Bedrock Geology of the Vipond Park 15-Minute, Stine Mountain 7½-Minute, and Maurice Mountain 7½-Minute Quadrangles, Pioneer Mountains, Beaverhead County, Montana

U.S. GEOLOGICAL SURVEY BULLETIN 1625

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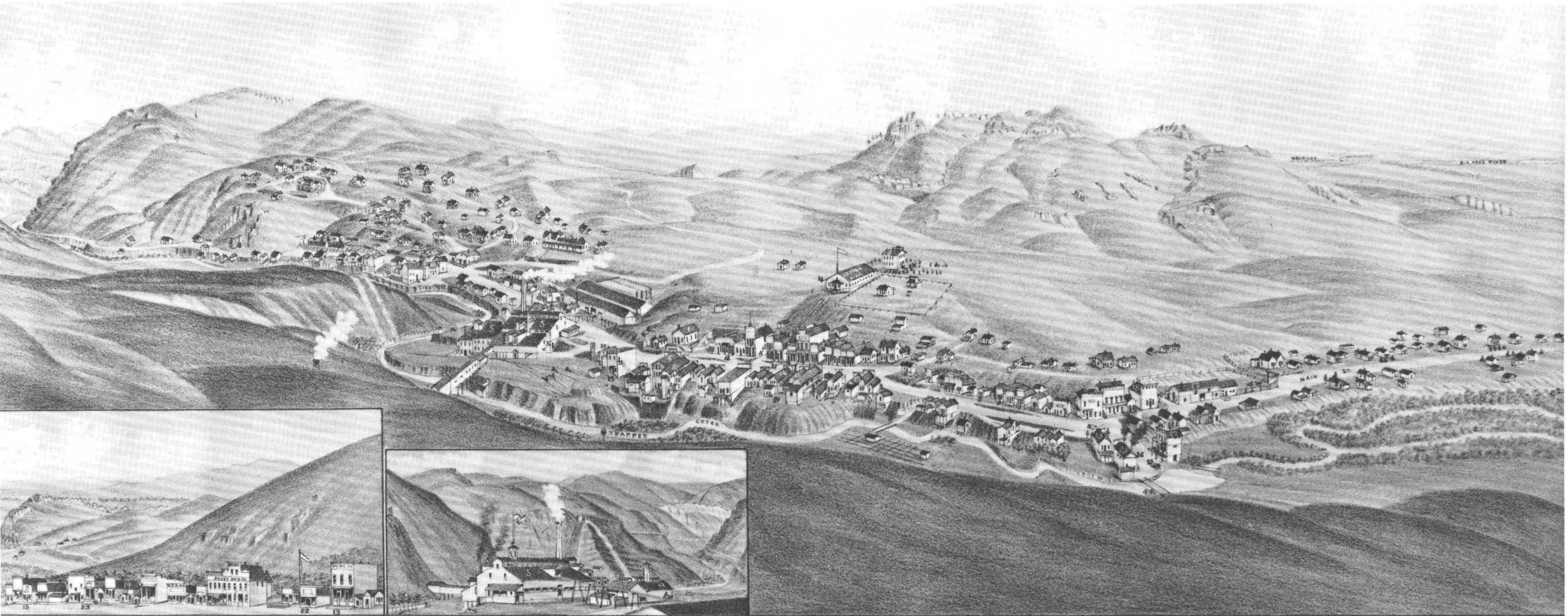
BEDROCK GEOLOGY OF THE VIPOND PARK 15-MINUTE,
STINE MOUNTAIN 7½-MINUTE, AND MAURICE MOUNTAIN
7½-MINUTE QUADRANGLES, PIONEