be divided into two groups: (1) An early group of voluminous, predominantly intermediate composition lavas erupted from widely scattered small vents in the period from 51 m.y. to 49 m.y. ago and (2) a later group of predominantly silicic, cauldron-derived, ash-flow tuffs erupted from about 48 m.y. to 45 m.y. ago. These two groups overlap, and silicic ash-flow tuffs locally are interbedded with intermediate lavas. A similar sequence of volcanic rocks also seems to be present in the general region of the Lemhi Range where the volcanic rocks are predominantly andesites and latites, with a few thin silicic ash-flow tuffs stratigraphically higher in the sequence. In the Salmon, Idaho, area, Anderson (1961) reported a lower latite-andesite member overlain by silicic ash-flow tuffs. In the Gibbonsville-North Fork area, the same general sequence is present (Lopez, 1982).

In the central Lemhi Range, Challis Volcanics are predominantly of intermediate composition, with minor amounts of associated rhyolitic and basaltic rocks (Ruppel, 1968, 1980; Ruppel and Lopez, 1981). The largest volume of these rocks occurs in a complexly faulted area that extends from the Sawmill Canyon area (Ruppel and Lopez, 1981), northwestward into the South Fork of Big Creek and northward across the southwest quarter of the Leadore quadrangle (Ruppel, 1968) (fig. 5). Another area of volcanic rocks, in the northern part of the Patterson quadrangle (Ruppel, 1980), apparently is a separate, small volcanic field not connected originally with the sequence farther south.

VOLCANIC STRATIGRAPHY

In the Gilmore and Leadore quadrangles, eight stratigraphic units were recognized, the cumulative thickness of which is about 3,350 m, although the areas of maximum thickness of individual units do not coincide. The stratigraphic relations of the units in the central Lemhi Range are shown in figure 6*A*. Representative thin section modes are compared in figure 6*B*. Chemical analyses of representative volcanic rocks and related intrusive rocks are given in table 3 and are graphically compared in figure 6*C*.

Hornblende Andesite (Tca).—The stratigraphically lowest unit consists of porphyritic hornblende andesite flows and flow breccias; it is the most widespread and voluminous of the mapped units (unit Tca of Ruppel and Lopez, 1981; unit Tcb of Ruppel, 1968). It is thickest in the west-central Gilmore quadrangle, where it reaches about 1,500 m; it thins both northward and southward. This same unit is present as far north as the Middle

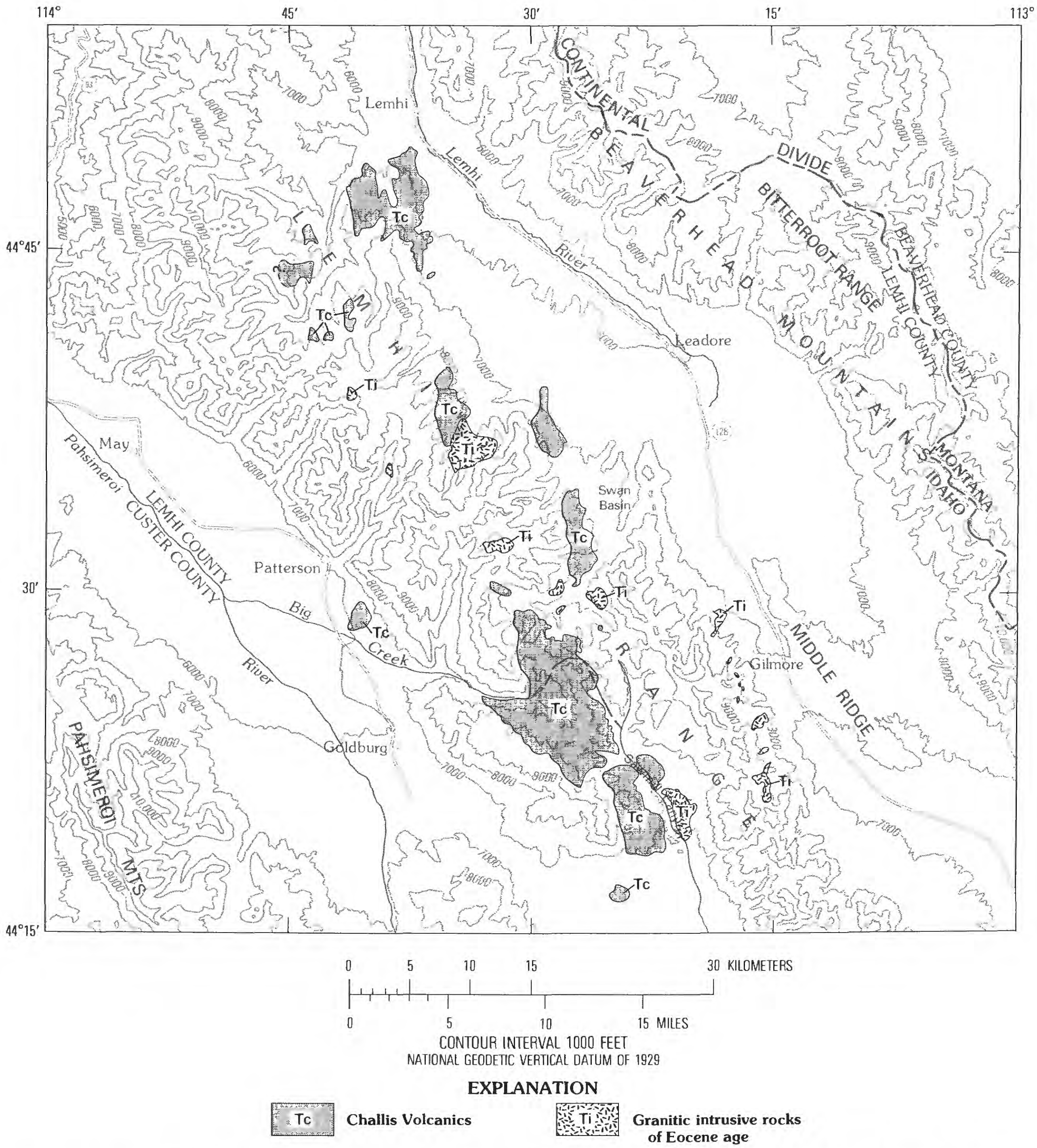


FIGURE 5.—Map showing the distribution of Challis Volcanics in and near the central part of the Lemhi Range, Idaho.

Fork of Little Timber Creek in the Leadore quadrangle (Ruppel, 1968). The unit also extends westward from the Gilmore quadrangle into the Donkey Hills quadrangle where it is present in a fault block in the South Fork of Big Creek. The contact with Paleozoic and Precambrian sedimentary rocks is everywhere poorly exposed and is typically a fault, except along the northwest flank of Bear Mountain where it is apparently depositional.

Typically this rock is medium-dark-gray to dark-gray, porphyritic hornblende andesite that weathers gray and brownish gray. Needle-shaped hornblende crystals

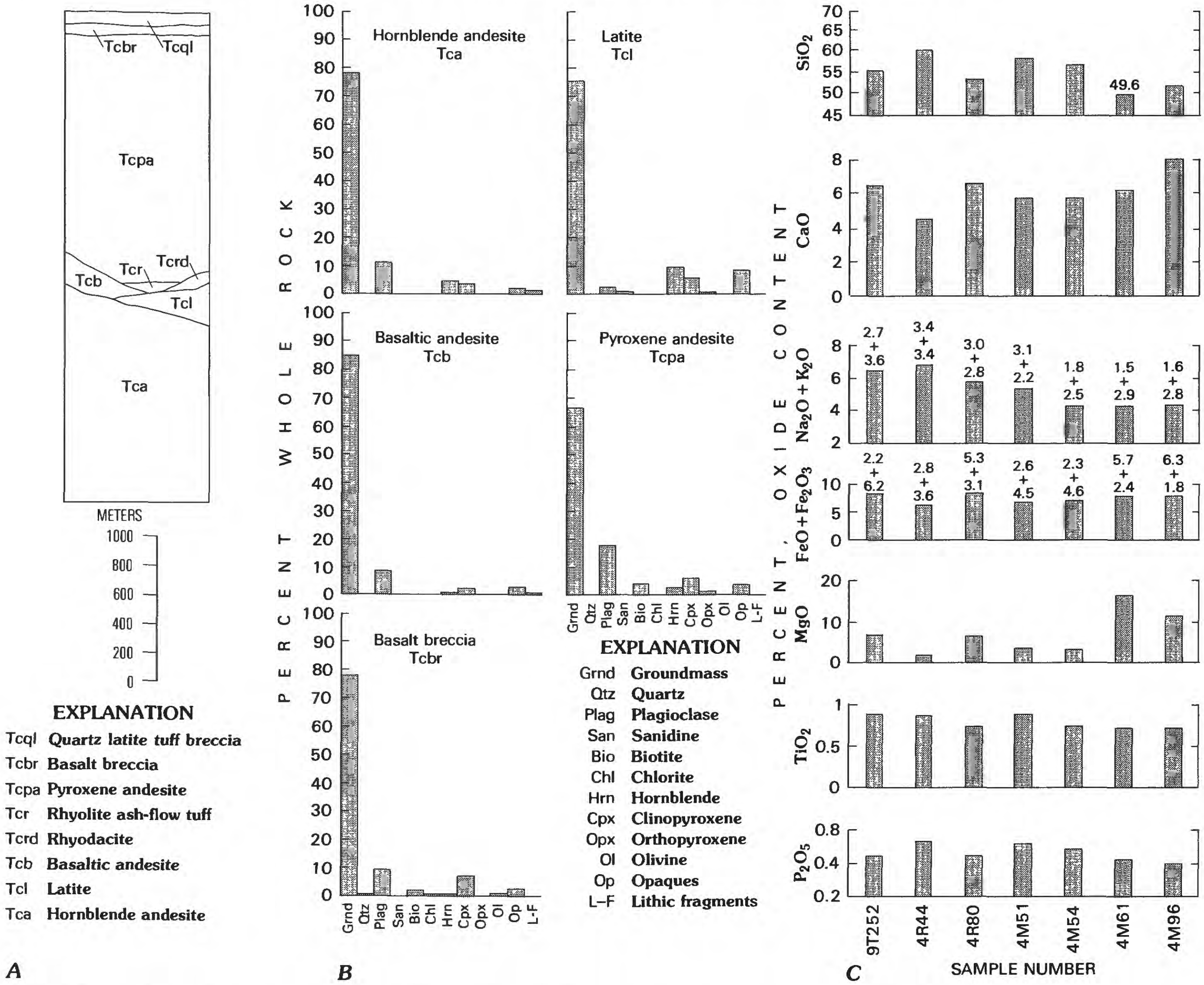


FIGURE 6.—Challis Volcanics in central Lemhi Range, Idaho. *A*, Generalized stratigraphic column of Challis Volcanics; *B*, Average modal compositions, Challis Volcanics; and *C*, Comparison of oxides in Challis Volcanics.

2–5 mm in length are the predominant phenocrysts, and are most easily recognized on weathered surfaces. Minor amounts of transparent to translucent, euhedral to subhedral plagioclase phenocrysts are present locally. The groundmass is intersertal to pilotaxitic and contains abundant plagioclase microlites and locally minor pyroxene microlites. Plagioclase composition averages An_{58}. Figure 6*B* (unit Tca) illustrates the average modal composition of the hornblende andesite.

Two small intrusive bodies in the Moonshine Creek area and a dike in the Timber Creek Pass area, are identical in appearance and composition to the hornblende andesite, and the hornblende andesite is thickest in this area. In addition, the dike in the Timber Creek Pass area is associated with abundant andesite breccia. These relations suggest that the intrusives were the vents for the extrusive hornblende andesite.

Latite (Tcl).—This unit has limited distribution, occurring only in the Gilmore quadrangle and pinching out to the north and south. The maximum exposed thickness is about 300 m in the Horse Lake area. Most of the unit is pale-red to light-brownish-gray porphyritic latite. Phenocrysts of sanidine, plagioclase (An_{35}), hornblende, and biotite make up about 25 percent of the rock (fig. 6*B*, unit Tcl). The intersertal groundmass contains microphenocrysts of augite as well as microlites of plagioclase. Locally, the latite unit contains enough magnetite (as much as 1 percent) to deflect a compass

TABLE 3.—*Chemical analyses of Challis Volcanics and related intrusive rocks in the central Lemhi Range*

[Rapid-method chemical analyses analyzed by Paul Elmore, Sam Botts, Lowell Artis, Gillison Chloe, H. Smith, J. Kelsey, J. Glenn, and C.L. Parker. Quantitative spectrographic analyses for minor elements analyzed by Norma Rait, A.L. Sutton, Jr., J.C. Hamilton, and W.B. Crandell; results are reported in percent to the nearest number in the series 1, 0.7, 0.5, 0.3, 0.2, 0.15, and 0.1, and so forth, which represent approximate midpoints of group data on a geometric scale. The assigned group for semiquantitative results will include the quantitative value about 30 percent of the time. Elements looked for but not detected are: Ag, As, Au, Bi, Cd, Eu, Ge, Hf, Hg, In, Li, Mo, Nd, Pd, Pr, Pt, Re, Sb, Sm, Ta, Te, Th, Tl, W, and Zn. 0, looked for but not detected; leaders (--), not looked for]

Field No. Lab. No. Rock type[1]	9T252 W173257 latite dike	4R44 D116004W variolitic quartz latite dike	4R80 D116005W trachybasalt dike	4M51 D116012W rhyodacite	4M54 D116013W rhyodacite	4M61 D116014W latite	4M96 D116016W olivine latite
Locality[2]	Morse Creek	Big Creek	Devils Basin Patterson Creek	Dairy Creek	Big Eightmile Creek	Dairy Creek	Stroud Creek
			Rapid-method chemical analyses (percent)				
SiO_2	55.4	60.7	53.1	58.3	57.1	49.6	51.7
Al_2O_3	13.7	14.6	14.0	16.5	13.5	10.2	11.9
Fe_2O_3	6.2	2.8	3.1	4.5	4.6	2.4	1.8
FeO	2.2	3.6	5.3	2.6	2.3	5.7	6.3
MgO	6.5	1.9	6.1	3.5	3.2	17.0	11.6
CaO	6.7	4.7	6.9	6.0	6.0	6.6	9.1
Na_2O	2.7	3.4	3.0	3.1	1.8	1.5	1.6
K_2O	3.6	3.4	2.8	2.2	2.5	2.9	2.8
H_2O^-	0.71	0.26	0.20	0.81	2.4	0.84	0.27
H_2O^+	.79	1.6	2.3	.69	1.6	1.8	1.6
TiO_2	.88	.88	.75	.88	0.76	.74	.74
P_2O_5	.45	.63	.42	.60	.51	.40	.36
MnO	.13	.15	.14	.13	.16	.14	.15
CO_2	<.05	1.2	1.7	.11	3.6	.09	.06
Sum	100	99.82	99.81	99.92	100.03	99.91	99.98
			Quantitative spectrographic analyses for minor elements				
Titanium (Ti)	--	0.3	0.3	0.3	0.3	0.3	0.3
Manganese (Mn)	--	.03	.05	.03	.05	.05	.05
Barium (Ba)	0.2	.1	.1	.1	.1	.1	.07
Beryllium (Be)	0	.0015	.0007	.0003	.0003	.0005	.0003
Cerium (Ce)	.05	0	0	0	0	0	0
Cobalt (Co)	.002	.0015	.003	.002	.0015	.005	.005
Chromium (Cr)	.015	.0005	.02	.005	.005	.1	.15
Copper (Cu)	.007	.005	.005	.0015	.0005	.002	.003
Gallium (Ga)	.0015	.0015	.0015	.002	.002	.0015	.001
Lanthanum (La)	.02	.007	.003	.007	.005	.005	.005
Niobium (Nb)	0	0	0	.001	.001	.001	1.001
Nickel (Ni)	.007	.0015	.007	.003	.002	.05	.02
Lead (Pb)	0	.0015	.0015	.0015	.0015	0	0
Tin (Sn)	.003	0	0	0	0	0	0
Scandium (Sc)	0	.002	.005	.002	.002	.003	.005
Strontium (Sr)	.2	.1	.07	.1	.05	.07	.07
Vanadium (V)	.015	.015	.02	.02	.015	.015	.02
Yttrium (Y)	.0005	.0005	--	--	--	--	--
Zirconium (Zr)	0	.015	.01	.015	.01	.01	.01

[1]Named in accordance with Rittman's classification (1952).
[2]All localities are in Patterson quadrangle (Ruppel, 1980).

needle. In the Mill Creek-Bear Creek area, the unit has been intensely hydrothermally altered adjacent to a rhyodacite porphyry stock. Alteration products include clay minerals, sericite, alunite, and goethite.

No local source is known, but the unit does not occur outside the Gilmore quadrangle, thinning both northward and southward. These relations suggest the source may have been within the Gilmore quadrangle, possibly in the Mill Creek-Bear Creek-Horse Lake area where the thickness appears to be the greatest. The rhyodacite porphyry stock in that area may be a vent of the latite unit, in which the rocks have been altered by late-stage hydrothermal solutions.

Basaltic Andesite (Tcb).—This unit is widespread in the western part of the Gilmore quadrangle; it thickens toward the center, and pinches out near the northern border of the quadrangle. A maximum thickness of 200 m occurs near the center of the Gilmore quadrangle, just north of Iron Creek. Typically, this unit is made up of dark-gray to dark-reddish-brown, vesicular to scoriaceous, basaltic andesite flows. Flow breccias commonly occur at the tops and bases of individual flows. A zone about 15 m thick in the center of the unit contains abundant globular to ellipsoidal, banded-agate amygdules from 0.5 cm to 15 cm in diameter. Sparsely scattered phenocrysts (0.1–0.2 mm in diameter) of augite and plagioclase (An_{55}) are common and olivine crystals are locally present (fig. 6*B*, unit Tcb).

The basaltic andesite flows probably were fed from a source near the area of maximum thickness near the center of the Gilmore quadrangle. In this area, between Redrock Creek and Iron Creek on the west side of Sawmill Canyon, a fault block of latite (unit Tcl) is intruded by a system of radiating basaltic dikes lithologically similar to the basaltic andesite flows. These dikes are interpreted to have been feeders to an eroded volcano from which the basaltic andesite was erupted.

Rhyodacite (Tcrd).—These flows occur only in the southern part of the Gilmore quadrangle about 3 km west of the mouth of Sawmill Canyon, and have a maximum thickness of about 70 m. This unit consists of light-olive-gray, vesicular, phenocryst-poor rhyodacite. Plagioclase, sanidine, and quartz occur as phenocrysts in a pilotaxitic groundmass that commonly contains microphenocrysts of biotite and hornblende. Commonly, biotite and hornblende are nearly completely replaced by hematite.

Rhyolite Ash-Flow Tuff (Tcr).—Rhyolite ash-flow tuff occurs only in the Cub Creek area near the mouth of Sawmill Canyon, and has a maximum thickness of 90 m. It underlies pyroxene andesite in this area, but to the north it is absent between the underlying basaltic andesite and the overlying pyroxene andesite. The tuff is very light gray to light greenish gray and moderately welded. It is phenocryst poor, containing crystals of quartz and sanidine in a glass-shard matrix. The source of the tuff is unknown.

Pyroxene Andesite (Tcpa).—This unit is widespread in the Gilmore quadrangle and extends northward across the western part of the Leadore quadrangle (unit Tcl of Ruppel, 1968). Locally in the Leadore quadrangle the unit includes thin interbeds of tuff, lapilli tuff, and tuff breccia of andesitic composition. Near the base, this unit includes lenticular flows as much as 60 m thick of partly scoriaceous dark-gray, fine-grained basalt. The maximum exposed thickness is about 1,500 m in the Adams Creek area. The predominant rocks are dark- to medium-dark-gray, porphyritic pyroxene andesite. Flow breccia is common at tops and bases of flows. Typically, the pyroxene andesite is phenocryst poor (10–20 percent phenocrysts). Phenocrysts include transparent to translucent plagioclase (An_{35}) 1–2 mm in length, augite about 0.5 mm in diameter, and locally minor amounts of biotite and hornblende (fig. 6*B*, unit Tcpa). The groundmass is glassy to pilotaxitic with microlites of plagioclase. The source of the pyroxene andesite is unknown but probably is local.

Basalt Breccia (Tcbr).—A unit of basalt breccia overlies the pyroxene andesite in the northwest quarter of the Gilmore quadrangle where it forms cliffs and steep rocky knobs and slopes. The maximum exposed thickness is about 60 m. The basalt breccia is very dark red to dark gray and monolithologic; it contains 2–30 cm angular clasts and bombs. The matrix is fine-grained, vesicular to scoriaceous basalt. Very small phenocrysts of olivine are present in places. The lithology of the basalt breccia suggests a local source, but none has been found.

Quartz Latite Tuff Breccia (Tcql).—Quartz latite tuff breccia occurs in only two small exposures, one on the ridge between Squirrel and Cabin Creeks and another in a fault sliver just north of Iron Creek. The maximum exposed thickness is about 80 m. The breccia is pale reddish brown to pale brown. The matrix is crystal ash-flow tuff containing crystals 0.1–0.5 mm long of sanidine, plagioclase, smoky quartz, and biotite. Crystals are generally broken and subhedral. Breccia clasts range from 0.25 to 2 cm in maximum dimension, and are composed of porphyritic andesite. Phenocrysts in the andesite are plagioclase and biotite. The clasts are typically angular but near the base they occur as streaks, apparently having been stretched plastically during flow.

VOLCANIC ROCKS IN THE NORTHERN PART OF THE PATTERSON QUADRANGLE

Volcanic rocks exposed in the northern part of the Patterson quadrangle cannot be correlated with the volcanic rocks farther south in the central Lemhi Range. These northern rocks are predominantly basalt flows and interbedded flow breccias and basaltic pyroclastic rocks that are medium dark gray to dark gray, fine grained, and porphyritic. They contain subhedral to anhedral phenocrysts of olivine 2–3 mm in diameter and subhedral to euhedral plagioclase as much as 1 cm long. Locally, porphyritic andesite is interbedded with the basaltic rocks, and is brownish gray to medium dark gray and fine grained, with phenocrysts of hornblende and plagioclase. The sources of this basaltic unit probably are basaltic intrusive bodies in the area of the East Fork of Hayden Creek, and the basaltic dike that cuts the Big Eightmile granitic stock (p. 67) (Ruppel, 1980).

CENOZOIC SEDIMENTARY ROCKS AND SURFICIAL DEPOSITS

Tuffaceous sedimentary rocks of Tertiary age partly fill both the Lemhi and Pahsimeroi valleys, but in most places they are concealed beneath glacial and alluvial deposits of Pleistocene and Holocene age. The Tertiary rocks are mainly reworked, water-laid, tuffaceous and partly bentonitic mudstones and fine-grained sandstones, but they include some vitric air-fall tuffs, and much fine- to coarse-grained clastic material eroded from older rocks on the mountainous flanks of the Tertiary basin. They also include major landslide plates of Tertiary and older rocks that slid from adjacent mountain flanks during and after the main period of mountain uplift in Miocene time (Beutner, 1972; Ruppel, 1982, p. 14).

A few isolated areas underlain by locally derived, coarse conglomerate in the southern part of the Gilmore quadrangle suggest renewed uplift of mountain blocks, which accompanied the rise of the Gilmore summit in late Pliocene and earliest Pleistocene time, after deposition of the Tertiary tuffaceous rocks (Kirkham, 1927, p. 11; Ruppel, 1967). These conglomerates probably are closely related to the Donkey Fanglomerate exposed a short distance west of the Gilmore quadrangle on the Donkey Hills Summit (Ross, 1947, p. 1122–1124). Other gravels, near Coal Kiln Canyon (Hait, 1965, p. 52–54) along the east front of the Lemhi Range just east of the Gilmore quadrangle, and similar gravels in the Patterson quadrangle, include rocks that cannot have come from the Lemhi Range; the gravels probably are of early Tertiary age.

Great sheets of outwash gravel extend far out into the Lemhi and Pahsimeroi valleys in front of the moraine-choked, glaciated valleys in the Lemhi Range, and reflect, as do the moraines, the episodes of Pleistocene glaciation that carved the range into its present, jagged form. A final, relatively minor episode of glaciation that was confined to high, sheltered cirques was accompanied by the growth of protalus ramparts and rock streams in the lower parts of earlier glaciated and oversteepened mountain valleys, and of alluvial fans at the mountain front. Mass-wasting on the oversteepened slopes has continued since then, with recurrent growth of protalus ramparts and rock streams and the accumulation of widespread taluses, landslide deposits and snow avalanche deposits, which, with alluvial deposits, continue to form today.

TERTIARY TUFFACEOUS SEDIMENTARY ROCKS

The Lemhi and Pahsimeroi Valleys are partly filled with Tertiary tuffaceous sedimentary rocks that probably were joined in a more continuous blanket that surrounded the upland areas in early and middle Tertiary time, before major uplift formed the present mountain ranges. These rocks also must have been joined to the extensive deposits of similar rocks, of similar ages, in southwest Montana (fig. 7). The Tertiary rocks are almost entirely restricted to the valleys now, and extend back into the main mass of the Lemhi Range only in the Swan Basin west of Leadore and in the Hayden Basin north of the Patterson quadrangle (fig. 7, locs. 17, 11) (Anderson, 1961, p. 34–35). In the Beaverhead Mountains, east of Leadore, they are preserved in and south of Bannock Pass, where they form massive landslide deposits at the head of Railroad Canyon (loc. 16); these rocks extend northward into Horse Prairie, and so also are a part of the much more widespread tuffaceous rocks in the Tertiary basins of southwest Montana. Farther north in the Beaverhead Mountains, Staatz (1979) reported Tertiary sedimentary rocks in Cow Creek (loc. 10), about 25 km northwest of Leadore. They probably are also present in and beneath the extensive landslides in the upper part of Little Eightmile Creek (loc. 13) (Staatz, 1973), which is a short distance north of the Leadore quadrangle. Any Tertiary tuffaceous sedimentary rocks originally on the mountain flanks appear to have been removed almost completely by massive landsliding into the valleys during mountain uplift (Ruppel, 1982, p. 14), and by accelerated erosion that accompanied uplift and repeated glaciation of the ranges.

The evidence for interpreting the Tertiary tuffaceous sedimentary rocks to have been more widespread at one time, and later to have been stripped from the edges of mountain blocks by landsliding and erosion, is found partly in the Lemhi Range and Beaverhead Mountains and in the Lemhi Valley, and partly in southwest Montana. The rocks do occur in a few places in the ranges,

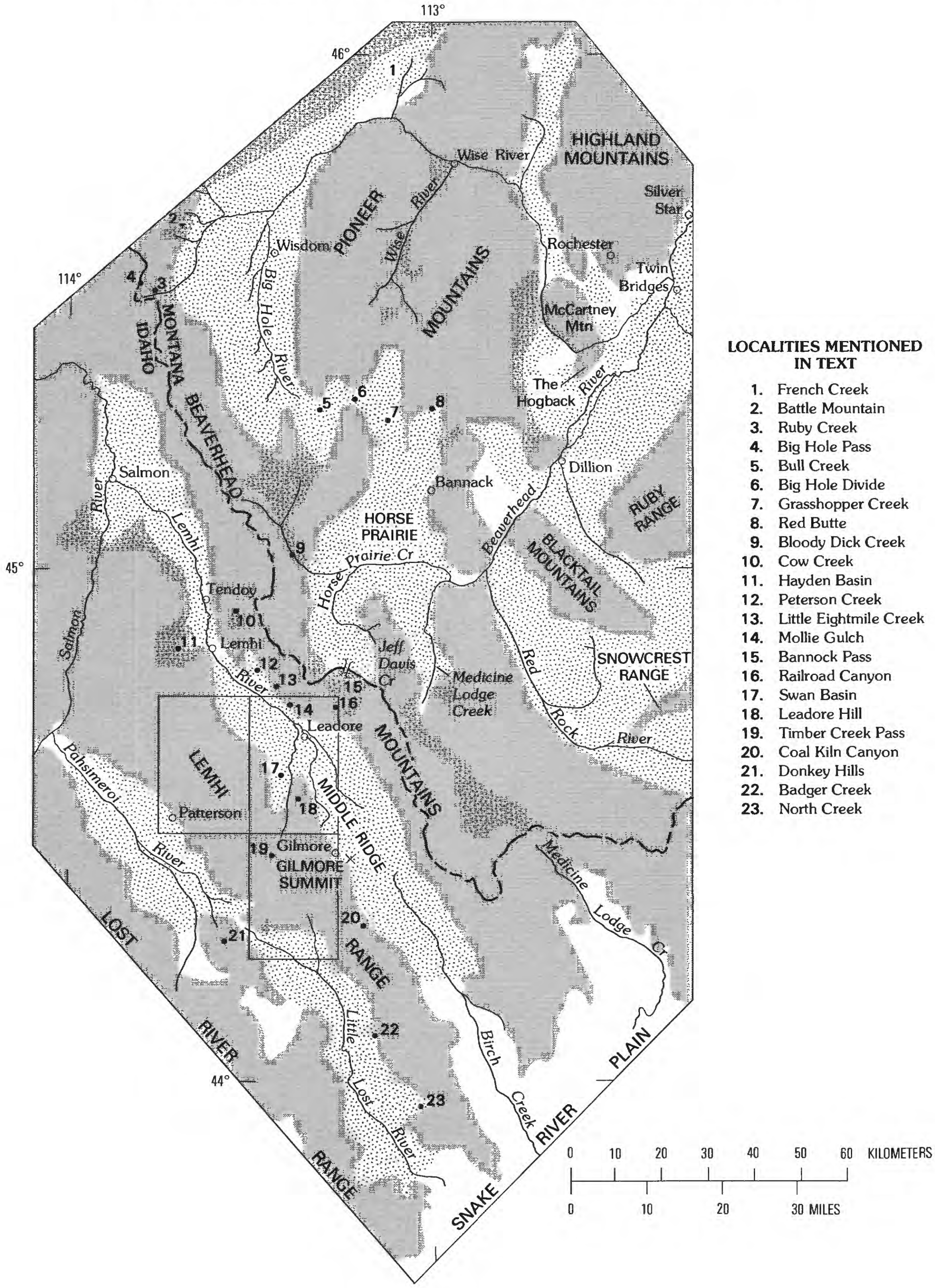


FIGURE 7.—Map showing distribution of Tertiary sedimentary rocks in east-central Idaho and southwest Montana. Tertiary sedimentary rocks patterned.

where they are in small structural basins and seem likely to have been preserved as a result of downdropping on adjacent faults—as in the Swan and Hayden Basins. Gravitational sliding of major plates of older rocks off the Lemhi Range and into the adjacent Little Lost River Valley has been described at Badger Creek and North Creek, along the southeast flank of the Lemhi Range (Beutner, 1972). Similar plates of rock, still hidden, seem likely to form a significant part of the valley fill (Ruppel, 1982, p. 13–14), and would have carried with them any Tertiary rocks originally deposited on the flanks of the mountain areas.

A well drilled southeast of Gilmore, in the head of the Birch Creek valley, reportedly penetrated coal or lignitic rocks at relatively shallow depth; lignitic rocks appear to be confined to the lower part of the Tertiary sequence. The well reportedly was bottomed in Mississippian limestone at a depth almost 1,000 m above the valley bottom, which is determined from gravity information to be at a depth of about 3,000 m (Kinoshita and others, 1969). A possible interpretation of the relations in the well is that the lignitic rocks and the Mississippian rocks are landslide plates derived from the Lemhi Range during block uplift in Miocene and Pliocene time, interleaved with Tertiary sediments deposited in the valley.

The process of stripping Tertiary tuffaceous rocks from mountain flanks by massive landsliding can be observed today in several areas: in the upper part of Railroad Canyon, in and near the northeast part of the Leadore quadrangle, where Tertiary tuffaceous rocks are sliding from the adjacent mountain flanks; in Cow Creek (Staatz, 1979), where both older Tertiary Challis Volcanics and Tertiary tuffaceous rocks are sliding off the Precambrian rocks of the valley walls, on slopes steepened by young faulting; and in the valley of French Creek, north of the Big Hole River in southwest Montana, where all of the Tertiary tuffaceous rocks are sliding off bedrock on the adjacent mountain flanks, to fill the valley with overlapping landslide deposits (Ruppel and others, 1983).

Elsewhere in southwest Montana, Tertiary tuffaceous rocks in depositional contact with older rocks are preserved deep in the ranges in several places (Ruppel and others, 1983). At Ruby Creek (fig. 7, loc. 3), west of the Big Hole Basin, lignitic tuffaceous rocks extend westward in the Beaverhead Mountains almost to Big Hole Pass (loc. 4) and the Continental Divide; the lignite was used by placer miners in 1862, some of the earliest coal mined in Montana. A short distance north of Ruby Creek, partly conglomeratic Tertiary rocks extend westward across the top of Battle Mountain (loc. 2), where they overlie deeply weathered granitic rocks. On the south flank of the Pioneer Mountains, on the east side of the Big Hole Basin, Tertiary sedimentary rocks are exposed in the upper part of Bull Creek (loc. 5), and reach eastward from there in an almost continuous blanket across the Big Hole Divide (loc. 6) into the upper part of Grasshopper Creek (loc. 7). The Tertiary sedimentary rocks in Grasshopper Creek are separated by younger faults and by granitic rocks of the Pioneer batholith from tuffaceous rocks in the upper part of Wise River Valley, in the central part of the Pioneer Mountains; these two areas of Tertiary rocks originally may have been continuous. At Red Butte (loc. 8), north and east of Grasshopper Creek, Tertiary rocks overlie deeply weathered Precambrian quartzites; Red Butte draws its name from the deep-red-colored regolith preserved at the base of the Tertiary sedimentary rocks. South of the Big Hole Basin, partly conglomeratic tuffaceous rocks crop out in Selway Creek and Bloody Dick Creek (loc. 9) extending southward almost as far as exposures of similar rocks in Horse Prairie.

In east-central Idaho, the Tertiary tuffaceous rocks are separated from the mountain blocks by steep faults in most places, but the regional relations—particularly in southwest Montana—suggest that at least the lower Tertiary sedimentary rocks were deposited on a landscape of relatively low relief, and lapped much farther back into the mountains than the present exposures suggest. Probably the early Tertiary basins in east-central Idaho were interconnected, as those in southwest Montana still are, but were separated by middle and late Tertiary and Quaternary block uplifts. The Tertiary rocks were removed from the edges of the mountain blocks by landsliding and accelerated stream and glacial erosion.

The maximum possible thickness of Tertiary sedimentary rocks in the Lemhi Valley is about 3,000–3,500 m, the approximate depth to older bedrock in the valley between Leadore and Gilmore on the basis of gravity information (Kinoshita and others, 1969); the amount of duplication due to Miocene and younger landsliding is unknown. As a guess, the original thickness of the Tertiary sedimentary rocks in the Lemhi Valley might have been about half that suggested in the upper part of the valley, perhaps about 1,500 m. This is about the same thickness as that of similar rocks in the Pahsimeroi Valley, which is only half as deep as the Lemhi Valley (Kinoshita and others, 1969).

Nomenclature.—The Tertiary sedimentary rocks in the north part of the Lemhi Valley were divided into the Kenney, Geertson, and Kirtley Formations, in ascending order (Anderson, 1957, p. 16–19; 1961, p. 31–35), but later studies have shown that these formations cannot be recognized and mapped consistently even in their type areas (Staatz, 1979, p. A23; Harrison, 1982). Harrison (1982) suggested that the Kenney and Geertson Formations are lithologically indistinguishable, and that the Geertson and Kirtley Formations are confused by problems with their original definition and use. Accordingly, the names Kenney Formation, Geertson Formation, and Kirtley Formation are abandoned in this report on the grounds that they are not clearly recognizable and mappable lithologic units; they cannot be recognized and mapped in their type areas or in any adjacent area in the Lemhi Valley.

The Tertiary tuffaceous sedimentary rocks overlie Challis Volcanics with angular unconformity in the Salmon area (Anderson, 1956, p. 31, pl. 1) and near Lemhi Pass (Staatz, 1979, p. A23), and overlie Precambrian and lower Paleozoic rocks with angular unconformity in the Swan Basin in the upper part of the Lemhi Valley. The Tertiary rocks are overlain and widely concealed by Quaternary glacial and alluvial deposits and by a widespread blanket of windblown silt as much as 1 m thick.

Description.—Tertiary tuffaceous sedimentary rocks are well exposed in many places in the upper part of the Lemhi Valley, but equivalent rocks in the Pahsimeroi Valley are almost entirely concealed (Mapel and others, 1965; Mapel and Shropshire, 1973). In the upper part of the Lemhi Valley, the oldest Tertiary tuffaceous rocks are exposed in the northeast part of the Patterson quadrangle and the northwest and west-central parts of the Leadore quadrangle. Younger Tertiary tuffaceous rocks are exposed on Middle Ridge in the south part of the Leadore quadrangle and northeast corner of the Gilmore quadrangle.

The older Tertiary rocks, north and west of Leadore, and in the upper part of Railroad Canyon, are dominantly yellowish-gray to very light gray, tuffaceous, calcareous siltstone and fine- to medium-grained sandstone, although they include an appreciable proportion of conglomerate that contains a large variety of rock fragments. The coarsest and most abundant conglomerates are interbedded with tuffaceous siltstone and sandstone along the flank of the Beaverhead Mountains near Mollie Gulch, northwest of Leadore. Conglomerate forms 20–40 percent of the exposed tuffaceous sequence here in lenticular beds 0.2–1 m thick, that range from beds composed almost entirely of angular fragments a centimeter or less in diameter, to beds composed of subangular to subrounded boulders and cobbles 25–40 cm in diameter in a sandy, tuffaceous matrix. The rock fragments were derived locally, mainly from Precambrian and Paleozoic rocks in the adjacent Beaverhead Mountains, but also from Challis Volcanics to the west along the front of the Lemhi Range. The texture and composition within any single conglomeratic unit is fairly constant, but both change abruptly from one bed to another.

The Tertiary tuffaceous rocks exposed in the northeast part of the Patterson quadrangle and southward into the Swan Basin are very pale orange to yellowish- or light-gray, fine-grained tuff and tuffaceous sandstone and interbedded siltstone and pale-brown bentonitic mudstone, all commonly in beds 1–6 m thick. These rocks locally include small lenses of well-sorted, rounded to subangular fragments, as much as 5 mm in diameter, of Kinnikinic Quartzite and less commonly of Precambrian quartzites and Challis Volcanics. More persistent beds of conglomerate, such as those near Mollie Gulch, are rare. The few persistent beds of conglomerate that are present are composed of well-rounded fragments of quartzite as much as 20 cm in diameter, in a tuffaceous, sandy matrix.

The rocks that underlie Middle Ridge, east of Gilmore, are the youngest Tertiary sedimentary rocks known in the Lemhi Valley. Their relation to the rocks near Mollie Gulch is unknown, because the intervening area is covered with younger alluvial gravels, but they are almost continuous with exposures of Tertiary sedimentary rocks around the north end of Leadore Hill into the Swan Basin, and northward from there into the northeast corner of the Patterson quadrangle. These exposures suggest that the Tertiary rocks between the sequence at Mollie Gulch and the sequence on Middle Ridge probably are mostly light gray or yellowish-gray, tuffaceous, calcareous fine-grained sandstone and siltstone, of unknown thickness. The rocks that underlie Middle Ridge are principally pale orange to yellowish-gray and very light gray, friable, bentonitic, vitric tuffs, in beds commonly less than 1 m thick but in places as much as 5 m thick. About 20 percent of the sequence is composed of interbeds, also commonly less than 1 m thick, of light-grayish-brown, tuffaceous, poorly sorted, fine- to medium-grained sandstone and siltstone that is partly conglomeratic. The basal 6–25 cm of each interbed commonly includes lenses of abundant, well-rounded pebbles, 1–5 cm in diameter, of Kinnikinic Quartzite, in a tuffaceous matrix. The sequence also includes a few interbeds of very light gray, finely crystalline fresh-water limestone, in beds a few centimeters thick and in groups of thin beds, separated by shaly partings, as much as 1 m thick. The limestone beds in places contain abundant fossils of fresh-water gastropods.

Throughout the upper part of the Lemhi Valley, erosional surfaces cut on the tuffaceous rocks commonly are veneered with a lag gravel made up of small, angular fragments of quartzite, mostly of Kinnikinic Quartzite, derived from weathering of quartzitic pebbles, cobbles, and boulders in the tuffaceous rocks.

Description of tuffaceous rocks in the Salmon area.—Regional mapping in the Salmon area (Ruppel and others, 1983) suggests that the Tertiary tuffaceous sedimentary rocks there can be subdivided into two stratigraphic units. The lower unit, overlying the Challis Volcanics with angular unconformity, is dominantly olive gray and yellowish-gray to grayish-orange, tuffaceous, bentonitic mudstone and shale, which contains brownish-gray layers and beds, commonly 2–15 cm thick, of lignite, lignitic mudstone, and beds with abundant woody fragments. The lignite and lignitic mudstones are characteristic of the unit, and do not occur higher in the Tertiary sequence. The lower unit also includes some interbeds as much as 1 m thick of tuffaceous fine-grained to very fine grained, prominently crossbedded sandstone, and lenticular interbeds of conglomerate from 1 to 8 m thick. The conglomerate is composed mainly of fragments of quartzite and siltite from the Yellowjacket Formation, most commonly in beds of closely packed pebbles a centimeter or so in diameter, but including a few beds of cobbles and boulders as much as 0.5 m in diameter, in a sandy, tuffaceous matrix. Conglomerate beds in the basal part of the lower unit also include cobbles and boulders derived from the underlying Challis Volcanics. The uppermost beds of the unit appear to be mainly grayish orange to yellowish-gray, tuffaceous, fine-grained to very fine grained, thick-bedded to massive sandstone. The greatest unfaulted thickness of the lignitic unit is about 250 m; the total thickness is unknown.

The lower lignitic sequence is gradationally overlain by a much thicker and more widespread group of light-olive-gray and yellowish-gray to yellowish-brown or yellowish-orange, tuffaceous, bentonitic mudstone and shale, fine- to medium-grained, tuffaceous sandstone, and lenticular interbeds of conglomerate with a sandy, tuffaceous matrix. This upper unit probably is at least 600–700 m thick. The lower part of the unit is mainly mudstone and shale, with widely spaced interbeds and lenses of sandstone and conglomerate; the sandstone and conglomerate beds and lenses commonly are 0.3–8 m thick, separated by as much as 30 m of shale and mudstone. The upper part of the unit includes more sandstone and conglomerate, separated by thinner sequences of mudstone and shale; in some places, it is entirely sandstone and conglomerate. Throughout the unit, the conglomerates are composed mostly of fragments of the Yellowjacket Formation, with subordinate fragments of other Proterozoic sedimentary rocks.

The conglomerates in the upper part of the upper unit appear to grade westward from coarse conglomerates at the eastern margin of the Salmon basin into interbedded red and yellow tuffaceous conglomerate, sandstone, mudstone, and shale in the central part of the basin. At the east margin of the basin, overlying the Yellowjacket Formation, the tuffaceous rocks are dominantly moderate reddish brown, coarse, bouldery conglomerate and interbedded tuffaceous red shale and sandstone, all in beds as much as 50 m thick. The conglomerate contains unsorted or poorly sorted, subangular to subrounded pebbles, cobbles, and boulders from 1 cm to 1 m in diameter. The conglomerate beds thin westward, and become finer grained. In the central part of the Salmon basin, these rocks contain well-rounded to subrounded or, less commonly, subangular pebbles and cobbles 3–10 cm in diameter. In this area, red-colored mudstone and sandstone beds and less abundant conglomerate beds are interbedded with gray and yellow tuffaceous mudstones and sandstones.

The relations suggest that the red, coarse conglomerates and the interbedded red mudstones and sandstones are alluvial fan deposits on the periphery of the Salmon basin that interfinger westward in the central part of the basin with lighter colored, dominantly tuffaceous, finer grained rocks. Harrison (1982, 1985) described similar relations in the northern part of the Salmon basin, where the northernmost outcrops represent basin margin talus and proximal fan facies, which grade southward into distal fan facies and ultimately into a complex fluvial and lacustrine system.

Age and regional relations.—Tuffaceous sedimentary rocks in the Lemhi Valley range in age from Eocene(?), or Oligocene(?) to Pliocene, but only Miocene and Pliocene rocks are exposed in the south part of the valley. The oldest rocks are those of the lignitic unit that forms the lower part of the unnamed sequence above the Challis Volcanics near Salmon. These rocks have yielded plant remains that at one time were considered to be of Oligocene, perhaps middle Oligocene age (Ross, 1962a, p. 97), but which recently have been reinterpreted to be tentatively of Eocene age (Harrison, 1982). Tuffaceous sedimentary rocks in the valley of Cow Creek, near Lemhi Pass, are lithologically similar to some of the rocks near Salmon, and contain a fossil flora similar in some respects to the flora in the Salmon rocks. The oldest possible age of the Cow Creek flora is late Eocene, and the plant assemblage does not extend into the Miocene (Staatz, 1979, p. A24–A25). Staatz considered the Cow Creek rocks to be most probably of Oligocene age.

In the central part of the Lemhi Valley, tuffaceous sedimentary rocks at Peterson Creek, a short distance northwest of the Leadore quadrangle, contain a distinctive vertebrate fauna that establishes the age of these rocks as early Miocene (Nichols, 1975, 1979). About 10 km farther southwest, in the northwest corner of the Leadore quadrangle, tuffaceous sandstones and siltstones interbedded with conglomerate along Mollie Gulch and Little Eightmile Creek contain vertebrate fossils of middle Miocene age (Schultz and Falkenbach, 1947, p. 186–187; Nichols, 1975, p. 10). Tuffaceous sandstones near Bannock Pass, 8–10 km north of the northeast corner of the Leadore quadrangle, contain fossil bone fragments of camels, oreodonts, and small horses (G. Edward Lewis, written commun., 1965); these include:

- *Merychippus* (*Merychippus*) sp. (large sp.), 6 upper and 18 lower cheek teeth or fragments thereof.
- *Hypohippus* sp., crown of upper cheek teeth.
- ?oreodont, genus and species undetermined, but reminiscent of *Brachycrus*.
- *Merychippus* sp. (size of *M. primus*), subgenus and species undetermined, three upper cheek teeth and four incisors.
- *Merychippus* sp. (large sp.), crowns of two and one-half upper cheek teeth and one lower cheek tooth (all milk teeth).
- ?camelid of *Aepycamelus* size, fragments of cranium and two cheek teeth.
- ?oreodont of *Brachycrus* size, crown of cheek tooth.
- ?*Procamelus* sp., fragmentary of left ramus with molars.

Lewis commented that these fragmentary fossils all suggest a late Miocene age.

The tuffaceous rocks on Middle Ridge, at the head of the Lemhi Valley east of Gilmore, contain vertebrate fossils of Pliocene age (Nichols, 1975, p. 10). The fresh-water limestone interbeds in the tuffs contain diatoms and fresh-water gastropods probably of late Tertiary age (Knowles, 1961, p. 73; Dwight W. Taylor, written commun., 1964).

The age and stratigraphic position of these rocks near Salmon have been the subject of uncertainty and controversy in many early reports. Both Anderson (1956, p. 30–31; 1961, p. 33–34) and Ross (1962a, p. 96–102) discussed the age of the Tertiary rocks near Salmon, based on the fossil flora collected there, but reached different conclusions. This was partly because of uncertainties about where the fossils were collected, partly because of differing interpretations and reinterpretations of the age of the fossil flora, and partly because the vertebrate fossils from the central and upper parts of the Lemhi Valley were lumped together and used to date the rocks near Salmon. Anderson (1961, p. 32–35) finally concluded that all of the rocks in the Salmon area were of Miocene age. Ross (1962a, p. 97) concluded that they most probably were Oligocene, perhaps middle Oligocene, and suggested that the range of ages represented by the vertebrate fauna was so wide that the fossils must have come from a number of different stratigraphic horizons not represented in the Salmon area.

The range of ages of the tuffaceous sedimentary rocks exposed at different places throughout the length of the Lemhi Valley suggests that a fairly complete stratigraphic sequence is present. The oldest rocks are the lignitic unit near Salmon, tentatively of Eocene age, and the somewhat similar rocks in the valley of Cow Creek, which could be as old as late Eocene; both the Salmon and Cow Creek rocks are certainly at least as old as Oligocene. These are overlain by tuffaceous sandstone, siltstone, and mudstone, and interbedded conglomerate that at Peterson Creek, Mollie Gulch, and Railroad Canyon are of early to late Miocene age. The youngest rocks are the vitric tuffs and interbedded limestones on Middle Ridge at the head of the Lemhi Valley, which are of Pliocene age. The Miocene rocks in the Peterson Creek and Mollie Gulch areas are similar lithologically to the upper, conglomeratic unit in the Salmon area, and the rocks near Salmon probably also are of Miocene age, although they have not yielded any fossils. Recent studies of the vertebrate fossils from the central and upper part of the Lemhi Valley by Nichols (1975, 1979) make clearer the ages of the Tertiary sedimentary rocks in different parts of the Lemhi Valley, and help resolve the earlier conflicts in age assignments.

The lithologic characteristics of the Tertiary sedimentary rocks in the Lemhi Valley are broadly similar to those of Tertiary rocks in the Jefferson and Three Forks basins of southwest Montana, and suggest that the lower unit of mostly fine grained sedimentary rocks and lignite in the Salmon area is about equivalent to the Renova Formation of Oligocene age, and that the overlying, more conglomeratic rocks are about equivalent to the Sixmile Creek Formation of Miocene and Pliocene age (Kuenzi and Fields, 1971; Thompson and others, 1981; Hughes, 1981; Robinson, 1965, 1967). The lignitic rocks of the Lemhi valley are similar to lacustrine, lignitic rocks in Horse Prairie that have yielded a fish fauna of early to middle Oligocene age (Cavender, 1977). Some of the tuffaceous sandstones, siltstones, and conglomerates above the lignitic unit in the Salmon area and in the central part of the Lemhi Valley are lithologically similar to tuffaceous rocks in the Big Hole Basin, which also are of Miocene age (Hanneman and Nichols, 1981). The upper Miocene rocks in the vicinity of Bannock Pass extend northward into Horse Prairie, and have been correlated with the Medicine Lodge beds of Scholten and others (1955, p. 369–370) and tentatively with the "Blacktail Deer Creek basalts" of the Lima region (Scholten and others, 1955, p. 370). Other formations described in the Lima region (Scholten and others, 1955, p. 368–370)—the "Sage Creek basalts," the Cook Ranch Formation, and the "Muddy Creek beds"—are similar in age to some of the Tertiary rocks in the Lemhi Valley, but the Lemhi rocks lack the lava flows and other volcanic materials associated with the formations in the Lima region.

The occurrence, in the Lemhi Valley, of Tertiary rocks that are similar in lithologic characteristics and age to those in the basins of southwest Montana suggests the conclusion reached by Thompson and others (1982, p. 106), that "***the episodes of basin filling (in this region) cannot have occurred in response to local tectonic or volcanic events, but must record regionally synchronous events."

OTHER TERTIARY CONGLOMERATES AND RELICT GRAVELS

Donkey Fanglomerate and correlative(?) deposits.—The Pliocene(?) Donkey Fanglomerate of Ross (1947, p. 1122-1124) was named for and described at exposures in and near the Donkey Hills, which form the divide betweeen the Little Lost and Pahsimeroi Rivers a few kilometers west of the southwest corner of the Gilmore quadrangle (fig. 7). Similar conglomerates, probably roughly correlative with the Donkey Fanglomerate, cap the ridge crest on Timber Creek Pass at the head of Sawmill Canyon in the northwest part of the Gilmore quadrangle and crop out in a thin remnant farther southeast at the mouth of Sawmill Canyon. The fanglomerate is as much as 300 m thick in the Donkey Hills, 120-150 m thick on Timber Creek Pass, and only a few meters thick at the mouth of Sawmill Canyon. In all these areas, it overlies Precambrian and lower Paleozoic rocks with angular and erosional unconformity; in the valley of the South Fork of Big Creek, just west of the Gilmore quadrangle, it similarly overlies Challis Volcanics. Its relation to Tertiary tuffaceous sedimentary rocks is not so clear, although it apparently overlies tuffaceous rocks near Goldburg School in the upper part of the Pahsimeroi Valley (Ross, 1947, p. 1122). A deposit of similar, coarse conglomerate also is present about 25 km farther south, near the west flank of Hawley Mountain (Mapel and Shropshire, 1973), where it overlies upper Paleozoic rocks and Challis Volcanics.

The fanglomerate that caps Timber Creek Pass at the head of Sawmill Canyon is a flat-lying deposit of very coarse conglomerate composed of well-rounded boulders of Kinnikinic Quartzite as much as 3 m in diameter, 75-90 percent; well-rounded boulders of Middle Proterozoic quartzites 0.5-1 m in diameter, 10-25 percent; and subordinate amounts of dolomite from the Saturday Mountain and Jefferson Formations, and of granitic rocks like those in the nearby Lake Creek stock, all in well-rounded cobbles and boulders 0.3-1 m in diameter. The matrix between the boulders and cobbles is pebbly sand derived from quartzitic and granitic rocks. The conglomerate is grayish red to brownish gray, or less commonly light olive gray, and is prominently and coarsely crossbedded. All of the rocks in the conglomerate are exposed nearby, and it is interpreted to be a coarse alluvial fan or valley fill deposit derived from the high, eastern crest of the Lemhi Range a short distance farther north and east, mixed with a small percentage of locally derived boulders and cobbles of Paleozoic dolomitic rocks.

The small deposit of conglomerate at the mouth of Sawmill Canyon is composed mainly of cobbles of Challis Volcanics and of Middle Proterozoic quartzites; it is only a few meters thick. Its composition suggests that it was derived from farther northwest in Sawmill Canyon, and not from the east, where Kinnikinic Quartzite is widely exposed.

The conglomeratic rocks in Sawmill Canyon are lithologically similar to the Donkey Fanglomerate (Ross, 1947, p. 1122-1124). Like the Donkey Fanglomerate, however, they do not contain any fossils and their stratigraphic position is not so clear that they can be directly and closely dated. The presence of fragments of Tertiary intrusive rocks and of Challis Volcanics in the conglomerate indicates that it is younger than the Eocene age of these rocks, and because the conglomerate has been scoured by glacial ice and is overlapped by glacial deposits, it is older than at least some Pleistocene deposits. The single outcrop near Goldburg School in the Pahsimeroi Valley suggests that the Donkey Fanglomerate overlies Tertiary tuffaceous sedimentary rocks, but the relations are not well exposed. Nevertheless, Ross (1947, p. 1123) concluded that the Donkey Fanglomerate probably is of Pliocene(?) age, and a similar age for the conglomeratic rocks in Sawmill Canyon seems most likely, a conclusion also reached by Hait (1965, p. 52-54).

The deposits in Sawmill Canyon, those in the Donkey Hills, and those near Hawley Mountain, are all on or near the axis of the broad, northeast-trending arch that extends through the Gilmore and Donkey Hills summits (Kirkham, 1927, p. 11; Ruppel, 1967, p. 657-658; 1982, p. 17). This suggests that they were deposited in alluvial fans and in mountain valleys as a result of arching or of climatic change in late Pliocene and early Pleistocene time (Pierce and Scott, 1982).

Relict gravels near Coal Kiln Canyon and Patterson Creek.—Hait (1965, p. 52-54) described relict gravel deposits near Coal Kiln Canyon, southeast of the Gilmore quadrangle along the east face of the Lemhi Range, that are unlike the Sawmill Canyon and Donkey Hills deposits. These gravels cannot have been derived locally, because they are composed of cobbles and small boulders, from 2 to 40 cm in diameter, of Kinnikinic Quartzite and Proterozoic quartzites that are not present on the east flank of the Lemhi Range in this area.

In addition to the Kinnikinic and Proterozoic quartzites described by Hait, the gravels near Coal Kiln Canyon also include small, well-rounded cobbles of very coarsely feldspathic quartzite that is not present in the Lemhi Range. These unique rocks are known to occur only in the Clayton Mine Quartzite in central Idaho (Hobbs and others, 1968, p. J15–J17), almost 100 km west of Coal Kiln Canyon and now separated from the canyon by several major mountain ranges. A small deposit of somewhat similar, deeply weathered and exfoliated, well-rounded cobbles of feldspathic quartzite is thinly scattered on the high divide between the West Fork of Big Eightmile Creek and the East Fork of Patterson Creek, in the central part of the Patterson quadrangle; it is probably a relict from an old conglomerate or gravel deposit. It is possible that it also was derived from the Clayton Mine Quartzite or from the Cash Creek Quartzite of central Idaho (Hobbs and others, 1968, p. J18–J19). Such exotic gravels must have been deposited by streams that flowed east or southeast from central Idaho in early Tertiary time, before any uplift of mountain ranges ancestral to the present ones. The early Tertiary, Paleocene or Eocene, pre-Challis Volcanics age assigned to similar gravels by Anderson (1959, p. 21) and Ross (1962a, p. 86), and to the gravels near Coal Kiln Canyon by Hait (1965, p. 54), seems appropriate.

GLACIATION AND GLACIAL DEPOSITS

The high peaks and deep U-shaped valleys of the central part of the Lemhi Range show the cumulative effects of valley glaciation in abundant and sometimes spectacular detail (figs. 2, 8). Sculpturing by glacial ice has left the range a seemingly jumbled mass of sharp, ice-carved peaks and jagged ridges. Deep valleys reach far into the range, and once carried major ice streams that were collected from small tributary glaciers, from wide compound cirques, and from a mountain ice field, to form trunk glaciers that in places flowed out onto the floor of the Lemhi Valley on the east side of the range.

Tills and outwash gravels that accumulated as a result of glaciation suggest that these surviving deposits are a result of three major episodes of glaciation and a later, final episode of ice accumulation that was limited to high, sheltered cirques. The oldest of these, probably of early Pleistocene age, is represented by a few deposits of boulders and of alluvial gravels of similar composition that are remnants of outwash gravels. Deposits of the intermediate and latest glaciations, both of later Pleistocene age, are more widespread, and more clearly of glacial origin. Moraines of intermediate age are rounded and dissected by later streams, but retain much of their glacial form. Outwash deposits of intermediate age locally overlap older outwash and in turn are overlapped by outwash deposited during the youngest major glaciation. The moraines of the youngest major glaciation cover the floors of most of the mountain valleys; the terminal moraines left by these great valley glaciers commonly are nestled just within those of the intermediate age glaciers where ice flowed out of the mountain valleys onto the floor of the Lemhi Valley.

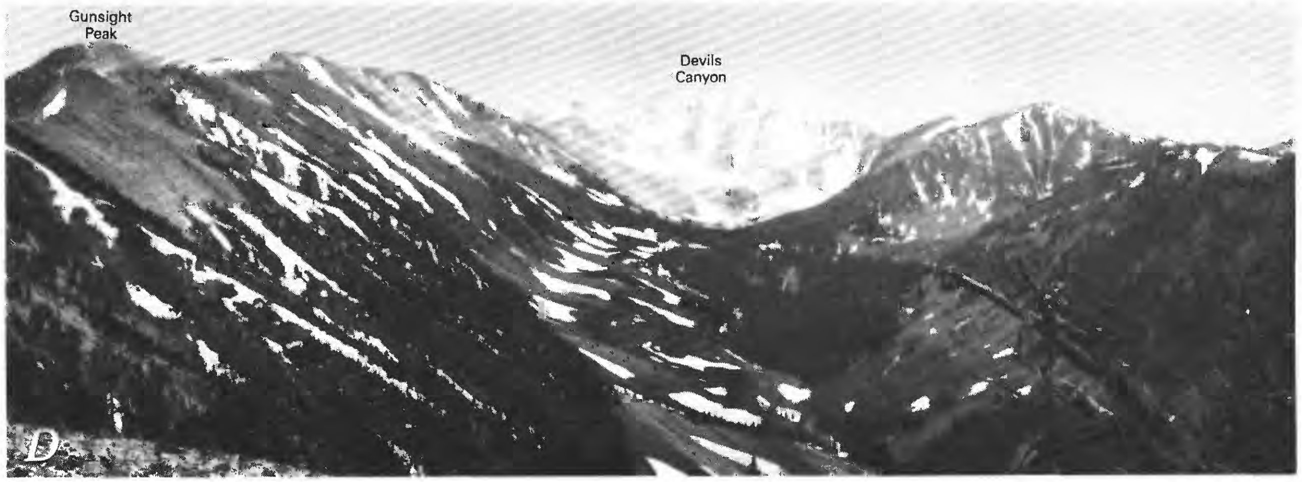
The distribution of glacial erosional features and of glacial deposits indicates that the central part of the Lemhi Range was the principal center of ice accumulation in the range (fig. 9), and that only a few small valley glaciers were present in other parts of the range. The extent of glacier ice in the early and intermediate episodes of glaciation is uncertain, because almost no erosional features remain in the mountains that can be attributed clearly to these earlier episodes. Their deposits of till are farther from the mountain front than those of the last major glaciation, which suggests that the early and intermediate glaciers were somewhat larger than the later glaciers. In the last major glaciation, ice accumulated to form a thick and nearly complete mountain ice field in the axial part of the range; only peaks and ridges that now are higher than about 3,000 m rose above the ice in this area. The central ice field and other glaciers in adjacent, high cirques fed major trunk glaciers that occupied all the valleys draining the central part of the range. The ice thickness in most of the valleys was about 200 m, but in the deepest valleys, like that of the North Fork of Big Creek in the south part of the Patterson quadrangle, these major ice streams were as much as 300 m thick, judging from the present distribution of ice-carved features, till, and glacial erratic boulders.

The ice flowed more or less radially from the axial part of the range, but in valleys on the west and southwest sides of the range it melted so rapidly that these valleys were glaciated only in their upper reaches, principally in upper tributaries that face north or east; the lower parts of these valleys are deep, V-shaped gorges, washed clean of till and of most alluvial deposits by meltwater; no moraines are present on the floor of the Pahsimeroi Valley along the west flank of the Lemhi Range. In contrast, the glaciers that flowed east into the Lemhi Valley, and north into the Hayden Basin, everywhere reached the mountain front, and between Big Eightmile Creek and Mill Creek, northwest of Leadore, they reached 3–5 km east of the mountain front onto the valley floor. In both the Hayden Basin and the Big Eightmile-Mill Creek area, the mountain glaciers coalesced at the mountain front to form piedmont glaciers.

FIGURE 8.—Glacial erosional features in the central Lemhi Range.

A, Horseshoe Gulch and the head of Spring Mountain Canyon. Thrust-faulted dolomite of the Saturday Mountain and Jefferson Formations exposed on the cirque walls. Looking south. Gilmore quadrangle. 1970.

B, Cirque wall of south-dipping Kinnikinic Quartzite, head of Meadow Lake Creek. A protalus rampart is in the central part of the picture. Looking west. Gilmore quadrangle.



C, Panoramic view southeast along the east flank of the Lemhi Range. U-shaped glaciated valley is North Fork of Little Timber Creek. Beaverhead Mountains in background. 1963.

D, Devils Canyon, a north-trending glaciated tributary of Big Eightmile Creek, showing extensive rock streams and avalanche deposits on the canyon floor, derived from Proterozoic and Ordovician quartzite on the valley walls. Looking south. Patterson quadrangle.

E, U-shaped, glaciated canyon of East Fork of Hayden Creek looking south. Ice flowed across divide west of Mill Mountain from glacier in Mill Creek farther south. Rocks exposed on Mill Mountain and on cliffed peaks west of East Fork of Hayden Creek are Swauger Quartzite cut by multiple west-dipping imbricate thrust faults. Patterson quadrangle. 1967.

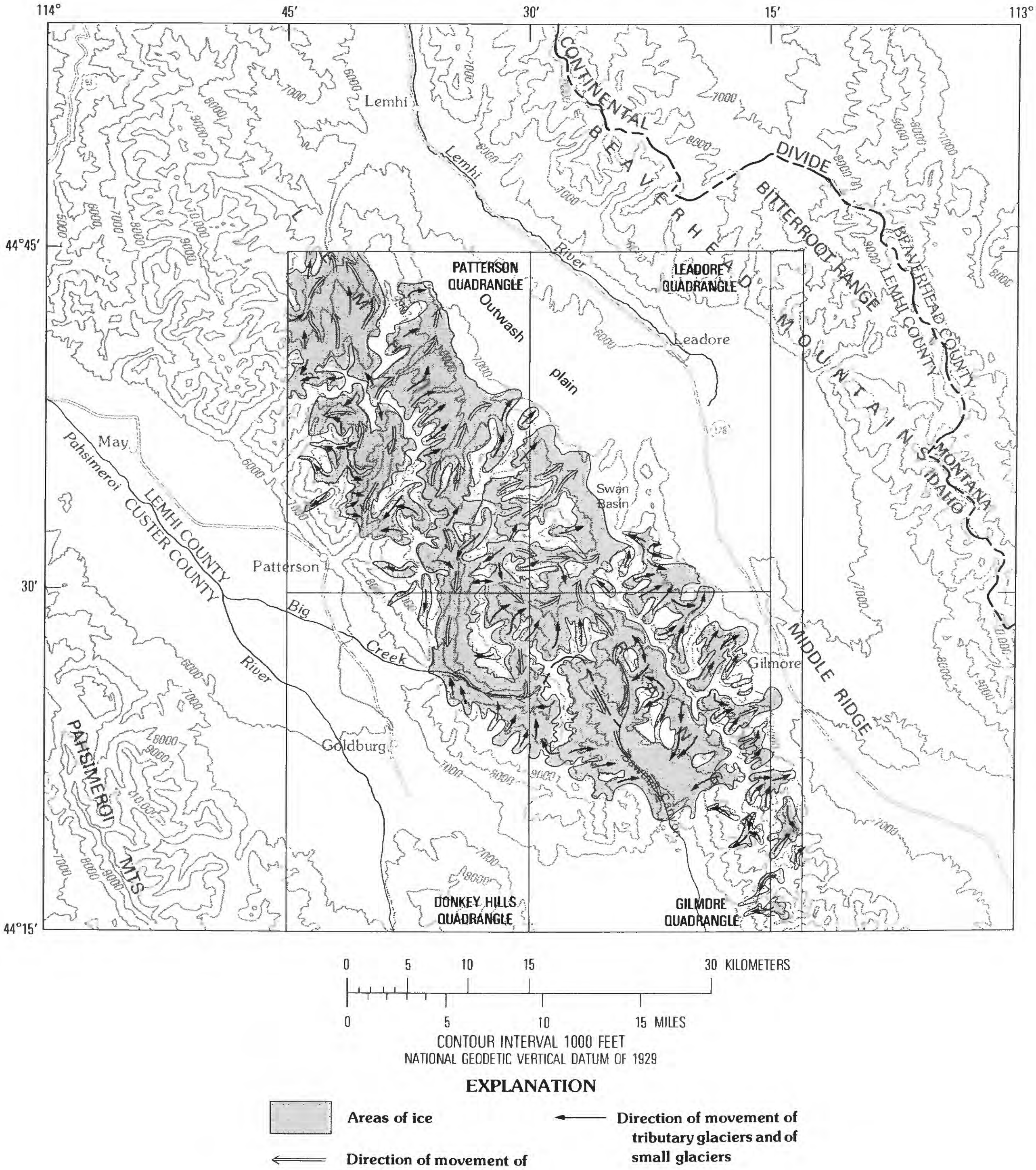


FIGURE 9.—Map showing distribution of glacier ice in central Lemhi Range during last glacial stage, interpreted from distribution of till, glacial erratics, and ice-carved features.

A smaller piedmont glacier formed on the west side of the Swan Basin from the merging of the glaciers in the different forks of Little Timber Creek and those on the east flank of Rocky Peak. The Swan Basin piedmont glacier probably also included ice from the Rocky Creek and Big Timber Creek glaciers, but later erosion and postglacial faulting have separated the deposits of till, and the prefaulting relations are not clear. Neither the

Rocky Creek nor the Big Timber Creek glaciers have terminal moraines, and the area below where they originally merged in Big Timber Creek is cut by north-trending faults that have moved since glaciation; the valley of Big Timber Creek below Rocky Creek is a relatively narrow, alluvium-filled stream-cut valley that links the upper glaciated canyons to the till-filled Swan Basin (fig. 10).

The peaks and ridges that rose above the ice in the axial part of the range are mostly bare rock or frost-riven outcrop. Although many ridges are sharp, nearly knife-edged aretes, some high ridge crests are broad, gently sloping or nearly flat surfaces of hummocky alpine tundra underlain by loose rock waste that is creeping downslope. In other areas in central Idaho, similar high surfaces have been interpreted to be remnants of an old erosional surface (Ross, 1934a, p. 93). In the central part of the Lemhi Range, however, these surfaces invariably are controlled by gently west dipping thrust faults; the mechanically disaggregated rocks most commonly are from breccia zones just above the thrust fault surfaces. Similar thrust fault surfaces in many places on cirque walls and along glaciated canyons have been completely cleaned by glaciation and creep, leaving bare, silicified and iron-oxide-coated surfaces that clearly show their relation to thrust faults (fig. 11).

Minor features of glacial erosion are common in the cirques and ice-carved canyons and canyon walls of the region. Small mountain lakes, or their successor ponds and bogs, occupy ice-cut rock basins in many of the cirques, although the floors of most of the smaller cirques now are covered by talus, protalus ramparts, and snow-avalanche deposits. Valleys that were occupied by small glaciers tributary to the main ice streams commonly are now hanging, typically far above the floor of the main valleys. In the central part of the range, rock steps as much as 100 m high are present above the junctions of some tributary glaciers that contributed substantial amounts of ice to the trunk glacier, but near the margins of the range these steps are absent, perhaps because no single tributary glacier there could contribute enough ice to significantly affect the main ice stream. Ice-rounded, streamlined, fluted, and striated outcrops of bare rock form the floors of many of the larger, compound cirques. Polished outcrops are preserved also on some canyon walls and floors, including the upper canyon of the North Fork of Big Creek, where glistening, mirrorlike polish is preserved on a cliff of Proterozoic quartzite more than 30 m high.

The till deposited by the glaciers is preserved in terminal moraines near the mountain front on the east side of the Lemhi Range, in medial moraines where major ice streams merged, in lateral moraines along the edges of some glaciated valleys, and in recessional and ground moraines on the floors of most glaciated canyons. The deposits representing each of the preserved major stages of glaciation have been separated on the basis of (1) differing degrees of destruction of original glacial landforms such as rounding of moraine and filling of kettles; (2) integration of drainage across the deposits; (3) degree of weathering and destruction of rock fragments of similar composition in each of the tills; and (4) amount of soil development and vegetative cover. Alluvial gravels interpreted to be outwash are correlated with the tills on the basis of similarities in composition and weathering characteristics.

Deposits of the early glaciation.—Coarse boulder deposits, interpreted to have been deposited by early Pleistocene glaciers, form an arcuate fringe of low, rounded knobs and hills at the north edge of the Swan Basin, and also are present in a small area just north of the mouth of Big Eightmile Creek. The boulder deposits overlie Tertiary tuffaceous rocks, and are overlain by till of the intermediate and youngest glaciers, which also fills channels cut into the old boulder deposits. The boulders are thinly scattered across the weathered surface of the rounded hills; most of them are partly buried in deep soil. The boulder deposits may be as much as 30 m thick, but in most places they are only a thin veneer—perhaps at most a few meters thick. The arcuate form of the deposits in the Swan Basin suggests the shape of a broad terminal moraine, but the deposits do not have any other topographic features that suggest moraine, and drainage across them is completely integrated.

The boulder deposits consist predominantly of Kinnikinic Quartzite in unsorted, subangular to subrounded pebbles, cobbles, and boulders, which are commonly from 1 cm to 1.5 m in diameter, but may be as much as 5 m in diameter. Quartzite and siltite from Proterozoic formations commonly form about 1–5 percent of the boulder deposit, but may be as much as 25–30 percent in a few areas where these rocks are in nearby outcrops. The Proterozoic rocks mostly are in frost-shattered fragments that are 10 cm or less in diameter, but include a small percentage of subround cobbles and boulders as much as a meter in diameter. The vitreous Kinnikinic Quartzite boulders are not significantly weathered, but commonly are wind faceted, and may be covered by desert varnish and stained yellow to a depth of about 1 cm. The Proterozoic quartzites and siltites are more deeply weathered, commonly to a depth of several centimeters. The deposits also contain a small percentage of small angular fragments of dense volcanic rocks, but they do not contain any carbonate or granitic rocks, even though these are common in the source areas of the boulder deposits.

FIGURE 10.—The valley of Big Timber Creek below the former location of the Big Timber Creek and Rocky Creek glaciers, and the upper part of the Swan Basin. Looking southwest from Leadore Hill. Leadore quadrangle.

FIGURE 11.—Thrust fault surfaces.

A, Nearly flat interstream divide controlled by an imbricate thrust fault in the Apple Creek Formation, west of Golden Trout Lake. Looking north. Patterson quadrangle. 1963.

B, Surface of a thrust fault that places Gunsight Formation on top of Kinnikinic Quartzite, stripped by glacial scour and creep on the south wall of the valley of the North Fork of Little Timber Creek. Patterson quadrangle. 1963.

Dort (1962, p. 9–10) described somewhat similar bouldery deposits on Middle Ridge, east of Gilmore, and concluded that they were deposited by early Pleistocene glaciers from the Gilmore area, in Long Canyon, Meadow Creek, Negro Green Creek, and Nez Perce Creek. The deposits are now fault bounded on the west, and appear to have been uplifted about 100 m since early glaciation. Although the deposits are similar in composition and weathering characteristics to other bouldery gravels interpreted to be till, their glacial origin is uncertain. They are almost 3 km from the mountain front at Gilmore and from the terminal moraines of younger glaciers, a relation that suggests that the regional snow line would have to have been significantly lower than it was during the younger glaciations (K.L. Pierce, written commun., 1985). In contrast, the bouldery deposits interpreted to be old moraine in the Swan Basin are only about a kilometer from the moraines of younger glaciations. An alternative and perhaps more likely interpretation of the Middle Ridge deposits is that they are alluvial gravels deposited by high-gradient streams, and that the gravels have been exposed as a result of subsequent uplift and erosion.

Alluvial gravels similar in composition and weathering characteristics to the boulder gravels in the Swan Basin and at the mouth of Big Eightmile Creek are widely distributed on the crowns of the flat-topped hills that flank the Lemhi River northwest of Leadore; these gravels are interpreted to be remnants of a once much more extensive blanket of alluvial gravels that may be outwash related to the earliest glaciation in the Lemhi Range. The deposits overlie Tertiary tuffaceous rocks, and in turn, are overlain by outwash gravels of the younger glaciations. In many places along the Lemhi River, however, tributary streams have cut through the old outwash into the underlying tuffaceous rocks, and outwash gravels of the intermediate and youngest glaciations occupy successively lower positions in the tributary valleys and gulches (Ruppel, 1967, p. 653). The old alluvial gravels are poorly sorted, well-rounded fragments of Kinnikinic Quartzite and Proterozoic quartzite and siltite that range in diameter from less than a centimeter to as much as 20 cm, in a matrix of sand, silt, and clay that forms 50–70 percent of the deposit. The matrix typically includes tuffaceous material derived from the underlying rocks.

Deposits of the intermediate glaciation.—Till in smoothed and rounded moraines, outwash of composition similar to the till in sheets in the Lemhi Valley, and alluvial fans in the valley of the Pahsimeroi and Little Lost Rivers represent the intermediate glacial episode. The intermediate-age moraines are present at the mouths of most of the canyons along the east flank of the central Lemhi Range, where they form a narrow and discontinuous fringe of low, soil-covered, smoothed and rounded terminal moraines just outside the much fresher terminal moraines of the late glacial stage. Drainage is well integrated across the moraines, and most originally closed basins or kettles are either filled, or are drained by streams that dissect the moraines.

The till predominantly consists of Kinnikinic Quartzite and Proterozoic quartzite and siltite in differing proportions depending on the availability of these rocks in the source areas, but it also includes smaller amounts of carbonate and granitic rocks where these rocks are present in the source areas. The Kinnikinic quartzite fragments are subangular to subrounded without visible weathering rinds. These fragments are most commonly pebbles, cobbles, and boulders less than 1 m in diameter but may be as much as 5 m in diameter. Proterozoic rock fragments commonly are a meter or less in diameter, and typically are rounded with weathering rinds about 1 cm thick. Deeply weathered and pitted, rounded cobbles and small boulders of dolomite and limestone, generally less than 3/4 m in diameter, are common but very minor components of the till; smaller cobbles and pebbles of carbonate rocks apparently have been dissolved. Similarly, granitic rocks are represented by rounded cobbles and small boulders that have weathering rinds several centimeters thick. Fragments less than 10–20 cm in diameter have disintegrated, although they are evident as weathered ghosts composed of loose grains, in fresh exposures of the till. Volcanic rocks in angular to rounded fragments from a few to about 30 cm in diameter are abundant in till near the source areas of these rocks, but only dense, basaltic rocks are present away from the source areas.

Alluvial gravels compositionally similar to these tills, and interpreted to be outwash deposits of the intermediate glaciation, form broad blankets in the Lemhi Valley, and widespread alluvial fans in the Pahsimeroi Valley and the upper part of the Little Lost River Valley. The deposits in the Lemhi Valley overlap the older gravels of the early glaciation near the mouth of Big Eightmile Creek, but in most other places northwest of Leadore and in the Swan Basin, they overlie tuffaceous rocks on benches cut by streams that dissected the early glacial gravels and cut into the underlying tuffaceous rocks during the interval between the early and intermediate glaciations. The gravels interpreted to be of intermediate glacial age in the Pahsimeroi Valley and the valley of the Little Lost River are in broad, smoothed alluvial fans formed by outwash streams from the glaciers on the west side of the Lemhi Range. These fans are large compared to the outwash blanket on the east side of the range, probably because they received virtually all of the rock eroded by ice in the upper part of glaciated valleys, and none is preserved as till. The gravels are predominantly composed of Kinnikinic Quartzite and Proterozoic quartzite and siltite, in rounded fragments typically less than 30 cm in diameter, with weathering characteristics like those of similar rocks in the till. They may contain as much as 10–20 percent limestone, dolomite, and granitic rocks, in deeply weathered pebbles and small cobbles. The granitic rocks commonly are so deeply weathered that they are a loose aggregate of sand, which contributes to the sand, silt, and clay that form the matrix of the gravel deposits.

Deposits of the late glaciation.—The youngest episode of glaciation is represented by till that blankets most of the valley bottoms in the central part of the Lemhi Range, extends high on the flanking valley walls, and forms massive lateral, medial, and terminal moraines and broad sheets of ground moraine, and by outwash gravels. The distribution and extent of the glaciers that left these deposits are shown on figure 9. The moraines have been only slightly modified by postglacial erosion; in most places, original depressions and kettles in the moraine are undrained, and are either unfilled or contain shallow ponds or bogs. Soils developed on the till are thin and rocky. Well-developed terminal moraines, and combined terminal and lateral or medial moraines are present at the mouths of many of the valleys on the east side of the range (fig. 12). Where valley glaciers merged into broad piedmont glaciers at the mountain front, the till is partly ground moraine, separated by sharp medial moraines that commonly are bedrock-cored in their upper parts, and terminates in an irregularly arcuate end moraine. The principal streams draining the glaciated valleys have cut sharp channels through the moraines, but there is no other drainage integration; the moraines remain much as they were deposited originally except for the thin mantle of rocky soil.

The till of the young glaciation is composed of whatever rocks are in the source area or areas, in fragments that are essentially unweathered. Quartzites derived from both the Kinnikinic Quartzite and Proterozoic formations are the principal constituents, the Kinnikinic occurring in angular to subangular blocks as much as 5 m in diameter, and the Proterozoic quartzites and siltite in subangular to subrounded, somewhat smaller blocks typically a meter or less in diameter. Limestone and dolomite are common in the till, in unweathered subangular fragments as much as 1.5 m in diameter. Granitic rocks are abundant in some areas, and include very large joint blocks as much as 6 or 7 m across. The granitic rocks have weathering rinds about a centimeter thick, and large blocks typically are partly surrounded by exfoliation shells. Small, angular fragments of volcanic rocks are common in the till in the Swan Basin, in the area between Big Eightmile Creek and Mill Mountain, and in the upper part of Hayden Creek, where these rocks crop out.

FIGURE 12.—Glaciated valley and compound terminal moraine in Meadow Lake Creek, north of Gilmore, Idaho, showing fresh, undissected younger moraine, and older, dissected and rounded intermediate moraines. Looking south. Liberty Gulch on left. Photograph by Warren B. Hamilton.

Alluvial gravels related to the young glaciation are widespread along the flanks of the Lemhi Range in outwash sheets in the Lemhi Valley, and in alluvial fans in the Pahsimeroi Valley and the valley of the Little Lost River. The outwash sheets in the Lemhi Valley partly overlap older glacial outwash gravels, and partly were deposited in stream valleys cut through these gravels between glacial episodes, so that the young outwash overlies tuffaceous rocks on benches just above the present stream bottom, below the benches veneered with intermediate glacial outwash gravel, and well below the outwash remnants of the early glaciation, which cap the adjacent hills. The alluvial fans on the west and southwest sides of the Lemhi Range overlap and are somewhat smaller than the fans deposited during the intermediate glaciation. The deposits on both sides of the Lemhi Range retain original depositional features on their upper surfaces. The composition of the gravel mirrors that of the young till, but the almost unweathered, subrounded to rounded fragments are smaller, typically pebbles, cobbles, and boulders less than 0.5 m in diameter. As in the till, limestone, dolomite, granitic rocks, and volcanic rocks are relatively abundant in the alluvial gravels, depending upon the source area.

Deposits of cirque glaciers.—The youngest glacial deposits in the central part of the Lemhi Range occupy high, northeast- and north-facing, sheltered cirques. These fresh, unweathered, and undissected deposits are composed of coarse, angular boulders standing about at the angle of repose. The deposits, which mark the outer limit of ice that accumulated during this episode of cirque glaciation, are confined to the cirques and do not extend beyond the cirque lip. In more exposed cirques, and lower in the valleys, where ice did not accumulate, this period of climatic change is represented by large accumulations of mass-wasted rock fragments in talus, protalus ramparts, and rock streams, all of which continue to accumulate in the near-glacial climate that prevails in the central Lemhi Range today.

Alluvial deposits related to cirque glaciation are incorporated in young alluvial fans on the flanks of the range and in alluvial deposits associated with present streams.

MASS-WASTING DEPOSITS

The landscape of the central Lemhi Range is dominated by the effects of glacial erosion. The stripped rock surfaces and oversteepened valley walls and cirque headwalls eroded by glacier ice yield abundant rock waste that forms talus, protalus ramparts, and rock streams. These deposits show that a major agent of alpine erosion since glaciation (in addition to transport in solution) has been frost-wedging and downslope gravitational movement of angular rock fragments from the ice-carved cliffs that bound the cirques and form the valley walls, a process that is still active in the rigorous climate that prevails in this region today. Frost-wedged rock waste also contributes to the debris carried into the valleys by snow avalanches from the oversteepened valley walls, and forms a major part of the rockslides or debris falls that have partly filled many cirques. Other commonly much larger landslides have formed on steep, glaciated valley walls covered by a veneer of till; in unglaciated areas underlain by tuffaceous volcanic rocks or sediments; and in a few areas that were not glaciated but were so close to glaciers that the exposed rocks were deeply fractured and weathered as a result of nearly glacial extreme climate.

Talus, protalus ramparts, and rock streams.—Deposits of fresh, angular rock waste in talus, protalus ramparts, and rock streams are common throughout the central Lemhi Range. Talus is particularly abundant in cirques, where it deeply fringes the base of the headwall cliffs and is nearly as abundant along the bottoms of cliffs on the flanks of glaciated canyons; it is present throughout the region wherever rocks are exposed in cliffy outcrops. Protalus ramparts are present in many of the glaciated canyons, most commonly on their north-facing flanks, where snow is protected from the sun and can accumulate to greater depths along cliff bottoms. They typically form an outer rim, outside the taluses, and generally are in three distinct sets; an old, partly dissected set that is rounded and covered by thick soil and vegetation; an intermediate set that is only slightly dissected and covered by thin and rocky soil; and a young set that is fresh, undissected, and standing at the angle of repose. The sets are successively overlapped, and in some places only the youngest set is present; the older sets, if they ever were deposited, have either been buried or destroyed. All three sets overlap till of the youngest major glaciation, and so are post-glacial. Their relation to the later episode of cirque glaciation is not clear, but a possible interpretation is that the oldest ramparts accumulated during the recessional stage of the last valley glaciers, as the valleys were cleared of ice, and that the intermediate ramparts accumulated during the later cirque glaciation. The youngest, fresh ramparts still receive rock waste during severe winters.

Rock streams and talus flows are present in many of the glaciated valleys, and grade almost imperceptibly upstream into massive taluses and protalus ramparts that provide abundant supplies of waste rock. A few, like the rock streams in Inyo Creek east of Patterson, blanket most of the adjacent valley floor, but most of them are rather small, lobate, flow-wrinkled deposits on the south sides of major valleys where they are fed by large taluses and protalus ramparts (fig. 13). They overlie glacial deposits, and their characteristic steep frontal slopes and lack of vegetation suggest comparatively recent movement.

Snow avalanche deposits.—Deposits of debris carried down by snow avalanches and from the precipitous chutes that feed them are present everywhere in the central Lemhi Range, and show that snow avalanches are a major hazard to winter travel in this remote region (fig. 14). The deposits overlap to cover large areas in some valleys, like those of Patterson Creek and Inyo Creek, and in others include single cones of avalanche

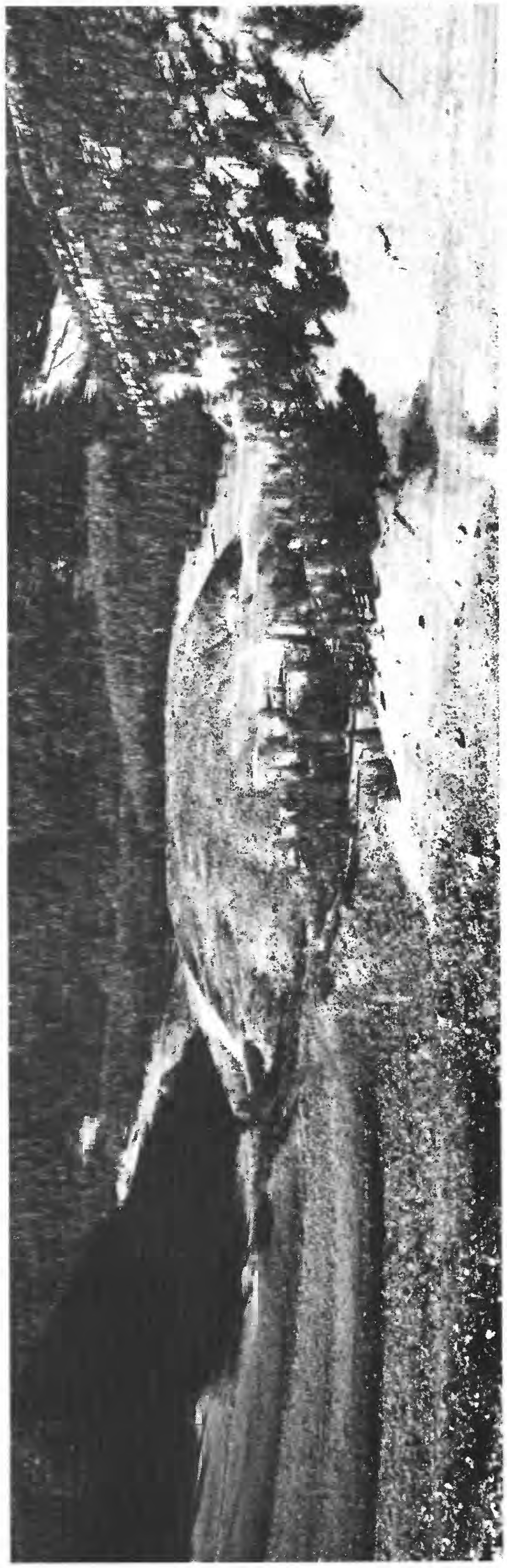


FIGURE 13.—Rock stream in the Middle Fork of Little Timber Creek. Looking north. Patterson quadrangle.

debris, or larger fringes of coalesced deposits that in places clog or dam the present drainage. All of them show the destructive power of snow avalanches: most of the deposits are littered with broken trees, and the edges of the chutes are lined with trees broken or heavily scarred and stripped of branches on their upstream sides. Large, angular boulders and tree trunks, oriented with their long axis in the direction of avalanche flow, are mixed, unsorted, into the cone-shaped deposits of mostly smaller, flow-oriented, angular fragments of rock waste. The destructive impact of the snow commonly extends beyond the edge of the avalanche deposit, into a tangled mass of broken and blown-down timber on the opposite valley side.

Landslides.—Landslides are nearly as characteristic of the central Lemhi Range as snow avalanche deposits, and like the avalanches, many of them are active today. Small debris slides are common in cirques and on some mountain flanks where they consist of an open rubble of angular rock fragments, retaining the fresh, chaotic and hummocky appearance of recent landslides. Larger and far more extensive landslides have formed where till, earlier plastered on steep, glaciated canyon walls, is sliding off the walls into the canyons below. The depressions on these hummocky landslide surfaces commonly are boggy or water filled. These deposits retain their original soil cover and generally support at least moderately abundant vegetation—including trees leaning at different angles or with curved trunks, that show continuing movement of the slide.

Some landslides, like the debris flow near Grove Creek in the upper part of the Swan Basin, have moved so recently that the torn and displaced sod cover has not yet regrown. Most landslides seem to be creeping slowly, but are almost stable under present conditions. Any change in those conditions can lead to more rapid or even catastrophic movement, like that on the Hayden Creek landslide, on the north flank of Mogg Creek (secs. 16, 21, and 29, T. 16 N., R. 23 E.), which failed catastrophically a few years after logging operations, nearly damming Hayden Creek and destroying the upper part of the Hayden Creek road. Other recently active or still creeping, large landslides are common where glacial till or other surficial deposits overlie tuffaceous volcanic or sedimentary rocks. The large landslides near the mouth of Sawmill Canyon, and the landslide on the east face of Gunsight Peak, are composed partly of glacial till and partly of weathered volcanic rocks, sliding on tuffaceous units in the Challis Volcanics. All of them have the hummocky, locally boggy appearance of landslides that have moved recently; they appear to be stable or nearly so under present conditions.

Young faults, marked partly by linear scarps, cut many of the landslides, or are partly buried by them.

FIGURE 14.—Snow avalanche chute and associated rubbly deposit on the south flank of Gunsight Peak, damming the valley of the North Fork of Little Timber Creek. Looking north. Patterson quadrangle. 1962.

The common association of faults and landslides suggests that much of the landsliding has been caused by the earthquakes that accompany faulting.

ALLUVIAL DEPOSITS

Alluvial gravel blankets much of the Lemhi Valley, in a series of broad, coalescing alluvial fans of several ages. These fans include some gravels that almost certainly were deposited before glaciation, but now are mixed with other gravels deposited at least partly during glaciation. Holocene gravels form small alluvial fans where present streams leave their mountain canyons, and gravels, sand, and silt form alluvium in the beds of the present streams.

Alluvial gravels not clearly of outwash origin, but deposited as a result of increased snowmelt during glacial episodes and partly reworked later (Pierce and Scott, 1982, p. 698–699; Funk, 1976), form broad, coalescing, smoothed alluvial fans in the upper part of the Lemhi Valley north of the Swan Basin and east and northeast of Gilmore, and along the west flank of the Beaverhead Mountains. These gravels overlie Tertiary tuffaceous rocks, and were deposited on a surface in places cut below the boulder gravel of the early glaciation, probably of early Pleistocene age. They mainly consist of Proterozoic, Ordovician, and Pennsylvanian quartzite and siltite, Challis Volcanics, and a small proportion of limestone, dolomite, and granite. Proterozoic and Pennsylvanian quartzitic rocks are weathered to a depth of about 1 cm; Ordovician quartzites are essentially unweathered. Carbonate rocks, volcanic rocks, and granite tend to be absent in the finer grained, distal parts of the fans, but are present as pebbles, cobbles, and boulders in the coarser deposits near the mountain front. The coarsest boulders commonly are about a meter in diameter, and are subangular to well rounded; most fragments less than 2 cm in diameter are angular, probably broken fragments of originally more rounded pebbles and cobbles. The matrix of the gravel is sand, silt, clay, and tuffaceous silt derived from the underlying tuffaceous sedimentary rocks. The fans appear to include materials originally deposited before glaciation, perhaps including pediment gravels of early Pleistocene age. In and near Railroad Canyon they include deeply weathered cobbles and smaller fragments of Archean crystalline metamorphic rocks that have been reworked from still older gravels that cap Bannock Pass and were deposited there originally by south-flowing streams from areas underlain by these crystalline rocks in southwest Montana. But mostly the gravels seem to be a mixture of outwash from the major glaciations and of other gravels eroded simultaneously from unglaciated parts of the Beaverhead Mountains; all these gravels have been reworked by later streams that planed smooth the surfaces of old fans. Somewhat similar old alluvial fans are widespread in the Pahsimeroi Valley and the upper part of the Little Lost River Valley south of Sawmill Canyon, although the gravels in these fans are clearly related to glaciation in the Lemhi Range.

The old alluvial fans are overlapped by younger fans along the mountain fronts on both sides of the Lemhi Range and on the west side of the Beaverhead Mountains. The younger fans include coarser material, such as boulders as much as 2 m in diameter, particularly in areas near the mountain front. The younger fans are small compared to the older fans, and do not reach more than 2–3 km beyond the mountain front except in a few places where deposits of larger streams have coalesced to extend 8–10 km from the mountain front. Most of them have steep surface gradients, 50–100 m/km as compared to about 20 m/km on the older fans; their surfaces are characterized by depositional features, like linear boulder trains that mark the former courses of streams, which show that the alluvial materials have not been reworked and smoothed like those on the old alluvial fans. The younger fans are in two sets, one that is being dissected by present streams and seems to be composed of alluvial boulder gravels eroded during cirque glaciation, and a still younger set, not quite as coarse grained, that probably is being deposited by flash floods in the present streams. Both of them are accumulations of unweathered subangular to subrounded pebbles, cobbles, and boulders, in a matrix of sand, silt, and clay, that have been deposited partly or entirely by present streams in late Pleistocene through Holocene time. Similar materials form the subrounded to well-rounded gravel, and sand, silt, and clay, that make up the alluvium along the courses of present streams.

GRANITIC INTRUSIVE ROCKS

The sedimentary rocks in and near the central part of the Lemhi Range have been intruded by several varieties of granitic rocks of Eocene age, and by granitic rocks of early Paleozoic age in the Beaverhead Mountains. The Eocene granitic rocks in the central part of the Lemhi Range are in small stocks along or near the flanks of the range, and in sheets, peripheral to many of the stocks, that have been intruded into thrust faults. Only three other small granitic bodies of Eocene age are known in east-central Idaho: the Little Eightmile pluton, probably a sheet, intruded near the base of the Medicine Lodge thrust plate northwest of Leadore (Staatz, 1973); the Carmen stock, a complex granitic body northeast of Salmon, intruded into the Yellowjacket Formation and younger Proterozoic rocks (Mackenzie, 1949, p. 15–21; Anderson, 1959, p. 34–36; Kilroy, 1981); and the Bobcat intrusive, intruded into the Yellowjacket Formation on Napoleon Ridge in the Salmon River Mountains near North Fork, Idaho (B.B. Bunning and F.W. Burnet, Cominco American, Inc., written commun., 1981).

Other granitic rocks, which are widely distributed in the central and south parts of the Beaverhead Mountains, are of early Paleozoic age (Ruppel, 1968; Scholten and Ramspott, 1968, p. 18–21). These granitic and syenitic rocks are intrusive into Proterozoic and Ordovician sedimentary rocks in the Medicine Lodge thrust plate. The northernmost exposures are in a complexly faulted thrust slice along the front of the Beaverhead Mountains near Leadore (Ruppel, 1968). Farther south, these rocks include the granite of Bull Canyon in a thrust slice near Hawley Creek, east of Leadore (Lucchitta, 1966, p. 86–88; Staatz and others, 1972, p. B51–B56), and the Beaverhead pluton, a large, thrust-faulted pluton in the south part of the Beaverhead Mountains (Scholten and others, 1955, p. 370–372; Scholten and Ramspott, 1968, p. 18–20).

PETROGRAPHY

Granitic rocks in the Lemhi Range and in the Beaverhead Mountains include granite, monzogranite and quartz monzonite, granodiorite, quartz monzodiorite, and tonalite, quartz diorite, and diorite in some marginal zones and dikes. These rocks are classified and named in general accordance with the system recommended by the IUGS Subcommission on the Systematics of Igneous Rocks (Streckeisen, 1973) (fig. 15*A*). These names replace the more general names quartz monzonite and granodiorite used on the maps of this region (Ruppel, 1968, 1980; Ruppel and Lopez, 1981) (fig. 15*B*). Virtually all these rocks are porphyritic, in that they contain conspicuous phenocrysts; the term phenocryst is used without genetic connotation for those crystals that stand out from a finer grained groundmass. Chemical analyses of representative intrusive igneous rocks are given in table 4, and are graphically compared in figure 16. Representative thin-section modes are compared in figure 17. Stratigraphic, structural, and radiometric evidence for the ages of granitic rocks is discussed later in this report.

EARLY PALEOZOIC GRANITE IN THE BEAVERHEAD MOUNTAINS

Granitic rocks along the flank of the Beaverhead Mountains north of Leadore and in Railroad Canyon are moderate- to light-pink porphyritic granite and alkali granite. These hypidiomorphic to xenomorphic porphyritic rocks typically contain 35–42 percent quartz, 30–45 percent alkali feldspar (although some rocks contain as much as 60 percent alkali feldspar), and 15–30 percent plagioclase (although plagioclase is less abundant or absent in some rocks). Alkali feldspar and plagioclase occur as individual subhedral to anhedral crystals of microcline and albite, and as perthite in the groundmass. Some microcline grains have albite overgrowths. Quartz and microcline commonly are micrographically intergrown. Accessory minerals, commonly 1–3 percent of the granite, include biotite, opaque minerals, zircon, and apatite. The groundmass typically is fine grained (0.1–0.2 mm), but in places is somewhat coarser, as much as 0.5 mm, and in granulated rocks, as many of these are, groundmass crystals are 0.01–0.05 mm in diameter. Phenocrysts are quartz and microcline in subhedral to anhedral crystals commonly as much

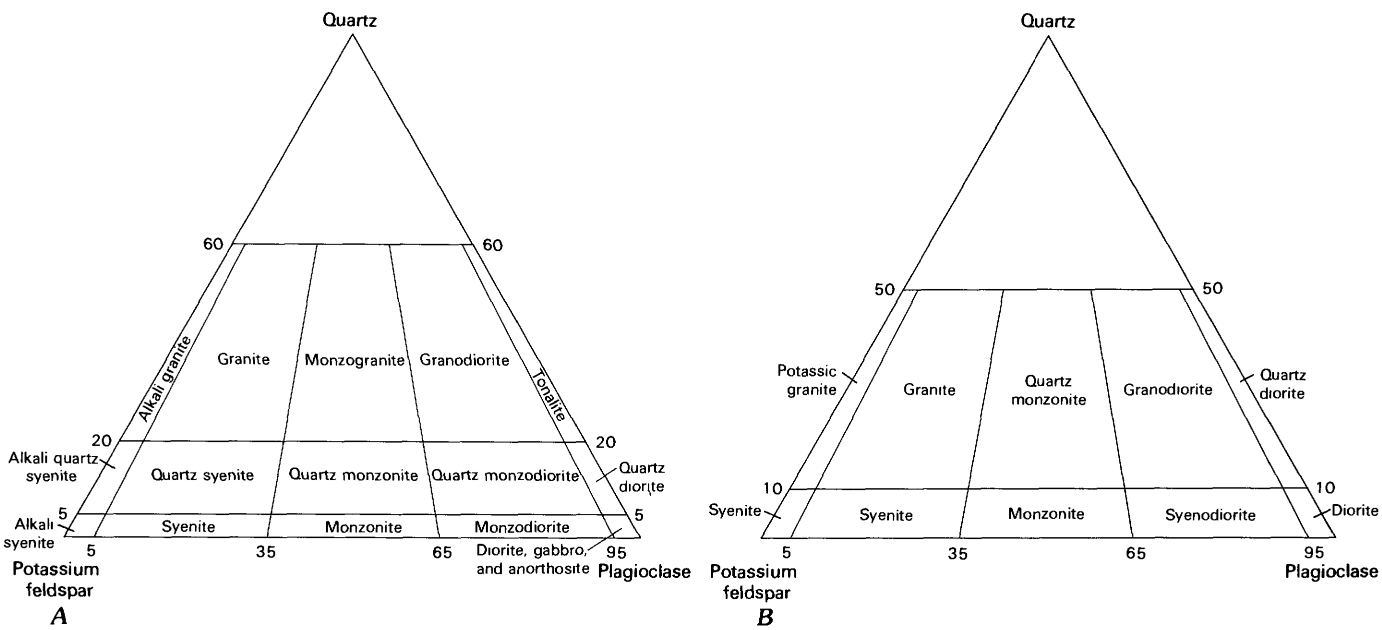


FIGURE 15.—Nomenclature of intrusive igneous rocks. *A*, Nomenclature used in this report; *B*, Nomenclature used on published maps of the central Lemhi Range, for comparison.

TABLE 4.—*Chemical analyses of granitic intrusive rocks in and near the central Lemhi Range, Idaho*

[K-Ar ages in biotite: monzogranite, field No. 9T327, 49.4±1.7 m.y.; quartz monzodiorite, field No. 9R84, 51.3±1.5 m.y. (John D. Obradovich, written commun., 1971). Field No. 15159, Rb-Sr ages on monzogranite: potassium-feldspar, 435±60 m.y.; whole rock, 430±40 m.y. (C.E. Hedge and F.G. Walthall, written commun., 1964, recomputed with λ_{87}Rb = 1.39×10^{-11}/yr). Rapid-method chemical analyses analyzed by Paul Elmore, Sam Botts, Lowell Artis, Gillison Chloe, H. Smith, J. Kelsey, J. Glenn, and C.L. Parker. Quantitative spectrographic analyses for minor elements analyzed by Norma Rait, A.L. Sutton, Jr., J.C. Hamilton, and W.B. Crandell; results are reported in percent to the nearest number in the series 1, 0.7, 0.5, 0.3, 0.2, 0.15, and 0.1 and so forth, which represent approximate midpoints of group data on a geometric scale. The assigned group for semiquantitative results will include the quantitative value about 30 percent of the time. Elements looked for but not detected: Ag, As, Au, Bi, Cd, Eu, Ge, Hf, Hg, In, Li, Pd, Pr, Pt, Re, Sb, Sm, Sn, Ta, Te, Th, Tl, U, W, and Zn. 0, looked for but not detected; leaders (--), not looked for]

	Big Eightmile stock			Big Timber stock		Gilmore stock		Park Fork intrusive		Early Paleozoic granite near Leadore
Field No.	4R24	4M42	9T327	15167	2R48	7R101	9R84	2M139	2M140	15169
Lab. No.	D116006W	D16011W	W173256	13834	161345	W169451	W173255	161346	161347	13835
Rock type	Quartz monzodiorite	Monzogranite	Monzogranite	Granodiorite	Diorite	Quartz monzodiorite	Quartz monzodiorite	Quartz monzodiorite	Monzogranite	Monzogranite
				Rapid-method chemical analyses (percent)						
SiO_2	58.8	56.0	65.1	62.05	60.8	61.4	57.3	53.8	66.4	76.87
Al_2O_3	15.3	16.9	15.3	16.02	16.2	14.9	13.9	14.4	15.8	12.86
Fe_2O_3	2.0	2.8	1.7	2.06	3.3	2.3	6.1	3.0	1.9	.45
FeO	4.8	4.7	2.6	3.28	3.2	4.1	.16	5.6	1.8	.16
MgO	4.3	4.6	2.1	2.73	3.2	3.8	4.0	6.4	1.4	.10
CaO	6.5	7.8	3.5	4.67	4.8	5.5	6.5	8.1	3.6	.20
Na_2O	3.3	3.2	3.5	3.89	3.8	3.4	2.7	3.2	4.0	2.63
K_2O	2.6	1.4	3.7	2.98	2.5	2.6	3.7	1.2	3.3	5.96
H_2O^-	.15	.16	0.26	.69	.18	.22	.95	.30	.06	.51
H_2O^+	.69	.58	0.94	.19	1.2	.51	1.1	1.2	.61	.14
TiO_2	.82	.60	0.53	.65	.77	.77	.70	.95	.42	.11
P_2O_5	.61	.61	.25	.22	.32	.28	.44	.90	.25	.01
MnO	.10	.51	.07	.10	.09	.12	.11	.13	.04	.01
CO_2	.06	.11	.05	.23	.06	<.05	2.4	.25	.09	.02
				(CHF.1)						(CHF.06)
Sum	100.03	99.97	99.6	99.86	100.42	100	100.06	99.43	99.67	100.09
				Quantitative spectrographic analyses for minor elements						
Titanium (Ti)	0.3	0.2	--	0	0.5	--	--	0.5	0.3	--
Manganese (Mn)	.05	.55	0.07	0	.07	--	0.07	.1	.02	--
Barium (Ba)	.15	.1	.2	.15	.2	0.2	.2	.1	.2	0.0002
Beryllium (Be)	.0005	.0003	.0001	0	.00015	.0001	0	0	.0002	--
Cerium (Ce)	--	--	--	--	.02	.02	0	0	.02	0
Cobalt (Co)	.002	.003	.0015	.002	.002	.002	.003	.003	.001	0
Chromium (Cr)	.015	.01	.001	.003	.007	.007	.01	.02	.005	.0002
Copper (Cu)	.005	.002	.007	.003	.0015	.003	.015	.015	.001	.0002
Gallium (Ga)	.002	.002	.002	.002	.003	.002	.0015	.003	.003	.002
Lanthanum (La)	.01	.007	.007	.007	.01	.015	.01	.007	.015	0
Molybdenum (Mo)	0	.0007	--	0	0	.001	0	0	0	0
Niobium (Nb)	.0015	.001	.003	0	.002	.002	.001	.001	.002	.005
Nickel (Ni)	.005	.005	.0015	.002	.005	.005	.007	.005	.002	0
Lead (Pb)	.0015	.0015	.003	.003	.05	.0015	.0015	.0015	.0015	0
Scandium (Sc)	.003	.003	0	.001	.0015	.002	0	.003	.003	0
Strontium (Sr)	.1	.1	.15	.1	.15	.15	.15	.2	.2	.003
Vanadium (V)	.02	.015	.01	.01	.02	.015	.015	.03	.03	0
Yttrium (Y)	.002	.002	.005	.002	.002	.003	.005	.002	.002	.002
Ytterbium (Yb)	--	--	.0005	.0002	--	.0003	.0005	--	--	.0003
Zirconium (Zr)	.01	.005	0	.007	.015	.02	0	.007	.007	.01
Neodymium (Nd)	.007	0	0	0	.015	.015	0	.01	.01	--

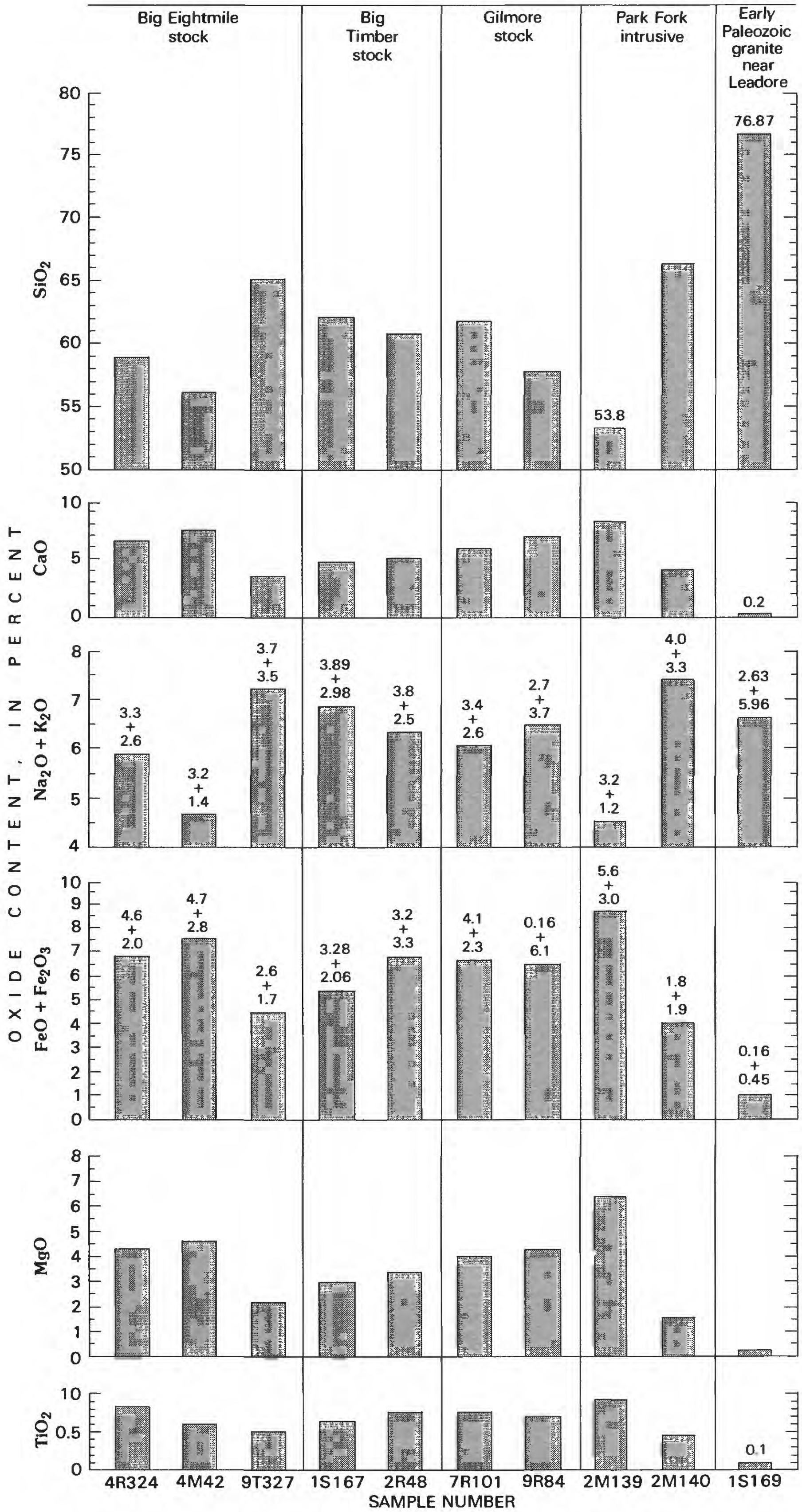


FIGURE 16.—Comparison of oxides in intrusive rocks, central Lemhi Range and Beaverhead Mountains, east-central Idaho. See table 4 for complete rapid-method chemical analyses and quantitative spectrographic analyses of rocks collected in this study.

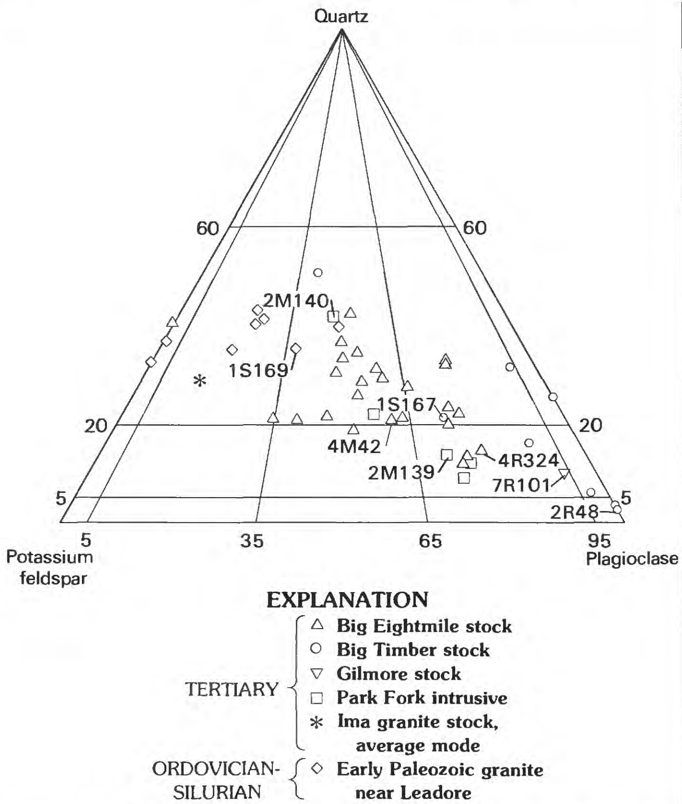


FIGURE 17.—Modal proportions of quartz and feldspar in granitic intrusive rocks in east-central Idaho. Mafic minerals were subtracted and the modes recalculated to 100 percent.

as 10 mm long, but as much as 2 cm long in a few places. Secondary minerals include sericite, saussurite, leucoxene, limonite, magnetite, hematite, chlorite, and quartz and calcite in small veinlets. A few samples contain microscopic euhedral grains of pyrite that probably are late-hydrothermal additions. Most of the granite has been moderately to strongly hydrothermally altered. Feldspars have been partly converted to sericite and saussurite; biotite commonly has been almost completely altered to chlorite, magnetite, and leucoxene.

Most of the granite is at least partly granulated, and on a larger scale brecciated, as a result of intense post-emplacement faulting. The granulated rocks are very fine grained (<0.01–0.05 mm), and typically are in subparallel layers, only a small fraction of a millimeter thick, composed of quartz and feldspar and their alteration products, separating less granulated but shattered layers of granite that contain fracture fillings of quartz and calcite.

Granitic rocks are widespread farther south in the Beaverhead Mountains in imbricate thrust slices, and are mineralogically similar to those north of Leadore. Staatz and others (1972, p. B52–B53) described the granite between Hawley Creek and Bull Canyon, about 10 km east of Leadore, as containing 20–45 percent quartz, 55–75 percent perthite, 0–5 percent plagioclase, as much as 2 percent biotite and magnetite, and less than 1 percent zircon, apatite, rutile, monazite, and allanite, and including chlorite, sericite, hematite, and limonite alteration products. The chemical composition of the granites is nearly identical to that of the granite north of Leadore (Staatz and others, 1972, p. B54). The Beaverhead granite pluton (Ramspott, 1962, p. 103–156; Scholten and Ramspott, 1968, p. 18–21), 20–25 km farther south in the Beaverhead Mountains, includes a small leucosyenite mass, but is mainly one- and two-feldspar granite that contains about 30 percent quartz, 60–65 percent microcline in one-feldspar granite and perhaps 30–45 percent microcline and 15–30 percent albite in two-feldspar granite, and as much as 5 percent accessory biotite, zircon, and apatite. Secondary alteration minerals include sericite, hematite, leucoxene, chlorite, magnetite, and quartz. No complete chemical analyses are available for rocks of the Beaverhead pluton, but at least the content of CaO, Na_2O, and K_2O is similar to that of the analyzed rocks farther north (Ramspott, 1962, p. 154–155).

EOCENE GRANITIC ROCKS IN THE LEMHI RANGE

Granitic rocks in small stocks and sheets in the central Lemhi Range are mainly monzogranite, granodiorite, and quartz monzodiorite, but also include granite—in the Ima stock alone, quartz monzonite, and tonalite, quartz diorite, and diorite. These rocks are medium light gray to medium dark gray, fine to medium grained, and porphyritic. In most intrusive bodies, the rocks are homogeneous except that border zones tend to be darker, more basic, and in some places contain abundant inclusions of the wall rocks. The major part of most stocks is monzogranite and texturally similar granodiorite. Border zones most commonly are quartz monzodiorite and granodiorite, except in the Big Timber and Gilmore stocks, where border zones and nearby related sheets and dikes are tonalite, diorite, and quartz diorite.

Granite.—The light-gray to grayish-orange-pink, coarse-grained biotite granite of the Ima stock, near Patterson, occurs only in this stock, and is not exposed or known to be present any place else in east-central Idaho. It is not exposed at the surface in Patterson Creek, but is intersected in mine workings and drill holes (Callaghan and Lemmon, 1941, p. 8; Hobbs, 1945, p. 5). The granite is a coarse-grained (10–25 mm) hypidiomorphic granular aggregate of large, subhedral crystals of quartz, alkali feldspar, and subordinate plagioclase and biotite. Quartz forms about 25–27 percent of the rock; alkali feldspar about 50–85 percent, principally orthoclase but including 3-5 percent microcline; plagioclase about 8-10 percent that is An_{13-18} (oligoclase); biotite about 7-15 percent, and a trace of accessory zircon. Most of the granite has been extensively hydrothermally altered with alteration products forming more than 30 percent of some samples; they include abundant sericite, saussurite, chlorite, zoisite, epidote, magnetite, secondary muscovite after biotite, and clay minerals. Pyrite, chalcopyrite, and molybdenite are widely disseminated throughout the granite.

Partial chemical analyses of the granite suggest that it contains about the following proportions of elements (Ora H. Rostad, written commun., 1982):

Major elements as oxides	percent
SiO_2	75.4-76.4
K_2O	5.2-5.4
Na_2O	2.1-2.8
CaO	0.46-0.55
MgO	0.17-0.24
Minor elements	parts per million
Mn	260-290
Ti	490-660
Ba	130-190

Monzogranite and quartz monzonite.—The dominant granitic rocks in the larger stocks and sheets are medium-light-gray to medium-dark-gray, porphyritic biotite monzogranite and a slightly less abundant variety of the monzogranite that contains a little less quartz, and locally is porphyritic biotite-quartz monzonite (fig. 17). In these rocks, subhedral phenocrysts, typically 4-6 mm long, but in places as much as 15 mm long, of plagioclase, alkali feldspar, quartz, biotite, and hornblende, are set in a fine-grained (0.1-0.5 mm, typically 0.2-0.3 mm) hypidiomorphic to xenomorphic granular groundmass. In places the groundmass is very fine grained (0.02-0.08 mm). These rocks typically contain 24-30 percent quartz; 25-31 percent alkali feldspar, less commonly as much as 45 percent; 37-42 percent plagioclase; 4-11 percent biotite; and 1-5 percent accessory hornblende, apatite, zircon, sphene, magnetite, rutile, and, in a few places, hypersthene and augite. The alkali feldspar is orthoclase, in interstitial anhedral to subhedral grains and in subhedral phenocrysts. Graphic intergrowths of alkali feldspar and quartz are present locally. Plagioclase occurs both as subhedral to euhedral phenocrysts, and as subhedral laths in the groundmass. Its composition most commonly is An_{25-33} (oligoclase and andesine), but some is An_{40} (andesine); some crystals are zoned, and some are enclosed by a thin envelope of more sodic plagioclase. Plagioclase phenocrysts commonly show patchy extinction; both phenocrysts and groundmass plagioclase have been widely saussuritized. Biotite is ubiquitous as ragged subhedral crystals in the groundmass, and as euhedral phenocrysts typically only a few millimeters across. Hornblende is present locally in subhedral phenocrysts as much as 1.5 cm long, commonly replacing pyroxene, and being replaced by biotite; both biotite and relict hornblende have been partly altered to chlorite. Other alteration products include sericite, leucoxene, calcite, epidote, hematite, and limonite.

Porphyritic monzogranite in the central part of the Big Eightmile stock has been hydrothermally altered to alkali granite (fig. 17), which contains about 35 percent quartz, 50 percent alkalic feldspar, 1 percent biotite, and 14 percent alteration products. Virtually all the plagioclase and most of the mafic minerals have been destroyed.

Granodiorite.—Although the modes of monzogranite and granodiorite suggest that they form an almost continuous series (fig. 17), the granodiorite is generally somewhat finer grained and somewhat darker colored, mostly medium dark gray as a result of more calcic plagioclase; it also contains phenocrysts that are mainly plagioclase, hornblende, and biotite. The rock is a porphyritic biotite-hornblende granodiorite that is distinct and separable from biotite monzogranite in the Big Eightmile stock, where together with porphyritic quartz monzodiorite it forms a significant part of the stock. In other stocks, the granodiorite most commonly forms border zones that were not mapped separately. The groundmass of the granodiorite is hypidiomorphic granular and fine grained (0.05-0.1 mm). Phenocrysts typically are 2-3 mm long but in a few places are as much as 6 mm long. The rock is composed of 13-23 percent quartz, 10-17 percent alkali feldspar, 37-49 percent plagioclase, 3-12 percent biotite, 4-12 percent hornblende, 1-2 percent hypersthene and augite, and 1-3 percent accessory magnetite, enstatite, apatite, zircon, and sphene. Quartz and alkali feldspar occur in the groundmass, enclosing and surrounding plagioclase and mafic minerals. Plagioclase occurs both as elongate grains in the groundmass, and as relatively sparse phenocrysts; its composition is An_{30-50} (andesine and labradorite). Some grains are zoned, with more sodic rims, and some apparently late plagioclase is myrmekitic. Hornblende and pyroxenes are principally in subhedral to euhedral phenocrysts, with hornblende replacing pyroxene, commonly almost completely, and being replaced by biotite. Both biotite and hornblende also occur in small, subhedral grains in the groundmass. The mafic minerals have been partly altered to chlorite and antigorite, and feldspars have been partly sericitized, but in general, the granodiorite has been less altered than adjacent monzogranite.

Quartz monzodiorite.—Medium-dark-gray to medium-gray, porphyritic biotite-hornblende-quartz monzodiorite is closely associated with the porphyritic biotite-hornblende granodiorite in most stocks and sheets, and is the principal rock in the border zone of the Park Fork intrusive. As discussed earlier it forms, with the granodiorite, a significant part of the Big Eightmile stock, and probably is a principal component of many thin sheets and of the border zones of most stocks. The groundmass of the monzodiorite is hypidiomorphic granular and very fine grained to fine grained (0.05–0.5 mm, typically about 0.3 mm). Phenocrysts of plagioclase, hornblende, biotite, and pyroxene typically are 1–3 mm long, but exceptionally are as much as 8 mm long. The rock is composed of 8–12 percent quartz; 7–18 percent alkali feldspar (as little as 4 percent alkali feldspar locally in the border zone of the Gilmore stock); 45–55 percent plagioclase (as much as 63 percent in a few places); 4–10 percent biotite; 3–11 percent hornblende; 8–12 percent augite and hypersthene (some rocks contain as little as 1 percent or as much as 16 percent pyroxene); and 2–6 percent accessory magnetite, zircon, apatite, and rutile. Quartz and alkali feldspar occur in the groundmass, enclosing and surrounding plagioclase and mafic minerals as interstitial filling, and in places are micrographically intergrown. Plagioclase occurs both as small, elongate grains in the groundmass and as long, lathlike phenocrysts; its composition is An_{34-55} (andesine and labradorite). Biotite occurs in small crystals in the groundmass and as small, euhedral crystals surrounding and replacing hornblende. Hornblende includes both dark-green, primary subhedral phenocrysts and more abundant pale-green uralitic hornblende replacing pyroxene. Augite and hypersthene phenocrysts are abundant in some places, but most commonly have been partly or entirely replaced by hornblende, which in turn has been partly replaced by biotite in small crystals around the phenocryst margin. Alteration products include chlorite, sericite, hematite, epidote, actinolite, and sphene, but the monzodiorite has not been as strongly hydrothermally altered as the monzogranite.

Tonalite, quartz diorite, and diorite.—These basic, porphyritic rocks are exposed and relatively abundant only in the border zones of the Big Timber and Gilmore stocks; they are rare or absent in the other stocks and sheets in the central Lemhi Range. They are fine grained to very fine grained (less than 0.01 to 0.1 mm) xenomorphic granular or felted, porphyritic rocks which contain subhedral to euhedral phenocrysts of plagioclase commonly 1–3 mm long, fairly abundant rounded inclusions of the enclosing quartzitic rocks and embayed and rounded, disaggregated quartz xenocrysts derived from the quartzites. The groundmasses of some of these rocks are nearly irresolvable in thin section, because of the very fine grain size, or because of alteration. This type of groundmass forms 50–75 percent of most of these rocks, and commonly is a felted mass of very small grains of plagioclase and quartz in about equal proportions, with a few very small grains of biotite and magnetite. These rocks contain 1–23 percent quartz, excluding the disaggregated grains from the enclosing quartzites; 0–3 percent alkalic feldspar; 25–52 percent plagioclase, as much as 70 percent in marginal dikes of the Gilmore stock; 5–16 percent hornblende; 0–13 percent biotite; and 2–5 percent accessory magnetite, augite, apatite, zircon, and sphene. The quartz is all in very fine euhedral grains in the groundmass, as is the alkali feldspar where it is present at all. Plagioclase is in very small laths in the felted groundmass, and in phenocrysts; its composition is An_{40-55} (andesine and labradorite), and a few phenocrysts are zoned, with more sodic (An_{32}) rims. Hornblende is in subhedral to euhedral phenocrysts, and in sparse euhedral grains in the groundmass. Most of it appears to be uralitic, replacing pyroxene, and has been almost completely altered. Biotite is in very small, altered grains in the groundmass. All these basic rocks have been strongly altered, with plagioclase, hornblende, and alkali feldspar, where present, partly or largely converted to sericite, chlorite, hematite, leucoxene, epidote, and clay minerals.

DESCRIPTIONS OF INTRUSIVE MASSES

EARLY PALEOZOIC GRANITE IN THE BEAVERHEAD MOUNTAINS

The early Paleozoic granite along the flank of the Beaverhead Mountains north of Leadore and in Railroad Canyon (Ruppel, 1968) is so completely broken by faults that its intrusive relation to the surrounding sedimentary rocks is not clear. The granite is part of an imbricate thrust slice that includes Ordovician sedimentary rocks, and which has been broken into fault-bounded slivers by younger steep faults along the range front. The Ordovician Kinnikinic Quartzite appears to have been intruded by granite in a few places, for example, west of the mouth of Italian Gulch, but the intrusive relation is not clear and unequivocal. The quartzite is almost completely brecciated and silicified, cut by multiple thin quartz veins, and heavily stained with limonite. These widespread changes suggest that the granite did intrude and alter the quartzite. The dolomite of the Ordovician part of the Saturday Mountain Formation is in a separate imbricate thrust slice, and was not affected by emplacement of the granite.

Dolomite of the Devonian Jefferson Formation also is in thrust contact with the granite, as are younger Paleozoic sedimentary rocks, but these sedimentary rocks all are younger than the granite.

The intrusive relations of the granite with the country rocks are clearer in the thrust slice of granite and enclosing sedimentary rocks between Hawley Creek and Bull Canyon, about 10 km east of Leadore (Staatz and others, 1972, p. B51–B52; Lucchitta, 1966, p. 86–88). Here the granite was intruded into quartzite and siltite of the Lemhi Group and the lower part of the Kinnikinic Quartzite. Contact relations between granite dikes and the sedimentary rocks, described by Staatz and others (1972, p. B51), are clearly intrusive, and the contact is gradational as a result of incorporation and assimilation of quartzitic rocks in the granite. The contact relations of the Beaverhead granite pluton, 20–25 km south of Bull Canyon in the Beaverhead Mountains, have not been described, except that the granite intruded, or can reasonably be inferred to have intruded, the Kinnikinic Quartzite (Scholten and Ramspott, 1968, p. 19), and it contains inclusions of quartzitic rocks of the Lemhi Group (Evans, 1981, p. 95).

In summary, the early Paleozoic granite of the Beaverhead Mountains intruded and metamorphosed Middle Proterozoic rocks of the Lemhi Group, and in a few places either can be seen or can be reasonably inferred to have intruded the Ordovician Kinnikinic Quartzite but not any younger sedimentary rocks. These granites everywhere are in imbricate thrust slices in the Beaverhead Mountains, and were thrust across younger Paleozoic sedimentary rocks. Their original site of emplacement—the location of their unthrusted roots—is not known, but must lie buried beneath the thrust plates some unknown distance farther west.

IMA STOCK AND PATTERSON CREEK AND FALLS CREEK INTRUSIVES

The Ima biotite granite stock is not exposed at the surface, and its relations to the enclosing Middle Proterozoic rocks are known only from places where the contact is cut in underground workings of the Ima mine, and in holes drilled for mineral exploration. These relations are discussed by Callaghan and Lemmon (1941, p. 8), Hobbs (1945, p. 4), and Ora H. Rostad (written commun., 1971), the sources of much of the following summary. They suggest that the granite exposed in the Ima mine is an irregular, elongate, east-trending, steep-sided cupola about 610 m long, above a larger granitic mass at greater depth. The somewhat irregular surface of the cupola dips southward at an angle of 15°–30° beneath the valley of Patterson Creek, and near there, or only a short distance farther south, probably intersects the steep south edge of the stock. The north edge of the cupola has not been crossed in mine workings, but apparently also is steep, because no granite has been cut in deep drill holes reaching southeastward from the General Electric tunnel, which is northwest of the Ima mine.

The Ima stock intruded feldspathic quartzite and siltite of the Big Creek, Apple Creek, and Gunsight Formations, which had earlier been cut by a complex network of interlaced imbricate thrust faults and strongly sheared (Ruppel, 1980). The Yellowjacket Formation is inferred to be present, and intruded by the stock, at relatively shallow depth beneath the thrust zone, but it is not exposed any place nearby, and has not been cut in mine workings or drill holes. Its presence is inferred from regional geologic relations, which suggest that the gently west dipping sole zone of the Medicine Lodge thrust system should be relatively shallow here, and from aeromagnetic evidence (U.S. Geological Survey, 1971) that suggests a gently west dipping surface, interpreted to be the sole zone, beneath the Lemhi Range. The marginal parts of the stock are at least locally aplitic, and are cut by quartz-orthoclase mica pegmatite veins that contain molybdenite, pyrite, and chalcopyrite, and that also extend outward from the granites into the enclosing sedimentary rocks. Otherwise the stock appears to be fairly uniform, coarse-grained biotite granite that contains disseminated pyrite, chalcopyrite, and molybdenite. The emplacement of the stock does not seem to have been accompanied by any significant metamorphism of the enclosing rocks, other than a thin, marginal baked zone and some sericitization of feldspar in the quartzites; at least no significant metamorphism is mentioned in the reports of Callaghan and Lemmon (1941), Hobbs (1945), or Rostad (written commun., 1971), and none is evident in surface exposures above the stock. The enclosing rocks and the outer shell of the stock are cut by multiple north- to northwest-trending quartz veins that contain fluorite, pyrite, huebnerite, scheelite, rhodochrosite, tetrahedrite, sphalerite, galena, siderite, and chalcopyrite. Most of these veins terminate upward against a gently west dipping breccia zone, interpreted by Hobbs (1945, p. 6–8) to be a premineralization thrust fault.

The Ima stock has been interpreted to be intrusive into an outer ring of a ring-fracture complex (Ruppel, 1982, p. 9, 14) that curves north from the Ima area into the upper part of Patterson Creek, and from there curves northwest and finally southwest to control the valley of Morse Creek. The ring-fracture complex also includes two small monzogranite and granodiorite stocks in the North Fork of Patterson Creek and in Falls Creek. Andesitic and basaltic dikes related to the Challis

Volcanics are present in the outer ring of the complex, which includes the granitic intrusives, and in an inner ring in the center of the complex. The presence of these intrusive rocks in the curving faults of the ring-fracture complex suggests that other granitic stocks probably have been emplaced elsewhere in different rings of the complex, but, like the Ima stock, are not exposed.

BIG EIGHTMILE (BLUE JAY) STOCK

The complex Big Eightmile stock underlies an area of about 15 km^2 in the vicinity of the Blue Jay mine at the mouth of Big Eightmile Creek (Ruppel, 1980). Only the south half of the stock is well exposed, on the glaciated south wall of the canyon; the north half is mostly concealed beneath till. The stock intruded thrust-faulted, Middle Proterozoic feldspathic quartzite and siltite of the Big Creek, Apple Creek, and Gunsight Formations, and quartzite of the Swauger Formation. It probably intruded the Yellowjacket Formation, beneath these thrust-faulted rocks, at relatively shallow depth. These enclosing rocks have been slightly bleached but otherwise not appreciably metamorphosed in most places adjacent to the granitic rocks; in a few places, however, feldspathic quartzite has been converted to spotted hornfels in a narrow zone adjacent to the contact. The granitic rocks typically were chilled in a zone only a few centimeters thick at the contact. Inclusions of wall rocks are abundant in the stock along its south margin but are sparse elsewhere.

The Big Eightmile stock is a composite stock of porphyritic biotite monzogranite, porphyritic biotite-hornblende granodiorite, and porphyritic biotite-hornblende-quartz monzodiorite that is closely associated with the granodiorite; the granodiorite and monzodiorite were mapped together. The monzogranite core of the stock has been hydrothermally altered, the plagioclase saussuritized, and the mafic minerals destroyed, so that in places the composition of the altered monzogranite is near that of alkali granite (fig. 17). The granitic rocks are cut by a thick, tabular body of autobrecciated basaltic andesite that probably is a dike related to the Challis Volcanics.

The intrusive relations between monzogranite and granodiorite-quartz monzodiorite are not completely clear, but it appears likely that the granodiorite and quartz monzodiorite were emplaced first. These rocks are consistently finer grained than the monzogranite, which suggests that they were intruded into still-cool country rocks. They are appreciably more basic, perhaps representing an earlier, more basic magma. They also contain much more abundant inclusions of bleached and hornfelsed quartzitic rocks, but no inclusions of the monzogranite, which also suggests that they were emplaced first. The angular, northeast-northwest pattern of the granodiorite-quartz monzodiorite body (Ruppel, 1980) suggests that it might have been intruded as a thick dike along older northeast- and northwest-trending basement faults (Ruppel, 1982, p. 21–22), but, if so, such faults do not break the exposed adjacent sedimentary rocks of the Medicine Lodge thrust plate. Porphyritic biotite monzogranite, which forms the bulk of the stock, is texturally and mineralogically homogeneous throughout the stock. Its contact with the granodiorite and quartz monzodiorite is sharp; the monzogranite shows no textural or mineralogic changes at the contact. Although a zone along the south margin of the stock, above Devils Canyon, is choked with quartzitic inclusions, most of the monzogranite contains only rare inclusions of bleached, light-gray hornfels derived from the enclosing feldspathic quartzites, and no inclusions of granodiorite or quartz monzodiorite.

The central part of the monzogranite is partly concealed by glacial deposits and colluvium, but exposures in the vicinity of the Blue Jay mine and scattered outcrops through the till indicate that it has been extensively altered and mineralized by late hydrothermal solutions in a more or less circular area of 3–4 km^2. The monzogranite in this area has been sericitized and chloritized, and in places almost completely altered to alkali granite (fig. 17) that is about 34 percent quartz, 50 percent alkali feldspar, 2 percent somewhat chloritized biotite, and 15 percent alteration products including sericite, antigorite, chlorite, magnetite, hematite, limonite, and clay minerals. Original plagioclase has been completely sericitized, and hornblende has been converted to antigorite and sericite. The altered monzogranite contains pyrite, chalcopyrite, and molybdenite, disseminated through the rock and with quartz in abundant veinlets. The altered rocks have been much more closely fractured than the surrounding, relatively unaltered granitic rocks, and in places have been brecciated or cut by layers of clayey gouge. Some alteration and closely spaced fracturing of the monzogranite, accompanied by sulfide mineralization, persists to a depth of at least 150–175 m in holes drilled beneath the valley floor of Big Eightmile Creek. As a result of intense fracturing and pervasive alteration, this core zone of the Big Eightmile stock is deeply weathered. The weathered zone seems to be more than 60 m deep in exploratory holes drilled at the Blue Jay mine, but is absent in holes drilled near the valley bottom, where it has been removed by glacial erosion. At the Blue Jay mine, this zone is capped by a strongly fractured, siliceous, limonitic gossan that contains abundant malachite and azurite.

The south edge of the hydrothermally altered and mineralized core of the Big Eightmile stock is cut by an east-trending, steeply south dipping composite dike of basaltic andesite, which probably fed the widespread basaltic flows and pyroclastic rocks in the Challis Volcanics north of Big Eightmile Creek. The granitic rocks enclosing the dike have been sericitized and limonite stained, but otherwise do not appear to be much changed. The margins of the dike, which are aphanitic or very fine grained, were probably chilled against the granitic rock; in other words, the dike was intruded into a granitic mass that had already cooled, even though radiometric age determinations suggest that the granitic rocks and the Challis Volcanics might be about the same age (p. 72). The dike is separable into three parts from north to south: a lower zone, 15–25 m thick, of medium-dark-gray, aphanitic, generally nonporphyritic basaltic andesite, in contact with deeply disintegrated, altered monzogranite of the stock core; an intermediate zone, 10–15 m thick, of bleached, autobrecciated basaltic andesite, with a prominent breccia zone at its top; and an upper zone, 90–125 m thick, of medium-dark-gray, very fine grained to aphanitic (at its contact with granitic rocks), porphyritic basaltic andesite that contains abundant phenocrysts, 2–5 mm long, of euhedral to subhedral plagioclase. The groundmass of the basaltic andesite in the thick upper zone is a microcrystalline, partly felted or microlitic aggregate of altered plagioclase, hornblende, chalcedony, and nontronite; about one-fourth of it is glass and devitrified glass. The groundmass constitutes about 50–60 percent of the rock, and encloses phenocrysts of plagioclase (An_{25-50}, oligoclase and andesine), about 25 percent; of hornblende, about 5 percent, which is partly pseudomorphic after clinopyroxene; and of zircon in trace amounts. About 1 percent of the rock is made up of silica in chalcedonic or microcrystalline aggregates pseudomorphic after olivine and rimmed with minute magnetite grains. The rock, which has been strongly altered, contains about 10 percent quartz, chalcedony, chlorite, sericite, nontronite, magnetite, hematite, leucoxene, and carbonate. All the quartz and chalcedony is secondary. In thin section, the rock clearly is autobrecciated, and consists of angular fragments that have been repeatedly broken and rehealed. The rock originally was probably an olivine basalt or basaltic andesite porphyry. It later was repeatedly brecciated, probably reheated, as is suggested by alteration of the olivine, and hydrothermally altered. The zoning in the dike, the extensive autobrecciation, and the indications in thin sections of repeated brecciation, rehealing, and pervasive alteration, all suggest that the dike was the site of a vent that repeatedly fed the compositionally similar basaltic andesite flows in the Challis Volcanics north of Big Eightmile Creek.

PARK FORK INTRUSIVE

The Park Fork intrusive underlies an area about 2 km long and 1–1.5 km wide in the head of the Park Fork of Big Creek, about 6 km south of the Big Eightmile stock. Only the upper part of the intrusive is exposed; much of it is thinly mantled with glacial debris and with creeping colluvial deposits, as much as 2 m thick, which are derived at least partly from the underlying, deeply disaggregated and mechanically weathered granitic rocks. In the creeping surficial cover, granitic rocks are abundant as rounded boulders which are joint blocks rounded by mechanical weathering. Granitic rocks in place are well exposed on the ice-stripped valley flanks and in a small cirque on the north flank of the Park Fork. These exposures show a thickness of only about 150 m of granitic rock beneath a nearly flat or gently west dipping roof of siltite of the Apple Creek Formation. The enclosing siltite has been converted to calc-silicate hornfels in a few places adjacent to the granitic rock. More commonly, the feldspar in the siltite has been sericizited, and the rock hardened and darkened to medium to dark gray, losing its characteristic grayish-green color, in a metamorphosed zone about 1 m thick at the contact.

The smoothly curving roof is broken only at its northwest corner, where granitic rocks extend upward into an overlying thrust fault, in the form of a small cupola that is increasingly choked with inclusions towards the top, and in which the granitic rock grades almost imperceptibly upward into brecciated siltite in the thrust fault. The gently dipping roof is about parallel to imbricate thrust faults in the overlying sedimentary rocks, and probably is controlled by a preintrusive thrust. Although the intrusive mass was originally thought to be a stock (Ruppel, 1980), its roof relations, and the relations of stocks and sheets to their enclosing rocks elsewhere in the Lemhi Range, suggest that it more probably is a sheet intruded into a thrust fault. It may have been fed from the Big Eightmile stock, which contains similar rocks.

The intrusive is composed mainly of porphyritic biotite monzogranite, fringed on part of its north side by porphyritic biotite-hornblende-quartz monzodiorite. The monzogranite is texturally and compositionally uniform throughout most of the intrusive, and is in sharp contact with quartz monzodiorite along the northwest margin of the intrusive. The quartz monzodiorite zone is about 300–400 m wide, and its groundmass ranges from very fine grained to nearly aphanitic, consistently much finer grained than the monzogranite groundmass. Plagioclase phenocrysts, which form 25–30 percent of the rock, are strongly aligned, generally with the long axis subparallel to the strike of the contact.

The quartz monzodiorite appears to be an early phase of the intrusive, preserved mainly in the sharp northwest corner and cupola of the intrusive body; the more widespread monzogranite is a later, main phase.

The quartz-monzodiorite chilled zone is cut by a single thin dike of very fine grained to aphanitic andesite porphyry similar to the basaltic andesite dike in the Big Eightmile stock, but not autobrecciated or altered. Groundmass constitutes 70 percent of the andesite, and is a felted to microlitic mass of minute calcic plagioclase crystals and partly devitrified glass. Phenocrysts of augite as much as 2 mm long form about another 20 percent of the rock; the remaining 10 percent is biotite, aragonite, and minor quartz, magnetite, hematite, and clay minerals.

BIG TIMBER STOCK

The Big Timber stock is a northwest-trending, elongate intrusive, about 3 km long and 1-2 km wide, at the junction of Lake Creek and Big Timber Creek (Ruppel, 1968; Ruppel and Lopez, 1981). The stock has intruded a northwest-trending fault, probably near its intersection with a concealed east-trending fault, and was emplaced in and just above the sole zone of the Medicine Lodge thrust plate, which is exposed just south of the mouth of Lake Creek. A sill extends outward from the stock into the Saturday Mountain Formation; it is exposed on the flanks of Big Timber Creek west of Lake Creek. A probable sheet, in an imbricate thrust fault in Proterozoic rocks, extends south to the Smithie Fork of Sawmill Creek, where granitic rocks are present in a single small outcrop surrounded by till.

The rocks enclosing the stock are mainly feldspathic quartzite and siltite of the Gunsight and Apple Creek Formations, and feldspathic quartzite of the Yellowjacket Formation beneath the sole zone of the Medicine Lodge thrust plate. The enclosing rocks were thermally metamorphosed for about a meter adjacent to the stock, the quartzitic rocks were bleached, and the siltite was both darkened to medium or medium dark gray instead of the grayish-green color typical of the Apple Creek Formation, and slightly coarsened, to very fine grained quartzite. Inclusions form 10-20 percent of the granitic rock in the upper part of the stock exposed north of Lake Creek, and include Proterozoic quartzite and siltite, white quartzite of the Kinnikinic Quartzite, and sparse blocks, as much as 4 m long, of bleached white dolomitic marble, probably metamorphosed Saturday Mountain dolomite. The inclusions were recrystallized to coarser grain sizes than in surrounding unmetamorphosed rocks, and bleached. Most of them are angular fragments from 2 cm to 1 m in diameter, but some, of brecciated Kinnikinic Quartzite, are as much as 100 m in diameter and probably are remnants of roof pendants. Two inclusions of strongly foliated quartzose schist, each several meters in diameter, also are present in the Big Timber stock in this contact zone. The schist has been intricately contorted and sheared, and cut by thin veins of massive white quartz and associated specular hematite, which also is disseminated in the schist. No comparable metamorphism is present in any rocks adjacent to the stock or elsewhere in the Lemhi Range, and the two inclusions of schist probably are basement crystalline metamorphic rocks carried upward in the stock during its emplacement. Dolomite of the Saturday Mountain Formation that encloses the sill west of the main stock has been strongly metamorphosed, partly to coarsely crystalline white marble, but more commonly to light-brownish-gray to dark-gray, epidote-rich, calc-silicate hornfels.

The central part of the Big Timber stock is homogeneous porphyritic biotite monzogranite and porphyritic biotite-hornblende granodiorite like that in other stocks in the central Lemhi Range. Finer grained and much more basic border and roof zone rocks form much of the exposed part of the stock and its tabular offshoots, however, and include porphyritic hornblende-quartz monzodiorite, tonalite, quartz diorite, and diorite (fig. 17). The border zone rocks are very fine grained to nearly aphanitic at the contact with enclosing sedimentary rocks. They commonly are choked with inclusions, and include abundant deeply embayed quartz grains from disaggregated quartzite, and abundant mafic-rich segregations of quartz, hornblende, and biotite surrounded by mafic-poor zones. Phenocrysts of hornblende in the border zone rocks are in euhedral to subhedral crystals that mostly are only slightly larger than grains in the groundmass, but which include a few much larger crystals, as long as 6-8 mm. Away from the contact, the granitic rocks become progressively coarser grained, and contain fewer inclusions; they appear to grade into the monzogranite-granodiorite core of the stock.

GILMORE STOCK

The Gilmore stock with its peripheral sheets and dikes is the largest granitic body in the Lemhi Range (Ruppel and Lopez, 1981). The stock, which forms the core of this intrusive complex, is concealed by till, alluvial fan gravels, and pediment veneer gravels at Gilmore, but its presence here is indicated by widespread thermal metamorphism of the sedimentary rocks, and by geochemical, gravity, and magnetic evidence (Ruppel and others, 1970, p. 14, 29, 34-35). The concealed stock is more or less northwest-trending, about 5 km long and 1-3 km wide, and centered at Gilmore. It extends northwest beneath the mouth of Meadow Lake Creek, south to the granitic rocks exposed at the mouth of Silver Moon Gulch, and east of Gilmore beneath pediment veneer gravels, where it has been cut in a single drill hole. North of Meadow Lake Creek, granitic rocks are exposed in a thick sill that extends westward into dikes that are exposed on the flanks of Deer Creek, and northward into a group of dikes that are in the roof of the sill east of the Hilltop mine. South and west of Gilmore, granitic rocks are exposed in multiple dikes and sills around the head of Liberty Gulch, and in a sill at the mouth of Silver Moon Gulch. South of Long Canyon, they form a thick, complex sheet, intruding a zone of thrust faults, that reaches south to Spring Mountain Canyon and Horseshoe Gulch, a distance of about 8 km. South of Spring Mountain Canyon, this thick sheet dies out in a series of dikes that extend another 6 km farther south, to the southernmost exposures of granitic rocks in the Lemhi Range.

The sedimentary rocks that enclose the Gilmore stock and most of its peripheral sheets and dikes are the Kinnikinic Quartzite, the Saturday Mountain Formation, the Jefferson Formation, and undivided Mississippian limestones in a thrust breccia near Gilmore. The white quartzite of the Kinnikinic is in contact with granitic rocks on and near the north end of Portland Mountain, and north of Spring Mountain Canyon, and is nearly unchanged except for a slight coarsening of its grain size. The carbonate rocks that overlie the Kinnikinic, and which enclose most of the intrusive mass, were much more extensively metamorphosed, to finely to coarsely crystalline, white dolomitic marble, garnet-epidote-calcite skarn, and calc-silicate hornfels, in zones from a few meters to several tens of meters wide. Marble is present almost everywhere along the edges of the intrusive complex; and in places near the contact, the marble includes lenses and beds of coarse-grained magnetite-bearing garnet-epidote-calcite skarn. In the contact zone in the upper part of Bruce Canyon, marble and skarn in metamorphosed Saturday Mountain dolomite include a lens, about 60 m long and as much as 15 m wide, of coarse (1–3 mm) euhedral to subhedral crystals of magnetite, quartz, and sparse dolomitic marble. Magnetite forms almost 90 percent of the lens, and as much as 50 percent of the adjoining skarn. Malachite and azurite, derived from primary bornite, are abundant in the deposit, which, by assay, contains 1–2 percent copper (Bell, 1914, p. 152).

Calc-silicate hornfels is present mainly adjacent to some of the dikes that intrude the Jefferson Formation near the Hilltop mine, northwest of Gilmore, where the dikes are offshoots in the roof of a sill that is more fully exposed a short distance farther south. But even here, marble and skarn are the most widespread contact metamorphic rocks.

Middle Proterozoic sedimentary rocks in contact with the Gilmore granite intrusive are exposed in only two places: in Deer Creek, where granitic rocks intrude feldspathic quartzite of the Gunsight Formation at its contact with overlying Kinnikinic Quartzite; and in Squaw Creek, west of Spring Mountain Canyon, where the granitic rocks intrude feldspathic quartzite of the Yellowjacket Formation, beneath the sole zone of the Medicine Lodge thrust. The quartzitic rocks are hardened at the contact, but otherwise are little changed.

The stock that is the core of the Gilmore granite intrusive is not exposed, but probably is composed almost entirely of porphyritic biotite monzogranite and porphyritic biotite-hornblende granodiorite, an inference drawn from the composition of other stocks in the central Lemhi Range, and from the composition of the stock where it has been penetrated in a drill hole. Geochemical evidence suggests that it contains a small amount of molybdenite, pyrite, and probably chalcopyrite (Ruppel and others, 1970, p. 29). Monzogranite and granodiorite also are the principal rocks in thick sills and sheets both north and south of Gilmore, but the marginal parts of the sheets and sills, and all the thin sills and dikes, are much more basic porphyritic quartz monzodiorite, quartz diorite, and diorite. These fine-grained basic rocks commonly contain a few percent alkali feldspar and less than 10 percent quartz in the groundmass; as much as 60–70 percent plagioclase, andesine and labradorite, in groundmass and phenocrysts; and as much as 20 percent biotite, hornblende, and augite in groundmass and phenocrysts. Plagioclase phenocrysts are dominant in a few dikes that contain exceptionally high percentages of this mineral, where they commonly occur as rounded crystals that resemble rolled oats, forming "oatmeal" porphyries. In most of these rocks, however, phenocrysts are both plagioclase and hornblende. The hornblende replaces augite, and in turn is replaced by biotite. A few dikes are nonporphyritic, or contain only sparse phenocrysts, commonly of plagioclase.

Only slight hydrothermal alteration is apparent in the exposed rocks of the Gilmore intrusive, primarily minor sericitization of feldspar along grain boundaries and fractures.

SAWMILL CANYON SHEET

The Sawmill Canyon sheet is an elongate body of deeply weathered granitic rocks, about 4 km long and 2 km wide, that underlies the low, rolling hills in the lower part of Sawmill Canyon (Ruppel and Lopez, 1981). The sheet seems to have been intruded into a thrust fault that had earlier placed Jefferson dolomite over the Kinnikinic Quartzite and the Swauger Formation, but structural interpretation is difficult as only the roof of the sheet is exposed. The exposed upper part of the sheet is very irregular. North of Squaw Creek, it comprises a group of sills, conformable with steeply dipping beds in the enclosing Jefferson Formation, that coalesce downward into the main body of the sheet. South of Squaw Creek, abundant, isolated outcrops of dolomite of the Jefferson Formation, surrounded by granitic rocks, appear to be pendants—remnants of an original roof like that farther north, but eroded to a slightly deeper level.

The present southwest edge of the sheet is a fault, and the only granitic rocks exposed west of the fault are in a dike that cuts thrust-faulted and intensely brecciated rocks of the Kinnikinic Quartzite and Swauger Formation that are interpreted to be the floor of the Sawmill Canyon sheet. The brecciated sedimentary rocks are imbricate thrust slices below an imbricate slice of Jefferson dolomite exposed on Dragons Head, at the mouth of Sawmill Canyon. The Sawmill Canyon sheet thus is interpreted to be intrusive into the thrust zone between the quartzites and the Jefferson Formation. The intense brecciation and deformation in the Kinnikinic and Swauger quartzites suggest that they are just above the sole zone of the Medicine Lodge thrust system, like similarly brecciated rocks exposed above the sole zone in Squaw Creek, 3–4 km east of the Sawmill Canyon sheet.

The northeast margin of the Sawmill Canyon sheet is partly concealed under glacial deposits. The glacial deposits do not contain any granitic rocks, however, and so probably conceal only the dolomitic rocks that farther south form the roof of the sheet.

The north side of the sheet is bordered by Challis Volcanics, which may conceal other granitic rocks. The weathered surface of the Sawmill Canyon sheet is littered with compositionally homogeneous, small, angular fragments of Challis Volcanics. Between Squaw Creek and Mill Creek, strongly bleached and weathered granitic rocks are overlain by very small, thin outcrops of volcanic rocks that are erosional remnants of a once more continuous blanket. These relations of Challis Volcanics and granitic rocks show that the rolling topography now present on the granitic rocks in Sawmill Canyon and at Dragons Head is an old topography now being exhumed from beneath the volcanic rocks, and that the granitic rocks had been exposed by erosion, and deeply weathered, before eruption of the Challis Volcanics. Granitic rocks and marble extend northward, unchanged in texture and composition, to the edge of the volcanic field at Mill Creek, and almost certainly extend farther northwest beneath the volcanics.

A small area of granitic rock, similar to that in the Sawmill Canyon sheet, is exposed through till in the valley of Bear Creek, about 6 km farther north. Its borders are concealed, but it seems likely to be a sheet emplaced in an imbricate thrust fault that cuts Proterozoic rocks just to the west.

Neither the Sawmill Canyon sheet nor the Bear Creek sheet can be related to any nearby, stocklike core, although a parent stock could be concealed beneath the surrounding widespread blanket of glacial deposits, beneath the Challis volcanics, or beneath sedimentary rocks. The dike at Dragons Head does not seem likely as a feeder for the Sawmill Canyon Sheet, because it is chilled, fine-grained granitic rock and is not surrounded by any metamorphosed or altered rocks that might suggest the passage of large volumes of magma. Perhaps the most likely possibility is that the Sawmill Canyon and Bear Creek sheets are tabular offshoots of the Gilmore stock, like the compositionally similar sheet that intruded a thrust-faulted zone south from Gilmore to Spring Mountain Canyon.

The dolomite of the Jefferson Formation, intruded by the Sawmill Canyon sheet, has been thermally metamorphosed to light-gray and white, coarsely crystalline dolomitic marble in most places. Fine-grained calc-silicate hornfels occurs in a few places at the contact. A few thin beds of dark-gray dolomite retain their color and are characteristically fetid, even though they were recrystallized to a slightly coarser grain size than is typical of unmetamorphosed Jefferson.

Most of the Sawmill Canyon sheet is porphyritic biotite monzogranite and porphyritic biotite-hornblende granodiorite. These rocks appear to grade into thin, marginal zones that are much finer grained and more basic porphyritic biotite-hornblende-quartz monzodiorite and diorite, rocks that also form the Bear Creek sheet.

ALDER CREEK INTRUSIVE

The Alder Creek intrusive is a small, apparently thin and sheetlike body of hornblende-biotite-quartz monzodiorite intruded into the sole zone of the Medicine Lodge thrust plate north of Mill Mountain (Ruppel, 1980). It is the northernmost known exposure of granitic rocks in the Lemhi Range. The top of the intrusive dips gently west, beneath sheared and brecciated, thrust-faulted feldspathic quartzite of the Big Creek Formation, which forms its roof. Sheared feldspathic quartzite and siltite of the Yellowjacket Formation apparently are beneath the intrusive, but the contact is concealed by surficial deposits. A thin dike of monzodiorite similar to that in the sheet is exposed about 0.5 km farther southeast in the gulch of Alder Creek, intruding the Yellowjacket Formation. The feldspar in the sedimentary rocks that enclose the sheet and dike has been sericitized next to the contact, and the rocks have been hornfelsed in a thin zone adjacent to the contact in a few places.

The hornblende-biotite-quartz monzodiorite in the Alder Creek intrusive is uniformly fine grained and porphyritic, including as much as 50 percent phenocrysts of hornblende in euhedral to subhedral crystals about 1 mm in diameter, and in some places including conspicuous subhedral crystals 1-5 mm long of myrmekitic plagioclase feldspar. Near the intrusive contact with the enclosing rocks, hornblende phenocrysts and platy inclusions of Yellowjacket quartzite commonly are aligned parallel to the contact.

The east edge of the Alder Creek intrusive is cut by a younger, northeast-trending dike of rhyolite porphyry related to the Challis Volcanics. The rhyolite porphyry is very light gray to medium dark gray. Its groundmass is aphanitic, and encloses 20-30 percent euhedral to subhedral phenocrysts of clear and smoky quartz, sanidine, biotite, and hornblende. Phenocrysts of quartz and sanidine are 1-5 mm long, and those of biotite and hornblende are 1-2 mm long and typically, strongly chloritized.

AGE OF GRANITIC ROCKS

EARLY PALEOZOIC GRANITE IN THE BEAVERHEAD MOUNTAINS

Radiometric age determinations on the widespread granite in the Beaverhead Mountains suggest that it is of Ordovician and Silurian age. The range in isotope ages is fairly large, however, and some uncertainty remains as to whether the range in ages is real, or might instead reflect different dating techniques, some more modern than others; changes due to contamination of the magma; or changes induced by thrust-faulting. The first isotope ages reported for these rocks were remarkably similar: the granite in the Beaverhead pluton of Scholten and Ramspott (1968, p. 21) yielded a single K-Ar age of 441±15 million years; similar granite in Railroad Canyon yielded Rb-Sr determinations of 435±60 m.y. on potassium feldspar and 430±40 m.y. on whole rock (recomputed with $\lambda=1.39 \times 10^{-11}yr^{-1}$) (Ruppel, 1968; 1978, p. 18). The Beaverhead pluton subsequently was restudied by Scholten and Evans, and yielded a ^{207}Pb-^{206}Pb age on zircons of about 483 m.y. (Evans and Zartman, 1981; Evans, 1981, p. 105), an age close to K-Ar ages of 467±16 m.y. and 482±16 m.y. on hornblende and biotite from a meladiorite dike associated with another alkalic pluton farther south in the Beaverhead Mountains (R.F. Marvin, 1980, oral commun., in Evans, 1981, p. 105). Somewhat similar syenitic granites intruding the Yellowjacket Formation in the Salmon River Mountains west of Salmon, Idaho, have yielded a slightly discordant ^{207}Pb-^{235}U age of 492±10 m.y. for the Arnett Creek pluton, and slightly discordant ^{207}Pb-^{206}Pb ages of 521 m.y. and 541 m.y. for the Deep Creek pluton (Evans, 1981, p. 53-60, 94-105; Evans and Zartman, 1981). A reversed time sequence for these rocks was suggested by Rb-Sr whole-rock age determinations of 526 m.y. for the Arnett Creek pluton and 465 m.y. for the Deep Creek pluton (Evans, 1981, p. 100-105).

The isotope ages of the granite in the Beaverhead Mountains thus range from about 430-440 m.y. (in Railroad Canyon granite intrusive and the Beaverhead pluton) to 467-483 m.y. (Beaverhead pluton and meladiorite in southern Beaverhead Mountains); these ages are partly in conflict with geologic relations of the granite and its enclosing rocks. The granite intrudes the Kinnikinic Quartzite, of Middle Ordovician age, and Scholten and Ramspott (1968, p. 21) suggested that it was emplaced as a shallow, epizonal pluton, at a depth of 600 m or less. Its relation to younger Ordovician and Silurian rocks is not known, but the Kinnikinic clearly was a strongly indurated orthoquartzite when the granite was emplaced in it, which suggests at least a Late Ordovician age. The continuous deposition of marine carbonate rocks of the Saturday Mountain Formation from Middle and Late Ordovician to Early Silurian time suggests that emplacement of the granite was even later, perhaps in the Silurian. A period of Early Silurian uplift and erosion is shown by the channeled, irregular surface cut on top of the Saturday Mountain Formation beneath the Silurian Laketown Formation, and could be interpreted to be a result of uplift accompanying the emplacement of granitic rocks in the Early Silurian. If so, the first-reported isotope ages of 430-440 m.y. probably are most nearly correct for the granite in the Beaverhead Mountains. The older isotope ages of 467-483 m.y. may be from granites that are compositionally similar, but are older intrusions not clearly identified in the sequence of sedimentary rocks, or the ages may be mixed, or, in the case of the zircon ages, may reflect radiogenic lead inherited from older zircons.

The significantly older ages reported for the Deep Creek and Arnett Creek syenitic granite intrusions in the Salmon River Mountains suggest that these rocks may truly be older than the granite and related rocks in the Beaverhead Mountains. Their isotope ages, all on zircons, suggest that they were emplaced in Late Cambrian or Early Ordovician time. They are intrusive into the Yellowjacket Formation, which is interpreted to be autochthonous, and so differ from the granite in the Beaverhead Mountains, which was thrust faulted and tectonically transported eastward an unknown distance. If the Late Cambrian and Early Ordovician ages are correct, and not a result of inherited older radiogenic lead not detected in the zircon, the Deep Creek and Arnett Creek intrusions in the Salmon River Mountains represent an earlier episode of granitic intrusions in this region, an episode distinct from that represented by the thrust-faulted granite in the Beaverhead Mountains.

Finally, the abundant isotope ages suggesting early Paleozoic ages for granite in the Beaverhead Mountains and elsewhere in east-central Idaho, and the evidence on the geologic relations of granites in the Beaverhead Mountains, show that earlier questions regarding the validity of the isotope ages have been resolved (Ruppel, 1978, p. 18, 1981). These questions, raised by the initial results of radiometric studies (Ruppel, 1968; Scholten and Ramspott, 1968, p. 21), stressed the absence, at that time, of geologic evidence to confirm the early Paleozoic ages, or even to understand the intrusive or structural relations of the granite. They also pointed out the possibility that the granites could be Archean Dillon Granite Gneiss, which some granite of the Beaverhead Mountains closely resemble, with ages mixed by tectonism. The granite now is known to have been thrust faulted, to have intruded rocks as young as the Kinnikinic Quartzite, but not any younger rocks, and apparently not to have included any thrust slices of Dillon Granite Gneiss. The regional geologic evidence mostly supports the early Paleozoic radiometric ages; none of the known geologic relations suggest a different age.

TERTIARY GRANITIC ROCKS IN THE LEMHI RANGE

Granitic rocks that form the stocks, sheets, and sills in the central Lemhi Range cut and partly intruded the thrust faults. These rocks are cut by dikes related to the Challis Volcanics, of Eocene age; the Big Eightmile stock had cooled enough to chill the Challis dike that cuts it. At least the Sawmill Canyon sheet was unroofed by erosion before eruption of the Challis Volcanics. Geologic evidence therefore shows that these granitic rocks are younger than the thrusting but older than the Challis Volcanics. Regional evidence suggests that thrusting began in late Early Cretaceous time, perhaps about 100 m.y. ago, and ended by about 70–75 m.y. ago (Ruppel and Lopez, 1984, p. 32–33). The Challis Volcanics in this region are mainly of Eocene age, ranging from about 40 to 49 m.y. old (Staatz, 1979, p. A22–A23; Axelrod, 1966, p. 497–498; Armstrong, 1974, p. 12). On geologic evidence alone, the granitic rocks in the Lemhi Range are younger than about 70–75 m.y. and older than 40–49 m.y.

Radiometric age determinations suggest that these rocks are closest in age to the Challis Volcanics—and, indeed, that the ages of granite and volcanic rocks overlap slightly. Age determinations on two granitic intrusives in the central Lemhi Range have been made by John D. Obradovich, U.S. Geological Survey (written commun., 1971), yielding ages that probably are representative of all of the monzogranite-granodiorite-quartz monzodiorite and related rocks. Biotite from granodiorite in the Big Eightmile stock gave a K-Ar age of 49.4±1.7 m.y. Biotite from a quartz monzodiorite dike in Long Canyon, related to the nearby Gilmore stock, gave a K-Ar age of 51.3±1.5 m.y. The granite in the Ima stock at Patterson yielded a K-Ar age of 48.5±2.2 m.y., presumably on alkali feldspar (Ora Rostad, oral commun., 1981). Another K-Ar age on muscovite from the Ima granite stock was 41.3±1.4 m.y. (Armstrong and others, 1979), but the muscovite probably is a secondary, alteration product. These ages from the central Lemhi Range are essentially the same as ages determined for the compositionally similar granodioritic rocks in the Bobcat Gulch intrusive, near North Fork, Idaho, which have yielded $^{40}Ar/^{39}Ar$ ages of 49.1±0.6 m.y. on biotite, and an $^{40}Ar/^{39}Ar$ age of 48.5±0.8 m.y. on whole rock (B.B. Bunning and F.W. Burnet, written commun., 1981).[1]

The geologic relations of granitic rocks and Challis Volcanics indicate that the granitic rocks had cooled and been exposed by erosion, at least locally, before eruption of the volcanic rocks. The Challis Volcanics are as old as 54 m.y. near Challis, Idaho, but they were erupted throughout much of Eocene time, from 54 m.y. to 39 m.y. ago. The small fields of volcanic rocks in the Lemhi Range are satellitic to the main field near Challis, and seem most likely to have been erupted late in the volcanic episode, perhaps 40–45 m.y. ago.

STRUCTURAL CONTROLS ON THE EMPLACEMENT OF EOCENE GRANITIC ROCKS

The location and emplacement of Eocene granitic stocks and sheets in the Lemhi Range are interpreted to have been structurally controlled, both by steep

[1]Whole rock sample 80PO61. Sample is drill core from 1,497–1,538 ft, DDHBC–1 of Cominco American. Location, NE cor. sec. 5, T. 23 N., R. 21 E.; elevation, 6,240 ft. Rock is very fine, even grained, dark greenish gray; it contains clinopyroxene, brown hornblende, biotite, and sodic plagioclase. The plateau age on a whole rock sample was determined to be 48.5±0.8 m.y. by the $^{40}Ar/^{39}Ar$ age-spectrum technique by Lawrence W. Snee at Ohio State University.

Biotite sample 80PO62. Sample is drill core from 1,140–1,160 ft, DDHBC–1 of Cominco American. Location (same as above). Rock is silicic porphyry that contains andesine, quartz, potassium-feldspar, and biotite phenocrysts in fine-grained quartz-potassium-feldspar-rich groundmass. Biotite is reddish brown, slightly ragged, and very slightly altered. The $^{40}Ar/^{39}Ar$ total-fusion age of 49.1±0.6 m.y. on biotite was determined by Lawrence W. Snee at Ohio State University.

faults, which provided conduits for magmatic emplacement, and by the base of the Medicine Lodge thrust plate, in which magma spread laterally to form sheets like those exposed in this region (Ruppel, 1978, p. 18–20; 1982, p. 19–21; Ruppel and Lopez, 1984, p. 35–36). Nearly all the granitic stocks are at or near the edges of the mountain block, and all of them are emplaced in fault or fracture zones, most commonly at the intersections of northwest- and east-trending fault zones that are interpreted to reflect recurrent movement on ancient faults that bound basement blocks at depth. Only the Ima stock and the small stocks in Patterson Creek and Falls Creek, in the Patterson ring-fracture complex (Ruppel, 1982, p. 9, 11–14), depart from this regional association of stocks and steep faults, and even there the controlling ring-fracture zone, centered on the range front northwest of Patterson, seems likely to have formed above a northwest-trending basement fault near its intersection with a group of east-trending faults (Ruppel, 1982). All the stocks and sheets in the central Lemhi Range are roofed in or near imbricate thrust faults where geologic and geophysical studies indicate that the sole zone of the Medicine Lodge thrust system is at relatively shallow depth. No stocks seem to have penetrated far into the allochthonous block (Ruppel, 1978, p. 18–19; Ruppel and Lopez, 1984, p. 35–36). The Gilmore and Big Timber stocks and peripheral sheets show these relations most clearly. Both of them are at the intersections of zones of northwest- and east-trending faults or zones of faults, and both of them either demonstrably are, or can reasonably be inferred to be, at or near the base of the Medicine Lodge thrust plate. The emplacement of granitic rocks was early in the Cenozoic, however, before block uplift. The granitic magma clearly spread laterally in imbricate thrusts at the base of the thrust plate, which suggests that the steep, northwest- and east-trending faults had not then broken the overlying rocks of the Medicine Lodge thrust plate and that no conduits to higher levels in the thrust plate existed. The movement that broke the rocks of the Medicine Lodge plate took place later, and reflects renewed movement on old faults. The faults are interpreted to be shallow extensions of major, ancient reverse faults that cut basement rocks at depth, a conclusion also suggested by the occurrence, in the Big Timber stock alone, of inclusions of schistose crystalline metamorphic rocks.

The association of northwest- and east-trending faults and granitic intrusives is repeated to some degree in all of the major Tertiary granitic bodies in the Lemhi Range and in east-central Idaho, and suggests that northwest- and east-trending faults controlled granite emplacement in much the same way as in the Gilmore and Big Timber stocks. The Sawmill Canyon sheet is just northeast of the intersection of a zone of northwest-trending faults and a major, east-trending fault at the range front, and a small granitic dike is almost continuous with an offset segment of the east-trending fault; the sheet could have been fed through a still-concealed, east-trending dike, although the exposed dike does not appear to have been a feeder. The Little Eightmile stock and sheet (Staatz, 1973) is in a somewhat similar structural setting, at the intersection of east-trending faults at the fault-bounded front of the Beaverhead Mountains northwest of Leadore. Much farther north, in the Beaverhead Mountains northeast of Salmon, Idaho, the Carmen stock is intruded at the junction of northwest- and east-trending fault zones, mainly into autochthonous rocks of the Yellowjacket Formation, and was later broken by faults of both trends (Ruppel and others, 1983; Kilroy, 1981, p. 40, 44, 81). The Bobcat intrusive, also intruded into the Yellowjacket Formation, is in the Salmon River Mountains directly west of the Carmen Creek stock across the north end of the Salmon Basin, possibly reflecting a concealed east-trending fault at the north end of the Salmon Basin (Kilroy, 1981, p. 35). All the Tertiary granitic intrusives are either in the lower part of the Medicine Lodge thrust plate, essentially at its base, or in the Yellowjacket Formation beneath the thrust plate. All of them exhibit the same relations as the Gilmore and Big Timber stocks, and suggest the same conclusion—that their emplacement was controlled by northwest- and east-trending faults that cut the Yellowjacket Formation but did not reach upward into the overlying Medicine Lodge plate at that time. Later, during mid-Cenozoic block uplift, recurrent movement on the faults broke the rocks of the Medicine Lodge plate, and granitic rocks as well, to form the present northwest- and east-trending faults of this region.

No Tertiary granitic rocks are exposed in the southern parts of the Lemhi Range and Beaverhead Mountains, or in the Lost River Range farther west, all areas where the base of the Medicine Lodge thrust plate is deeply buried. Small, hydrothermal mineral deposits, which are widely distributed in this region, suggest that granitic rocks may be present at depth, but, if so, they seem likely to be confined to the lower part of the Medicine Lodge thrust plate, like the exposed Tertiary granitic rocks elsewhere in east-central Idaho.

STRUCTURE

The rocks of the central Lemhi Range are intricately folded and broken by multiple faults of different kinds and different ages. Thrust faults and nearly isoclinal overturned folds associated with them are present throughout the region. They are cut by steep faults that mostly trend about north and northwest, but which include east- and northeast-trending faults in some places. Most of the steep faults are interpreted to be Precambrian faults that moved recurrently, after thrust faulting, throughout much of middle and late Cenozoic time, and are still active today, but much of the evidence for faulting older than the thrust faults is obscured by the thrust plates themselves, which were not transported into this region until Late Cretaceous time.

The complex faulting and folding in the Lemhi Range are a result of several episodes of deformation and mountain growth that are best understood by fitting them into the regional structural framework of east-central Idaho and southwest Montana. That structural framework, reconstructed from detailed geologic mapping in the central Lemhi Range and from regional mapping elsewhere in east-central Idaho and southwest Montana, has been described in several earlier reports that are companions to this one. Regional thrust faulting, of Late Cretaceous age, is discussed in two reports, one that redefines and describes the Medicine Lodge thrust system and thrust plate (Ruppel, 1978), and one that discusses the thrust belt from the Medicine Lodge thrust plate in the Lemhi Range and Lost River Range eastward to the unthrusted, cratonic region in southwest Montana (Ruppel and Lopez, 1984). The steep faults of east-central Idaho and southwest Montana, their Precambrian ancestry, and their recurrent movement during Cenozoic mountain growth are discussed in another report (Ruppel, 1982) that also touches briefly on strike-slip faulting, and on late Cenozoic regional warping, normal faulting, and resultant drainage reversals, all of which were discussed in somewhat more detail in earlier reports (Ruppel, 1964, 1967). These reports describe a regional structural framework that, in the central Lemhi Range and Beaverhead Mountains, is dominated by the effects of two major periods of folding, faulting, and erosion in the Precambrian, and by the much later, mainly Late Cretaceous, Medicine Lodge thrust system.

The effects of Medicine Lodge thrusting and later faulting and folding are abundantly exposed in the ice-carved and stripped peaks and cirques of the central Lemhi Range. The sole zone of the Medicine Lodge thrust system is exposed in a few places above rocks of the Proterozoic Yellowjacket Formation, which are interpreted to be autochthonous (p. 12) and not significantly cut or displaced by thrust faults. The Medicine Lodge thrust plate includes other Proterozoic sedimentary rocks, but not rocks of the Yellowjacket Formation. It also includes Paleozoic sedimentary rocks, all of which have been tectonically transported eastward from the Cordilleran miogeocline in central Idaho. The sedimentary rocks in the Medicine Lodge plate were repeatedly broken by closely spaced, interlaced, imbricate thrust faults that dip down into the basal, sole zone of the thrust plate. These intricately folded and thrust-faulted rocks were broken again in the middle Cenozoic by steep faults and drape folded over the edges of block uplifts. The uplifts, which form the present mountain ranges of east-central Idaho, are interpreted to be more or less vertically uplifted basement-cored blocks, bounded mainly by northwest- and east-trending basement faults of Precambrian ancestry.

In late Pliocene and early Pleistocene time the region was warped into two broad, northeast-trending domal uplifts (Ruppel, 1982, p. 17), which are nearly at right angles to the trend of the mountain blocks. The southernmost and best known domal uplift crosses the Lemhi Range between the Donkey Hills Summit, just west of the southwest corner of the Gilmore quadrangle, and the Gilmore Summit a short distance southeast of Gilmore. Small normal faults, reflecting extension along the long axis of the domal uplifts, and a nearly undetectable amount of rotation of the mountain blocks, down on the northeast, accompanied this regional warping. The Donkey Hills-Gilmore Summit domal uplift divided the intermontane valleys of the region, reversing the drainage in the northern part of previously southeast-flowing river systems into northwest-flowing tributaries to the Salmon River, which was formed at this time.

PRECAMBRIAN AND EARLY PALEOZOIC FOLDING, FAULTING, AND UPLIFT

The Precambrian rocks of east-central Idaho were folded and broken in at least two major episodes of deformation before any Paleozoic rocks were deposited. In latest Precambrian and early Paleozoic time, they were again broadly folded, in minor episodes of regional uplift. The earliest deformation followed deposition of the Yellowjacket Formation, and was a major period of tectonism that either preceded or accompanied emplacement of igneous rocks in the Salmon area 1.3–1.4 billion years ago (Evans, 1981, p. 45–48; Ruppel and Lopez, 1984, p. 23–26; Ruppel, 1986, p. 121–123). This period of tectonism produced major and lasting changes throughout east-central Idaho and southwest Montana. It was accompanied by folding and overturning to the southwest, and by movement on steep northwest-, northeast-, and east-trending faults. The Yellowjacket depositional basin was destroyed, the present cratonic region of Archean crystalline rocks in southwest Montana was uplifted and exposed by faulting, and the eastern facies of the Yellowjacket Formation was eroded. The uplifted region formed the south flank of the Belt uplift, and probably was a source of sediments for the later, Middle Proterozoic rocks deposited in the Cordilleran miogeocline (Ruppel, 1986, p. 122). The major steep faults of the region were clearly defined at this time; they still control the structural fabric of east-central Idaho and southwest Montana.

A second major period of folding, uplift, and erosion took place later in the Precambrian, probably late in Middle Proterozoic time, and is clearly shown in the Lemhi Range by the relations between the Middle Proterozoic rocks of the Lemhi Group and Swauger Formation, and lower Paleozoic rocks. The Lemhi Group and Swauger Formation are overlain with angular unconformity by the Ordovician Summerhouse Formation and Kinnikinic Quartzite in the central Lemhi Range. In places, for example east of Patterson Creek (Ruppel, 1980), the Swauger Formation was eroded away completely and the Kinnikinic overlies the eroded edge of the Gunsight Formation. Farther south in the Lemhi Range, the Gunsight Formation is angularly overlain by the Early Cambrian(?) and Late Proterozoic(?) Wilbert Formation (Ruppel and others, 1975, p. 28–29; McCandless, 1982). Farther west, in the Lost River Range, the Swauger Formation is conformably and gradationally overlain by the Middle Proterozoic Lawson Creek Formation (Hobbs, 1980), a formation not known to be present in the Lemhi Range. The absence of both the Lawson Creek Formation, and in places, the Swauger Formation, beneath Paleozoic rocks in the Lemhi Range shows that as much as 4,500 m of folded Middle Proterozoic sedimentary rocks was removed by erosion before deposition of the Early Cambrian and Late Proterozoic(?) Wilbert Formation. Only slight regional unconformities separate the Wilbert Formation from the overlying Early Ordovician Summerhouse Formation, and the Summerhouse from the overlying Middle Ordovician Kinnikinic Quartzite (Ruppel and others, 1975, p. 28–29), so the principal deformation in Proterozoic rocks must have taken place well before the Early Cambrian, most probably in late Middle Proterozoic time but perhaps continuing into the early Late Proterozoic. The angular unconformity between Middle Proterozoic rocks and younger rocks clearly represents a period of regional uplift and folding that ended sedimentation in the Middle Proterozoic miogeocline, and resulted in uplift of the Lemhi arch (Sloss, 1954; Ruppel, 1978, p. 12; 1986).

The trends of late Middle Proterozoic folds are unknown, because later deformation has intricately refolded and broken the Middle Proterozoic rocks. Ross (1947, p. 1128–1129; 1961a, p. 237) suggested that the old folds in the Lemhi Range might trend nearly north or northeast, in contrast to later, more or less northwest fold trends, but he also stressed the uncertainty of determining old trends in a region so complexly deformed during later tectonic events. The period of major deformation in the late Middle Proterozic probably included faulting as well as uplift and folding, but no faults clearly of this age are known in the thrust-faulted, allochthonous Lemhi Group or Swauger Formation in the central Lemhi Range; if they exist, any evidence that would indicate their age has been confused or destroyed by later thrust faulting, rotation, and tight folding.

Finally, the rocks of the Lemhi Group and Swauger Formation that show the major deformation late in Middle Proterozoic time, and the younger rocks that record the minor uplifts in the early Paleozoic, have been tectonically transported far east of the region where they were deposited and where the late Middle Proterozoic deformation took place. That deformation occurred in the Cordilleran miogeocline in what is now central Idaho, and not where the rocks are now; it destroyed the Middle Proterozoic depositional region of the Lemhi Group and Swauger and Lawson Creek Formations. The late Middle Proterozoic deformation formed the Lemhi arch, and all the lower Paleozoic rocks shoaled eastward onto a deeply eroded highland underlain by the Middle Proterozoic rocks. No similar shoaling pattern is present in the Lemhi Group or Swauger Formation. The original eastern depositional edges of these Proterozoic rocks, against the south flank of the older Belt uplift, are not exposed in east-central Idaho. Probably they were at least partly destroyed by erosion in the Late Proterozoic and early Paleozoic, when they were exposed on the Lemhi arch, and the resulting eroded edges were buried beneath thrust plates in the Late Cretaceous.

THRUST FAULTS AND RELATED FOLDS

Most of the sedimentary rocks in the central Lemhi Range are in the Medicine Lodge thrust plate, and are miogeoclinal rocks tectonically transported eastward perhaps as much as 150–170 km (Ruppel, 1978; Ruppel and Lopez, 1984). These rocks were thrust over the Yellowjacket Formation, which is interpreted to be autochthonous. Above the sole zone of the Medicine Lodge thrust, the sedimentary rocks are folded into tight, nearly isoclinal asymmetric folds, overturned to the east, and are broken by a pervasive network of closely spaced, multiple, interlaced imbricate thrust faults, the structures most characteristic of the central Lemhi Range and Beaverhead Mountains.

The Medicine Lodge thrust system includes the sole zone of the Medicine Lodge thrust plate, and the network of imbricate thrusts and overturned folds in the thrust plate. The sole zone of the Medicine Lodge thrust is exposed most widely north and west of Mill Mountain, both on the northeast flank of Mill Mountain and farther west near Cooper Creek and Wade Creek (Ruppel, 1980). Smaller exposures also are present much farther south near the junction of Lake and Big Timber Creeks, and in the upper reaches of Squaw Creek (Ruppel and Lopez, 1981). In all of these areas, the sole zone consists of crushed, brecciated, and in places mylonitized rocks that generally do not contain any clearly recognizable, discrete thrust surfaces. The sole zone grades upward into the Medicine Lodge thrust plate through a zone that shows a gradual decrease in shearing and brecciation. The sole zone and brecciated rocks above it may be as much as 300 m thick, although more commonly the combined zones are only a few tens of meters to 100 m thick. The Yellowjacket Formation beneath the sole zone commonly is also sheared, but not as conspicuously as the overlying rocks, nor in as thick a zone. On Mill Mountain (Ruppel, 1980), the Big Creek Formation is thrust over a coarse turbidite facies of the Yellowjacket Formation, and is strongly sheared and brecciated. Similarly sheared and brecciated quartzite of the Big Creek Formation is thrust across the quartzite and siltite of the Yellowjacket Formation near Cooper Creek, but 3–4 km farther west, near Wade Creek, the steeply overturned Swauger Formation is at the thrust sole and the entire Lemhi Group is missing.

The Cooper Creek and Wade Creek exposures are at the south edge of the Hayden Creek window, an erosional window through the Medicine Lodge thrust (Ruppel, 1978, p. 4–9). The window extends farther north, to include much of the area in the upper part of Hayden Creek and in Bear Valley Creek (Anderson, 1961), north of the Patterson quadrangle. At the north edge of the window, north of Bear Valley Creek, strongly brecciated and locally mylonitized quartzite of the Swauger Formation is above the sole zone of the thrust. The western part of the Hayden Creek window has not been mapped in any detail, but reconnaissance mapping suggests that most of the rocks above the thrust sole there are part of the Lemhi Group.

The small windows at Lake Creek and Squaw Creek, west and southwest of Gilmore (Ruppel and Lopez, 1981), expose the Yellowjacket Formation beneath the intensely brecciated Kinnikinic Quartzite, the Saturday Mountain Formation, and, at Lake Creek, the Gunsight Formation. In these places, the rocks are brecciated in wide areas surrounding the windows, and the sole zone of the Medicine Lodge thrust system probably underlies the brecciated rocks at shallow depth. The widespread distribution of shattered Kinnikinic Quartzite in Sawmill Canyon, west of the Squaw Creek window, suggests that this wide basin also is underlain at shallow depth by the sole zone of the Medicine Lodge thrust which was intruded by the Sawmill Canyon sheet and partly concealed beneath blanketing Challis Volcanics.

The sole zone of the Medicine Lodge thrust system, which is nearly flat where it is exposed in the Lemhi Range, probably lies at relatively shallow depth beneath the central part of the range. Aeromagnetic evidence (U.S. Geological Survey, 1971) suggests that a nearly flat to gently west dipping magnetic surface, interpreted to reflect the magnetite-bearing Yellowjacket Formation beneath the sole zone of the thrust plate, underlies the axial part of the central Lemhi Range.

The thrust plate was folded and broken by younger, steeply dipping faults, as a result of Cenozoic tectonism. At the edges of both the Lemhi Range and Beaverhead Mountains, it is monoclinally folded, and commonly dips into the adjacent valley at an angle of 20°–60°, although in places, the dip is nearly vertical (Ruppel and Lopez, 1981, section *E–E'*; Ruppel, 1982, p. 9–11).

The Proterozoic and Paleozoic rocks above the sole zone of the Medicine Lodge thrust are tightly folded and broken by multiple, interlaced imbricate thrust faults. The tight folds commonly are asymmetric, in places nearly isoclinal, overturned toward the east, and broken on their overturned limbs by imbricate thrusts. The imbricate thrusts dip 5°–15° southwest almost everywhere in the Lemhi Range and Beaverhead Mountains except on the range flanks, and dip down into the sole zone of the thrust system. These closely spaced, interlaced faults shuffle upright and overturned limbs of associated folds, and make the structure of this region complex, but mostly they are relatively small stretch thrusts that have nearly obliterated the folds they break, to produce a confused interlayering of upright and overturned rocks. The imbricate thrusts are discrete faults, which in quartzitic rocks commonly are overlain by 1–15 m of fault breccia, but which in carbonate rocks may be marked only by a few centimeters or less of clay gouge. Surfaces of imbricate thrusts are widely exposed in the Lemhi Range, where they have been stripped clean by glaciation and by alpine erosion processes that are particularly effective in the shattered rocks of the hanging walls. The fault surfaces have been smoothed and slickensided, silicified in quartzitic rocks, and heavily limonite stained. Most of the brecciated rocks above imbricate thrust faults along ridge crests of both the Lemhi Range and Beaverhead Mountains have been deeply weathered and disaggregated by alpine weathering and frost action, forming gently sloping, hummocky creep slopes. The nearly flat surfaces near the crest of the central Lemhi Range and the

Beaverhead Mountains thus are thrust surfaces being stripped by alpine erosion and creep, rather than remnants of an old erosion surface (p. 50).

The tight folding and imbricate thrusting was accompanied by widespread development of small, lenticular, curving gash veins, filled with massive, white quartz and specular hematite. Many of these have been prospected on the south slope of Grizzly Hill, north of Leadore, and they are present in most of the thrust-faulted rocks of the central Lemhi Range, as well. Thrusting also was accompanied by the intrusion of rare pebble dikes, above imbricate thrusts, that are most common, although still rare, in the vicinity of Gilmore, especially near Lemhi Union Gulch and Sheephorn Peak. The pebble dikes suggest that fluidization of tectonic breccias occurred during thrusting.

The time of Medicine Lodge thrusting is only approximately known; starting and ending points remain somewhat uncertain. Local evidence provides only a minimum date of about 50 m.y., which is based on the age of granitic rocks that cut and intrude thrust-faulted rocks in the central Lemhi Range.

Regional evidence indicates, however, that the Medicine Lodge and other thrust plates of the thrust belt in east-central Idaho and southwest Montana were in their present position 75–76 m.y. ago, the time of emplacement of the oldest granitic rocks in the Pioneer Mountains in southwest Montana, which cut the Grasshopper thrust plate (Zen and others, 1975, p. 367–370; Ruppel and Lopez, 1984, p. 33). Stratigraphic evidence from the syntectonic Beaverhead Formation in southwest Montana (Ryder and Scholten, 1973, p. 783; Lowell and Klepper, 1953, p. 240), indicates that thrusting began no later than late Early Cretaceous (Albian) time, perhaps about 100 m.y. ago. The combined stratigraphic and radiometric evidence thus suggests that thrust fault movement in this region began no later than late Early Cretaceous (Albian) time, perhaps 100 m.y. ago, and was completed by about 75 m.y. ago. The Medicine Lodge thrust plate is inferred to be older than the Grasshopper thrust plate and other parts of the thrust belt farther east (Ruppel and Lopez, 1984, p. 33), but how much older is not known.

STEEP FAULTS, BLOCK UPLIFTS, AND MONOCLINAL FOLDS

The mountain ranges of east-central Idaho were first outlined in mid-Cenozoic time, in a major period of movement on steeply dipping faults, block uplifting, and monoclinal folding over the edges of the rising mountain blocks (Ruppel, 1982). Most, perhaps all, of the steep faults of the region seem to have moved at this time. The faults are interpreted to reflect renewed movement on ancient faults that bound blocks of basement crystalline rocks at depth. The system of basement faults, first formed in the Precambrian, moved recurrently in mid-Cenozoic time and imposed, from beneath, an ancient structural pattern on these already complexly deformed rocks that had been thrust faulted into this region. The reasons for interpreting the ranges to be basement-controlled block uplifts are discussed in greater detail in Ruppel (1982).

Many of the steep faults are marked by scarps in glacial deposits and younger alluvial deposits, which shows that mountain growth, controlled by ancient but still active faults, is continuing today. The block uplifts, bounded mainly by northwest- and east-trending steep faults, are interpreted to be cored by basement crystalline rocks, like the basement-cored ranges of southwest Montana. The crystalline cores are not exposed in the Idaho ranges, however, where they and the steep faults of Precambrian ancestry inferred to cut them remain concealed beneath the sequence of sedimentary rocks enormously thickened by thrust faults in the Medicine Lodge thrust plate.

The thrust-faulted and complexly folded rocks of the region were folded again over the edges of the rising basement blocks, to form the flanks of the Lemhi Range and Beaverhead Mountains. These mountain ranges once were considered to be broad anticlines partly bordered by normal faults, but are better described as flat-topped block uplifts with monoclinal, drape-folded flanks. The monoclinal folds are broken by normal faults that mostly are relatively minor, secondary normal faults formed by gravitational collapse of the monoclinal shoulders of the uplifted blocks. A few small, but conspicuous, normal faults reflect regional extension related to late Cenozoic regional warping rather than to mid-Cenozoic block uplift.

The mid-Cenozoic age of block uplifting, and of recurrent movement on controlling basement faults, is based on relations in the Tertiary rocks that fill the adjacent valleys. Conglomeratic zones in Tertiary rocks along the range flanks suggest that block uplift may have started as early as late Oligocene, but occurred mainly in Miocene and early to middle Pliocene time. The youthful Holocene scarps that mark many of the faults show that uplift and mountain building are still active.

Steep faults.—Steeply dipping faults break virtually all of the rocks and surficial deposits of the central Lemhi Range and adjacent Beaverhead Mountains. Their trends fall more or less into four groups—northwest, north, east, and northeast, in general order of abundance, but with local variations in trend that nearly box the compass. Their dips range from 30° to

35°, on part of the prominent, east-trending fault along the base of the Beaverhead Mountains north of Leadore, to vertical on the north-trending faults. Most of the northwest-, northeast-, and east-trending faults probably dip at angles of 60°–80°, judging from their fairly straight traces and from measured dips on a few well-exposed faults. They include reverse faults, normal faults, and strike-slip faults. All of them break the thrust faults of the Medicine Lodge thrust system, and so reflect movement later than thrusting. The north-trending, vertical faults break those of other trends. Northwest-, northeast-, and east-trending faults break and offset each other, but generally do not break the north-trending faults. The relations suggest that the northwest-, northeast-, and east-trending faults formed at about the same time, and that the north-trending faults formed later. All the faults are marked in places by young scarps that show continuing movement almost to the present time.

The faults at and near the margins of the mountain blocks previously have been considered to be major normal faults reflecting regional extension. The conclusion that the mountain ranges of east-central Idaho are drape-folded, basement-cored block uplifts, however, suggests that many of these faults are slip surfaces of gravitational slide blocks controlled by folded thrust surfaces (Ruppel, 1982). Other faults, with small displacements, reflect minor extensional movement related to later regional warping, and not to mid-Cenozoic block uplifting (Ruppel, 1982, p. 15–17).

Displacement on the steep faults commonly is difficult to determine because different imbricate thrust slices, of uncertain or unknown relation to each other, may be faulted together. Displacements on northwest-, east-, and northeast-trending faults within the ranges most commonly do not exceed a few hundred meters, where they can be determined at all, and are mainly or entirely dip slip. The apparent faults of similar trends at the range margins are at least partly slip surfaces of slide blocks, and partly are younger, small normal faults. The north-trending faults are largely strike and oblique slip; lateral displacement on some of them may be measured in kilometers, not meters.

The long, linear mountain blocks of east-central Idaho trend northwest, but are repeatedly broken by east-trending segments that are especially prominent in the Lemhi Range and Beaverhead Mountains (Ruppel, 1964, 1982). The northwest trends are parallel to those of most reverse and normal faults in the ranges. Northwest-trending reverse faults break and repeat the Jefferson Formation south of Gilmore (Ruppel and Lopez, 1981). A much larger, parallel, reverse fault is exposed near the range crest west and southwest of Gilmore where it is intruded by the Big Timber stock at the base of the Medicine Lodge thrust plate. All northwest-trending reverse faults are interpreted to be shallow extensions of ancient basement faults that were overridden by the Medicine Lodge thrust plate, and to reflect recurrent movement on these ancient faults that broke the thrust plate during mid-Cenozoic block uplift (Ruppel, 1982, p. 11).

Most of the other northwest-trending faults in the axial part of the central Lemhi Range are normal faults that define small grabens or half-grabens. Probably they reflect both stretching during block uplift and drape folding, partly controlled by recurrent movement on basement faults at depth, and later Cenozoic normal faulting related to regional warping (Ruppel, 1982, p. 17). The prominent northwest-trending faults at and near the range flanks near Patterson and Gilmore also are mainly a result of late Cenozoic regional warping. Those at Patterson are nearly parallel to adjacent drape-folded thrust faults and probably include older slip surfaces of gravitational slide blocks, controlled by folded thrust surfaces, and integrated into more linear, normal faults during late Cenozoic faulting.

East-trending faults are most conspicuous along the southwest front of the Lemhi Range at the mouth of Sawmill Canyon and in the upper part of the Pahsimeroi Valley, and along the southwest front of the Beaverhead Mountains near Leadore (fig. 2) (Ruppel and Lopez, 1981; Ruppel, 1968; Staatz, 1973). These faults are long, continuous fractures, broken in places by younger, north-trending faults. Most of them are in groups of closely spaced faults, like those north of Leadore (Ruppel, 1968), although the fault at the mouth of Sawmill Canyon is a single, well-defined fracture (Ruppel and Lopez, 1981). All of them stop abruptly against north-trending faults. Their dips commonly are very steep, 70° to nearly vertical, judging from fault traces and a few measured dips; the frontal fault of the east-trending group of faults north of Leadore, however, dips only 35° where it is cut underground in the Leadville mine. This flat dip seems anomalous, considering the remarkably straight and well-defined trace of the rest of the fault, and the steep dips of other east-trending faults nearby. It coincides with a small, southward bulge in the frontal zone, and may reflect some later local, gravitational sliding. All the east-trending frontal faults are marked by scarps that extend across young surficial deposits, and which reflect continuing recurrent movement. The faults in bedrock, behind the frontal fault, generally do not show such evidence of recent movement. The east-trending faults, like those that trend northwest, are interpreted to reflect recurrent movement on basement faults at depth beneath the Medicine Lodge thrust plate (Ruppel, 1982, p. 5–6).

The relatively few northeast-trending faults in the central Lemhi Range all seem to be steeply dipping normal faults. They probably also are controlled by faults of Precambrian ancestry, because major faults of this trend near Salmon, Idaho, are known to have formed first in the Precambrian, at about the same time as the other basement faults (Lopez, 1981, p. 108–112; Ruppel, 1982, p. 6–7; Hahn and Hughes, 1984, p. 65; O'Neill and Lopez, 1985, p. 437–439).

North-trending faults break the other steeply dipping faults, and are vertical faults that are mainly strike slip and oblique slip (Ruppel, 1964; 1982). They occur in closely spaced groups of parallel faults that are most common where the mountain ranges change trend from northwest to east, and in the east-trending mountain segments. They are both remarkably long, some of them clearly traceable for 10–15 km, and remarkably straight, almost everywhere straight north. The straight, principal faults are linked by short curving faults that appear to transfer movement from one north-trending fault to another. These curved, linking faults are most common where one fault or group of faults horsetails and dies out, and at least some of the movement is transferred to a parallel fault or group of faults in complexly faulted areas, like those west of Summerhouse Canyon, in the Swan Basin, and north of Leadore.

Lateral displacement along individual north-trending faults, or across closely spaced groups of faults, ranges from a few tens of meters to several kilometers, but generally is difficult to determine because parts of different thrust plates are faulted together. The estimates of displacement are based mainly on offsets of other faults assumed to have been continuous originally, like the offset northwest- and east-trending fault zones near the mouth of Sawmill Canyon, and north and east of Leadore.

The distribution of north-trending faults, and their relation to steep faults of other trends, suggest that these faults, too, are products of block uplift. But, unlike the steep faults of other trends, the north-trending faults cannot be related to similarly trending faults of Precambrian ancestry, because no such Precambrian faults are known (Ruppel, 1982, p. 7). Despite their length and substantial displacements, the north-trending faults do not completely cross either the Lemhi Range or Beaverhead Mountains. Rather, they are concentrated where the ranges change trend and along the east-trending, fault-bounded range segments. They die out, both in bedrock and valley fill, by horsetailing or by merging into steep faults of other trends. East-trending fault zones abruptly end at north-trending faults, both northeast of Leadore and at Sawmill Canyon, and no offset parts of the east-trending zones are present elsewhere; the relations suggest that the inferred underlying east-trending basement fault, inferred to have moved vertically to form the east-trending zones, must also end. The north-trending faults also break, but do not terminate so abruptly, the steep faults of other trends. They clearly are most closely associated with the east-trending faults, both in terms of distribution and unique relations. The north-trending faults seem to be explained best by interpeting them to be tear faults in the covering blanket of sedimentary rocks, above different basement blocks that moved laterally or tilted differentially during block uplift. This interpretation fits the observed relations reasonably well, and makes understandable both the concentration of north-trending faults where block uplifts change trend, and so where different basement blocks intersect, and the occurrence of oblique slip and strike slip faults in a region of mid-Cenozoic vertical tectonism. It suggests that tear-faulting probably should occur fairly late in block-uplift, as basement blocks adjust to new positions, and so explains why north-trending faults cut all the others. It suggests that the north-trending faults should be larger and more continuous at greater depth, nearer the edges of the basement blocks, as they are in the Yellowjacket Formation in the Salmon area (Ruppel and others, 1983), and suggests that they should splay upward into more complex fault zones like those in the central Lemhi Range and in the Beaverhead Mountains north of Leadore. Also it explains why the north-trending faults, like the northwest- and east-trending faults, are marked by young scarps, because late regional warping and continued mountain growth could reasonably be accompanied by continued adjustments between basement blocks.

Monoclinal folds.—The Lemhi Range and Beaverhead Mountains have been described as domal or anticlinal uplifts with faulted limbs (Umpleby, 1913, p. 49; Hait, 1965, p. 164–181; Scholten and Ramspott, 1968, p. 27–40; Lucchitta, 1966, p. 92–131), but mapping in and around the central Lemhi Range shows that they are broad, northwest-trending, nearly flat-topped uplifts, flanked by monoclinal limbs that dip steeply into the adjacent valleys (Ruppel, 1968, 1980, 1982; Ruppel and Lopez, 1981). The structural flatness of the axial part of the central Lemhi Range is suggested by the attitudes of thrust faults there, which dip 5°–15° southwest into the nearly flat sole zone of the Medicine Lodge thrust. In addition, the remnants of upper Tertiary fanglomerate in Sawmill Canyon, deposited in the core of the range on deeply eroded, thrust-faulted rocks, remain essentially horizontal, neither folded nor tilted.

The flat thrust surfaces are abruptly and steeply folded at the crests of the range flanks. On the west flank of the Lemhi Range at Patterson (Ruppel, 1980; 1982, p. 12–13), imbricate thrust slices and thrust faults abruptly steepen westward at the crest of the valley wall and dip west almost as steeply as the younger normal faults along the west flank of the range. On the east flank of the range, imbricate thrust faults that are flat or gently west dipping farther west are abruptly folded and dip 20°–80° eastward (Ruppel, 1980; 1982, p. 8, 12–13; Ruppel and Lopez, 1981). Similarly, in the Beaverhead Mountains, imbricate thrust slices dip south and southwest at angles of 10° or less at the top of Grizzly Hill, but abruptly steepen at the crest of the valley wall, to dip about 35° southward into the Lemhi Valley.

The structural configuration of flat-topped blocks with monoclinal flanks suggests that the thrust-faulted rocks of the region were folded more or less passively over vertically rising basement blocks (Ruppel, 1982). The structurally flat, central parts of the mountain blocks are interpreted as being above the central parts of uplifted basement blocks. The structural flatness suggests that uplift was essentially vertical, without significant tilting of the basement blocks. The monoclinal folds are interpreted as drape folds over the edges of rising basement blocks, a folding process facilitated by the multiple slip surfaces in imbricate thrust faults and by the sole zone of the Medicine Lodge thrust system.

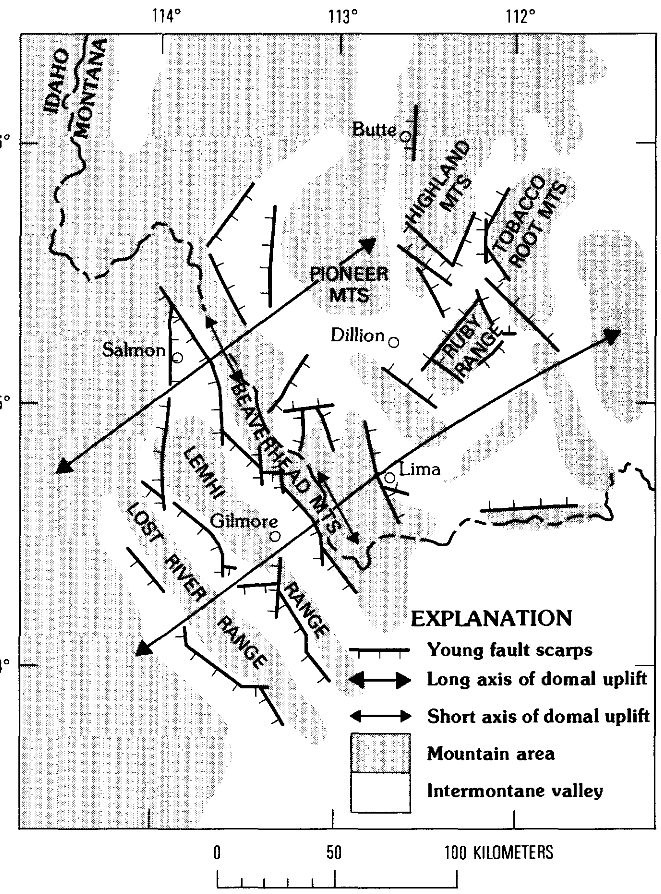


FIGURE 18.—Map showing the effects of Pliocene to Holocene arching and faulting north of the Snake River Plain, Idaho-Montana (modified from Ruppel, 1982, p. 17).

LATE CENOZOIC REGIONAL WARPING AND YOUNG NORMAL FAULTS

The development of broad, northeast-trending domal uplifts, across the grain of earlier block uplifts, was first suggested by Kirkham (1927, p. 11), who considered the uplift that crosses the Lemhi Range south of Gilmore to be a result of regional warping parallel with and related to the Snake River downwarp, principally in Pleistocene time (Kirkham, 1931, p. 456–482). Subsequent studies have suggested that a second linear dome, broader and lower than the one near Gilmore, but parallel to it, is defined by drainage patterns farther northwest (fig. 18), and that regional warping was accompanied by relatively minor but conspicuous normal faulting and by renewed movement on all faults (Ruppel, 1967; 1982, p. 15–17).

The axis of the northeast-trending dome near Gilmore is marked by the Gilmore Summit, southeast of Gilmore, and the Donkey Hills Summit, at the head of Summit Creek west of Sawmill Canyon. These broad summits, which were noted by Kirkham (1927, p. 13) in his discussion of regional upwarping, divide the drainages in the northwest-trending structural trenches that flank the Lemhi Range. The drainage originally was through going, to the southeast, but the patterns of drainage in the valleys of the now northwest-flowing Lemhi and Pahsimeroi Rivers suggest that both of them occupy valleys that were reversed by regional upwarping (Ruppel, 1967, p. 657). The former presence of a southeast-flowing river is shown, too, by the distribution of deposits of placer gold in the Lemhi Valley, which extend upstream from the source along the present river, rather than down, and by the presence of boulders of augen gneiss in high-level terrace gravels southeast of Salmon, Idaho. The gneiss could have come only from the area northwest of Salmon, an area that is north of the present Lemhi River (Ruppel and others, 1983). Indeed, much of the placer gold seems likely to have been reworked from the same gravels that contain augen-gneiss boulders, and originally could have been derived from lode deposits like those at the Shoo Fly and Queen of the Hills mines northwest of Salmon (Anderson, 1956, p. 58–53; 1959, p. 73–81).

The regional warping that formed the northeast-trending domal uplifts apparently began in the late Pliocene, and had reversed the drainage in the Lemhi and Pahsimeroi Valleys before glaciation in the early Pleistocene (Kirkham, 1927, p. 26; 1931, p. 474–482; Ruppel, 1967, p. 657–658). Young faults, thought to be related to the domal uplifts, cut glacial deposits and Holocene alluvial deposits; these faults are marked by scarps that have been only slightly rounded and dissected, suggesting that regional warping has continued to the present time.

The young faults that formed with the northeast-trending domal uplifts are steep, northwest-trending normal faults with as much as 120-m displacement along the southwest fronts of the Lemhi Range and Beaverhead Mountains (Ruppel, 1982, p. 16–17). These small, northwest-trending faults are interpreted to be a result of extension, or stretching parallel to the long axes of the domal uplifts. Many older faults, particularly those that trend north and east, are marked by scarps as young as those on the northwest-trending faults, and moved recurrently at the same time. The recurrent movement on older faults probably reflects uplift and tilting of basement blocks, during regional warping, like the tilting that appears to have rotated the Lemhi Range block up on the southwest and down on the northeast by about the same amount, a little more than 100 m, in late Pliocene through Holocene time (Ruppel, 1982, p. 17).

SOME REGIONAL CONSEQUENCES OF CENOZOIC BLOCK UPLIFTING, WARPING, AND FAULTING

Many of the consequences of Cenozoic tectonism and mountain building have been touched on briefly in other sections of this report, but because these consequences are so widespread and diverse it seems useful to summarize them here. A principal conclusion is that the mountain ranges of east-central Idaho are flat-topped block uplifts flanked by monoclinally folded limbs, formed by nearly vertical uplift of basement blocks in late Oligocene to early or middle Pliocene time. The basement blocks are interpreted to be bounded by major faults of Precambrian ancestry, which moved recurrently during mid-Cenozoic block uplift to fold and break the overlying Medicine Lodge thrust plate, and continue to control the structural framework of the region today. The rising block uplifts were the sources of the flood of boulder gravels that now are conglomerates in Tertiary tuffaceous rocks along the mountain fronts. They also were the sources of major landslides into the flanking Tertiary basins, as increasingly unsupported masses of bedrock failed and slid from the monoclinal shoulders of the uplifts, partly on ready-made slip surfaces provided by folded and steepened thrust faults. Both Challis Volcanics and younger Tertiary tuffaceous rocks appear to have been stripped from most of the Lemhi Range and Beaverhead Mountains by gravitational sliding during block uplift. The valley fill seems likely to be a complexly interlayered sequence of tuffaceous rocks in place, mixed with other, partly older tuffaceous rocks in landslides, and with interlayered major landslide plates of Challis Volcanics and Paleozoic and Proterozoic sedimentary rocks.

In late Pliocene and early Pleistocene time the northwest-trending block uplifts were broadly cross-folded by northeast-trending, linear uplifts parallel to the Snake River Plain. The original southeast-flowing rivers in the structural trenches between block uplifts were partly reversed by this regional warping, to form the present northwest-flowing Lemhi and Pahsimeroi Rivers. The Salmon River was formed by piracy and faulting that integrated bits and pieces of other river systems into one major system that drains much of central Idaho (Ruppel, 1967, p. 660–661). Integration of the Salmon River must have been very rapid, because there are no indications that the Lemhi or Pahsimeroi Rivers were ponded after reversal, and the valley fill along the lower reaches of the Lemhi River has been deeply excavated. Glacial valleys in the Beaverhead Mountains east of the Salmon River are very steep—so steep that the glaciers must have been virtually ice cascades—and are about at grade with the Salmon River, which suggests that the river had excavated its valley nearly to its present depth before the later Pleistocene glaciations.

In contrast to the deeply excavated Salmon River Valley, the Big Hole basin region in southwest Montana, east of the Beaverhead Mountains and adjacent to the Salmon River drainage, has not been excavated. This headwaters region of the Big Hole River, which also was formed by drainage reversal and piracy in the late Pliocene and early Pleistocene, is nearly ponded behind a bedrock lip at the north end of the Big Hole basin, and streams tributary to the upper Big Hole River meander through swamps and bogs in long, gently sloping, glaciated canyons on the east flank of the Beaverhead Mountains. The steep, rapidly eroding tributary streams of the Salmon River now interfinger with the Big Hole tributaries along the Continental Divide; some of them east and southeast of Gibbonsville, Idaho, have steep, eroded headwalls that have been cut nearly into the tuffaceous fill of the Big Hole basin. The Big Hole basin clearly is destined for capture by the Salmon River drainage system, beheading the Big Hole River and pirating its entire headwaters region and principal source of water.

The development of northeast-trending domes in Pliocene and early Pleistocene time was accompanied by relatively minor normal faulting, by a slight amount of regional tilting down on the northeast of earlier uplifted mountain blocks, and by renewed movement on many of the older steeply dipping faults in the region. Faulting has continued through the Pleistocene and Holocene to the present time, as shown by faults that cut even the youngest glacial and alluvial deposits and commonly are marked by scarps that are only slightly rounded and dissected. Many faults cut Holocene landslide deposits, and the association of young landslides and young faults is so common that it suggests that many of the landslides are a result of earthquakes accompanying fault movement. The abundance of little-dissected, very young fault scarps in and near the central parts of the Lemhi Range and Beaverhead Mountains shows that these faults are still active, and are capable of renewed movement accompanied by earthquakes and by earthquake-induced landsliding. A recent example was the Borah Peak earthquake on the west side of the Lost River Range on October 28, 1983. Accumulating evidence indicates that mountain growth has been almost continuous in this region since the late Oligocene, and that it continues today.

MINERAL DEPOSITS

The mineral deposits in the central Lemhi Range include base- and precious-metal veins and bedding replacement deposits in carbonate rocks including a few vein deposits that contain gold or silver with few if any base metals, tungsten-copper-silver quartz veins and copper-silver quartz veins, disseminated deposits of copper and molybdenum in granitic stocks, deposits of secondary copper minerals commonly associated with the Apple Creek Formation, a deposit of lead disseminated in granite, a deposit of copper-bearing magnetite skarn, small gash veins filled with quartz and specular hematite, and a few nonmetallic deposits. The most valuable deposits have been the base- and precious-metal veins and replacements, particularly those mined in the Texas district at Gilmore, and the tungsten-quartz veins in the Blue Wing district at Patterson.

Most of the mineral deposits occur in reasonably well defined areas; the prospects and mines that explore them are loosely organized into mining districts (fig. 19). The largest area of lead-silver mineralization is along the east flank of the Lemhi Range, from Gilmore south to Horseshoe Gulch and Spring Mountain Canyon (Ruppel and Lopez, 1981). This area, 18 km long and 1-4 km wide, includes the Texas and Spring Mountain districts.

The Texas district at Gilmore is one of the main mining districts in the Lemhi Range. For many years it was the principal source of lead-silver ores in Lemhi County. The mines of the district explore lead-silver veins and replacements, a gold vein, and, in the north part of the district, lead-silver veins that also contain some gold. The principal mines in the central part of the district are numbered on figure 19: the Pittsburgh-Idaho (United-Idaho) (loc. 5), the Latest Out (loc. 6), and the Allie (Martha) gold mine (loc. 7) and in the north part of the district the Hilltop (loc. 8), and the Democrat (fig. 19, loc. 9). The Silver Moon mine (loc. 10) explores a silver deposit in the south part of the district.

The Lemhi Union mine (loc. 19), farther south, probably is best included in the Spring Mountain district, which for the most part encompasses a group of small mines and prospects that explore lead-silver veins. The district also includes a deposit of bornite-bearing magnetite skarn at the Sims (Bruce estate) prospect (loc. 23).

Only a few prospects are known in the area west of the Texas and Spring Mountain districts. Most of them along the east flank of Sawmill Canyon explore lead-silver or barite veins; a few prospects near the junction of Lake and Big Timber Creeks explore the contact zone of the Big Timber stock (Ruppel and Lopez, 1981).

Lead-silver ores similar to those of the Gilmore area were also mined in the Junction district (fig. 19), a small district in the Beaverhead Mountains north of Leadore (Ruppel, 1968). The only mine in the Junction district known to have produced any significant quantity of lead-silver ore is the Leadville (Sunset) mine (loc. 1). The district also contains a number of other lead-silver prospects, most of them located on the upper part of Grizzly Hill north of the Leadore quadrangle, a deposit of lead minerals disseminated in granite at the Kimmel mine (loc. 1), several prospects that explore small deposits of secondary copper minerals, and many prospect pits dug on gash veins of quartz and hematite. The district extends about 8 km northwest of the Leadville mine to include prospects in the Mineral Hill mining area at the mouth of Mollie Gulch. The adjoining Little Eightmile mining area at the mouth of Little Eightmile Creek (Staatz, 1973; Thune, 1941), north of the Leadore quadrangle, includes lead-silver veins similar to those in the Leadville mine, and calcite-quartz-stibnite veins, the only reported occurrence of this antimony mineral in the mineral deposits of the Lemhi Valley.

Two other major mineralized areas are known in the central Lemhi Range, both in the Patterson quadrangle (Ruppel, 1980), both characterized by mineral deposits

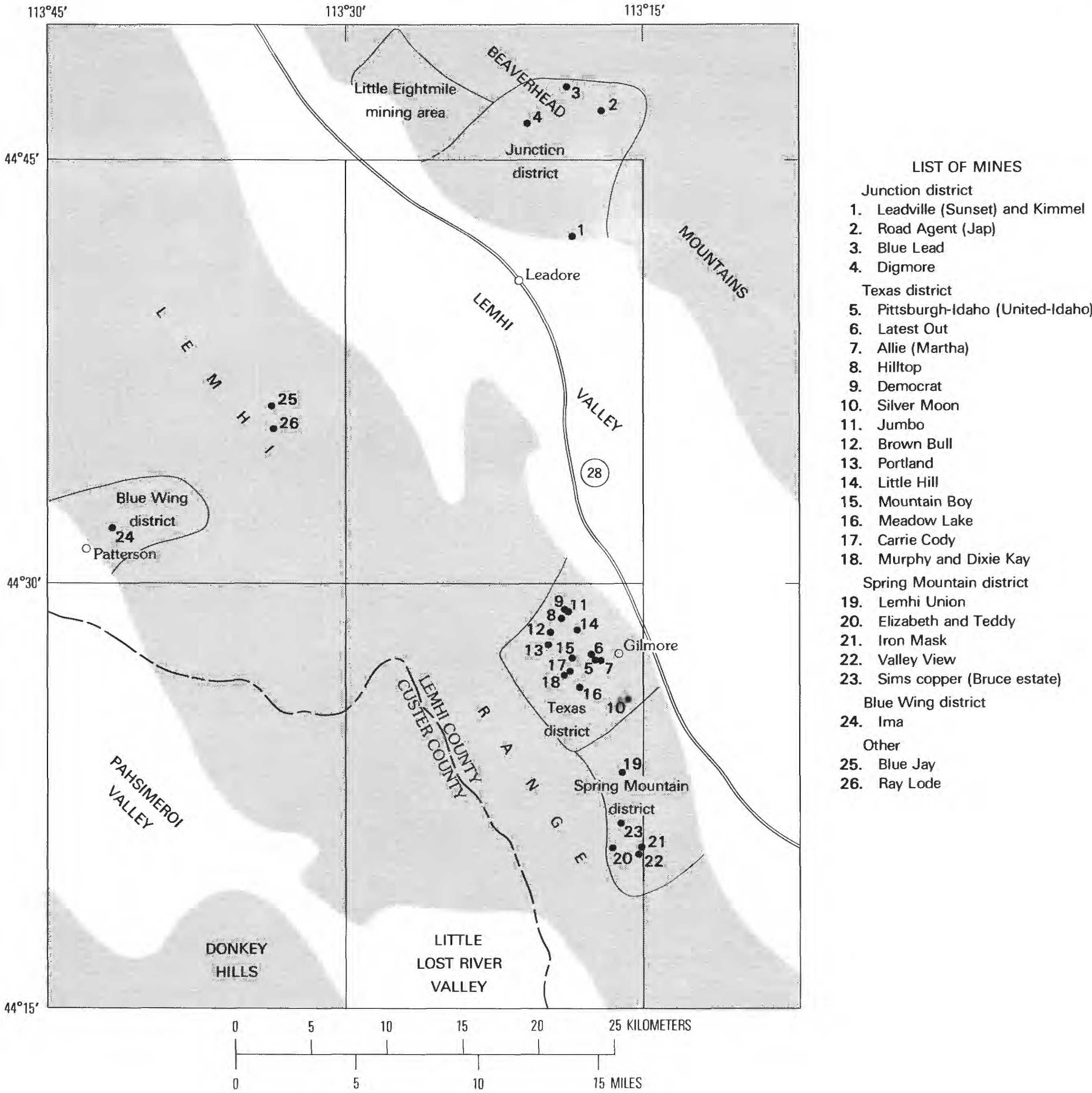


FIGURE 19.—Map showing locations of mining districts and of principal mines and prospects in the Junction, Texas, and Spring Mountain mining districts, Lemhi County, Idaho.

different from the lead-silver deposits of the Texas, Spring Mountain, and Junction districts, and each dominated by a single deposit. The larger of these is in the Blue Wing district, at Patterson, where tungsten-copper-silver quartz veins are explored in the Ima mine. The other is the Blue Jay mine, which explores a deposit of secondary and disseminated copper in the Big Eightmile stock, in the east-central part of the quadrangle.

The only nonmetallic materials mined or explored in the area are gravels, mainly of Kinnikinic Quartzite, that have been excavated from alluvial fans, crushed, and used locally for road metal, and bentonitic tuffaceous rocks that have been mined for use as ditch lining.

HISTORY OF MINING[2]

The discovery and mining of lead-silver ores on the upper Lemhi valley was fairly late in the history of exploration and settling in the valley, because early explorers and settlers were interested mainly in furs or agriculture, and because early explorers and prospectors looked for gold—not for lead at a few cents a pound, or for silver at less than a dollar an ounce. Lewis and Clark, the earliest white explorers, entered the valley through Lemhi Pass on August 12, 1805. They were followed in 1823 and 1824 by Hudson Bay Company trappers, Big Finan MacDonald and party in 1823, and Alexander Ross in 1824, who was accompanied by 140 persons including 25 women and 64 children. In June 1855, a group of Mormon pioneers settled and started farming at Fort Lemhi, near the present Tendoy, Idaho, only to be driven out by Indians in March 1858, after an attack on February 25, 1858, that preceded more widespread Indian uprisings. These uprisings nearly closed the region; only 6 or 8 white settlers were known to be in the region between Soda Springs and Coeur d' Alene, Idaho in 1860. The Mormon pioneers had prospected for copper in the north part of the Lemhi Range, and reportedly discovered some of the copper deposits of that region in 1856 (Ross, 1925, p. 2).

A few years later, in July 1862, gold was discovered in southwest Montana in Ruby Creek and Grasshopper Creek, and in May 1863, in Alder Gulch, followed, in 1866, by the discovery of gold in Napias Creek, which led to the founding of Leesburg, Idaho. Salmon City, the present Salmon, Idaho, was founded in 1867. The town of Junction, from which the Junction mining district draws its name, was established earlier, in 1864, at the junction of the wagon road in Lemhi Valley and an Indian trail in what is now Railroad Canyon, a major trail used by Montana settlers and prospectors. The search for placer gold led ultimately to a search for lode deposits as the placers were claimed up or exhausted. The earliest discoveries in the central Lemhi Range were in the Spring Mountain district, where Preston Sheldon and Jack Hicks located 27 claims in 1880, and where a small and unsuccessful smelter was constructed in 1882 (Oberg, 1970, p. 61). These discoveries followed an earlier report of "***several pieces of good galena from the Lemhi, but no lode mines" (Raymond, 1873).

In 1881, the Viola mine was located at Nicholia, in the Beaverhead Mountains southeast of Gilmore. The Viola was a rich deposit that yielded about 400,000 oz of silver and perhaps 18,500 tons of lead, valued at about $2 million, between 1882 and 1887. The Viola ore was almost entirely lead carbonate that contained about 20–60 percent lead and about 5–15 oz silver per ton (Shenon, 1928, p. 19–22). The orebody was about 300 m long and 0.3 to about 20 m thick. It extended about 60 m down dip, to a flat fault that cut off the orebody. No extension has been found, but comparison with similarly faulted orebodies in the Pittsburgh-Idaho, Latest Out, and Ima mines suggests that the Viola orebody was displaced westward on the flat fault, and that its extension beneath the fault is some distance to the east.

The Viola ores were shipped to smelters in Omaha and Kansas City until 1886, when a lead smelter was blown in at Nicholia, by then a town of nearly 1,000 people. By then, too, the community of Junction had a population of about 200. Prospectors from these briefly thriving communities located many of the mining claims in the surrounding region, in the search for another Viola. Probably most of the initial discoveries in the Texas and Spring Mountain districts were made at this time, and some mines shipped ore to the Nicholia smelter. R.N. Bell, the Idaho Inspector of Mines in 1904, reported that hundreds of promising claims were discovered in the Texas and Spring Mountain districts during the period of active mining at the Viola, from 1882 to 1887, and that about two dozen of these made shipments to the Nicholia smelter of from one to several loads of high-grade lead and silver ores. The Meadow Lake mine (fig. 19, loc. 16), for example, is rumored to have shipped $6,000–7,000 worth of ore to Nicholia at this time; some of the old mines near Sourdough Creek, north of Gilmore, also were among the initial discoveries. Shipments to the Nicholia smelter from the Texas district are reported as early as 1886 (Minerals Resources for the year 1886), and Umpleby (1913, p. 91) reported that about 700 tons of lead and more than 100,000 oz of silver were recovered from Texas district ores treated in the Nicholia smelter. The ore in the Viola mine was exhausted by 1887, and the smelter was closed in 1889. Without a nearby market, and under the influence of low metal prices, the majority of the early prospects and mines in the Lemhi Range apparently were abandoned, as was the town of Nicholia.

[2]The historical information in this section has been drawn from many sources. Chief among these have been the Annual Reports of the Idaho Inspector of Mines on the mining industry of Idaho for the years 1903 through 1921; Minerals Resources and Minerals Yearbooks for the period 1885–1914; and the discussions given by J.B. Umpleby in his U.S. Geological Survey report on the geology and ore deposits of Lemhi County in 1913, a period of major mining activity. The early history of the Gilmore district was summarized by Edgar C. Ross, who purchased the discovery claim from James Forrester in 1902, and started development of the rich lead-silver ores of the Texas district. Mr. Ross' personal narrative is used with permission of the Gilmore Mercantile Company, and was obtained through the courtesy of the late G. Grover Tucker, who arrived in Gilmore in 1911 and remained there as local manager of the Gilmore Mercantile Company until 1965, the last resident of the mining camp.

Mining in the Lemhi region entered a new phase in 1902, when Edgar C. Ross visited the region, sent west to restore his health by his doctor in Dubois, Pa. Ross, apparently acting as an agent for F.G. Lauer of Dubois and accompanied by C.T. Mixer, a mining engineer, purchased a lead-silver prospect, the original five claims that have yielded most of the ore in the Texas district, from James Forrester for $3,500. Under Mixer's supervision, the prospect was further explored, and in 1903 "17 carloads" (an uncertain amount, probably 200–250 tons) of ore were shipped that reportedly averaged 55 percent lead[3] and contained appreciable silver. The ore was hauled about 135 km to the railroad at Dubois, Idaho, by four wagons in train, pulled by 10–16 horses, with an average load of one ton per horse. This ore was discovered in an adit intended to intersect the discovery vein at a depth of about 60 m, but which, instead, cut a previously undiscovered and much richer vein, about 30 m from the adit portal, which yielded the shipping ore from a drift stoped 20–25 m to the surface. At this time, when the fledgling mine apparently was operated by the Gilmore Mining Company, Ross staked an additional 18 claims surrounding the initial 5 claims. The original five claims, the Texas, Never Sweat, Sixteen-To-One, Silver Dollar, and Silver Dollar Extension (Umpleby, 1913, pl. 15) were sold to the Pittsburgh-Idaho Company on January 1, 1906, for $80,000, and subsequently, probably about 1920, passed to the United-Idaho Mining Company. The principal mine in the district, on the initial five claims, was known as the Pittsburgh-Idaho through most of the active life of the district, but was renamed the United-Idaho mine when control passed to the United-Idaho Company. The old name, Pittsburgh-Idaho, is used in Umpleby's early report (1913), and is retained in this report to avoid confusion.

Twelve of the 18 additional claims were sold in 1912 to the Gilmore Mercantile Company, successor to the Gilmore Mining Company. The six remaining claims were owned by Ross' Allie Mining Company, incorporated in 1931 as the Delaware-Idaho Gold Mining Company. The "Martha fissure" vein, on the Andy (Allie Mining Company) and Martha (Gilmore Mercantile Company) claims, was discovered in 1910. It is the only gold vein discovered in the Texas district.

One other claim, the Latest Out, in the central part of the Texas district, explored major orebodies. This claim, which apparently was first staked in 1880 (Umpleby, 1913, p. 105), shipped 1,200–1,500 tons of ore to the Nicholia smelter. It was purchased by Ralph Nichols, a mining engineer and manager at the Viola mine, who reopened the mine and started development of its large orebodies in 1908.

The discovery of the rich orebody in the Texas district in 1902–1903 was accompanied by renewed prospecting elsewhere in the region. Dozens of prospects in the Spring Mountain district were actively explored, and aerial trams were constructed to carry hoped-for ores down the precipitous cirque walls of Horse Shoe Gulch (the Iron Mask and Valley View Mines) and the head of Spring Mountain Canyon. A lead smelter was started in 1908 by the Lemhi Smelting Company at Hahn, a small mining community at the mouth of Spring Mountain Canyon. It operated for only 10 days in 1909, smelting about 80 tons of ore, and for about 17 days in 1910, smelting perhaps 400 tons of ore, from the Lemhi Union, Red Warrior, Iron Mask, Valley View, Teddy, and Elizabeth mines in the Spring Mountain district; the smelter burned down in 1912. The Leadville (Sunset) mine in the Junction district was discovered in June 1904, and by 1907 had shipped about 260 tons of ore.

Tungsten minerals were identified in the Blue Wing district in 1901, by Professor J.E. Talmidge of Salt Lake City, but the veins had been prospected for silver and gold as early as 1881 (Umpleby, 1913, p. 109). The tungsten-bearing veins were prospected by 1911 in two adits, the Idaho Tungsten Company adit about 215 m long, and the Ima Mining Company adit about 275 m long. A concentrating mill was installed in 1912.

Although many promising prospects had been found in the 1880's, and these and others were actively explored in the flurry of prospecting after 1902, the absence of shipping and transportation facilities seriously impeded development until 1910. The Nicholia smelter had provided a local market only until 1889. Probably a very small amount of hand-concentrated ore was shipped after that—indeed, James Forrester was hand-panning lead-silver concentrates from the Texas district when Edgar C. Ross met him and bought his claims. Between 1902 and 1906, ores mainly from the Gilmore Mining Company claims in the Texas district, but including a small amount from the Junction and Spring Mountain districts, were hauled by wagon trains to the railroad at Dubois, Idaho. In 1906, the Dubois and Salmon Transportation Company started hauling ore in a train of four steel wagons and a sheep-wagon tender drawn by a steam tractor. The capacity of each wagon was 15 tons, but the average train load was about 40 tons. Four train loads were hauled in 1906, and about eight more in 1907—Umpleby (1913, p. 90) noted that only a dozen trips were completed before the tractor-drawn train was abandoned, a victim of rough roads and broken axles.

[3]This figure, and others elsewhere in this report that give ore grades for these early shipments, must be for ores that were nearly free of waste rock—the only kind likely to be shipped on so long a wagon haul. No concentrating mills are known to have been built this early, but the ore must have been carefully hand-picked, to remove as much waste as possible, and to make certain that only the highest grade ores went into the wagons.

In 1906, the Pittsburgh-Idaho Company purchased the principal mine in the Texas district, and over the next couple of years succeeded in promoting a railroad to carry Texas district ores to the Union Pacific-Oregon Short Line railhead at Armstead, Mont. This railroad, the Gilmore and Pittsburgh Railroad,[4] reached the Lemhi Valley in 1910, with tracks reaching Leadore, a new town and switching center near Junction, in January, and Gilmore in September (Myers, 1981, p. 165). With the promise of rail transportation, major development and mining began in 1909, after years of desultory development when only a little ore could be hauled by horse or steam-tractor to Dubois. The major mining period in the Texas district continued until about 1925, probably with declining ore shipments after 1919 or 1920, although no complete records are available. The Pittsburgh-Idaho mine was closed in 1929, after a machinery failure and fire in surface installations. The original orebodies were largely exhausted, and deeper workings—by then down to the 950-level[5] in the Pittsburgh-Idaho mine—required pumping and had not disclosed new orebodies sufficient to sustain mining at depression metal prices. Regular train service to Gilmore was discontinued in 1935, and was replaced by railbus service. The Gilmore and Pittsburgh Railroad stopped its service to the Lemhi Valley in 1939, and its tracks, fish plates, and even spikes were salvaged for scrap in 1940.

The rich lead-silver deposits at the Viola and Pittsburgh-Idaho mines overshadowed the smaller deposits in other mines in the central Lemhi Range; events in these two large mines strongly influenced—indeed, nearly controlled—the history of mining in this region. The mining-out of the Viola orebody in 1887 stopped mining and most prospecting in this region for 15 years until the new discovery at Gilmore in 1902–1903. The loss of the Viola destroyed the town of Nicholia so completely that almost no traces of it remain today (Oberg, 1970, p. 68–72). The new town of Gilmore, founded in 1903, grew up around the Pittsburgh-Idaho and Latest Out mines, and faded with the dwindling ore shipments between 1925 and 1935, when the last train out was filled with departing residents (G. Grover Tucker, oral commun., 1962). The Pittsburgh-Idaho orebody made rail transportation feasible, however, and after 1910 made possible the development of other, smaller mines like the Leadville mine in the Junction district, which between 1906 and 1919 yielded about 4,000 tons of lead-silver ore. When rail transportation stopped in 1935, truck transportation was possible and smelters were available nearby in Montana. The underground workings of the Gilmore mines had been tied together by a long drainage and haulage tunnel, the Transportation Tunnel, started in 1912. This tunnel tapped the Pittsburgh-Idaho mine on its 400-level, provided access to the Martha gold vein, and by 1917 was more than 1,800 m long, reaching nearly under the Latest Out workings. After the Pittsburgh-Idaho was closed in 1929, lessees continued to mine above the Transportation Tunnel, extracting by-passed ore in the Pittsburgh-Idaho until about 1940. The deeper levels, below the 500-level, in the Pittsburgh-Idaho mine were allowed to flood after 1929 and have never been reentered. Mining continued in the Latest Out mine until about 1952, mainly from new orebodies in the deeper levels of this mine but still above the Transportation Tunnel.

After 1935 two other mines began to produce major quantities of ore. The largest of these was the Ima mine, in the Blue Wing district (fig. 19, loc. 23) where tungsten-quartz veins had been recognized in 1901. The first efforts to recover the tungsten minerals were started in 1911 and continued in 1912 with construction of a mill (Callaghan and Lemmon, 1941, p. 4; Umpleby, 1913, p. 109). Small shipments of tungsten concentrates were made in some subsequent years, particularly during World War I, but the principal development began in 1936, after a period of preparatory work in 1934 and 1935, and continued until 1957 when price supports for tungsten concentrates were removed. The mine was a major source of tungsten in the United States during World War II. The mine was closed in 1958 but has been the focus of extensive geologic and geochemical prospecting since then, which has indicated more completely the widespread presence of molybdenum as well as tungsten (Ora H. Rostad, oral commun., 1970).

The second large mine, the Hilltop (fig. 19, loc. 8) developed a lead-silver-gold replacement orebody in the north part of the Texas district (Phillip N. Shockey, written commun., 1963). The main period of mining was from 1943 to 1961, but the surface exposures of the orebody were known earlier and probably were discovered in the 1880's. Adjacent and nearby mines, the Hughes, Grooms, McCellan, Jumbo, and Alex Stevens (now the Hilltop mine) all seem likely to have shipped small amounts of ore to the Nicholia smelter, and to have been among the earliest discoveries in the Texas district (Report of Idaho Inspector of Mines for year 1904). The Hilltop mine explored two major ore shoots, yielding about 9,000 tons of ore. The ore shoots were thought to have bottomed in the deepest level of the

[4]Also known locally as the Get Out and Push Railroad, for its torturous switch-back route up Railroad Canyon to Bannock Pass.

[5]Throughout this report, mine levels are given in feet, as they are in various earlier reports and on available mine maps. They are not converted to meters, on the grounds that to do so would unnecessarily complicate understanding of the earlier documents, and because a numbered mine level is as much a name as it is a measure.

mine, a 21-m winze below the 500-level, but a new orebody was discovered in 1979–1980 in the new Democrat mine, about 90 m beneath the Hilltop 500-level.

STRUCTURAL AND STRATIGRAPHIC CONTROLS ON MINERALIZATION

The emplacement of mineral deposits in and near the central Lemhi Range appears to have been closely controlled by preexisting faults—both by the ancient steep faults that bound the block-uplifted ranges and by the Medicine Lodge thrust system. All mineral deposits in and near the central Lemhi Range are in narrow mineralized belts along the mountain fronts or in single deposits, also at or near the mountain front, like those at the Ima and the Blue Jay mines. Most of them are closely associated with granitic intrusive rocks of Tertiary age that were the source of mineralizing solutions. All the stocks and sheets in the central Lemhi Range, and all the associated mineral deposits, either are demonstrably in the lower part of the Medicine Lodge thrust plate or can be reasonably inferred to be there. The distribution of granitic rocks and associated mineral deposits along the fronts of both the Lemhi Range and Beaverhead Mountains seems to be a result of primary structural controls exerted on the emplacement of granitic rocks by the steep northwest- and east-trending faults that are interpreted to bound these block-uplift ranges at depth (Ruppel, 1982). The steep faults, of Precambrian ancestry, were utilized as conduits by granitic magma that rose to the basal part of the Medicine Lodge thrust plate. The thrust plate had overridden the steep faults, but had not been broken by renewed movement on them at the time of granitic intrusion, so that the magma spread laterally in the sole zone of the Medicine Lodge thrust and in imbricate thrusts, just above the sole zone, to form sheets and sills. The best example is the thick sheet that extends southward from the Gilmore stock through the full length of the Spring Mountain district (Ruppel, 1978, p. 18–20; Ruppel and Lopez, 1981; 1984, p. 35–38).

Large mineral deposits, like those in the Texas district, are adjacent only to complex, major intrusive centers. Small mineral deposits, like those in the Spring Mountain district, are associated with sheets, which are fairly simple offshoots from stocks in the intrusive centers and probably were emplaced in one or a few intrusive pulses. All of these igneous rocks and associated mineral deposits are on the flanks of the Lemhi Range and Beaverhead Mountains. No substantial orebodies have been found and none are likely to have been deposited in the axial parts of these ranges, because the seemingly necessary deep structural controls are not present there (Ruppel, 1982, p. 21). In the few places in the Lemhi Range where granitic stocks are present away from the range front, emplacement of the intrusives is nonetheless controlled by major structural features. The Big Timber stock is at the intersection of northwest- and east-trending major fault zones, and the small stocks in Patterson and Falls Creeks, including the Ima stock, are in curving faults and fractures thought to be part of a large ring fracture complex (Ruppel, 1982, p. 9).

The lead-silver deposits in the Junction district are not associated with any exposed Tertiary granitic rocks, but their similarity to the lead-silver deposits in the Lemhi Range, and to those in the adjacent Little Eightmile mining area where Tertiary granite is exposed, suggests that a Tertiary intrusive is present at depth. All of these deposits, and others farther south in the Beaverhead Mountains, are in or near granite of early Paleozoic age, but most are replacement deposits in rocks that are younger than the granite of early Paleozoic age. These older granites are thrust-faulted and, in the Junction district at least, have been hydrothermally altered and mineralized along with adjacent sedimentary rocks, younger than the granite. The alteration clearly is later than thrusting; such alteration suggests even more strongly the presence of a still-buried Tertiary granitic intrusive. The close association of hydrothermally altered granite and sedimentary rocks with the ore-bearing veins of the Junction district strongly indicates that the hydrothermal alteration is related to that mineralization. In addition, the pervasively altered granite in the Kimmel mine includes disseminated galena, apparently introduced during alteration. For these reasons, alteration and mineralization in the Junction district are thought to be related to a buried granitic intrusive of Tertiary age, like the stocks in the central Lemhi Range. This conclusion applies as well to other lead-silver deposits farther south in the Beaverhead Mountains where no Tertiary granitic rocks are exposed, including those at the Viola Mine. For these reasons, the granite of early Paleozoic age cannot have been the source of any of the known mineral deposits in the Beaverhead Mountains.

In addition to the primary structural controls, the localization of mineral deposits has been influenced by the presence of limestone and dolomites that are especially favorable hosts for mineralization. In the Texas district, for example, all large orebodies are replacement deposits in the impure, partly sandy, thin-bedded dolomites in the lower part of the Jefferson Formation, as are many of the ore deposits in the Beaverhead Mountains (Shenon, 1928). But to conclude, from this, that some part of the Jefferson Formation is *the* most favorable host for mineral deposits is to make a leap of faith not quite justified, because other carbonate rocks in the region also contain mineral deposits. The Mountain Boy mine, and other small mines in both the Texas and Spring Mountain districts explore mineral deposits in dolomite of the Saturday Mountain Formation, which is compositionally similar to but cleaner than part of the Jefferson. The mines and prospects in the Junction district and Little Eightmile mining area explore deposits in Mississippian limestone. A few prospects in the Texas district, the Brown Bull mine, for example, explore veins in the Kinnikinic Quartzite, an almost pure siliceous quartzite. In the Texas district, the Saturday Mountain Formation beneath the deepest workings on the 950-level of the Pittsburgh-Idaho mine has never been explored. It contains mineral deposits where it is exposed in the outer parts of the Texas district and in the Spring Mountain district, and probably should be considered to be as likely a host for mineral deposits as the Jefferson. In short, given the apparently necessary primary structural controls, and the presence of Tertiary granitic rocks, most of the carbonate rocks of the region can be suitable host rocks for lead-silver-gold replacement mineral deposits.

In general, the quartzitic rocks are not particularly favorable hosts for mineral deposits. The small silver-lead deposit at the Brown Bull mine, which is partly in the Kinnikinic Quartzite, is clearly an anomaly in this district of replacement deposits in carbonate rocks. The tungsten-copper-silver quartz veins at the Ima mine occupy fractures in Proterozoic feldspathic quartzite and in the granite of the parent Ima stock; however, they represent a different kind of mineralization than the replacement lead-silver orebodies elsewhere in the central Lemhi Range. A few small quartz veins have been found in quartzitic rocks south of the Big Timber stock, and have been prospected for their gold and silver content. One of these, the Ray Lode mine, reportedly yielded a small amount of gold ore.

In summary, the primary and critical controls on mineralization have been structural, the steeply dipping faults that were intruded by granitic magma, and the nearly flat, basal part of the Medicine Lodge thrust system, which formed the roof of the granitic intrusives and diverted the magma into flat sheets in zones of imbricate thrusts. Where carbonate rocks were present in the lower part of the thrust plate, near the major stocks, base- and precious-metal sulfides accumulated in replacement deposits. Where quartzitic rocks were present in the roof or walls of the stocks, only thin quartz veins with a small precious-metal content were deposited. Adjacent to the Ima stock which is composed of a granite different in composition from all other stocks in the Lemhi Range, tungsten-copper-silver quartz veins have yielded a large quantity of ore.

AN ALTERNATIVE HYPOTHESIS: STRATABOUND LEAD, SILVER, AND ZINC IN THE JEFFERSON FORMATION

Skipp and others (1983) concluded that the lead-silver-zinc mineral deposits in the Italian Peaks area, and elsewhere in the south part of the Beaverhead Mountains, are partly remobilized, but essentially stratabound deposits in the Jefferson Formation, similar to stratabound deposits of the Mississippi Valley type. In contrast, earlier interpretations of the mineral deposits in the Beaverhead Mountains suggested that they were hydrothermal replacement veins and irregular replacements along bedding surfaces (Shenon, 1928, p. 11–24).

The hypothesis that the ores are stratabound in the Jefferson Formation seems to explain why so many mineral deposits are present in this formation in both the Beaverhead Mountains and in the Lemhi Range, and suggests that similar mineral deposits in other formations are made up of metallic sulfide minerals derived from the Jefferson Formation and redeposited in other carbonate rocks. Briefly stated, Skipp and others (1983, p. 7) concluded that the metallic elements in the mineral deposits of the Italian Peaks area were derived from an original source much farther west, and were carried in solution eastward through the Kinnikinic Quartzite, which is thus interpreted to have remained an unsilicified and permeable sandstone long after its deposition in Middle Ordovician time. The metallic elements were deposited where the transporting solutions encountered the carbonate rocks of the Jefferson Formation, sealed beneath the overlying shales of the Devonian Three Forks Formation and the Lower Mississippian McGowan Creek Formation. The chief evidence cited in support of the hypothesis, other than the concentration of known mineral deposits in the Jefferson, is drawn from geochemical studies of the Italian Peaks area, which indicate widespread anomalous amounts of metallic elements in the sedimentary rocks of the area, and particularly in the Jefferson Formation. The time of metallization is not closely fixed, but it is post-Mississippian, because it followed deposition of the Lower Mississippian McGowan Creek Formation, and some Mississippian carbonate rocks above the McGowan Creek are weakly mineralized, and pre-late Early Cretaceous, because the presumed stratabound deposits are cut by thrust faults of that age.

The hypothesis that the ores of this region are stratabound seems to answer only one question—why so many of the known mineral deposits are in host rocks of the Jefferson Formation. That question can also be answered, however, by concluding that these rocks are particularly favorable hosts, but not the only hosts for hydrothermal replacement deposits, as indeed they are in the central Lemhi Range. In addition, the widespread weakly anomalous metal content in the Jefferson, cited as evidence for stratabound metallization, is to be expected in the outer, mineralized aureole of a hydrothermal system, as the presence of barium suggests the Italian Peaks deposits could be. The suggestion that mineralizing solutions traveled through the Kinnikinic Quartzite seems unlikely, because the Kinnikinic is now essentially impermeable and probably has been since it was buried by the later Ordovician and Silurian dolomites of the Saturday Mountain and Laketown Formations. Pebbles of siliceous quartzite derived from the Kinnikinic are present in places in the basal part of the overlying Saturday Mountain Formation (James and Oaks, 1977, p. 1498). If mineralizing solutions had traveled through the Kinnikinic, and reacted with the Jefferson carbonate rocks where they overlap the eastern erosional edge of the Kinnikinic Quartzite, it seems likely that the solutions would also have reacted with ankeritic sandstone that is common in small lenses in the Kinnikinic, as well as with dolomite in the Saturday Mountain Formation that overlies the quartzite everywhere west of the Beaverhead Mountains. There is no evidence to suggest that the Saturday Mountain Formation contains anomalous amounts of metals.

The stratabound hypothesis is based principally on data from a very small area, the Italian Peaks in the south Beaverhead Mountains, and does not consider regional relations of similar mineral deposits, which mostly are reasonably considered to be of hydrothermal origin. As discussed in this report, the mineral deposits in and near the central part of the Lemhi Range occur in Precambrian (the Blue Wing or Ima district), Ordovician (the Spring Mountain district), Devonian (the Texas district), and Mississippian rocks (the Junction and Little Eightmile Districts). The largest orebody in the south part of the Lemhi Range, at the Wilbert mine, is in quartzitic dolomite that is Late(?) Proterozoic and Cambrian in age (Anderson, R.A., 1948; Ross, 1933; 1961a, p. 250). In southwest Montana, only the Argenta district is particularly notable for mineral deposits in Devonian rocks (Shenon, 1931); major base- and precious-metal deposits are widespread in Cambrian (the Hecla district, for example) and Mississippian carbonate rocks (the Bannack district, for example) (Geach, 1972), even where Devonian rocks are also present.

In summary, the stratabound hypothesis does not seem to be unequivocally supported by the available evidence, which more strongly suggests a hydrothermal origin for the base- and precious-metal mineral deposits in the Beaverhead Mountains (Shenon, 1928) and Lemhi Range. The carbonate rocks of the Jefferson Formation clearly are favorable hosts for mineralization, but so are most other Paleozoic carbonate rocks, given the appropriate structural and magmatic-hydrothermal setting. This is not to suggest that stratabound mineral deposits could not exist in the Jefferson, because its depositional setting certainly permits such deposits, as Gilmour (1972) points out. The evidence cited by Skipp and others (1983), however, can be used just as well to support a hydrothermal interpretation of the mineral deposits in the Beaverhead Mountains. The hydrothermal interpretation seems to be the better interpretation, when the mineral deposits of the Beaverhead Mountains are compared to similar deposits that are known elsewhere in the region. The source of metallic elements remains unknown, but sources more likely than the Jefferson are the Yellowjacket Formation, which is known to contain stratabound mineral resources (Hahn and Hughes, 1984), and older Precambrian crystalline rocks at great depth.

DESCRIPTIONS OF MINERAL DEPOSITS

The metallic minerals found in and near the central Lemhi Range are in a number of different kinds of deposits. Those of major economic interest have been base- and precious-metal replacement deposits and tungsten-copper-silver-quartz veins. Other deposits, which so far have yielded little if any ore, include a few copper-silver bearing quartz veins, disseminated deposits of molybdenum and copper in granitic rocks, small deposits of secondary copper minerals, a deposit of lead minerals disseminated in granite, and a deposit of copper-bearing magnetite. Many of these deposits have been described in some detail in earlier reports. The lead-silver deposits and other deposits of the Texas, Spring Mountain, and Junction districts are described by Umpleby (1913, p. 63–69, 85–109, 112–118), whose report is still a useful and necessary reference, because it was written when the major mines were open and the relations of mineral deposits could be observed underground. The tungsten-quartz veins of the Ima mine in the Blue Wing district are described by Umpleby (1913, p. 73–74, 109–112), Callaghan and Lemmon (1941), Hobbs (1945), and A.L. Anderson (1948). Umpleby (1913, p. 66–69, 74–82) also discussed the mineralogy of ores in Lemhi County and oxidation and secondary enrichment of the mineral deposits.

BASE- AND PRECIOUS-METAL REPLACEMENT VEINS AND BEDDING REPLACEMENT DEPOSITS

Most of the mineral deposits in the Texas, Spring Mountain, and Junction districts are lead, zinc, and silver deposits in replacement veins and in replacements adjacent to the veins. A few of these deposits contain an appreciable amount of gold, and one, the Martha vein in the Texas district, contained only gold. The deposits all are in dolomite or limestone host rocks, in the lower half of the Jefferson Formation in most of the Texas district, in the Saturday Mountain and Jefferson Formations in the Spring Mountain district, and in Mississippian limestone in the Junction district. The orebodies of this group in the Texas district have yielded more lead-silver ore than all other mines in the Lemhi Valley combined, almost all of it from the Pittsburgh-Idaho and Latest Out mines. The lead-silver deposits are much the same throughout the three mining districts, however; except for size, they differ mainly in depth and extent of oxidation and in the proportions of primary sulfide minerals and secondary carbonate minerals. In addition, gold is a prominent constituent of the lead-silver veins in the Democrat and Hilltop mines, but not in other lead-silver orebodies, either elsewhere in the Texas district, or in the Spring Mountain or Junction districts.

The base- and precious-metal replacement veins occur in fractures and small faults that break the enclosing carbonate rocks. Their walls commonly are well defined, but in places the veins bulge outward along intersecting joints and fractures, or as replacement deposits in adjacent beds or along bedding surfaces. The orebodies are lenticular and range in thickness from a few centimeters or less at their disappearing ends to 1–8 m, most commonly about 4–5 m, in the exceptionally thick orebodies in the Pittsburgh-Idaho and Latest Out mines. At the Leadville mine, in the Junction district, the two orebodies mined were 2 ft and 4 ft (0.6 and 1.2 m) thick (Umpleby, 1913, p. 118), and most of the veins mined in the Spring Mountain district probably were thinner, perhaps 0.2–0.5 m. The orebodies have ranged from small pods to large shoots, for example, in the Pittsburgh-Idaho mine, where some were stoped for more than 150 m along strike (the more common strike length was 30–95 m), and extended upward from the 700-level to the surface. The most productive mines developed several orebodies in groups of closely spaced, partly intersecting veins, most commonly of similar trend but different dips. In the Pittsburgh-Idaho mine, most of the veins trend about N. 10°–15° E.; these are separated into a set of steep veins that dip about 70°–90° W. and are the most prominent and productive veins, and another set of flatter veins, called flat veins in contemporary reports, which dip 40°–60° W. Both sets of veins contained ore deposits, but the largest orebodies tended to be at and above intersections of flat and steep veins, suggesting that the flat veins fed the steep veins. Umpleby (1913, p. 93–95) noted also that some ore in the Pittsburgh-Idaho mine occurred at the intersection of the major veins and steeply dipping east-trending fractures, in rocks brecciated by earlier movement on the east-trending fractures and susceptible to replacement. Umpleby pointed out that the east-trending fractures generally do not contain ore minerals elsewhere and apparently did not carry ore-forming solutions. Similar sets of steep and flat veins are present in the Latest Out mine, where at least three fairly distinct orebodies were developed along them.

The Hilltop mine included several orebodies, apparently on a single, nearly vertical, northwest-trending vein, but other, parallel veins, some of which have yielded lead-silver ore, were explored in adjacent mines and prospects. No flat veins, like those in the Pittsburgh-Idaho and Latest Out mines, have been recognized in the Hilltop mine or in the newer Democrat workings at greater depth beneath the Hilltop, but the similar geologic settings of the various mines suggest that flat veins may be present in all.

Most of the other mines and prospects in the Texas, Spring Mountain, and Junction districts explore single, small veins; only a few of them have yielded more than two or three hundred tons of ore. The Leadville mine in the Junction district, probably the largest of these, developed two lead-silver ore shoots apparently on a single, long and well-defined east-trending vein.

Most of the lead-silver ores mined have been secondary carbonate ores consisting of cerussite, the principal ore mineral, and lesser amounts of anglesite, smithsonite, and hemimorphite, in a gangue of earthy hematite, limonite, and manganese oxides. Umpleby (1913, p. 76) suggested that cerargyrite is widely present as very small fragments in the oxidized ores. Galena was relatively rare in the deeply oxidized ores above the 700-level in the Pittsburgh-Idaho mine and in the Latest Out mine, occurring only as residual kernels rimmed with anglesite, but is much more abundant in other mines both in the north part of the Texas district and in the Spring Mountain and Junction districts. Ore minerals commonly constitute half or more of the vein filling.

The average grade of lead-zinc-silver ores mined in the Texas district appears to have been about 23–25 percent lead, 4–5 percent zinc, 10–12 oz silver per ton, and little more than a trace of gold. The ratio of about 1 oz of silver to 2 percent lead is fairly constant in all lead-silver ores of the region. Umpleby (1913, p. 103, 106–107) noted that some of the richer, secondary ores above the 500-level in the Pittsburgh-Idaho mine averaged about 37 percent lead and 15–16 oz silver per ton, about the same as ores from the upper levels in the Latest Out mine, which contained about 34 percent lead, 5 percent zinc, and 18 oz silver per ton. Ores mined from the 600-level at the Latest Out mine, between 1948 and 1952, averaged about 25-27 percent lead, 3-4 percent zinc, 13 oz silver per ton, and 0.04 oz gold per ton (Ralph Nichols, written commun., 1968). The deepest veins encountered in the Pittsburgh-Idaho mine, on the 950-level, were apparently entirely primary sulfides, galena, probably pyrite, sphalerite, and a small amount of chalcopyrite, probably in a siderite gangue; they reportedly assayed 12.75 percent lead, 4.75 percent zinc, and about 10 oz silver per ton across vein widths of about 0.4 m. The relation of these two veins to the richer, oxidized veins mined in the higher levels of the Pittsburgh-Idaho mine is uncertain, because they are separated from the higher veins by a flat, east-dipping fault, the Gilmore fault. The two lower veins probably are different veins rather than fault-separated extensions of the higher veins.

The orebodies in the Hilltop and Democrat mines, in the north part of the Texas district, differ from those in the central part of the district in that they are more siliceous, they include more abundant primary sulfide minerals in addition to secondary carbonate ore minerals, and they contain more gold. The average grade of ores mined from the Hilltop vein is about 15 percent lead, 0.9 percent zinc, 0.3 percent copper, 13-14 oz silver per ton, and 0.2 oz gold per ton, although a shipment of about 30 tons of somewhat richer ore, made in 1961, averaged about 23 percent lead, 13 percent zinc, 0.6 percent copper, 23 oz silver per ton, and 1.06 oz gold per ton (P.N. Shockey, written commun., 1963). Crude ore shipped from the nearby Jumbo mine yielded somewhat similar values, about 18 percent lead, 0.4 percent copper, 23 oz silver per ton, but less than 0.1 oz gold per ton. Ore from the recently opened orebodies in the Democrat mine, about a hundred meters beneath the Hilltop orebodies, averages about 23 percent lead, 15 oz silver per ton, and 0.14 oz gold per ton.

The two ore shoots mined in the Leadville mine in the Junction district, were largely primary sulfide minerals, galena, pyrite, and very rare sphalerite and chalcopyrite, with only sparse cerussite and anglesite, and with few gangue minerals (Umpleby, 1913, p. 65, 118). Smelter analyses of the richest, probably hand-sorted, ore from the upper levels in the mine indicate that it contained 54.5-56.5 percent lead, 0-1 percent zinc, 0.012 percent copper, 29-35 oz silver per ton, and a trace of gold. Some of these ores contained an appreciable amount of antimony, arsenic, and bismuth, which are not common elsewhere in the region, except that antimony is common in stibnite in the Little Eightmile mining area, north of the Junction district.

Gold is clearly an unusual element in the lead-silver ores except those of the Hilltop and Democrat mines, but it is the only metallic element of economic interest in a single vein in the central part of the Texas district, the Martha vein, in the Allie mine east of the Pittsburgh-Idaho mine. Umpleby (1913, p. 107) described the Martha vein as one that occupies a N. 10° E.-trending, 20° W.-dipping fractured zone, probably a fault. Later mining on the vein, after Umpleby's study in 1910-1911, showed that its dip was steeper at depth, about 65° W. The deeply oxidized vein was an earthy mass of iron and manganese oxides that contained well-defined casts of small pyrite crystals. It contained about 0.5-0.8 oz of gold per ton, about 0.3-0.4 oz of silver per ton, some copper in rounded chalcopyrite masses surrounded by copper oxide and carbonates, and no lead or zinc except for an isolated single, apparently small, kidney of galena. The Martha vein yielded about 30,000 tons of ore with a probable average grade of about 0.5 oz of gold per ton; it was much like the lead-silver veins of the Texas district, except for gold content. It also differs from the lead-silver veins in one other aspect—the earthy limonitic and manganiferous gangue of the Martha vein contained well-defined casts of pyrite crystals and rounded masses of chalcopyrite, and was siliceous and jasperoidal in places. Pyrite casts are not known to be present in the lead-silver veins; in any event, pyrite associated with primary lead-silver ores apparently is not auriferous. Chalcopyrite is very sparse in the primary lead-silver ores, and most commonly is represented in the oxidized ores only by relatively rare malachite and azurite; the rounded masses of chalcopyrite are unique to the Martha vein. The association of gold, pyrite, and chalcopyrite in the Martha vein suggests that gold-bearing sulfide minerals may have been deposited in some lead-silver veins late in the period of mineralization, or, perhaps more likely, in a later, separate period of silicic mineralization related to Challis volcanism and after the lead-silver veins had been partly oxidized.

The gold-bearing lead-silver orebodies in the Hilltop and Democrat mines are similar to the Martha vein, although they are not as completely oxidized. They contain about the same amount of gold, about 0.1-0.5 oz per ton, and more copper than is common in other lead-silver ores. Like the Martha vein, they also are more siliceous than most of the other lead-silver ores of the Texas district. The similarities suggest that the gold in the Hilltop and Democrat orebodies also was deposited, with pyrite and chalcopyrite, in a second period of mineralization, which followed lead-silver mineralization and some oxidation of the earlier vein filling.

Although Umpleby (1913, p. 66) described barite as an insignificant gangue mineral present only in the Spring Mountain district, it is actually much more widespread. In addition to its occurrence at the Teddy and Elizabeth mine, noted by Umpleby, it is a prominent gangue mineral in most of the small mines and prospects in the south part of the Texas district, in Silver Moon Gulch, and farther north in the small mines in the upper part of Meadow Lake Creek, on Portland Mountain, and in most of the prospects west of the Hilltop mine. These small mines and prospects are at the edges of the Spring Mountain and Texas districts, and the small, barite-lead-silver veins that they explore are in barite zones that mark the outer limits of mineralization around the Gilmore stock and the Spring Mountain sheet.

TUNGSTEN-QUARTZ VEINS AND COPPER-SILVER-QUARTZ VEINS[6]

Tungsten-quartz veins and related copper-silver-quartz veins in the central Lemhi Range are known only in the Blue Wing district, at Patterson (Ruppel, 1980). The tungsten-quartz veins are extensively developed in only one mine, the Ima mine (fig. 19, loc. 24). Smaller tungsten deposits have been explored in the General Electric (Miller) prospects north of the Ima, but they have not yielded much ore except from extensions of the main Ima veins deep beneath the General Electric claims. Small deposits of copper-silver minerals have been prospected and mined in a few places in the eastern part of the Blue Wing district, mainly from tetrahedrite-bearing quartz veins in the outer part of the mineralized zone. Most of the tungsten deposits are in Middle Proterozoic feldspathic quartzite and siltite of the Lemhi Group, the Ima deposits in the Gunsight and Apple Creek Formations, and the General Electric deposits above the Ima thrust in the Big Creek Formation. They extend downward into the outer shell of the Ima granite stock which is not exposed at the surface; the granite clearly was the source of the mineralizing solutions (Callaghan and Lemmon, 1941, p. 13). The stock is interpreted to have been intruded into the south edge of a ring-fracture zone (Ruppel, 1982, p. 9, 14), about 48–50 m.y. ago.

The sedimentary rocks are cut by multiple imbricate thrust faults that were first recognized underground as nearly flat fractures and breccia zones (Callaghan and Lemmon, 1941, p. 9). Hobbs (1945, p. 6–7) later suggested that one of the most prominent flat breccia zones, a major one called the Ima thrust fault in this report[7], is a premineralization thrust fault that had influenced the distribution of other premineralization fractures and the extent of mineralization itself. Hobbs noted that the Ima thrust separates distinctly different rocks which now are known to be feldspathic quartzite of the Gunsight Formation in the lower plate, and lighter colored, feldspathic quartzite of the older Big Creek Formation in the upper plate.

The fracture patterns above and below the Ima thrust are distinctly different; Hobbs suggested that the different quartzites had acted as independent structural units, each forming more or less independent fracture systems. The pattern of premineral fractures above the Ima thrust is not well exposed, but appears to be far less complex than that below the thrust. Later vein fillings in fractures in the rocks above the thrust are thin and discontinuous, but those in the rocks below the thrust are thicker, extend for hundreds of meters along strike, and contain the principal ore deposits in the Ima mine. Postmineral faults, mostly steep normal faults of small displacement, have irregular trends and no common orientation above the Ima thrust; below the thrust, they are much more abundant and are in two prominent sets in the Gunsight and Apple Creek rocks. One set strikes about northwest, parallel to the veins, and dips steeply east, opposite to the dip of the veins; a second set strikes about northeast and dips steeply either northwest or southeast. A third set of faults with both premineral and postmineral movement also cuts these lower rocks. These are thrust faults that strike about N. 75° E., and dip 5°–30° N., and bound the slice of Apple Creek siltites and fine-grained quartzites that is enclosed in Gunsight quartzites on the canyon flanks at the Ima mine (Ruppel, 1980). The postmineral movement on these flat faults clearly is to the west, but is minor and probably reflects late, gravitational sliding toward the Pahsimeroi Valley.

Fracturing in the rocks becomes less intense with increasing distance above the Ima stock. Because no granitic dikes intrude the fractures, but pegmatitic dikes and later veins do, Callaghan and Lemmon (1941, p. 10) suggested that most of the fractures were formed in the late stages of granite emplacement.

[6]This discussion of the tungsten-quartz veins in the Blue Wing district is drawn largely from a U.S. Geological Survey Open-File report by Hobbs (1945), on the geology of the Ima and General Electric tungsten deposits, and includes much information that is repeated directly from the Hobbs report. It also includes other information from an earlier report, by Callaghan and Lemmon (1941), and from a U.S. Geological Survey unpublished report by A.L. Anderson on the geology and mineralogy of the tungsten deposits, part of which is included in a later published report by A.L. Anderson (1948). The Hobbs report is a complete and thoughtful study of the tungsten deposits, and its conclusions have been widely and successfully used in guiding later exploration and development of the tungsten deposits. The earlier reports are useful supplements. We mapped the surface geology (Ruppel, 1980), which had not been done earlier, and we fit the mineral deposits into a different stratigraphic and structural framework than was known before. The mines of the Blue Wing district were inaccessible when we were there, but Hobbs described the tungsten deposits so completely that it would have been pointless to repeat his study even if the mines had been open.

[7]The Ima thrust fault is not named on the geologic map of the Patterson region (Ruppel, 1980), and the name is applied here informally, to simplify discussion, and to distinguish it from a lower thrust fault that is called the Water fault in the Ima mine. Many imbricate thrust faults cut and shuffle the Proterozoic rocks at the Ima mine on the north side of Patterson Creek. The Ima thrust extends diagonally across the center of unsurveyed sec. 14, T. 14 N., R. 23 E., between the Big Creek and Gunsight Formations. The Water fault probably is the thrust fault immediately beneath the Ima thrust.

Fracturing and faulting are discontinuous and less complex in the Big Creek Formation above the Ima thrust, partly because these rocks are far above the Ima stock, and perhaps partly because the quartzites of the Big Creek Formation commonly are more massive than the rocks of the Gunsight and Apple Creek Formations below the thrust. But the abrupt change in fracture density and persistence, and in fault patterns, across the Ima thrust, seems more likely to be a result of postmineralization, westward gravitational movement on the premineralization thrust fault, rather than a result of different mechanical responses across the thrust fault as suggested by Hobbs (1945, p. 6–7), so that fracture systems originally somewhat more widely separated have been brought together.

The Ima thrust fault clearly influenced the extent of mineralization (Hobbs, 1945, p. 7–8), but if there has been westward gravitational sliding of the rocks above the thrust, the veins would have been displaced, too. Although this possibility has not been explored on the Ima thrust, the main Ima orebodies clearly are displaced to the west above the Water fault, a flat fault that strikes N. 70°–80° E. and dips 20°–40° N., about 180–215 m below the parallel Ima thrust. Below the Water fault, the principal veins developed in the Ima mine persisted for lengths of more than 150 m, and through a similar vertical range, bottoming in the outer shell of the Ima stock. They are broken by several sets of steeply dipping normal faults of small displacement (Hobbs, 1945, p. 8). These veins were strongly mineralized below the Water fault, but could not be correlated with other, smaller veins above the fault. Rather, the main zone of mineralization appears to have been displaced westward perhaps as much as 1,000 m, where a zone of major mineralization about 250 m wide was found above the Water fault. The Water fault, like the Ima thrust fault, is clearly a premineralization thrust fault; the larger and richer ore deposits in veins beneath it suggest that it localized mineralization. The fault zone contains crushed and fractured vein material that indicates renewed movement after mineralization, but the movement was westward, the opposite of the original thrusting to the east, and probably was gravitational sliding during block uplift (Ruppel, 1982). As a result, the main ore zone was displaced westward above the Water fault. The similar relations of veins above and below the Ima thrust suggest that the main ore zone, if it persists this far upward, is also displaced westward above the Ima thrust.

The earlier reports on the geology of the Blue Wing district describe the tungsten deposits and the Ima stock, as being in the axial part of a northwest-trending anticline. More complete geologic mapping suggests, however, that the fold is a postmineralization monoclinal fold along the edge of the Lemhi Range (Ruppel, 1980, section *E-E'*; Ruppel, 1982) and that the eastward-dipping rocks on the east flank of the supposed anticline were folded earlier, and partly overturned, during thrust-faulting. The gravitational sliding of rocks and mineral deposits above the Ima and Water thrusts, and above other flat faults in the Ima mine, is a result of instability induced by steepening of the thrust surfaces during monoclinal folding—they now dip 20°–40° W., instead of 10°–15° W. as is typical of unfolded imbricate thrust faults farther east in the central part of the Lemhi Range.

The tungsten-quartz veins in the Blue Wing district generally strike about northwest and dip steeply southwest, but changes in attitude are common, and in places abrupt; strikes range from almost east-west to almost north-south, and dips range from vertical or steeply west or east, to gently west. The veins typically pinch and swell along both the strike and dip and may be as much as 6 m thick and 200 m or more long, and persist through a vertical range of as much as 150 m. In general, the veins and orebodies grouped close to the borders of the Ima stock, perhaps within 70–80 m of the contact, and near the Water fault are relatively large, and those more distant from the granite or from the Water fault are smaller and less continuous, but more numerous. Small offshoots from the veins are common in fractures in wall rocks, but die out within 5–10 m; the veins branch and split along premineral fractures as well. The veins die out with increasing depth in the Ima stock, and most commonly do not persist for more than 100 m or so in the granitic rocks, although thick veins may be present in the outer shell of the stock at or near its contact with the enclosing quartzite.

The principal minerals in the tungsten-quartz veins are pyrite, chalcopyrite, molybdenite, huebnerite, scheelite, tetrahedrite, galena, and sphalerite. Gangue minerals, which make up 90–95 percent of the veins, are abundant quartz, and less abundant orthoclase, fluorite, rhodochrosite, muscovite, sericite, siderite, and calcite. Huebnerite, tetrahedrite, and locally molybdenite, fluorite, and rhodochrosite are comparatively abundant, but the other minerals, except for the predominant quartz, are present in differing but generally sparse amounts. The distribution of minerals is zoned from the granite contact outward; muscovite, orthoclase, rhodochrosite, and molybdenite are most abundant near the contact. Fluorite is also most abundant near the contact, but is present somewhat farther away from the contact than the other contact-associated minerals, decreasing in crystal size and abundance with increasing distance from the contact. Huebnerite and most of the other ore minerals are distributed more or less throughout the full extent of the mineralized zone, which extends at least 600–650 m from the Ima stock. Secondary minerals are sparse in the tungsten-quartz veins, but include minor amounts of anglesite, azurite and malachite, cerargyrite, chalcocite and covellite, erythrite, limonite, manganese oxide, molybdite, pyromorphite, and tungstite.

Huebnerite, the most widely distributed and most important ore mineral, is the source of most of the tungsten in the ore. The ores from the major veins and shoots, mainly those 0.6 m or more thick, have contained about 0.6–0.7 percent WO_3. The huebnerite occurs as well-defined, bladed, tabular crystals commonly 1 cm or less long, although in places they are as much as 7–8 cm long. Most huebnerite crystals are more or less widely scattered through the veins, embedded in quartz, and in and across grains and small masses of sulfide minerals. Scheelite, a very minor tungsten ore mineral in the Ima deposit, is in small specks scattered throughout the veins and, most commonly, in thin crusts lining open spaces and in thin shells and films coating huebnerite crystals.

Tetrahedrite, one of the most abundant and widely distributed sulfide minerals in the tungsten-quartz veins, contains most of the silver and copper in both tungsten and copper-silver ores. Silver, mainly or entirely from tetrahedrite, averages about 1.75 oz per ton. Tetrahedrite and less abundant pyrite, sphalerite, chalcopyrite, and galena together constitute only 4–5 percent of the ore, but mill concentrates typically yield about 42–60 oz silver per ton of concentrate, 4–6 percent copper, 4–7 percent lead, and 2–5 percent zinc (Callaghan and Lemmon, 1941, p. 6; Hobbs, 1945, p. 11). The tetrahedrite and associated sulfide minerals are widely but not uniformly distributed through the quartz veins in small rounded grains and small irregular masses. The tetrahedrite partly replaces both pyrite and sphalerite, and in turn, is cut by veins of chalcopyrite and galena.

Molybdenite also is relatively abundant, but is more restricted in its distribution around the Ima stock than other sulfide minerals. It is found mainly in and near the border of the Ima stock, decreasing in amount and grain size with increasing distance from the stock margin. It forms 0.1–0.4 percent of the ore, most commonly 0.1–0.3 percent. Molybdenite occurs disseminated in the granite and as a thin coating or narrow fracture filling in the quartz veins and in adjacent fractured granite and quartzite wall rocks, where it is smeared in thin films over slickensided walls or around small fragments of vein and wall rock material. It also occurs less commonly in disseminated grains and in pods within the veins and in and near the granite, where in places it is in grains about a centimeter in diameter. The molybdenite disseminated in granite partly is in zones as much as 10–12 m wide, in which it forms 0.3–0.4 percent of the granite, separated by zones in which it is less abundant. It commonly is associated with minute pyrite crystals and microscopic grains of chalcopyrite. Molybdenite in and near the granite is accompanied by muscovite and fluorite.

Quartz is the most abundant and widely distributed gangue mineral, forming 90–95 percent of the veins. Rhodochrosite and fluorite are most abundant near the granite contact, decreasing in abundance and grain size with increasing distance from the contact. Both rhodochrosite and fluorite commonly occur in small grains or granular aggregates embedded in quartz; some rhodochrosite also occurs in small seams and veinlets. Quartz, fluorite, and rhodochrosite all appear to have been deposited in several generations. Muscovite is found only with molybdenite in and near the granite, commonly in granular aggregates of crystals or in seams and veins 1–2 cm thick, although some muscovite crystals are as much as 2–3 cm in diameter. Away from the granite, muscovite seams and aggregates become smaller and less abundant, grading into thin seams of finely crystalline muscovite or coarse sericite. Sericite is most abundant, however, in altered, quartzitic wall rocks; it is widespread, but not particularly abundant, in the gangue of quartz veins. Siderite is among the least abundant gangue minerals, occurring as small, platy crystals lining fractures in the veins, and in thin crusts in open spaces in the veins.

Hobbs (1945, p. 11–13) suggested that mineralization in the Ima deposit took place in two closely related stages. The first stage of mineralization is represented by extensively distributed, irregularly shaped, orthoclase-bearing quartz veins that locally are pegmatitic. These veins contain mica, pyrite, molybdenite and some chalcopyrite, and are best developed in and near the granite, although narrow stringers and veinlets are more widely distributed. They are localized along the northwest trends of the main vein zones, but are irregular in strike and dip and do not contain ore minerals in economic amounts. The second stage of the mineralization, during which most of the economically important ore was formed, is represented by a series of more persistent and regular veins that generally follow the trend of the first-stage veins. The veins of the second stage are two types: (1) massive unbanded, milky-white quartz veins containing large crystals of pyrite, fluorite, and huebnerite, and (2) well-banded quartz veins that contain these minerals plus rhodochrosite, tetrahedrite, sphalerite, galena, and late chalcopyrite. The two types of veins in the second stage are closely related, the banded ore being formed wherever the unbanded veins were fractured and sheared, and additional minerals were introduced. The repeated reopening of the veins and shearing along the vein walls has resulted in the incorporation of fragments of wall rock into the quartz. Many of these have been partly or completely altered to sericite, or to sericite and quartz aggregates. The several successive reopenings of the veins are best shown in the banded quartz veins, where later bands truncate earlier formed bands. At least three such reopenings are recognizable, and others probably exist that cannot be differentiated. The mineralization in any one vein has occurred, therefore, in a series of stages in which different minerals, or at least different proportions of minerals, were deposited at different times.

Deposition of disseminated molybdenite in granite of the Ima stock probably took place a little before or at the same time as the first, pegmatitic, stage of mineralization described by Hobbs (1945, p. 11-13).

DISSEMINATED DEPOSITS OF COPPER AND MOLYBDENUM IN GRANITIC ROCKS

In addition to the Ima stock, two other stocks in the central Lemhi Range, the Big Eightmile stock (Ruppel, 1980) and the Gilmore stock (Ruppel and Lopez, 1981; Ruppel and others, 1970, p. 28) are known to contain metallic sulfide minerals. The south half of the Big Eightmile stock is well exposed on the glaciated canyon wall of Big Eightmile Creek, but the north part is largely concealed beneath glacial deposits. The core of the stock has been hydrothermally altered and sericitized; most mafic minerals have been strongly altered or destroyed. Metallic sulfide minerals, pyrite, chalcopyrite, and molybdenite, are distributed in these hydrothermally altered rocks, and are rare or absent elsewhere in the stock. The exposed top of the altered and mineralized granitic rocks is capped by a brecciated, siliceous, limonitic gossan that is heavily encrusted with malachite and azurite, the principal minerals prospected for in the Blue Jay mine. Beneath the gossan, the granitic rocks have been oxidized to a depth of 75-100 m, and the metallic minerals have been converted to carbonates and oxides. At greater depth, sulfide minerals are dominant and occur disseminated in the granitic rock, in thin quartz veins along fractures and in fracture zones, and in thin films coating fracture surfaces. The combined sulfide minerals most commonly form less than 1 percent of the rock, and are mainly chalcopyrite, a trace of covellite, and pyrite, with subordinate molybdenite. In a few strongly fractured zones, where thin quartz veins are more abundant, the combined sulfides form as much as 3 percent of the rock; in these zones molybdenite is relatively common in thin films on fracture surfaces.

The Gilmore stock is almost completely concealed by glacial and pediment gravels, which form a wedge of surficial material that thickens eastward from a few meters at the mountain front to at least several hundred meters near Middle Ridge, where the gravels are interrupted by a fault. The granitic rocks exposed in the Gilmore area are dikes, sheets, and sills around the edges of the Gilmore stock; these thin intrusive bodies do not contain appreciable amounts of metallic sulfide minerals. Geochemical studies suggest, however, that the stock itself contains some molybdenum (Ruppel and others, 1970, p. 29). Samples of granitic rocks from the single drill hole known to have reached through the gravel blanket into the stock contain from less than 1 percent to as much as 10 percent pyrite, apparently in thin quartz veins in fractures and fracture zones. The main part of the stock has not been penetrated in drill holes, however, and the geochemical study (Ruppel and others, 1970) was designed to look for extensions of lead-silver replacement veins beneath the blanket of glacial deposits rather than for disseminated or stockwork deposits of copper and molybdenum. The potential for disseminated copper-molybdenum sulfide resources in the Gilmore stock therefore remains nearly untested; the scant evidence available suggests that the stock probably does contain some pyrite, chalcopyrite, and molybdenite, like that in the Big Eightmile stock.

DISSEMINATED DEPOSITS OF LEAD IN GRANITE AT THE KIMMEL MINE

The Kimmel mine is in the Junction district, just west of the Leadville (Sunset) mine. It explores a deposit of lead minerals, both galena and cerussite, disseminated in granite. The granite is of early Paleozoic age and has been strongly hydrothermally altered, sericitized, and silicified, like younger Paleozoic (Mississippian) limestone that occurs both in small, faulted blocks in the granite and in adjacent fault slivers. The alteration and associated mineralization clearly are much younger than the granite. They are similar, however, to alteration and mineralization in and near the lead-silver replacement veins in the adjacent Leadville mine, a relation that suggests that they represent the same mineralizing episode, in early Tertiary time. The granite was brecciated before alteration, as a result of thrust faulting that carried it over the adjacent limestone; the scattered limestone blocks probably were incorporated in it during thrusting. It is essentially a fault breccia, probably near the base of an imbricate thrust slice that has been fragmented further by postalteration steep faulting along the range front of the Beaverhead Mountains (Ruppel, 1968). The mineralizing solutions that formed the Leadville replacement deposits seem likely also to have pervasively entered the shattered granitic rocks and to have been responsible both for the nearly complete hydrothermal alteration and for introduction of the disseminated metallic sulfide minerals.

The shattered and hydrothermally altered granite and limestone contain disseminated fine-grained pyrite and even finer grained steel galena and secondary cerussite and limonite. Assays of channel samples (T.H. Kiilsgaard, written commun., 1970) suggest that the lead content ranges from 0.1 percent or less to as much as 5.5 percent, averaging about 1 percent. Silver is not as evenly distributed as lead, probably because of secondary alteration. Most commonly only 0.05–0.5 oz of silver per ton is present, but in a few places the silver content ranges from 1 to 3 oz per ton. Assays on sulfide concentrates (Bell, 1920, p. 61) suggest similar ore values, about 0.5–0.6 oz of silver for each 1 percent of lead, and suggest that the traces of gold also present are associated with pyrite, because they were lost in concentrates of lead minerals alone.

DEPOSITS OF SECONDARY COPPER MINERALS

Small deposits of secondary copper minerals, mainly malachite and azurite but including chalcocite and covellite in a few places, are common in the area between Jakes Canyon and Mollie Gulch, north of Leadore, and in the area west of Big Eightmile Creek in the central Lemhi Range. They occur both as thin seams lining the walls of fractures, and as thin irregular stringers, nodules, and nodule coatings in zones of gouge and breccia in thrust faults. The largest deposits are those in thrust faults; a few of these have yielded small amounts of secondary copper ores and specimen quality samples for mineral collectors. All the deposits are either in the Apple Creek Formation or in thrust faults that cut the Apple Creek nearby. This consistent association suggests that the copper has been leached from the green siltite and fine-grained quartzite of the Apple Creek Formation, transported in fractures and fault zones by ground water, and redeposited as secondary copper minerals where the fractures and fault zones intersect the surface. The Apple Creek Formation is locally pyritic, and the consistent association of small deposits of secondary copper minerals with the formation suggest that it also contains stratabound copper (Harrison, 1972, p. 1232–1234).

DEPOSIT OF COPPER-BEARING MAGNETITE

A deposit of copper-bearing magnetite is present at Sims Copper (Bruce estate) prospect (fig. 19, loc. 23), in the south part of the Spring Mountain district, on the north wall of the small cirque at the head of Bruce Canyon. The deposit is in the lower part of the Saturday Mountain Formation, above a short sill of granitic rocks that intruded the contact of the Saturday Mountain Formation and the Kinnikinic Quartzite at the margin of the Spring Mountain sheet. The dolomite of the Saturday Mountain Formation is almost surrounded by granitic rocks and has been strongly metamorphosed to white, finely to coarsely crystalline marble that locally contains clusters of radiating crystals, probably tremolite, as much as a centimeter in diameter. The contact of granitic rocks and dolomite typically is an almost gradational, pervasively intruded zone as much as 3 m thick, in which the dolomite has been converted to coarse-grained, garnet- and epidote-rich dolomite skarn. Inclusions of skarn are particularly abundant in granitic rocks near the contact, but are also present, and only slightly less abundant, throughout most of the sill.

The largest magnetite body is an irregularly shaped and layered, lenticular, northeast-trending body about 45–50 m long and 8–15 m wide. The layers, which commonly are 2–15 cm thick, are gradational and not well defined. They consist either of almost pure magnetite in crystals from less than 1 cm to about 3 cm in diameter, or of magnetite with a small amount of hematite and magnetite with 40–60 percent calcite. The magnetite lens is enclosed in skarn and marble that also contain abundant magnetite in veins and in disseminated crystals and clusters of crystals. Typically, the skarn contains about 10 percent magnetite; the marble contains as much as 30 percent magnetite. Locally abundant secondary copper minerals, mainly malachite, azurite, and cuprite, are derived from primary bornite. Small shipments of ore from the Sims copper prospect have yielded 1.5–2 percent copper, 60 percent iron, and a trace of silver and gold. A similar but smaller magnetite body in garnet-epidote-phlogopite skarn of metamorphosed dolomite of the Jefferson Formation, at the Tempest claim on the east margin of the Spring Mountain sheet, east of the Sims copper prospect, has yielded a small amount of ore containing 8–10 percent copper and as much as 10 oz of silver per ton. Other small deposits of copper-bearing magnetite are common along the margins of the Spring Mountain sheet elsewhere in Bruce Canyon and farther west into the upper canyon of Squaw Creek.

SUGGESTIONS FOR PROSPECTING

The principal known mineral deposits in the Lemhi Range are associated with intrusive igneous rocks that were emplaced in and along fault zones and controlled primarily by older structures. Steeply dipping northwest- and east-trending fault zones, which are inferred to control the block uplifts in this region (Ruppel, 1982), apparently acted as conduits for magma rising from deep within the crust and thereby controlled the emplacement of stocks and associated mineral deposits. The magma spread laterally into imbricate thrust faults at and near the bottom of the Medicine Lodge thrust plate, to form sheets like those exposed in the central Lemhi Range. The major mineral deposits of the region were deposited around and above necklike, central stocks that fed the sheets; only small deposits of metallic minerals are associated with the sheets themselves. The search for new mineral deposits, therefore, might be concentrated most usefully in areas known or inferred to be above the fault-bounded edges of buried blocks of crystalline rocks, along the edges of the linear mountain ranges of this region. Substantial orebodies have not been found and are not likely to have been deposited in the structurally flat central parts of the block uplifts, because the seemingly necessary deep structural controls are not present there.

The primary structural controls so evident in the localization of mineral deposits in the Lemhi Range seem likely also to have controlled localization of mineral deposits in the Beaverhead Mountains. Only the small antimony-lead-silver deposits of the Little Eightmile district are associated with exposed Tertiary granitic rocks, but other lead-silver deposits, all of them small except the Viola deposit at Nicholia, are scattered farther south along the southwest flank of this mountain range. The regional association of lead-silver deposits with Tertiary granitic stocks intruded into the lower part of the Medicine Lodge thrust plate suggests that buried stocks and associated mineral deposits similar to those at Gilmore could be present at depth in the south part of the Beaverhead Mountains. The concealed stocks, if they are present, may also contain disseminated deposits of molybdenum and copper minerals, like those in the Big Eightmile stock, the Ima stock, and probably the Gilmore stock.

The primary structural controls on the emplacement of granitic igneous rocks and associated hydrothermal mineral deposits probably also explain why so few mineral deposits have been found in the Lost River Range, west of the Lemhi Range. Only a few small, widely scattered quartz veins and even fewer lead-silver replacement veins are known in the Lost River Range, the main ones occurring along its western flank (Ross, 1947, p. 1156–1158). None have yielded much, if any, ore. The reason may be that the basal part of the Medicine Lodge thrust plate is deeply buried beneath this mountain range, much more deeply than it is beneath even the south part of the Beaverhead Mountains. If there are any mineral deposits, they are largely confined to that basal part of the plate, and therefore, are deeply buried.

In summary, most of the known mineral deposits of the three mountain ranges of east-central Idaho are found in consistent patterns that suggest the importance of primary structural controls in their emplacement, and the importance of understanding those controls in a search for other mineral deposits. Nearly all deposits of any economic significance are on or near the edges of the ranges, where their emplacement, and that of their parent granitic igneous rocks, has been controlled by ancient, steep faults that flank these block-uplifted ranges. Nearly all known mineral deposits are in or near the lower part of the Medicine Lodge thrust plate, which is exposed in places in the central part of the Lemhi Range where there are major mineral deposits. The lower part of the thrust plate is buried beneath the south part of the Beaverhead Mountains, where the exposed mineral deposits are relatively small, and is even more deeply buried beneath the Lost River Range, where for all practical purposes no mineral deposits are exposed at all. Mineral deposits seem likely to be present at depth, however, along the flanks of both the Beaverhead Mountains and the Lost River Range, as well as in places along the flanks of the Lemhi Range. The most likely places to explore are those where small base- and precious-metal veins or quartz veins, widespread metallization of country rocks, or widespread hydrothermal alteration suggests the presence of buried intrusive rocks and associated hydrothermal systems.

The ring-fracture zone that localized the concealed Ima stock, at Patterson (Ruppel, 1982, p. 9), also contains two exposed, small granitic stocks farther northeast. In addition, this zone cuts across three thick (3–8 m), east-trending hematite-bearing quartz veins, one of which contains interlayered veins of siderite, 0.3–1 m thick, on the east flank of Patterson Creek 1–2 km north of its junction with its south fork (Ruppel, 1980; Ruppel, 1982, p. 14). Neither stock contains any apparent, disseminated metallic sulfide minerals, and the quartz veins are similarly barren, although they do show that the ring-fracture zone has been intruded in several places by granitic rocks. Conceivably, other granite stocks like the Ima, with its associated tungsten-quartz veins, copper-silver quartz veins, and disseminated molybdenite, may be present but concealed elsewhere in the ring-fracture zone.

A wide zone of quartz veins is exposed in the south-central part of the Patterson quadrangle (Ruppel, 1980), on the remote and unprospected ridges that flank the headwaters of the West Fork of Big Creek and Mill Creek, west of Big Creek. The zone of quartz veins trends about east and dips 70° S. to vertical; it includes several groups of closely spaced veins and single veins. The veins are separated by country rock of the Inyo and

West Fork Formations that has been bleached locally, strongly sheared parallel to the vein system, and in places sheared parallel to bedding. The veins mostly are in a zone about 1 km wide and almost 3 km long; more widely separated, single veins are present as much as 3 km farther north. The veins range from 1 cm or less to almost 2 m thick, and consist of massive white quartz that in a few places is sheared and recemented by a later generation of fine-grained gray quartz, siderite, and locally calcite. The thickest group of quartz veins is 12–15 m thick. Stringers and coarse crystals of specular hematite are common to abundant in many of the veins and in adjacent wall rocks; pyrite and chalcopyrite are sparsely distributed through the veins east of the West Fork of Big Creek. As is typical of all of the groups of closely spaced veins, the principal veins are about east trending and nearly vertical; they are connected by a complex network of smaller veins of different trends and dips, many of them about north trending and east dipping more or less parallel to bedding. The abundant quartz veins, and the bleaching and alteration of the adjacent quartzite and siltite, suggest hydrothermal alteration and mineralization above a granitic stock intruded into a wide, east-trending and nearly vertical zone of shearing. No granitic rocks are exposed nearby, although they could be concealed beneath the widespread glacial deposits in the valley of Big Creek, or at greater depth beneath bedrock. The east-trending sheared zone probably is an extension of the east-trending faults that seem partly to have controlled the emplacement of the Big Timber stock, in the north part of the Gilmore quadrangle (Ruppel and Lopez, 1981).

Finally, given the presence of the seemingly required primary structural controls and of an appropriate magmatic/hydrothermal setting, the presence of carbonate rocks suitable as hosts for base- and precious-metal replacement deposits is a useful prospecting guide. Many of the lead-silver replacement deposits of the region are in the dolomitic rocks of the Jefferson Formation—an association so common that for many years the suspicion that the ores are stratabound has influenced prospecting. Clearly the Jefferson Formation is a favorable host for replacement deposits, but the case for stratabound ores is not particularly compelling, at least on the basis of available evidence. Other carbonate rocks also contain lead-silver replacement deposits, both the dolomite of the Saturday Mountain Formation and the Mississippian limestones, and these suggest somewhat broader opportunities for mineral deposits than the Jefferson Formation alone. The Saturday Mountain Formation beneath the productive orebodies of the Texas district has not been explored, but could well be a source of ores at greater depths than have been reached in these mines. Carbonate rocks, both the Saturday Mountain and Jefferson Formations, probably are present at or near the north margin of the Big Eightmile stock, but are buried beneath younger Challis Volcanics and glacial deposits (Ruppel, 1980). The depth of burial is not known, but the area between Big Eightmile Creek and Everson Creek offers some chance for discovery of lead-silver replacement orebodies beneath the blanket of volcanic rocks and surficial deposits.

DESCRIPTIONS OF MINES

The mines and prospects in the central Lemhi Range and in the adjacent Junction district in the Beaverhead Mountains number in the hundreds, but only a few mines have yielded more than two or three hundred tons of ore. Most of the lead-silver ore has come from two mines, the Pittsburgh-Idaho and Latest Out mines, with much smaller amounts from the Hilltop and Leadville mines. Tungsten ore has been mined only in the Blue Wing district, almost entirely from the Ima mine. Gold ore has been mined from only one vein, the Martha vein in the Allie mine in the Texas district. These mines, and a small number of other mines and prospects in the central Lemhi Range, are described briefly in the following pages. Their locations are shown on figure 19, and with the location of other mines and prospects, on the geologic maps of this region (Ruppel, 1968; 1980; Ruppel and Lopez, 1981). The descriptions are grouped by district, from the Junction district in the Beaverhead Mountains on the north, to the Texas and Spring Mountain districts in the Lemhi Range farther south, to the Blue Wing district on the west flank of the Lemhi Range, and finally to other small mines and prospects that are not in the main mining districts. Within each district, the principal mines are described first, and then smaller mines and prospects, which illustrate the kinds of ore deposits explored, are described in geographic order from north to south.

Because most of the mines now are caved and inaccessible, their descriptions were drawn from many sources, particularly from Umpleby (1913), and from mining periodicals, reports of the Idaho Inspector of Mines, and a few informal reports (see footnote, p. 84). Many of the mines and prospects in the Texas and Spring Mountain districts were examined by V.E. Scheid, in a U.S. Geological Survey study in 1940 and 1941, and some of Scheid's descriptions and other information from his notes and maps, not published elsewhere, are incorporated here. In addition, Scheid's notes included information drawn from mine maps and sections made by R.T. Walker in the Pittsburgh-Idaho mine in 1925, when the mine was almost completely developed. These maps were loaned to Scheid in 1941, by Milo Zook, then lessee in the Pittsburgh-Idaho mine. Known as the Walker Atlas, they included maps showing all of the workings in 1925, the mine geology and distribution of veins, and composite mine-plan maps and cross sections, in all an extraordinary assemblage of maps. Scheid's excerpts permit more complete understanding of the Pittsburgh-Idaho mine than would have been possible otherwise.

JUNCTION DISTRICT

The Junction mining district includes the mines and prospects on the south-facing flank of the Beaverhead Mountains, north of Leadore, Idaho (Umpleby, 1913, p. 112–118; Ruppel, 1968), and extends west and northwest from the mouth of Railroad Canyon to the Mineral Hill mining area and the mouth of Mollie Gulch, where it merges into the Little Eightmile mining area (Staatz, 1973) (fig. 19). It extends northward to include a number of mines on the flat top of Grizzly Hill, north of the Leadore quadrangle, among them the Road Agent (Jap) mine (loc. 2), the Blue Lead mine (loc. 3), and the Dignore mine (loc. 4). The principal mines in the district explore lead-silver replacement deposits in Mississippian limestone; a few small mines explore deposits of secondary copper minerals in fractures and in thrust fault breccias. Many small prospect pits have been dug on gash veins filled with massive white quartz and specular hematite.

Only one mine in the district has yielded much ore, the Leadville (Sunset) lead-silver mine (loc. 1). The Kimmel mine, adjacent to the Leadville on the west, explores a deposit of pyrite and galena disseminated in granite, but has not yielded any ore.

Leadville (Sunset) mine.—The Leadville mine (fig. 19, loc. 1) is located in the central part of the Junction district on the lower slopes of the Beaverhead Mountains between Italian Gulch and Baby Joe Gulch (NW¼, sec. 24, T. 16 N., R. 26 E.). The mine was located in June 1904. Active development began in 1905 (Umpleby, 1913, p. 115) and continued in subsequent years with the sinking of a single-compartment vertical shaft to 117 ft (36 m) in 1907; the shaft was deepened to 300 ft (91 m) in 1910, with drifts at different levels on the vein from the shaft. In 1909, a long crosscut adit was opened, which by 1917 reached from its portal in Tertiary tuffaceous sandstone, nearly on the banks of Canyon Creek, northward almost 900 ft (275 m) to intersect the Leadville vein. A drift run about 800 ft (245 m) east on the vein intersected the orebody at a depth of 510 ft (155 m). The adit was extended about 450 ft (138 m) to the northwest, under the adjacent Kimmel mine. The first shipments, totaling about 260 tons of hand-sorted ore, were made in 1907 and reportedly averaged 56 percent lead, 45 oz of silver per ton, and about 0.1 oz of gold per ton (Bell, 1908, p. 124–125). Later ore shipments continued regularly until 1912 but were of somewhat lower grade, perhaps averaging about 30 percent lead, 20 oz of silver per ton, and little or no gold. From 1912 to 1916, the mine apparently was under development and was only intermittently active; no ore is known to have been shipped. In 1917, ore shipments, partly from dumps and partly from ore left in the upper workings, were resumed and continued until about 1919, when the mine was closed. The shipments during the last 2 years of operation averaged about 7–8 percent lead, 4–6 oz of silver per ton, and only a trace of gold. A 25-ton concentrating mill with four tables was installed in 1918, and flotation machinery was installed in 1923–1924 to process ores from both the Leadville and Kimmel mines, but no shipments of concentrates after 1919 are known. The Leadville claims were prospected repeatedly by different mining companies between 1960 and 1980, and were extensively drilled and trenched. Apparently no new orebodies were found, and the mine remains idle. The original adit and shaft are caved and have been largely obliterated by more recent trenching. The portal of the lower working tunnel is open; the tunnel provides access to the lower part of the mine, but drifts and stopes above it are caving. The total production from the Leadville mine has been about 4,000–4,500 tons of ore, valued at about $100,000.[8]

The Leadville mine originally was explored through a short east-trending adit run under the surface exposures of the vein, but the principal development, later, was from a single-compartment vertical shaft; drifts run east and west from the shaft on the vein. The long, crosscut adit and connected drift, completed in 1917, were utilized as the main working tunnel during the last few years of mining. This working tunnel was connected with five level drifts in the upper workings of the mine, and with the original exploration adit and shaft, through a series of inclined raises and stopes. The longest drifts, containing the major stopes, are the original adit, about 500 ft (150 m) long, and two levels at more or less hundred-foot intervals beneath the adit, each of them about 400–450 ft (120–140 m) long. The lower three levels are more closely spaced, at 40–50 ft (12–15 m) intervals, and none of them is much more than 100 ft (30 m) long. The lowest of these levels, about 375–400 ft (115–120 m) below the original adit, is connected to the working tunnel through a short incline and a vertical raise from the tunnel. A single, deeper level explored the vein about 100 ft (30 m) below the working tunnel, at a depth, beneath the outcrop, of about

[8]Ore values are given in dollars at the time of production.

600–650 ft (183–200 m). Most of the ore came from the levels in the upper half of the mine, at depths of about 300 ft (90 m) or less. The grade of the ore, as reflected in the average grade of shipments, clearly declined with increasing depth, perhaps reflecting secondary enrichment of near-surface ores, as is also suggested by the higher gold content of near-surface ores. The smaller extent of the workings below about 300 ft suggests that the ore zone in the Leadville vein decreased in size at these depths, and that the ore zone contained only a few small shoots of milling-grade ore. In the deepest level in the mine, about 100 ft (30 m) below the working tunnel, the vein contained small lenses of galena.

The orebodies in the Leadville mine were described by Umpleby (1913, p. 116–118) as occurring in two shoots, separated by about 40 ft (12 m) of barren, crushed limestone. The shoots occupy an east-trending fault zone filled with bleached and hydrothermally altered fragments of Mississippian limestone, sericitized granite of Ordovician age, broken fragments of ore minerals, and clay gouge. The hanging wall of the fault zone is Tertiary tuffaceous sandstone, faulted against the vein zone in Miocene-Pliocene time. The footwall is brecciated, complexly faulted limestone and granite that has been bleached, hydrothermally altered, sericitized and silicified. The western shoot was about 180 ft (55 m) long and a few inches to 2 ft (0.6 m) thick at a depth of about 200 ft (60 m), decreasing rapidly in length and thickness at both higher and, especially, lower levels. The eastern shoot, at the same level, was about 110 ft long and ranged in width from a few, scattered crystals to as much as 4 ft (1.2 m) of galena. It decreased in length and thickness on the upper and lower levels in the mine, like the western shoot. The ore consisted mainly of fine-grained steel galena, less common pyrite, and rare sphalerite and chalcopyrite. The eastern shoot contained an appreciable amount of antimony, arsenic, and bismuth, although the minerals that contained these elements are unknown. Secondary ore minerals included cerussite and anglesite, but Umpleby (1913, p. 118) stressed that such oxidation products were rare in Leadville ores, and that most of the ore minerals were primary metallic sulfides. The ores from the uppermost levels of the mine had already been mined and shipped when Umpleby studied the mine, however, and, as noted above, their much higher grade suggests secondary enrichment of the near-surface ores.

Kimmel mine.—The Kimmel mine (fig. 19, loc. 1) is located just west of the Leadville mine and also is in the NW¼, sec. 24, T. 16 N., R. 26 E. The mine was opened about 1918, although its disseminated ores had been recognized earlier but ignored because of their low grade. Virtually all the underground workings at the mine were completed between 1918 and 1921, including a main adit that trends northwest at its portal, turns abruptly northeast about 100 ft (30 m) inside the mine to follow a similarly trending shear zone for about 100 ft (30 m), and then turns northward for an additional 350–400 ft (107–122 m). Three northwest-trending crosscuts from the adit, one of them an extension of the adit where it turns northeast, are each 100–200 ft (30–60 m) long with connecting crosscuts and stub drifts on a shear zone. The underground workings in all are about 1,600 ft (490 m) long. The deposit is also explored at greater depth by a crosscut extended about 450 ft (138 m) to the northwest from the Leadville mine working tunnel, about 200 ft (61 m) beneath the Kimmel workings. The concentrating mill and tables installed at the Leadville mine in 1918 were leased for use at the Kimmel mine in 1921, and in 1923–1924 a flotation plant was installed to concentrate ores from the Kimmel, Leadville, and Baby Joe mines; the Baby Joe adjoins the Kimmel on the west, and explores similar disseminated lead ores. Despite promising concentration tests and optimistic estimates of available ore in the Kimmel mine (628,571 tons) (Mining Truth, v. 9, no. 11, July 16, 1924), the mine has no recorded production. It apparently closed in 1924 or 1925 and has remained idle since then. Most of the original workings are still open.

As discussed on p. 95–96, the Kimmel mine explores deposits of disseminated fine-grained pyrite and still finer grained steel galena in hydrothermally altered and sericitized granite and limestone breccia. Assays suggest that the lead content ranges from 0.1 percent or less to as much as 5.5 percent and averages about 1 percent. The silver content commonly ranges from 0.05 to 0.5 oz per ton, although in a few places there is as much as 3 oz of silver per ton. Gold is present only in trace amounts, apparently associated with the pyrite.

TEXAS AND SPRING MOUNTAIN DISTRICTS

The Texas and Spring Mountain mining districts include a large number of mines and prospects along the east flank of the central Lemhi Range, extending from Deer Creek, north of Gilmore, southeast to Big Windy Peak and the head of Warm Creek, south of Spring Mountain Canyon (Umpleby, 1913, p. 85–109; Ruppel and Lopez, 1981) (fig. 19). The dividing line between the two districts is about at Long Canyon, south of Gilmore. Most of the mines and prospects in these districts explore lead-silver replacement deposits in carbonate rocks of the Jefferson and Saturday Mountain Formations. One mine, the Allie, explored a gold-bearing vein, and two others, the Hilltop and Democrat, have yielded lead-silver-gold ores. In general, however, gold is an unusual element in the ores of the districts. The Silver Moon mine has yielded ores valuable mainly for their silver content. The greatest part of lead-silver ores produced in Lemhi County has come from two mines in the central part of the Texas district, the Pittsburgh-Idaho and Latest Out mines. The Hilltop mine, in the north part of the Texas district, is the only other mine that has developed a large orebody. The Democrat mine, which explores a lead-silver-gold orebody directly beneath the older Hilltop mine, is a new mine, and the extent of the orebody exposed in the mine is not known.

The total values of ores from mines in the Texas district is about $17–17.5 million. Of this total, about $11.5–12 million was for ore from the Pittsburgh-Idaho mine, about $2.5 million for ore from the Latest Out mine, and $1–2 million for ore from the Hilltop and Democrat mines. The total value of ores from the Spring Mountain district is about $1 million.

Pittsburgh-Idaho (United-Idaho) mine.—The Pittsburgh-Idaho mine (fig. 19, loc. 5), a lead-silver mine, is in the central part of the Texas district, near Gilmore, at an altitude of about 7,800 ft (2,380 m). The mine was first opened, in 1902–1903, by a west-trending adit on the Silver Dollar claim, which intersected a blind vein, about 90 ft (27 m) from the portal at a depth of about 70–75 ft (20–22 m) beneath the surface. The vein, probably the East Vertical vein, which is the easternmost of the major veins developed in the Pittsburgh-Idaho mine, was subsequently explored by a drift southward on the vein about 400 ft (122 m), by a vertical winze, ultimately 400 ft (122 m) deep, with other levels and stopes at hundred-foot intervals, and by an inclined winze that may have been incorporated later into a deep, inclined shaft that by 1918 was the main production shaft. In 1909, the collar of the vertical winze squeezed shut, and the original adit, which had been mined out, was abandoned, although it remains the reference, or 0-ft, level in naming the lower levels in the mine. A new adit was connected from the surface to the 100-level in the mine, and from 1909 to 1918, this adit was the principal working adit (Umpleby, 1913, p. 99–103, for extent of mine workings in 1911). The deep shaft, inclined about 50° west-southwest, was completed to a depth of 700 ft in 1918. It connected the levels above the 400-level opened by 1909, the 500-level reached in 1910–1911, and the 600-level reached in 1912. Permanent water level and an increasing proportion of sulfide minerals in the ore were reached in a winze below the 600-level in 1913. In 1912, a long, approximately southwest trending adit was started from two headings, one on the 400-level of the mine, and one, the portal, at an altitude of 7,390 ft (2,254 m), to connect the undergound workings with a railroad heading. This adit, the Transportation Tunnel, was completed to a total length of about 6,000 ft (1830 m) by 1917, reaching nearly under the Latest Out mine, west of the Pittsburgh-Idaho, to which it was later connected through the Meister and Roberti raises. The Transportation Tunnel tied together most of the different mines in the central part of the Texas district; it was the main production tunnel for ore mined by lessees above the 400-level in the Pittsburgh-Idaho mine after 1917 and for ore mined from the Allie mine after 1916.

Umpleby's report (1913, p. 101–103) shows that by 1911 only three veins, the East Vertical vein, the East Flat vein, and the West vein were known in the Pittsburgh-Idaho mine, although other veins were known and partly explored on the adjacent claims—the Martha gold vein to the east, and the Latest Out and Roy Lauer veins to the west. As the Pittsburgh-Idaho mine was deepened and its levels extended in subsequent years, and the inclined shaft and Transportation Tunnel were completed, other veins were discovered; more than a dozen veins were finally explored, nearly all of them blind or poorly exposed at the surface. Substantial quantities of lead-silver ore were recovered from about half of these veins, the early-discovered three veins from the 700-level nearly to the surface, and the others, farther west, from between the 700- and 400-levels. After 1919, exploration continued from the inclined shaft at deeper levels in the mine—on the 800-level where small irregular stopes were mined on a few veins that were barren on the higher levels, and on the 950-level, the deepest reached in the mine, where several veins contained sulfide ores, but none on this level were mined. All work on the 700-level and below was hampered by heavy flow of water. Production after 1919 is not known, but comparatively little ore seems likely to have come from these deeper levels; what production there was came mainly from bypassed veins and ore on and above the 700-level.

The Pittsburgh-Idaho mine was closed in 1929, after a machinery failure and fire in surface installations. The deeper levels, below the 500-level, were allowed to flood. Mining and prospecting by lessees continued intermittently on the Transportation Tunnel level (400-level) and above until about 1941; after that time, the mine apparently was idle until 1953.

From 1953 to 1957, a previously unexplored area on the Transportation Tunnel (400) level containing the Silver Dollar vein and Neversweat flat vein was explored. In 1953 and 1954, almost 4,000 ft (1,220 m) of the Transportation Tunnel was rehabilitated, to a point west of the Silver Dollar vein. The Silver Dollar vein was explored in 1954–1957, in drifts extended about 400 ft (122 m) south of the Transportation Tunnel and 300 ft (92 m) north; these drifts disclosed only sparse lead minerals in iron- and manganese-stained gouge and breccia. A crosscut was extended westward from the north drift to explore for extensions of the Neversweat vein but did not encounter that vein; a crosscut from the south drift also failed to find any new orebodies. The mine was closed again in 1957 and has remained idle since then. Most of the workings above the Transportation Tunnel are caved and inaccessible, and those below the Transportation Tunnel are flooded, as they have been since 1929. The main, inclined shaft is caved at the collar.

In summary, the Pittsburgh-Idaho mine explored a dozen veins or more and developed substantial orebodies on about six veins. The mine included nine levels, and six sublevels between the 0-level, the discovery adit, and the 950-level at the mine bottom. The upper levels, above the 400-level, were originally connected by a vertical winze, but the deeper levels were mined through an inclined production shaft extending to the 950-level. The Transportation Tunnel intersected the 400-level and was the main production tunnel late in the mine's history for ores mined above the 400-level in the eastern part of the Pittsburgh-Idaho mine and for ores from the Allie mine. The Neversweat orebody, the westernmost major orebody in the Pittsburgh-Idaho mine, was mined through an inclined winze on six closely spaced levels below the Transportation Tunnel; the deepest level was the Neversweat 181, 181 ft (55 m) below the tunnel.

The mine yielded a total of about 282,000–290,000 tons of lead-silver ore, valued at perhaps $11.5-12 million. The average grade of the ore mined was about 27 percent lead, 5–10 percent zinc, 13 oz of silver per ton, and 0.03 oz or less of gold per ton. Umpleby (1913, p. 65–103) described the ore from the upper levels of the mine as containing about 37 percent lead, 9 percent zinc, 15.1 oz of silver per ton, and 0.03 of gold per ton, perhaps reflecting secondary enrichment of these ores.

In the Pittsburgh-Idaho mine, most of the veins trend about N. 10°-15° E., but occur as both steep veins, which dip 70°-90° W., and flat veins, which dip 40°-60° W. (p. 90). Both sets of veins contained orebodies, but the largest and most productive orebodies were in steep veins at and above their intersections with flat veins. The most productive veins in the mine, from east to west, were: the East Vertical (steep) vein, stoped along strike lengths as much as 125 ft (38 m), from the 400-level to the surface; the East Flat vein, stoped along strike lengths of 100–200 ft (30–61 m), from the 600-level to the surface; the West Vertical (steep) vein, stoped along strike lengths of 250–350 ft (75–107 m), from below the 700-level to just below the 200-level, where it intersected the East Vertical vein; the Reverse vein, a relatively minor, steeply east dipping vein stoped as much as 60 ft (18 m) along strike between the 400- and 100-levels; the No. 1 West Footwall vein, a steep vein stoped along strike lengths of 100–500 ft (30–152 m), between the 800- and 500-levels, the greatest strike lengths being between the 700- and 500-levels; and the Neversweat Flat vein, stoped along an uncertain strike length between the Neversweat 181-level and the Transportation Tunnel, which is the Neversweat 0-level. Smaller veins included the Texas Flat vein in the east part of the mine, which intersects both the East and West Vertical veins below their productive ore shoots, and contained small, irregular bodies of ore in a few places between the 800- and 600-levels; the Intermediate vein, west of the West Vertical vein; the Silver Dollar vein, which yielded a small amount of ore from a separate mine, the Silver Dollar tunnel, but did not contain ore where it was explored from the Transportation tunnel; and a few others, both steep and flat. The orebodies in the productive veins were lenticular masses that in a few places were as much as 8 m thick, but more commonly were 1–5 m thick.

All deposits are lead-silver replacement veins and irregular replacements in dolomite in the lower part of the Jefferson Formation (p. 90). Probably they extend into the dolomite of the underlying Saturday Mountain Formation, but the present mine workings are not deep enough to explore these rocks.

The veins are intersected by several steeply west dipping dikes in the mine; Umpleby (1913, p. 98) stated that the dikes are younger than the veins. Later mining suggested, however, that the veins continue through the dikes as unmineralized fracture zones (Milo Zook, oral commun., 1941, to V.E. Scheid). The conclusion that the veins are younger than the dikes is suggested also by the relation of mineralization to the Gilmore stock. The stock, which was only recently recognized (Ruppel, 1982; Ruppel and others, 1970), is compositionally similar to the dikes, and both the dikes and the Spring Mountain sheet are offshoots fed from the stock. The mineralizing fluids that formed the lead-silver replacement deposits were derived from the stock after it was emplaced, and represent a later, hydrothermal stage.

The veins and the dikes are broken repeatedly by northwest-trending faults that dip 35°–45° northeast, parallel to bedding in the Jefferson Formation. The displacement of the hanging-wall side is consistently down on the east. Commonly the displacement is only a few feet (a meter or less), but it ranges from a few feet to as much as 50 ft (15 m) on the Neversweat zone of east-dipping faults deep in the mine. The displacement probably is much greater on the still deeper Gilmore fault, exposed on the 950-level, because the veins below the Gilmore fault seem likely to be different veins than those more completely explored above the Neversweat fault zone on the 800-level. At any rate, the flat, east-dipping faults clearly are younger than both the dikes and veins. Probably they are gravitational normal faults, or slide blocks, formed during block-uplifting in mid-Cenozoic time (p. 78) (Umpleby 1913, p. 95; Ruppel, 1982). They are not clearly recognizable on the surface. Their principal effect has been to offset the Pittsburgh-Idaho veins and orebodies to consistently lower levels eastward, in much the same way that the tungsten-quartz veins at the Ima mine are consistently offset to the west on similar but west-dipping flat faults on the west side of the Lemhi Range. The veins on the 950-level of the Pittsburgh-Idaho, below the Gilmore fault, seem likely to be east of the main veins at this depth. If so, the deeper extensions of the main Pittsburgh-Idaho veins and orebodies, beneath the Gilmore fault, are west of the workings on the 950-level and have not been explored.

The ore minerals in the Pittsburgh-Idaho, as in most of the lead-silver replacement deposits (p. 89), were mainly secondary, chiefly cerussite, with lesser amounts of anglesite, smithsonite, hemimorphite, and cerargyrite in a gangue of earthy hematite, limonite, and manganese oxides. Primary sulfide minerals were rare in ore above the 700-level, occurring chiefly as scattered, small masses of galena rimmed with anglesite. Below the 700-level, increasing amounts of galena, pyrite, and sphalerite were present in the ore, and on the 950-level, the ore in the two small veins exposed was entirely galena, pyrite, and minor sphalerite and chalcopyrite, probably in a siderite gangue. The bottom of the oxidized zone is gradational, starting slightly above the 700-level—just below the level where abundant water was first encountered in the mine—and continuing downward to below the 800-level (Umpleby, 1913, p. 66–68, 98–103, for discussion of oxidation and secondary enrichment). The ore grade did not change significantly from the upper levels of the mine to the 700-level; some of the largest orebodies were between the 700- and 400-levels. Beneath the 700-level, the transition from oxidized ores to sulfide ores takes place, but the effect of the change on the ore minerals in the major veins is confused by displacement of the veins across the Neversweat and Gilmore flat, east-dipping faults. Comparatively little ore was mined from the zone of increasingly abundant sulfide minerals below the 700-level, and almost no ore was mined below the 800-level. The change from large, oxidized orebodies to thin, lean, sulfide-bearing veins takes place across the Neversweat and Gilmore faults and parallel, subsidary faults between them. The change seems likely to be more a reflection of faulting and displacement to the east, of the main ore-bearing structures above the Gilmore fault, than a reflection of changes in vein width and ore grade due to oxidation and secondary enrichment.

Latest Out mine.—The Latest Out mine (fig. 19, loc. 6), a lead-silver mine, is in the central part of the Texas district, just west of the Pittsburgh-Idaho mine, at an altitude of about 8,300 ft (2,530 m). The surface exposure of the Latest Out vein was discovered in 1880 (Umpleby, 1913, p. 104–106); a half interest in the prospect was purchased about 1889 by Ralph Nichols, reportedly for $300 and a barrel of whiskey (Oberg, 1970, p. 110). The prospect was partly developed through two inclined shafts and a shallow-level drift on the vein before 1889, and 1,200–1,500 tons of ore reportedly was shipped to the Nicholia smelter. It apparently was idle from 1889 until 1908, when Nichols purchased the remainder of the property and reopened it by driving a production adit from the east side of the Latest Out claim westward about 340 ft (104 m) to intersect one of the older inclined shafts on the vein (Umpleby, 1913, p. 104, for mine map, fig. 13). By 1911, the time of Umpleby's study, the mine had been deepened to about 410 ft (125 m), or to the 300-level beneath the production adit, and drifts on the vein had been extended from an inclined winze on the 100-, 200-, and 300-levels. These workings exposed the northeast end of what later was developed as the Main Vein oreshoot in the Latest Out mine, which here was 2–10 ft (0.6–3 m) thick and by 1912 had yielded perhaps 20,000–22,000 tons of ore, in addition to the early shipments made to the Nicholia smelter. The average grade was about 34 percent lead, 5 percent zinc, and 18 oz of silver per ton. A new shaft, inclined almost 60° west, was started about 1912, collared near the portal of the production adit. Most of the later mining and all of the deeper exploration were done from this shaft, which by 1913 had been deepened to the 400-level and by 1917 apparently had reached the 700-level, the deepest in the mine. The Latest Out 700-level is several hundred feet north of the Pittsburgh-Idaho Transportation Tunnel and about 150 ft (45 m) higher. The Transportation Tunnel later was connected to the Latest Out 600-level by raises. The mine finally included eight major levels at intervals of 50–150 ft (15–46 m), and four sublevels, connected by stopes on the major orebodies. It apparently was closed about 1926 and seems to have remained mostly idle, perhaps intermittently leased, until 1948 when it was leased to Milo Zook. Zook operated the mine from 1948 to 1952, mining principally on the 600-level. The mine was closed again in 1952 and has remained idle since then. The inclined production shaft now is caved at the collar, and the mine is inaccessible.

No complete production records are available for the Latest Out mine. The total production through 1913 seems to have been about 30,000–32,000 tons, judging from fragmentary reports in mining periodicals. In later years, the mine was credited several times with "handsome production"—presumably a well-rounded figure, but unspecified. Edgar C. Ross, in 1931 (footnote,

p. 84), credited the mine with a total production of about 90,000 tons of ore, having a value of about \$2.5 million, probably a reasonably close estimate. No records exist of any production from 1926 to 1948. From 1948 to 1952, about 500 tons of ore was mined, valued at about \$37,000. Representative ores from the 200-level and above (Nichols, 1913, p. 938) contained 35 percent lead, 7 percent zinc, 16 oz of silver per ton, and 0.025 oz of gold per ton. Representative ores from the 200-level to the 500-level contained 32 percent lead, 3 percent zinc, 12 oz of silver per ton, and 0.025 oz of gold per ton. Umpleby (1913, p. 65, 107) reported similar smelter analyses on ore from the upper levels of the mine: 34 percent lead, 5 percent zinc, and 18 oz of silver per ton. Ore from the deepest orebody in the mine, the Sandpit ore shoot, mined between 1948 and 1952, contained an average of 29 percent lead, 3.5 percent zinc, 0.3 percent copper, 13 oz of silver per ton, and 0.04 oz of gold per ton (Ralph Nichols, written commun., 1968).[9]

The lead-silver replacement veins and orebodies in the Latest Out mine are parallel to those in the Pittsburgh-Idaho mine, trending about N. 10°-15° E., and including both steep veins, dipping 70° or more W., and flat veins, dipping 40°-60°W. Three major orebodies were developed in the mine, the Main Vein ore shoot, a steep vein and the uppermost orebody in the mine; the Sandpit ore shoot, probably in the Cook flat vein near the bottom of the mine; and the East vein, a steep vein east of the other main orebodies. The Main Vein ore shoot was stoped from the 450-level to the surface, apparently along strike lengths as great as 200–250 ft (61–76 m) and thicknesses as much as 40 ft (12 m) above the 200-level; the shoot narrowed and thinned abruptly between the 200- and 450-levels. The Sandpit orebody was stoped from below the 600-level to the 450-level. It apparently was more than 200 ft (61 m) in strike length on and below the 600-level and perhaps as much as 40 ft (12 m) thick, and split into two somewhat smaller orebodies above the 600-level. The East vein orebody was stoped from the 600-level to just below the 200-level; its size is uncertain, but it clearly was a much smaller orebody than either of those in the other veins. The mine explored four principal steep veins from east to west, the East vein, East Split vein, Perpendicular vein, and Main Vein and a few other, shorter steep veins. It also explored five principal flat veins which are, from top to bottom, the Hatten Flat vein, East Flat vein, Cook Flat vein, Northwest Flat vein, and Latest Out Flat vein—and a few smaller flat veins and splits.

[9]Two "Ralph Nichols" are referred to in this report. The first was a mining engineer, manager of the Viola mine at Nicholia between 1882 and 1887, and owner of the Latest Out mine after 1908. The second, referred to here, is the grandson of the first, and is a vertebrate paleontologist and rancher.

Both Umpleby (1913, p. 106) and Nichols (1913, p. 937-939) described the Main Vein ore shoot as an irregular, lenticular replacement deposit filling a steep fissure, and extending in places into the adjacent walls. The other veins probably were similar. The deposits are in the same part of the Jefferson Formation as those of the Pittsburgh-Idaho mine, and the mineralogy of the ores also is similar. The Latest Out ore, even from the deepest orebody, consisted almost entirely of secondary minerals—cerussite, and lesser amounts of anglesite, smithsonite, hemimorphite, and cerargyrite, in a gangue of earthy hematite, limonite, and manganese oxides. Only small amounts of unoxidized sulfide minerals were present, chiefly small fragments of galena rimmed with anglesite.

The mineralized fracture zones in the Latest Out mine seem to have terminated at or near the flat, east-dipping Gilmore fault (p. 102); no exploration in the mine reached into the foot wall of the fault zone. The Latest Out orebodies, in the hanging wall, have been displaced eastward. Their possible extensions in the foot wall have not been looked for.

Allie (Martha) mine.—The Allie mine (fig. 19, loc. 7) is located immediately east of the Pittsburgh-Idaho mine in the central part of the Texas district. It explores the Martha vein, the only vein mined for gold in the Texas and Spring Mountain districts (Umpleby, 1913, p. 107). The mine apparently was first opened about 1908–1909, by a south-trending adit, the Martha tunnel. In subsequent years, until 1916, it was further developed by levels at 100 ft, 235 ft, 250 ft, 350 ft, and 400 ft, all connected by a winze from the Martha tunnel. The Martha vein was intersected by the Transportation Tunnel (p. 101), in 1912, at a depth of about 300 ft (91 m), and after 1912 was mined through the Transportation Tunnel. It had earlier been explored by a crosscut on the Martha 200-level from the Dorothy tunnel, east of the Martha tunnel, and by a crosscut, probably on the Martha 200-level, from the Pittsburgh-Idaho mine. The mine was inactive between 1916 and 1918. Between 1919 and 1920 it was reopened and mined, probably on the 300- and 350-levels, but it was apparently closed again from 1921 to 1939. The underground workings in 1921 totaled about 3,500 ft (1,066 m) in length. About 1939, a cyanide mill was installed to treat ore from the Martha vein, and the mine was reopened and deepened to the 500-level. The vein was mined between the 400- and 500-levels in 1940 and 1941. The last recorded shipments of gold-bearing concentrates were made in January 1942 (G. Grover Tucker, written commun., 1962). The mine has been idle since then and is inaccessible.

From 1910 through 1920, the Martha vein yielded about 13,000 tons of ore averaging about 0.6 oz of gold per ton and about 0.3–0.4 oz of silver per ton. The total gold contained in the ore was about 7,775 oz. The concentrates from ore mined in 1940–1941 yielded about 7,945 oz of gold, from ore that contained 0.2–1.86 oz of gold per ton and averaged about 0.4–0.5 oz of gold per ton, and also contained 0.3–0.4 oz of silver per ton. The Martha vein has thus yielded a total of about 15,720 oz of gold from ore that probably averaged about 0.5 oz of gold per ton. Some small pockets of ore contained as much as 25–30 oz of gold per ton, but the last ore mined, on the 500-level, contained only about 0.2 oz of gold per ton.

The Martha vein is in the Jefferson Formation, and trends about N. 10° E., and dips about 65° W. Umpleby (1913, p. 107) suggested that the uppermost part of the vein dipped 20° W., but most probably this dip was measured on the Dorothy flat vein, which is intersected in the Martha tunnel. The vein closely resembled the deeply oxidized lead-silver replacement veins of the Texas district, but contained more chalcopyrite than is typical of the lead-silver veins and almost no lead or zinc. Umpleby (1913, p. 107) described the ore as an earthy, iron- and manganese-stained mass showing casts of small pyrite crystals. On the 400- and 500-levels in the Allie mine, the ore contained free gold and pyrite in a gangue of hematite, limonite, and manganese oxides that was jaspery and siliceous in places. The ore occurred in irregular, lenticular shoots within the vein; a small shoot stoped between the 300- and 100-levels was probably about 75 ft (23 m) in strike length and up to 4.5 ft (1.4 m) thick; a shoot about 150 ft (46 m) in strike length and as much as 2–3 ft (1 m or less) thick was stoped between the 350- and 250-levels; and a larger shoot, perhaps about 200 ft (61 m) long and as much as 20 ft (6 m) thick was stoped from the 350-level to above the 200-level near the north end of the mine. The ore mined from between the 500- and 400-levels seems to have been in a separate shoot about 150 ft (46 m) long and as much as 6 ft (1.8 m) thick. The vein zone connecting the ore-bearing shoots typically was thin, barren, and filled with siliceous limonitic gangue.

Hilltop and Democrat mines.—The Hilltop mine (fig. 19, locs. 8, 9), a lead-silver-gold mine, is in the north part of the Texas district, at an altitude of about 8,380 ft (2,249 m) on the crest of the ridge at the head of Sourdough Creek (Ruppel and Lopez, 1981). The surface exposures of the Hilltop vein were discovered in the 1880's, and some ore from the mine, then known as the Alex Steven, reportedly was shipped to the Nicholia smelter. Small shipments of ore were made in 1901, 1918, and 1921, but major development of the Hilltop orebodies started about 1942, when the mine was opened by a shaft inclined 60°–70° west-southwest. By about 1955 the shaft had reached a total depth of 510 ft (155 m), and levels had been extended on the vein from the shaft at intervals of 75 to 95 ft (23–29 m), totaling more than 3,300 ft (1,005 m) of drifts and crosscuts. The deepest exploration was in a winze 70 ft (21 m) below the 500-level. The mine, idle from 1957 to 1959, was renovated in 1960; and a small amount of ore, totaling less than 100 tons, was shipped in 1961 and 1962, principally from exploration drifts on the 300- and 500-levels. Since 1962 the mine has been idle. The collar of the shaft is caved and the mine is inaccessible from the surface, but in 1980 the 500-level was reentered through a raise from the Democrat mine.

The Democrat mine is a new mine, started in 1963, to explore a group of lead-silver replacement veins east of the Hilltop mine, and to reach under the Hilltop mine to explore the Hilltop vein at depth. The portal of the mine is at an altitude of about 7,600 ft (2,013 m) in the upper part of the small basin of Sourdough Creek. From there a S. 50° W.-trending adit reaches about 3,000 ft (915 m), beneath the Hilltop mine, where rich lead-silver-gold ore was encountered in 1978 about 300 ft (91 m) beneath the Hilltop 500-level (Sandy Sims, oral commun., 1978). The orebody has been partly explored by drifts, stopes, a raise to the Hilltop 500-level, and, at greater depth, through a shallow winze and drill holes, but its extent is not yet known. These workings explore nine northwest-trending veins. The mine was idle in 1984. This Democrat mine is not the same mine as the Democrat mine reported in early mining periodicals and mentioned by Umpleby (1913, p. 108–109). The older Democrat mine is near the mouth of Sourdough Gulch (Ulrich Gulch in 1913), reportedly on the first claim staked in the Texas district in 1880.

The Hilltop mine has yielded about 9,000 tons of ore, not including the uncertain but certainly small amount of ore that was shipped from the Alex Steven mine to the Nicholia smelter. The ore contained an average of about 15 percent lead, 13.8 oz of silver per ton, 0.2 oz of gold per ton, 0.3 percent copper, and 0.9 percent zinc (A.J. Kauffman, Jr., U.S. Bureau of Mines, written commun., 1962). Most of the ore was mined in 1949–1953 from stopes between the 200-, 300- and 400-levels in the mine. The vein on the 500-level contained metallic minerals in places, but the small tonnage of ore produced after 1953 suggests that the orebodies developed at the higher levels had bottomed above the 500-level.

The Democrat mine has yielded about 2,100 tons of direct shipping ore that contained about 23.2 percent lead, 0.14 oz gold per ton, and 15.22 oz silver per ton[10]. Most of this ore was mined in 1979–1981 from stopes above the adit level, which is named the 800-level for

[10]Data on ore production and grade are from records of Democrat Resources, Inc., and are published by permission of that company.

its depth beneath the Hilltop mine. The relation of the veins explored in the Democrat mine to those in the Hilltop mine is unknown, because of faulting and because the intervening 300-ft-thick block between the 500-level in the Hilltop and the 800-level in the Democrat has not been completely explored.

The Hilltop vein trends about northwest, dips steeply southwest, and is a lead-silver-gold replacement vein that in places bulges out into replacement deposits in beds adjacent to the vein (P.N. Shockey, written commun., 1962). The vein and bedding replacements are in the steeply overturned west limb of a syncline in the lower part of the Jefferson Formation (Ruppel and Lopez, 1981) in most of the Hilltop workings—actually in the basal few hundred feet of the formation. As described by Shockey (written commun., 1962), the vein was explored for about 700 ft (213 m) along strike and 600 ft (183 m) down dip, and included two large ore shoots and several smaller ones. The principal ore shoot was a lenticular replacement in thin-bedded dolomite, about 50 ft (15 m) in strike length and as much as 6 ft (1.8 m) thick, that was stoped from the 400-level nearly to the surface. The other large shoot was similar but smaller and was broken into roughly equal parts by the Texas fault—a north-trending, 45°-65° west-dipping normal fault that displaced this orebody about 155 ft (47 m). The typical primary ore consisted of argentiferous galena, sphalerite, and very minor pyrite and chalcopyrite, in a gangue of jasperoidal hematite and white quartz. Most of the ore apparently contained some metallic sulfide minerals; secondary oxide and carbonate minerals were common throughout the mine, and are abundant in the parallel veins in the deepest workings in the subjacent Democrat mine. The secondary ore minerals are cerussite, anglesite, smithsonite, and malachite, in an gangue of earthy limonite and manganese oxide. Silver is more abundant in Hilltop ores than in most other lead-silver replacement veins in the Texas district, occurring in a ratio of about 1 oz of silver to 1 percent of lead. Gold is nearly as abundant in Hilltop ores as in the ore of the Martha vein (p. 104), about 0.2 oz per ton in Hilltop ore, and somewhat more abundant, about 0.5 oz per ton, in some ore from the Democrat mine at greater depth, although the typical gold content in Democrat ore probably is 0.08–0.1 oz per ton. As discussed on page 91, the pyrite and chalcopyrite-bearing, hematitic, siliceous, and in places jasperoidal gold-bearing ores seem likely to have been deposited in both the Martha vein and the Hilltop and Democrat veins in an episode of mineralization later than the episode of lead-silver mineralization, and perhaps after the lead-silver veins had been deeply oxidized.

No flat veins like those in the Pittsburgh-Idaho and Latest Out mines have been found in the Hilltop or Democrat mines. The similar geologic frameworks of the different mines and the similarities of the mineral deposits, suggest that flat veins are likely to be present, however.

Silver Moon mine.—The Silver Moon mine (fig. 19, loc. 10), a silver mine, is in Silver Moon Gulch in the south part of the Texas district at an altitude of about 7,450 ft (2,270 m). The silver deposit developed in the mine was discovered in the 1880's and reportedly shipped ore containing about 80,000 oz of silver to the Nicholia smelter (Umpleby, 1913, p. 108). In 1910–1912, the shallow, original workings were reopened and a vertical shaft was sunk to a depth of 70 ft (21 m), but no ore was found. In 1935–1937, the shaft was deepened to 140 ft (43 m), with levels at 90 ft (27 m) and 140 ft (43 m) that drifted on the vein. From 1936 to 1941, orebodies opened on these levels yielded several hundred tons of silver ore, some of it extremely rich. About 1939, the mine was deepened to 190 ft (58 m) through a winze from the 140 level, but no ore was found in drifts on the vein. From 1941 to about 1960 the mine was idle. In the 1960's, the mine was renovated and the orebodies on the 140-level were further explored. No ore is known to have been shipped at this time, and the mine has been idle since then.

The Silver Moon mine has yielded a total of about 140,500 oz of silver, probably from about 1,000 tons of ore. Of this total, 80,000 oz of silver was from an unknown amount of ore shipped to the Nicholia smelter, and 60,500 oz was from 460 tons of ore shipped during the period 1936–1940. Probably the ore produced between 1936 and 1940 is reasonably representative of the ores from the mine. It contained 15.5–223.2 oz of silver per ton, averaging about 117 oz per ton, and 0.0025–0.005 oz of gold per ton (although one shipment, in 1936, contained 0.1 oz of gold per ton), 1–36 percent lead, but most commonly about 3–6 percent lead, 0.15–2 percent zinc, and 0.15–0.3 percent copper. Most of this ore apparently was mined from a stope that reached from about 10 ft (3 m) below the 140-level to the 90-level, and was about 50 ft (15 m) in strike length and as much as 20 ft (6 m) thick. Specimen ores from the mine have contained 324–2,732 oz of silver per ton, but the ore shoots averaged about 117 oz of silver per ton, as noted above, and the veins away from the ore shoots commonly contain only 3–10 oz of silver per ton.

The Silver Moon mine explores four parallel veins that trend about north and dip about 65°–75° west. That part of the main Silver Moon vein described by Umpleby (1913, p. 108) as parallel to the east-dipping bedding of the enclosing carbonate rocks probably was a bedding replacement deposit in the crest of a small anticline adjacent to the vein. The veins are lead-silver replacement deposits in dolomite and limestone in the upper part of the Jefferson Formation and closely resemble the lead-silver replacement deposits elsewhere in the Texas and Spring Mountain districts. Most of the ore mined probably came from one or two shoots on the main Silver Moon vein, the westernmost of the four veins explored, and only small pockets of ore have been found on the other veins. The ore minerals are mainly secondary minerals, principally cerargyrite and lesser amounts of cerussite, smithsonite, hemimorphite, and rare galena, in a gangue of limonite, manganese oxides, clay gouge, and barite. Barite is a common gangue mineral in all other mines and prospects northwest of the Silver Moon mine, between Silver Moon Gulch and Liberty Gulch, but is not known to be present in the central part of the Texas district. The unusual abundance of silver in the Silver Moon ore suggests secondary enrichment.

The Silver Moon silver deposit is in the outer part of the mineralized aureole around the Gilmore stock. No other mines, significant prospects, or mineral deposits are known between Silver Moon Gulch and Lemhi Union Gulch, about 3 mi (5 km) farther south, where the Lemhi Union mine, related to the Spring Mountain granitic sheet, is the northernmost mine in the Spring Mountain district.

Jumbo mine.—The Jumbo mine (fig. 19, loc. 11) is in the north part of the Texas district, near the Hilltop and Democrat mines. The deposit was opened by a southwest-trending adit that intersects the Jumbo vein about 300 ft (91 m) from the portal; by an inclined winze, drift, and stopes on the vein; and by a winze from a short adit adjacent to the Jumbo adit, the Whiphouse winze. In all, the underground workings total about 1,500–1,600 ft (457–488 m). The mine has been inactive since 1916; the adit is caved and inaccessible. About 400 tons of ore was shipped from the mine before 1900, containing 37 percent lead, 48 oz of silver per ton, and about 0.2 oz of gold per ton (Umpleby, 1913, p. 108). No production is known from 1901 to 1909. From 1909 to 1916, the mine yielded 36 tons of ore that contained about 17.5 percent lead, 23.2 oz of silver per ton, 0.06 oz of gold per ton, and 0.4 percent copper (P.N. Shockey, written commun., 1962). The Jumbo vein trends about N. 60°–80° E., and dips 50°–65° S., and is in dolomite of the Jefferson Formation. The ore contained galena, cerussite, and probably other secondary minerals in a gangue of limonite, manganese oxides, calcite, and barite. Umpleby (1913, p. 108) described replacement deposits in dolomite adjacent to the vein as follows:

> Stringers of galena and cerussite following joint planes here and there merge into irregular lenses of ore which extend for several feet along bedding; in some places both ways from a given joint, in others in but one direction, either up or down the dip. Many of these lenses are made up of unaltered material, but some are composed entirely of secondary minerals.

Brown Bull mine.—The Brown Bull mine (fig. 19, loc. 12) is located at an altitude of 9,150 ft (2,789 m), about 0.8 mi (1.3 km) southwest of the Hilltop mine. The mine is in granitic rocks near their contact with the Kinnikinic Quartzite and explores several veins and the contact zone through a north-trending adit about 350 ft (107 m) long, and three appended drifts on veins. The total length of underground workings is about 700 ft (213 m). The history of the mine is unknown, but it seems likely to have been worked principally in the period from 1903 to 1912 and in 1917 when a small carload of lead-silver ore valued at $150.00 per ton was shipped. Since then it has been idle. The mine explores two east-trending quartz veins that dip 60°–80°N., parallel to both the jointing in the granitic rocks and the system of east-trending faults that break the Kinnikinic Quartzite-granodiorite contact south of the adit; a calcite-filled vein, parallel to the quartz veins; and the northeast-trending intrusive contact near the face of the adit. The quartz veins are 1–6 in. (2.5–15 cm) thick, heavily stained with limonite and manganese oxides, and enclosed in sheared and brecciated, hydrothermally altered granodiorite. Small stopes on the quartz veins suggest that they yielded a small amount of ore that contained galena and cerussite. The calcite vein contains 0.5–2 ft (15 cm–0.6 m) of white calcite, enclosed in sheared and altered granodiorite, but does not contain ore minerals. The contact of granodiorite and quartzite is a northeast-trending zone, 2–3 ft (0.6–0.9 m) thick, of clay gouge that contains quartz, sparse galena, and cerussite.

Several other small prospects in the vicinity of the Brown Bull mine explore similar deposits of galena and cerussite in thin quartz veins and in inclusions of marble in the granodiorite. Barite is a common gangue mineral in these prospects, but is not apparent in the gangue of the Brown Bull veins.

Portland mine.—The Portland mine (fig. 19, loc. 13) is an adit about 1,600 ft (488 m) long that explores the irregular contact of granodiorite and Kinnikinic Quartzite and a northwest-trending fault zone, at the north end of Portland Mountain. The portal of the mine, now caved and inaccessible, is at an altitude of about 9,500 ft (2,895 m). From the portal, the adit trends south and southwest for about 500 ft (152 m) and then curves to trend southeast to the face of the mine. Most of the adit is in granodiorite, but it closely follows the contact with the quartzite and in places crosscuts through spurs of quartzite where the contact is irregular. The southernmost 300 ft (91 m) of the adit is almost entirely in brecciated Kinnikinic Quartzite and clay gouge, along a northwest-trending fault zone that dips 70°–80° west. Thin quartz veins in the fault zone and in joints in the quartzite are heavily stained with limonite, manganese oxides, and malachite and azurite, and contain barite, as well as quartz, in the gangue. Only one shipment of ore from the mine has been reported—a single carload of 50 tons or less, in 1910, that is said to have contained 40 percent lead and 200 oz of silver per ton (Bell, 1911, p. 84). Most probably, this represents the total production from the mine, hand-sorted from small pockets of ore along the contact zone and in quartz veins like those in the Brown Bull mine.

Little Hill mine.—The Little Hill mine (fig. 19, loc. 14) is located southeast of the Hilltop mine, on the bedrock-cored east lip of a kettle in the broad sheet of moraine north of Meadow Lake Creek. The mine explored a northwest-trending, steeply west dipping lead-silver replacement vein in dolomite of the Jefferson Formation and yielded several hundred tons of oxidized lead-silver ore reportedly valued at about $200,000 (Ruppel and others, 1970, p. 21; Walker, 1924, p. 295–296). The mine includes shallow surface excavations, which yielded most of the ore, and two adits, one 125 ft (38 m) vertically below the other, with a raise from the lower adit to the surface. The lowermost adit is entirely in glacial gravel, and the mineral deposit was described by Walker (1924, p. 295–296) as being in a glacially transported block, completely enclosed by glacial gravels. The block was interpreted to have been carried a half mile (about 1 km) downstream by a glacier and deposited in the terminal moraine. The only dolomitic rocks upstream are in the Saturday Mountain Formation, however, and the Little Hill vein is in dolomite of the lower part of the Jefferson Formation, in a place where geologic mapping of the surrounding area suggests the lower Jefferson should be present (Ruppel and Lopez, 1981). An explanation more likely than Walker's is that the mine explores a vein in an ice- and melt-water-carved cliff buried beneath moraine, and that the lower adit was too far south to encounter bedrock.

Mountain Boy mine.—The Mountain Boy mine (fig. 19, loc. 15) is located at an altitude of about 8,560 ft (2,609 m) on the northeast flank of the moraine-covered ridge between Meadow Lake Creek and Portland Mountain. The mine apparently was first opened between 1905 and 1910, and was developed extensively between 1910 and 1916. It includes a lower working adit, probably more than 500 ft (152 m) long and trending about S. 15° E., an upper adit about 200 ft (61 m) vertically above the working adit, and a connecting raise with intermediate levels. The adits and drifts on the intermediate levels have an aggregate length of about 2,500 ft (762 m). Most of the ore known to have come from the Mountain Boy was mined between 1916 and 1918. One small shipment apparently was made about 1929. Since then, the mine has been partly reopened at times for further prospecting, but no ore is known to have been produced. The mine is caved and inaccessible.

The Mountain Boy mine developed lead-silver replacement veins in dolomite of the Saturday Mountain Formation. Two veins, one trending about north and dipping 65° W., and one trending about east and dipping 60° S., were explored; the principal orebodies reportedly were at the intersection of the two veins. The intersection was cut by the lower tunnel about 500 ft from its portal. The ore was almost completely oxidized and probably consisted almost entirely of cerussite, anglesite, smithsonite, and hemimorphite, with sparse galena if any, in a gangue of earthy limonite and manganese oxides. The only recorded shipments of ore were made between 1916 and 1918, and totaled about 570 tons of ore that contained an average of 24 percent lead, 6.5 oz of silver per ton, and 0.1 oz of gold per ton. The silver content ranged from about 2 to 10 oz per ton, but most commonly was near the average. The gold content ranged between 0.005 and 0.67 oz per ton, but most commonly was 0.05–0.1 oz per ton. The total production from the mine reportedly is more than 5,000 tons of ore of unknown grade, valued at about $200,000. The unrecorded ores, whatever their actual tonnage, were mixed with siliceous ores from other mines, to provide a higher iron content to the smelter flux.

Meadow Lake mine.—The Meadow Lake mine (fig. 19, loc. 16) is located on the south flank of the canyon of Meadow Lake Creek at an altitude of about 8,800 ft (2,682 m). The mine is a 300-ft (91-m) adit trending S. 50° E. in dolomite in the lower part of the Jefferson Formation. It explores several N. 20° W.-trending shattered zones, that dip about 45°–70° E., and multiple thin gouge zones along northeast-trending, 60°-northwest-dipping, small faults. The only mineralization appears to have been in a shattered zone about 25 ft (7.5 m) thick that was intersected in the adit 130 ft (40 m) from the portal, where raises reach 50 ft (15 m) above the adit. The mine is reported to have shipped lead-silver ore valued at $6,000–7,000 to the Nicholia smelter, its only known production. It is caved at the portal and inaccessible.

The mineral deposit explored in the Meadow Lake mine probably is similar to those explored in the Carrie Cody (fig. 19, loc. 17) and Murphy (Dixie Kay) (fig. 19, loc. 18) mines, opposite the Meadow Lake mine on the west wall of the canyon of Meadow Lake Creek (Ruppel and Lopez, 1981). Both of these mines explore northwest to nearly east trending, probably nearly vertical, lead-silver replacement deposits in dolomite of the Saturday Mountain Formation. The deposits include vein fillings and bedding replacements that contain cerussite, sparse galena, and pyrite in well-formed crystals, in an oxidized, siliceous gangue of earthy limonite, manganese oxides, quartz, jasperoidal and chalcedonic silica, and barite.

Lemhi Union mine.—The Lemhi Union mine (fig. 19, loc. 19) is located at an altitude of about 8,800 ft (2,682 m) in the north fork of Lemhi Union Gulch, about 5 mi (12.7 km) south of Gilmore. The mine was first developed through a vertical shaft in the 1880's, and reportedly shipped lead-silver ore valued at about $100,000 to the Nicholia smelter. The shaft had been deepened to 150 ft (46 m) by 1909, yielding several carloads of ore that were shipped to the Hahn smelter. The mine apparently was inactive between 1910 and 1920, but in the 1920's, a west-trending adit was driven about 1,800 ft (549 m) and, through a raise about 900 ft (275 m) west of its portal, connected to the original shaft. A number of short drifts were extended both north and south from the adit on small veins and stringers, and some ore reportedly was mined from the raise. The mine was partly reopened about 1951, but has been idle since then; it is caved and inaccessible. Known production between 1906 and 1951, from the mine and sorted from dumps, was no more than 150–200 tons of ore, that contained, at most, 23–39 percent lead and 7–20 oz of silver per ton. Ore sorted from a dump averaged about 5 percent lead, 1.4 percent zinc, and 1.6 oz of silver per ton.

The Lemhi Union mine explores a lead-silver replacement deposit in dolomite of the Saturday Mountain Formation. Replacement veins trend north to N. 25° E., and dip 40°–80° W., and typically are 1–6 in. (2.5–15 cm) thick. The veins widen in places, replacing adjacent beds of dolomite, to form small lenses of high grade ore. The principal ore minerals are cerussite, anglesite, smithsonite, hemimorphite, and probably cerargyrite, in an earthy gangue of iron and manganese oxides. No metallic sulfide minerals have been recognized in the deeply oxidized ores.

Elizabeth and Teddy mines.—The Elizabeth and Teddy mines (fig. 19, loc. 20) are near the top of Big Windy Peak, at the head of Spring Mountain Canyon (Umpleby, 1913, p. 89). The principal mine is the Teddy, a vertical shaft about 300 ft (91 m) deep. The Elizabeth is an east-trending adit that explores the same mineralized zone as the Teddy. The mineral deposit probably was discovered about 1880, but the earliest recorded ore shipments were in 1908, 1909, and 1910. The Elizabeth and Teddy and several other Spring Mountain mines supplied 80 tons of ore in 1909 and about 400 tons in 1910 to the Hahn smelter for its two brief runs. According to Umpleby (1913, p. 89) this ore contained about 20 percent lead, and 11 oz of silver per ton. The mine was idle from 1911 to 1917, when 150–200 tons of ore was shipped that contained 18–35 percent lead and 1 oz of silver for each 2 percent lead. The mines have been inactive since 1917, except for sporadic prospecting. Their total production is uncertain, but reportedly was valued at about $300,000, nearly one-third of the total value of ores mined in the Spring Mountain district.

The Elizabeth and Teddy mines explore lead-silver replacement deposits along a north-trending fracture zone in the basal part of the Saturday Mountain Formation (Ruppel and Lopez, 1981). The ore deposits have come from small, lenticular shoots, commonly 1–6 ft (0.3–1.8 m) thick, at the intersection of the fracture zone and dolomite beds favorable for replacement. The principal ore minerals are galena, cerussite, malachite and azurite, and associated silver, probably in cerargyrite, in an earthy, limonitic and jasperoidal gangue.

Iron Mask and Valley View mines.—The Iron Mask (fig. 19, loc. 21) and Valley View (fig. 19, loc. 22) mines are in the cirque at the head of Horseshoe Gulch, a tributary valley south of Spring Mountain Canyon. The Iron Mask is at the base of the cirque wall, at an altitude of 8,980 ft (2,737 m); the Valley View, appropriately named, is on the cirque wall at an altitude of 9,420 ft (2,871 m). Remnants of an aerial tram extend from the Valley View to the cirque floor; the tram apparently was built around 1902–1903. The properties were located in the 1880's, when the Iron Mask tunnel and some prospects on the surrounding cliffs were opened. The Iron Mask is reported to have shipped a small amount of ore in 1908, and both it and the Valley View may have contributed some ore in 1910 to the Hahn smelter. Most of the exploration and mining apparently was completed by about 1912, although the Valley View reportedly produced 6 tons of ore in 1937 and 38 tons in 1948. The mines were prospected sporadically until about 1950–1955. The Iron Mask is caved and inaccessible; the Valley View was open in 1976.

The Iron Mask tunnel extends southwest about 650 ft (198 m), and explores several lead-silver-copper replacement veins and adjacent bedding replacement deposits in dolomite of the Saturday Mountain Formation. The principal vein in the mine trends about northeast, dips 40°–55° NW., and is as much as 1.5 ft (0.5 m) thick. It is broken in a few places by north- to northwest-trending, 55°–60° west-dipping cross fractures that contain earthy, limonitic and jasperoidal gangue. The ore appears to have come from a few small, lenticular shoots as much as 6 ft (1.8 m) in diameter. The Valley View mine, a southwest-trending adit about 150 ft (46 m) long in dolomite of the Jefferson Formation, explores a northeast-trending, 40°–60° northwest-dipping vein similar to the Iron Mask vein. The vein zone is as much as 2 ft (0.6 m) thick, but only 4–6 in. (10–15 cm) on the hanging wall contained ore, principally as sparsely disseminated cerussite and small lenses of cerussite and galena in an earthy, limonitic, and jasperoidal gangue stained with malachite and azurite. The Valley View vein is cut off by a northeast-trending fault near the face of the mine.

BLUE WING DISTRICT

The Blue Wing district is a small mining district on the flanks of Patterson Creek, on the west side of the Lemhi Range near the community of Patterson (Umpleby, 1913, p. 109–112; Ruppel, 1980). The district includes a few small mines and prospects that explored copper-silver-bearing quartz veins, and one large mine, the Ima, that explored and developed a major deposit of tungsten-bearing quartz veins (p. 92). The only small mine known to have shipped any ore is the Castle Rock mine, which explored a group of east-trending quartz veins about 2 mi (3.2 km) east of the Ima mine, and is reported to have produced a small amount of copper-silver ore in 1913.

Ima mine.—The history of mining in the Blue Wing district is essentially the history of the Ima mine (fig. 19, loc. 24). Copper-silver-bearing quartz veins were discovered at the Ima as early as 1881 (Umpleby, 1913, p. 109; Callaghan and Lemmon, 1941; p. 4–6; Hobbs, 1945, p. 2–4), but active development of the prospects did not begin until 1900, when the Ima Consolidated Mining and Milling Company started exploratory work that continued until 1904. Tungsten minerals in the Ima veins were recognized in 1901 by Prof. J.E. Talmidge, but were not recovered until 1911, when the property was leased to the Idaho Tungsten Company. A 50-ton concentration mill was installed in 1912, and shipments of concentrates were made in 1913 and 1916; the 1916 shipment was reported to be 30 tons of huebnerite concentrates that contained 60 percent tungstic oxide obtained from ore that averaged less than 2 percent huebnerite and contained a variety of base-metal-sulfide minerals. This ore apparently came almost entirely from an inclined shaft near the portal of the lower of two adits that then explored the deposit, totaling about 2,000 ft (610 m) in length. In January 1918, the mine was closed. In 1921, the surface installations at the mine were rehabilitated by the Blue Wing Tungsten Mining and Milling Company, but no ore was mined. The mine remained inactive from then until 1934, except for a small amount of development work in 1927 and 1928. In 1934, the mine was taken over by the Ima Mines Corporation, and 2,300 pounds of concentrates averaging 56 percent tungstic oxide were produced. Development work continued in 1935, but no ore was milled. In 1936, about 800 tons of ore was mined and milled, and production was more or less continuous from then until 1957. The mine was closed in 1958. During this period, from 1936 to 1957, the Ima Mines Corporation, and after 1944 the Bradley Mining Corporation, mined in and extended the main workings on the north side of Patterson Creek. At the time the mine was closed, it included drifts and crosscuts on 12 levels and sublevels, the longest of which, the Zero-level, extended about 4,000 ft (1,220 m) northwest from its portal on the north wall of Patterson Creek. Much of the ore was mined below the Zero-level, reaching as deep as the 360- and 460-levels, which cut the top of the Ima stock (p. 65). Some also was mined above the Zero-level from a series of five short levels at about 100 ft (30 m) intervals, the "A" to "E" levels. The richest and most abundant ore was in the vicinity of the Water fault, a gently dipping, northwest-trending thrust fault that was crossed by the Zero-level about 1,600 ft (488 m) from its portal, and which offsets the Ima deposit perhaps as much as 1,000 m to the west. Some of these underground workings are shown on published maps of the Ima mine (Hobbs, 1945; Callaghan and Lemmon, 1941), made before the Water fault was found. Exploration and mining after 1945 extended the workings on the Zero-level and below to the northwest, across the Water fault, and explored above the fault on the "A" through "E" levels.

The total length of underground workings in the Ima mine is uncertain but must exceed 30,000 ft (9,144 m). The total production from the Ima mine has been about 280,000–310,000 short ton units of tungsten concentrates, and about 18,200 tons of sulfide concentrates, principally tetrahedrite and less abundant pyrite, sphalerite, chalcopyrite, and galena. These concentrates were recovered from about 800,000 tons of ore. The tungsten concentrates, mainly huebnerite, average about 65–70 percent tungstic oxide (Callaghan and Lemmon, 1941, p. 5). The concentrates of sulfide minerals contained about 42–60 oz of silver per ton, 4–6 percent copper, 4–7 percent lead, and 2–5 percent zinc (Hobbs, 1945, p. 11; Callaghan and Lemmon, 1941, p. 6). The total value of ores produced from the Ima mine is estimated to be \$14–15 million.

In addition to the mining and exploration in the Ima mine, exploration of adjacent prospects has been carried on intermittently. In 1938, the General Electric Company acquired the Miller group of claims, which are northwest of and more than 1,000 ft (305 m) higher than the Ima claims, and between 1940 and 1944, explored for tungsten deposits by means of a northeast-trending crosscut adit, the G.E. tunnel, more than 2,000 ft (610 m) long, by diamond drilling, and by numerous trenches (Hobbs, 1945, p. 4, pl. 1). Intercepts of veins in one diamond drill hole were later explored by Bradley Mining Corporation, in workings extended from the Ima mine deep beneath the General Electric claims; the exploration led to discovery of substantial tungsten ore in what seems to have been a continuation of the Ima vein zone. The south wall of Patterson Creek was explored for extensions of the Ima veins by the U.S. Bureau of Mines in 1941–1943, but the tunneling and drilling there did not disclose mineable bodies of ore.

Since 1957, the Ima mine has been reopened several times for further exploration and drilling, and the surface outcrops and soils have been extensively sampled for geochemical studies, particularly in the late 1960's by American Metals Climax, Inc., in a study of molybdenum associated with the Ima stock. The mine was open and being further explored by diamond drilling in 1981.

The geology and mineralogy of the tungsten-quartz veins and associated copper, silver, and molybdenum minerals in the Ima deposit are discussed on pages 92–95 of this report.

OTHER MINES AND PROSPECTS

Blue Jay copper mine.—The Blue Jay mine (fig. 19, loc. 25) is on the glaciated south wall of the canyon of Big Eightmile Creek (Ruppel, 1980), and explores a deposit of chalcopyrite and secondary copper minerals disseminated in the Big Eightmile stock. The deposit, first reported in mining periodicals about 1910, was explored underground by a south-trending adit about 1,000 ft (305 m) long in 1920–1921. The New Departure Copper Mining Company also constructed an aerial tram at this time, reaching from deeply oxidized outcrops of brecciated, copper-stained gossan, at an altitude of about 8,400 ft (2,560 m), to the valley floor. The copper deposit has been prospected repeatedly since 1921, and since 1960 has been further explored by drilling, apparently with discouraging results. Only the southern part of the potentially mineralized area has been explored. The northern part of the stock is largely concealed beneath glacial deposits, and the extent of copper mineralization remains unknown. The production from the Blue Jay mine is unknown.

Ray Lode mine.—The Ray Lode mine (fig. 19, loc. 26) is a small gold mine located south of the Big Eightmile stock, at an altitude of 8,950 ft (2,730 m) on the east wall of the valley of the east fork of Big Eightmile Creek (Ruppel, 1980). The mine includes three short east-trending crosscut adits and a shallow vertical shaft that intersect a N. 45° W.-trending, steeply west dipping limonite- and hematite-bearing quartz vein, and a drift along this vein for a short distance. The total length of underground crosscuts and drifts is less than 200 ft (61 m). All workings are caved and inaccessible. The mine is reported to have yielded a small amount of gold ore. The quartz vein is in a sheared and brecciated zone in quartzite of the Gunsight Formation. Other prospect pits east of the Ray Lode mine explore a parallel, limonite-stained sheared zone; a short adit near the ridge crest still farther east explores a N. 10° E.-trending, nearly vertical, limonite-stained, sheared and brecciated zone that contains thin, discontinuous quartz veins.

Prospects in Sawmill Canyon.—Only a few prospects are located in Sawmill Canyon, around the edges of the Sawmill Canyon granitic sheet, and farther north, near the Big Timber stock (Ruppel and Lopez, 1981). The Queen of the Hills prospect is located in the tributary canyon south of Squaw Creek, and explores a limonite-stained zone in the Saturday Mountain Formation. It may have yielded a small amount of lead-silver ore. Other, nearby prospect pits explore north-trending, brecciated fault zones in the Saturday Mountain Formation and Kinnikinic Quartzite. In the north-central part of the Gilmore quadrangle, near the mouth of the Smithie Fork of Sawmill Creek, several short adits and shallow pits explore flat-dipping veins of massive white quartz and barite in intensely sheared quartzite of the Gunsight Formation beneath an imbricate thrust fault. At the head of the east fork of Smithie Fork, a south-trending adit and several pits and trenches explore quartz veins in brecciated Kinnikinic Quartzite above an imbricate thrust fault. About 1 mi (1.6 km) southeast of Timber Creek Pass, a short adit, several pits, and a trench explore brecciated and limonite-stained dolomite of the Jefferson Formation. No metallic sulfide minerals or secondary minerals other than limonite are evident in any of these prospects, none of which are known to have yielded any ore, except perhaps the Queen of the Hills prospect.

MEASURED SECTIONS

Section 1.—*Saturday Mountain Formation*

[Measured on overturned limb of Rocky Peak anticline, east flank of Rocky Peak, Leadore quadrangle, Idaho, by E.T. Ruppel and M.H. Hait, Jr., 1960. Challis Volcanics consists of basaltic tuff breccia and bedded tuff, and overlies Saturday Mountain Formation with angular unconformity; contact of volcanic rocks and dolomite is probably about at top of dolomite as suggested by outcrops of Jefferson dolomite in valley of Grove Creek immediately north of measured section]

	Approximate thickness	
	Feet	*Meters*
Ordovician and Silurian		
Saturday Mountain Formation:		
56. Concealed	19.5	5.9
55. Dolomite, pale-grayish-red, pale-red- to pale-grayish-red-weathering, finely crystalline; sugary texture; in poorly defined beds 0.3-1 m thick. Reddish color may be a weathering phenomena	31.5	9.6
54. Dolomite, medium-light-gray; weathering same; finely crystalline, thick bedded to massive	39.3	12.0
53. Dolomite, medium-light-gray, light-gray or very light-gray-weathering, finely crystalline, massive to thick-bedded; weathered surface characterized by pits about 1/10 mm deep, 1 mm apart	15.0	4.6
52. Dolomite, medium-light-gray; medium-light-gray- to light-gray-weathering, finely to medium crystalline, locally mottled light-gray; in well-defined beds 0.15–0.6 m thick	17.4	5.3

	Approximate thickness Feet	Approximate thickness Meters
Saturday Mountain Formation—Continued		
51. Dolomite, medium-light-gray, medium-light-gray- to light-gray-weathering, finely to medium crystalline, locally mottled light-gray; sugary texture; mainly massive, but a few beds 0.15–0.3 m thick	44.1	13.4
50. Dolomite, medium-dark-gray, medium-light-gray-weathering, locally mottled light gray, thick-bedded to massive. Weathered surface characterized by cavities 3–7 cm long parallel to bedding; upper 0.6 m of unit is contorted laminated dolomite and intraformational conglomerate	17.3	5.3
49. Dolomite, medium-gray, medium-light-gray to light-gray-weathering, finely crystalline; massive; local irregular 1–7-cm-diameter vugs filled with brownish-orange, finely crystalline dolomite; 0.6-m-thick unit of 15-cm-thick beds in center of unit	11.3	3.4
48. Dolomite, medium-dark-gray, medium-light-gray-weathering, finely crystalline, in places mottled light-gray, massive to thick-bedded; contains 3–7-cm-long vugs parallel to bedding	45.2	13.8
47. Dolomite, medium-gray, medium-light-gray-weathering, finely crystalline, moderately fossiliferous, medium-bedded, mottled light-gray; contains white dolomite crystals in clusters as much as 2 cm in diameter	32.6	9.9
46. Dolomite, medium-gray, medium-light-gray-weathering, finely crystalline, thick-bedded to massive, moderately fossiliferous; contains white dolomite crystals in clusters as much as 2 cm in diameter	87.6	25.8
45. Dolomite, medium-gray, finely crystalline; mottled light gray, with mottles irregular in outline and typically less than 1 cm thick and 5–10 cm long; lower half of unit is thick bedded to massive; upper half of unit is in beds 0.15–0.6 to 2.0 m thick; increasingly fossiliferous toward the top of the unit, containing horn corals, hexacorals, and brachiopods. Weathered outcrop is vuggy; some vugs are lined with bluish-gray to white silica. (Fossil localities D–1044–CO and D–1045–CO)	91.4	27.9
44. Dolomite, medium-gray; weathering medium light gray with yellowish tint; finely crystalline; in beds 0.15–1.5 m thick; thinly laminated, crossbedded in some places	27.4	8.4
43. Dolomite, medium-light-gray, finely crystalline, thick-bedded to massive; contains oolitic bed 0.6 m thick about 1.2 m below top of unit	89.4	27.2
42. Dolomite, medium-gray, finely crystalline; mottled light gray, with mottles well defined, irregular, as much as 1 cm thick, 4–5 cm long. Sparsely fossiliferous, containing conspicuous horn corals and fossil fragments. Beds 1–1.5 m thick or massive; in places contains black chert in rounded nodules 1–2 cm thick, 5–10 cm long	65.4	19.9
41. Dolomite; sequence of thick-bedded to massive, light-gray dolomite faintly mottled medium light gray, containing abundant horn corals, much crinoidal material, sparse star-shaped algae(?) and sparse cephalopods, and containing black chert in rounded nodules about 4 cm in diameter; alternating with beds 0.3–0.6 m thick of medium-gray dolomite distinctly mottled light gray, sparsely fossiliferous, containing in places nodules of black chert 2–3 cm thick, 5–20 cm long; major part of unit is thick-bedded or massive dolomite. (Fossil locality D–1043–CO)	74.5	22.7
40. Dolomite, medium-gray, finely crystalline, massive; mottled shades of light gray, with mottles typically irregular 1 cm or less thick, 3–4 cm long; very fossiliferous, containing abundant horn corals and very abundant fossil fragments that form about 20 percent of the rock; contains light-gray, or less commonly, black chert in nodules that typically are 12–15 cm long and 3–5 cm thick	5.5	1.7
39. Dolomite, medium-gray, medium-light-gray-to light-olive-gray-weathering, finely crystalline, thinly laminated with laminae typically less than 1 mm thick; crossbedded	3.7	1.1
38. Dolomite, medium-gray, medium-light-gray-weathering, finely crystalline, in beds 0.1–0.5 m thick; mottled light shades of gray, mottles commonly less than 1 cm thick, 3–4 cm long sparsely distributed throughout unit. Fossil fragments are common thoughout the unit, and are almost the only constituent in several 0.1–0.5-m-thick beds—mainly fragments of star-shaped crinoid stems, corals common in upper 3 m of unit	48.8	14.9
37. Dolomite, interbedded medium-gray and medium-dark-gray, finely to medium crystalline; in beds 0.2–0.3 m thick	14.6	4.5
36. Dolomite, medium-gray; weathering medium light gray, with yellowish tint; finely crystalline, in beds 0.6–0.9 m thick, or massive; indistinctly mottled light gray on weathered surface; contains abundant fossil fragments in places	33.5	10.2
35. Dolomite, light-gray; in places tinted yellow on weathered surface; finely crystalline, in beds 0.1–0.3 m thick; upper 3 m of unit is medium crystalline	54.1	16.5
34. Dolomite, medium-light-gray, light-olive-gray-weathering; and fine-grained sandstone; thinly laminated, with laminae less than ½ mm thick; cross-laminated; in beds 0.2–0.5 m thick	15.5	4.7
33. Dolomite, medium-dark-gray, medium-gray-weathering, finely crystalline; crossbedded at base; contains abundant fossil fragments, many twiglike; contains lenses of contorted lighter gray dolomite that suggest intraformational conglomerate,		

	Approximate thickness Feet	Meters
Saturday Mountain Formation—Continued		
which typically are associated with thinly laminated beds of dolomitic sandstone; contains abundant horn corals; 0.3-m-thick dolomitic sandstone bed at top of unit	4.3	1.3
32. Dolomite, medium-dark-gray, medium-gray-weathering, finely crystalline; distinctly mottled light gray, with mottles irregular and parallel to bedding, typically 1 cm or less thick and 2–20 cm long; unit contains several cherty layers, and lenses typically less than 0.3 m thick and 3–4.5 long of medium-grained dolomite sandstone	11.0	3.4
31. Dolomite, medium-gray, medium-light-gray-weathering, finely crystalline; faintly and irregularly mottled shades of light gray with mottles commonly about 4 cm thick, 10 cm long; contains abundant fossil fragments; upper 3.7 m of unit contains black chert in irregular nodules 2–4 cm thick, 8–10 cm long; 1-m-thick distinctly mottled zone about 1.2 m below top of unit	64.8	19.8
30. Dolomite, medium-dark-gray, finely to medium crystalline, fetid; in beds 0.2–0.9 m thick; contains abundant black chert in irregular nodules 3 cm thick and as much as 10 cm long; contains abundant fossil fragments. Base of unit marked by thin band of black chert. Contact with overlying unit is irregular and appears to cut across bedding at a low angle	27.0	8.2
29. Dolomite, medium-light-gray, finely to medium crystalline, slightly fetid; in beds 0.2–0.3 m thick; contains abundant fossil fragments	3.5	1.1
28. Dolomite, medium-dark-gray to dark-gray, weathering medium gray to medium dark gray, medium-crystalline; in places mottled medium light gray; contains sparse fossil fragments	5.0	1.5
27. Dolomite, medium-gray; weathering same; finely crystalline; faintly and irregularly mottled light gray, in beds as much as 1 m thick; faintly wispy; contains sparse nodules of black chert as much as 4 cm thick, 10 cm long	5.3	1.6
26. Concealed; underlain by dolomite similar to unit below, but in beds as much as 1.5 m thick; platy parting not as pronounced as in underlying unit	24.7	7.5
25. Dolomite, medium-gray, light-gray-weathering, finely crystalline, wispy; in beds 0.1–0.3 m thick; locally contains thin layers of fossil fragments; platy parting pronounced in weathered outcrops	21.0	6.4
24. Dolomite, medium-gray, weathering light olive gray, finely crystalline; in beds 0.1–0.3 m thick; contains abundant chert in rounded, irregular nodules from less than 1 cm to 2 cm thick and from less than 1 cm to 6 cm long; contains abundant, very thin, irregular veinlets or wisps of white dolomite that are about parallel to bedding; chert is in thin bands 1–2 cm apart	20.9	6.4
23. Dolomite, medium-gray, finely crystalline, in beds 0.3–1.5 m thick; contains abundant fossil fragments, and in places contains abundant horn corals and brachiopods; characterized by abundant stringers and veinlets or wisps of white dolomite 0.5–1 mm thick, occurring in concentrations 2–5 mm thick, spaced at intervals of 5–15 mm. The wisps parallel to bedding are thin, and those perpendicular to bedding are thicker and do not have as pronounced wispy character. (Fossil localities D-1041-CO and D-1042-CO)	71.3	21.7
22. Dolomite, yellowish-gray to light-olive-gray, silty; in beds as much as 0.6 m thick; middle part of unit contains black chert in elongate nodules 1–8 cm long and as much as 2 cm thick; contains abundant branching, hairlike veinlets or wisps of white dolomite in bunches as thick as 2 mm	13.2	4.0
Lost River Member:		
21. Sandstone, dolomitic, and interbedded quartzite; sandstone is grayish orange to pale yellowish brown, fine to medium grained, faintly laminated in lower part and irregularly layered and crossbedded in upper part; near the base of unit, contains moderately abundant rounded fragments of earlier quartzite. Quartzite is medium gray, bluish tinted, fine to medium grained, in irregular lenses as much as 0.1 m thick, 0.6–0.9 m long; forms about 1 percent of the unit	11.3	3.4
20. Sandstone, quartzose, pale-brown, medium-grained; grains well rounded, well sorted, filmed with limonite; contains moderately abundant shale and mudstone chips and angular fragments as much as 3 cm in diameter, and sparse, rounded pebbles as much as 2 cm in diameter of white quartzite; unit includes 0.2 m of medium-light-gray, medium-grained, argillaceous sandstone 0.3 m above base	5.3	1.6
19. Shale and interbedded sandstone and quartzite; shale is dark gray, papery, in beds 3–27 cm thick, intricately interlayered with thin seams of medium-grained, well-sorted, limonitic quartzose sandstone in lenses 1–2 mm thick and 5–10 cm long; and medium-gray, blue-tinted, fine- to medium-grained quartzite in beds as much as 8 cm thick	1.5	0.5
18. Quartzite, medium-light-gray; pronounced bluish tint; fine- to medium-grained, clean; sand grains, well sorted, well rounded; beds 0.3–0.9 m thick in lower part of unit, 0.1–0.3 m thick in upper part of unit	10.0	3.0
17. Quartzite, light-olive-gray, fine-grained	0.2	0.06
16. Sandstone, medium-dark-gray, weathering dark gray, silty; with abundant streaks		

	Approximate thickness Feet	Meters
Saturday Mountain Formation—Continued		
Lost River Member—Continued		
of limonitic and carbonaceous sandstone in irregular layers 0.5-2 mm thick ...	3.5	1.1
15. Sandstone, similar to unit above, but not silty ...	1.5	0.5
14. Shale and mudstone, dark-yellowish-orange	3.0	0.9
13. Quartzite, medium-light-gray; bluish tint; fine to medium grained; sand grains well rounded and well sorted; in indistinct beds 0.4-0.9 m thick, in part crossbedded. Weathered outcrop irregularly streaked and blotched with Fe-oxide stain	14.5	4.4
12. Mudstone, pale-green to greenish-gray, blocky ...	0.25	0.08
11. Mudstone and shale, greenish-gray. Mudstone is blocky, in beds 8 cm thick separated by partings of paper shale 1.5-7 cm thick ...	2.0	0.6
10. Mudstone, dark-yellowish-orange	0.3	0.1
9. Sandstone, quartzose, light-olive-gray, rusty-weathering, medium-grained; sand grains well rounded and well sorted; in beds 1.5-9 cm thick separated by partings of light-olive-gray to grayish-orange paper shale typically 1-3 mm thick. Contains a small percentage of rounded grains of iron-oxide probably secondary after magnetite ...	2.0	0.6
8. Shale, pale-red, papery ...	0.1	0.03
7. Quartzite, heavily iron-stained, medium-grained; sand grains well rounded, well sorted, probably glauconitic ...	0.3	0.1
6. Shale, pale-red to grayish-red, very fissile, papery ...	0.2	0.1
Total thickness of Lost River Member	56.0	17.1
Total thickness of Saturday Mountain Dolomite ...	1,212.6	368.7
Ordovician Kinnikinic Quartzite (incomplete):		
5. Quartzite, very light gray to white, vitreous, medium-grained, very sparsely mottled, in beds 0.3-0.9 m thick; blocky fracturing	15.8	4.8
4. Quartzite, white to very light gray, vitreous; with abundant mottles of ferroan dolomite typically about 1 cm in diameter but some as large as 8 cm in diameter; includes a few lenses 0.9-1.5 m long and 0.03-0.6 m thick of quartzose sandstone cemented by ferroan dolomite; mottles are in bands about 3-7 cm apart; hackly weathering; weathered outcrop has honeycombed appearance ...	16.8	5.1
3. Quartzite, white to very light gray, vitreous; sand grains well rounded; abundantly mottled with 1 cm or less in diameter, rounded areas of iron-stained, ferroan dolomite-cemented quartz sandstone, that are increasingly abundant toward top of unit, and upper meter or so of unit is 20-25 percent mottles; platy fracturing in places 1-6 mm thick; weathers conspicuous moderate reddish orange ...	20.9	6.4
2. Quartzite, white to very light gray, medium-grained, well-sorted; in beds 0.5-0.9 m		
Ordovician Kinnikinic Quartzite (incomplete)—Continued		
thick, crossbedded in places; sparsely mottled near top of unit; blocky fracturing in blocks about 15 cm across	30.7	9.4
1. Quartzite, light-gray to white, weathers near moderate reddish orange, fine- to medium-grained, vitreous; in beds 0.3-0.6 m thick; in places conspicuously crossbedded; sparsely mottled with areas typically 0.5-3 cm across of ferroan dolomite-cemented quartz sandstone ...	9.5	2.9
Measured thickness of Kinnikinic Quartzite (partial)	93.7	28.6
Base of measured section.		

Section 2.—*Jefferson Formation*

[Measured on ridge south of Liberty Gulch, S½ sec. 19, T. 13 N., R. 27 E., Gilmore mining district, Lemhi County, Idaho, by Russell G. Tysdal, 1969]

	Approximate thickness Feet	Meters
Devonian Three Forks Formation:		
Yellowish-gray platy limestone, silty limestone, and siltstone.		
Devonian Jefferson Formation:		
Member 6:		
69. Covered, underlain by limestone sedimentary breccia for 60-90 more meters. Dip slope ... est.	250±	76±
68. Limestone breccia, similar to unit 64 ...	100	30.5
67. Limestone, similar to unit 65 ...	50	15.2
66. Limestone breccia, similar to unit 64 ...	30	9.2
65. Limestone, light-gray, thick-bedded, aphanitic ...	40	12.2
64. Limestone sedimentary breccia, medium-gray, weathers light gray ...	80	24.2
Approximate thickness member 6 ...	550±	168±
Member 5:		
63. Limestone, medium-gray to medium-light-gray and medium-dark-gray, thick-bedded to massive ...	103	31.4
62. Limestone sedimentary breccia, medium-gray to dark-gray, medium- and thick-bedded ...	2	0.6
61. Limestone, pale-yellowish-brown to light-gray, thin- and medium-bedded ...	15	4.6
60. Dolomite, similar to unit 56 ...	45	13.7
59. Sandstone, calcareous, pale-yellowish-brown, thin-bedded, fine-grained ...	5	1.5
58. Dolomite, similar to unit 56 ...	70	21.4
57. Limestone, medium-gray, medium- and thick-bedded ...	35	10.7
56. Dolomite, interbedded, dark-gray, medium- and thick-bedded; in units 0.6-3 m thick	33	10.1
55. Dolomite breccia, probably sedimentary, medium- and dark-gray, thick-bedded .	5	1.5
54. Dolomite, dark-gray, medium-bedded, sugary, finely crystalline, fetid, laminated	27	8.2
53. Sandstone, calcareous, pale-yellowish-brown, thin-bedded. Sand is quartz, medium grained, well rounded ...	7	21.1
52. Limestone breccia, sedimentary, probably redeposited ripped-up bottom clasts ..	3	0.9
51. Dolomite, medium-dark-gray, thick-bedded, fetid, sugary, laminated ...	8	2.4

	Approximate thickness Feet	Meters
Jefferson Formation—Continued		
Member 5—Continued		
50. Dolomite breccia, probably sedimentary	2	0.6
49. Concealed; float suggests similar to unit 48	45	13.7
48. Dolomite, interbedded, 10- to 20-ft-thick units similar to units 47 and 44	56	17.1
47. Dolomite, medium-gray, weathers light gray, medium-bedded, finely crystalline, sugary	4	1.2
46. Dolomite, dark-gray, medium-bedded, finely crystalline, sugary, fetid, laminated. Upper one-third includes interbedded sandy dolomite	175	53.4
Total thickness member 5	640	214.1
Member 4:		
45. Limestone, medium-gray, weathers light gray, medium-bedded, finely crystalline. Upper 3 m has interbedded sandy limestone, pale-yellowish-brown, thin-bedded; quartz sand consists of rounded medium grains	35	10.7
44. Dolomite, dark-gray, medium-bedded, sugary, finely and medium crystalline, fetid	5	1.5
43. Sandstone and dolomite interbedded. Sandstone, whitish, thin- and medium-bedded, fine- to medium-grained. Dolomite, medium-gray, medium-bedded, sugary, finely crystalline	60	18.3
42. Concealed	22	6.7
41. Quartzite, light-gray, medium-bedded, fine- to medium-grained	3	0.9
40. Concealed	10	3.0
39. Limestone, medium-gray, medium- and thick-bedded	70	21.4
38. Limestone breccia	30	9.2
37. Limestone, medium-gray, medium- and thick-bedded, interbedded with sandstone and sandy limestone	20	6.1
36. Sandstone, calcareous, moderate-red, thick-bedded	3	0.9
35. Dolomite, dark-gray, thick-bedded, sugary	2	0.6
34. Calcite dike. Rock on both sides of dike is brecciated	5	1.5
33. Limestone and sandy limestone, similar to unit 31	45	13.7
32. Dolomite, dark-gray, thin-bedded, fetid, sugary	20	6.1
31. Limestone and sandy limestone, light-gray to pale-yellowish-brown, and locally a very pale red, thin- and medium-bedded. Sand is quartz, well-rounded, medium to coarse	45	13.7
Total thickness member 4	375	114
Member 3:		
30. Dolomite, dark-gray, medium-bedded, sugary, fetid, vuggy; gives mottled appearance	25	7.6
29. Dolomite, medium-gray, medium-bedded, sugary, finely crystalline. Interbedded with dolomite, dark- gray, medium-bedded, finely crystalline, fetid, sugary. Units are 1.5–9.2 m thick	160	48.8
28. Limestone, sandy, pale-yellowish-brown, thin- and medium-bedded; quartz grains are medium to coarse and well rounded	6	1.8
27. Dolomite, dark-gray, thick-bedded, finely crystalline, sugary, fetid	5	1.5
26. Dolomite, similar to unit 24	15	4.6
25. Dolomite, similar to unit 23, with white-calcite hairline veinlets	24	7.3
24. Dolomite, medium-gray, thick-bedded, sugary, fetid, finely crystalline	21	6.4
23. Dolomite, dark-gray, medium- and thick-bedded, sugary, fetid, finely crystalline	95	29.0
22. Dolomite, light-gray, pale- and medium-brown, finely crystalline, sugary	10	3
21. Soil zone, light-brown-weathering, abundant calcite. Prospect pit. Probable shear zone or small fault, displacement probably negligible	6	1.8
20. Dolomite, interbedded 1.5–3-m-thick units of medium-gray and dark-gray dolomite similar to units 14 and 13. Sedimentary breccia, 1.8 m thick, in middle	130	39.7
19. Dolomite, similar to unit 14	10	3
18. Dolomite, similar to unit 13	10	3
17. Dolomite, similar to unit 14	33	10.6
16. Dolomite, shaly, pale-yellowish brown, thin-bedded	2	0.6
15. Dolomite, similar to unit 13	15	4.6
14. Dolomite, medium-gray, thin- and medium-bedded, finely crystalline, sugary	23	7
13. Dolomite, dark-gray, medium-bedded, finely crystalline, sugary, fetid	7	2.1
Total thickness member 3	597	182
Member 2:		
12. Dolomite, light-gray, thin- and medium-bedded, aphanitic	65	19.8
11. Sandstone, quartzose, fine-grained, calcareous and dolomitic, pale-yellowish-brown, medium-bedded	20	6.1
10. Dolomite, very pale yellowish brown, thin- and medium-bedded, aphanitic	10	3
9. Dolomite, dark-gray, medium- and thick-bedded, sugary, fetid, locally laminated. Unit contains much fossil material, including brachiopods and gastropods	40	12.2
8. Dolomite, similar to unit 6, interbedded with dolomite, medium-gray, weathering light gray, medium-bedded, finely crystalline	15	4.6
7. Dolomite, pale-yellowish-brown, medium- and thick-bedded	5	1.5
6. Dolomite, dark-gray, medium- and thick-bedded, sugary, fetid, finely to medium crystalline, laminated	20	6.1
5. Dolomite, sandy, light-brown to moderate-brown, medium-bedded; partly sedimentary breccia	5	1.5
4. Dolomite, medium-gray, weathers light gray, very finely crystalline, medium- and thick-bedded	5	1.5
3. Dolomite, dark-gray, medium- and thick-bedded, finely crystalline, sugary, fetid	40	12.2
Total thickness member 2	225	69

	Approximate thickness Feet	Approximate thickness Meters
Jefferson Formation—Continued		
Member 1:		
2. Sandstone, calcareous and dolomitic, yellowish-gray, thin- and medium-bedded	10	3
1. Cyclic bedded unit[11], includes pale-yellowish-gray, thin- and medium-bedded, dolomitic sandstone at base; medium-gray cyclic-bedded dolomite becomes conspicuous in formation 42.7 m above base	310	95
Total thickness member 1	320	98
Total measured and estimated thickness of Jefferson Formation	2,707	852

Contact of Devonian Jefferson Formation with Silurian Laketown Dolomite.

[11]See following measured section (no. 3, measured south of Latest Out Mine) for more detailed description of member 1.

Section 3.—*Jefferson Formation* (partial measured section)

[Members 1 and 2, measured on ridge south of Latest Out mine, S½ sec. 18, T. 13 N., R. 27 E., Gilmore mining district, Lemhi County, Idaho, by Russell G. Tysdal, 1969]

	Approximate thickness Feet	Approximate thickness Meters
Devonian Jefferson Formation:		
Member 2:		
Fault zone, top of measured section.		
60. Dolomite, light-gray to very light gray, aphanitic, thick-bedded	34	10.4
59. Dolomite, yellowish-brown, fine-grained, thick-bedded; contains sparse sand grains	7	2.1
58. Dolomite, light-gray, aphanitic, medium- and thick-bedded, thinly laminated; lower 0.3 m contains mud chips and is sandy	58	17.7
57. Sandstone, dolomitic, medium-light-gray to pale-yellowish-brown, fine-grained, thick-bedded, cross-laminated (same as unit 11 of Liberty Gulch section)	20	6.1
56. Dolomite, medium-gray, fine-grained, medium- and thick-bedded, sugary, fetid; top 0.3 m is flat pebble conglomerate	15	4.6
55. Dolomite, medium-gray, thick-bedded, irregularly laminated, sugary, locally cherty; upper 1.5 m sandy	14	4.3
54. Dolomite, medium-gray, very fine grained, medium- and thick-bedded	25	7.6
53. Dolomite, grayish-black, thick-bedded, laminated, sugary, fetid; upper half contains abundant fragments of fossil brachiopods and gastropods	31	9.4
52. Dolomite, medium-gray, medium-grained, medium- and thick-bedded, laminated, sugary, fetid	25	7.6
Measured thickness of member 2 (incomplete)	229	70
Member 1:		
51. Sandstone, pinkish-gray, medium-grained dolomitic, medium- and thick-bedded	4	1.2
50. Dolomite, medium-gray, medium- and thick-bedded, laminated, fetid	6	1.8
49. Concealed	4	1.2
48. Dolomite, medium-gray, very fine grained, thick-bedded	2	0.6
47. Concealed	4	1.2

	Approximate thickness Feet	Approximate thickness Meters
Jefferson Formation (partial section)—Continued		
Member 1—continued		
46. Dolomite, dark-gray, thick-bedded, laminated, sugary, fetid	2	0.6
45. Dolomite, pale-yellowish-brown, very fine grained, medium- and thick-bedded, thinly laminated	7	2.1
44. Dolomite, medium-gray, silty, medium- and thick-bedded, laminated; contains vugs lined with calcite	16	4.9
43. Dolomite, medium-gray, very fine grained, thick-bedded, laminated	6	1.8
42. Dolomite, dark-gray, medium-grained, sugary, fetid, laminated	2	0.6
41. Dolomite breccia	24	7.3
40. Concealed; zone of small faults that thin member 1 about 15–23 m	3	0.9
39. Dolomite, dark-gray, very fine grained, medium-bedded, thinly laminated, fetid	8	2.4
38. Concealed, probably underlain by sandstone similar to unit 35	3	0.9
37. Dolomite breccia	9	2.7
36. Dolomite, similar to unit 34	4	1.2
35. Sandstone, light-gray, dolomitic, fine-grained, medium-bedded; quartz grains well rounded	2	0.6
34. Dolomite, medium-dark-gray, thin- and medium-bedded and thinly laminated, very fine grained, fetid	13	4.0
33. Dolomite, similar to unit 23	2	0.6
32. Dolomite, similar to unit 24	1.5	0.5
31. Dolomite, similar to unit 23	4	1.2
30. Dolomite, similar to unit 24	1	0.3
29. Dolomite, similar to unit 24	1.5	0.5
27. Dolomite, similar to unit 23	10	3.0
26. Dolomite, similar to unit 24	2	0.6
25. Dolomite, similar to unit 23	5.5	1.7
24. Dolomite, light-gray, thin-bedded, silty; locally contains mud chips, partly laminated	1.5	0.5
23. Dolomite, dark-gray, thin- to thick-bedded, fetid, sugary, laminated	12	3.7
22. Dolomite, medium-dark-gray, medium-bedded, sugary	2	0.6
21. Concealed, probably similar to unit 19–20	20	6.1
20. Dolomite, dark-gray, medium-bedded, fetid, sugary, laminated, stromatolitic	2	0.6
19. Dolomite, medium-gray, very fine grained, medium- and thin-bedded, partly irregularly laminated	5	1.5
18. Sandstone, medium-gray, dolomitic, fine- to medium-grained	1	0.3
17. Dolomite, similar to unit 14, irregularly laminated, stromatolitic	7	2.1
16. Dolomite, very pale orange, medium-bedded; contains fine grains of quartz sand	5	1.5
15. Dolomite, medium-dark-gray, medium-bedded, very fine grained	4	1.2
14. Dolomite, dark-gray, medium-bedded, laminated, sugary, fetid, vuggy	2	0.6
13. Dolomite, medium-gray, medium-crystalline, medium-bedded, laminated, sugary, sandy	3	0.9
12. Dolomite, light-gray, thin-bedded, sandy	4	1.2

	Approximate thickness Feet	Meters
Jefferson Formation (partial section)—Continued		
Member 1—Continued		
11. Sandstone, pale-yellowish-brown to pale-reddish-brown, medium- to thick-bedded, cross-laminated	3	0.9
10. Sandstone, light-gray; composed of rounded, medium grains of quartz sand; calcareous and dolomitic, thin bedded	3	0.9
9. Dolomite, similar to unit 2	10	3.0
8. Sandstone, light-gray, thin-bedded, calcareous	1	0.3
7. Dolomite, similar to unit 2	5	1.5
6. Sandstone, light-gray to pale-red, fine- to medium-grained, calcareous, medium-bedded	4	1.2
5. Concealed, probably underlain by sandstone	7	2.1
4. Dolomite, similar to unit 2	2.5	0.8
3. Sandstone, similar to unit 1	1.5	0.5
2. Dolomite, medium-dark-gray, finely laminated, fetid, sugary, medium-bedded	5	1.5
1. Sandstone, light-gray, fine-grained, calcareous and dolomitic, medium- to thick-bedded; sand grains well rounded	3	0.9
Measured thickness of member 1, excluding fault zone	260	79
Silurian Laketown Dolomite:		
Light-gray, medium-crystalline, thick-bedded, vuggy dolomite	Not measured	

Section 4.—*Railroad Canyon Formation*

[Measured near Waugh Gulch in northeast corner of sec. 31, T. 17 N., R. 27 E., north of Railroad Canyon, Beaverhead Mountains, Lemhi County, Idaho, by E.T. Ruppel and H.B. McFadden, 1963]

	Approximate thickness Feet	Meters
Mississippian Bluebird Mountain Formation (incomplete):		
45. Quartzite, pale-red to light-brownish-gray, near-pinkish-gray-weathering, thick-bedded, fine-grained (0.2–0.3 mm), clean; sparse interbeds 12–30 cm thick of light-olive-gray, fine-grained, calcareous quartzitic sandstone that weathers grayish orange to dark yellow brown, with a honeycombed or fretted surface; contains minor chert and rock fragments		
Total measured thickness	30.0	9.1
Mississippian Railroad Canyon Formation		
44. Concealed. Float suggests is underlain by pale red to grayish-red platy to chippy silty limestone and yellowish-orange and yellowish-gray silty limestone	3.7	1.1
43. Mudstone, medium-dark-gray, medium-light-gray to pale-yellowish-brown-weathering, slightly calcareous, laminated, platy to papery	6.1	1.9
42. Concealed, probably underlain mainly by medium-gray and yellowish-gray thin-bedded (15–30 cm) limestone and sandy limestone; sandy rocks are thinly crossbedded	54.7	16.7
41. Limestone, medium-dark-gray and grayish-brown, medium-gray-weathering; contains abundant fossils	8.8	2.7

	Approximate thickness Feet	Meters
Railroad Canyon Formation—Continued		
40. Limestone, medium-dark-gray and grayish-brown, medium-gray, grayish-red, and yellowish-orange-weathering, very fine grained; contains abundant fossils and fossil fragments (Fossil collection 20221–PC)	21.5	6.6
39. Limestone, medium-gray, medium-dark-gray-weathering, thin-bedded and partly thinly crossbedded, very fine to medium-grained, fossiliferous	20.9	6.4
38. Limestone, medium-dark-gray, medium-gray-weathering, fine-grained, thick-bedded, sandy, silicic; contains abundant dark- gray chert lenses 3–10 cm thick, and as much as 1 m long in some beds in lower part of unit; fossiliferous (Fossil collections 20220–PC and 21930–PC)	64.3	19.6
37. Limestone, similar to unit 38 above, but with brownish-gray siltstone in interbeds about 10 cm thick increasing in abundance toward base of unit; limestone, thin bedded, fossiliferous (Fossil collection 21929–PC)	36.0	11
36. Siltstone and interbedded limestone, in alternate beds. Siltstone is brownish gray, weathers pale red to yellowish gray, in beds 15–30 cm thick, shaly. Limestone is medium dark gray, weathers medium gray, very fine grained, in beds 15–30 cm thick, partly laminated	91.8	28
35. Concealed	39.2	11.9
34. Concealed, probably underlain by medium-gray to brownish-gray to pale-red, very fine grained limestone	8.0	2.4
33. Limestone, medium-dark-gray, medium-light-gray- to grayish-orange-weathering, very fine grained, silty, fossiliferous	5.4	1.6
32. Concealed, similar to unit 36	37.0	11.3
31. Limestone, moderate-brown, pale-yellowish-brown-weathering, very fine grained, massive, sparsely fossiliferous (Fossil collection 21928–PC)	28.0	8.5
30. Conglomerate, very dusky purple, dark-greenish-gray to grayish olive-weathering; composed of small pebbles of limestone, in sand and silt matrix	10.2	3.1
29. Shale, brownish-gray and grayish-black, papery	6.6	2
28. Limestone, medium-light-gray and medium-brownish-gray, grayish-orange-weathering, fine-grained	6.6	2
27. Limestone, light-brownish-gray to pale-brown, medium-gray-weathering, very fine grained, massive, fossiliferous (Fossil collection 21927–PC, from lower part of unit)	40.3	12.3
26. Mudstone, shale, and limestone. Mudstone, olive-gray, light-olive-gray-weathering, shaly to flaggy. Shale, brownish-gray and grayish-black, papery. Limestone, olive-gray, grayish-orange- to moderate-yellowish-brown-weathering, very fine grained, massive, lenticular	17.3	5.3

	Approximate thickness Feet	Meters
Railroad Canyon Formation—Continued		
25. Conglomerate, medium-gray to medium-dark-gray, medium-gray-weathering; composed of well-rounded limestone pebbles and calcite sand, with some pebbles as large as 5 mm in diameter but most 3 mm in diameter or less; calcite matrix; contains abundant fossil fragments (Fossil collection 21926–PC)	23.5	7.2
24. Limestone, pale-red to pale-yellowish-brown, pale-red to grayish-orange-weathering, very fine grained, massive, muddy	10.7	3.3
23. Limestone, pale-red, pale-grayish-red-weathering, massive; contains abundant fossils and fossil fragments, which possibly include *Inflatia* and *Antiquatonia*	2.1	0.6
22. Limestone, medium-dark-gray, pale-grayish-brown-weathering, very fine grained, massive, fossiliferous (Fossil collection, no USGS collection number)	3.2	1
21. Limestone, dark-olive-gray, light-olive-gray and pale-red-weathering, very fine grained, massive, platy to flaggy, muddy; contains abundant hairlike veins of white calcite about 0.5 mm thick	9.6	2.9
20. Limestone, medium-gray, yellowish-gray-weathering, fine-grained, massive	2.6	0.8
19. Limestone, similar to unit 21, but fresh color is olive black in upper part of unit	62.4	19.0
18. Shale, brownish-gray, moderate-grayish-brown-weathering, calcitic, papery to shaly	5.7	1.7
17. Limestone, similar to unit 21	2.4	0.7
16. Mudstone, olive-gray, light-gray-weathering, flaggy to shaly	9.3	2.8
15. Limestone and mudstone; limestone is moderate to dark yellowish brown, yellowish-gray-weathering, fine grained, laminated, very fossiliferous (Fossil collection 21925–PC); mudstone is grayish black, yellowish gray weathering, thick bedded	3.3	1.0
14. Shale, brownish-black, brownish-black to dark-gray-weathering, partly laminated, papery; laminated parts are calcareous	19.9	6.1
13. Limestone, brownish-gray, light-olive-gray to dark-yellowish-orange-weathering, very fine grained	0.3	0.1
12. Shale, similar to unit 14	12.1	3.7
11. Mudstone, brownish-gray, dark-greenish-gray-weathering, calcareous, laminated	6.7	2.0
10. Shale, similar to unit 14	15.7	4.8
9. Limestone, moderate-yellowish-brown, near-dark-yellowish-orange-weathering, very fine grained, laminated	0.8	0.2
8. Shale, similar to unit 14	8.9	2.7
7. Mudstone, grayish-red-purple, moderate-reddish-brown to moderate-reddish-orange-weathering, shaly to flaggy; gradational into unit 8	14.2	4.3
6. Shale, pale-brown, papery, very poorly exposed; gradational into unit 7	46.4	14.2
5. Mudstone, grayish-red-purple, pale-red-purple-, moderate-reddish-orange-, and grayish-orange-weathering, laminated, partly cross laminated, shaly	53.8	16.4
4. Mudstone, moderate-yellowish-brown, dark-yellowish-orange-weathering, calcareous; gradational into unit 5	16.8	5.1
3. Mudstone, grayish-red-purple, dusky-red-purple-weathering, flaggy to slabby, slightly calcareous, laminated, partly cross laminated; gradational into unit 4	7.0	2.1
2. Mudstone, pale-yellowish-brown, grayish-orange-weathering, calcareous, shaly to flaggy; gradational into unit 3	6.0	1.2
1. Mudstone, grayish-red and yellowish-brown, shaly to flaggy	4.0	1.2
Base of Railroad Canyon Formation. Contact concealed and inferred from float. Float suggests rock beneath mudstone of unit 1 is medium-dark-gray massive limestone of Mississippian Scott Peak Formation.		
Total measured thickness of Railroad Canyon Formation	853.8	259.5

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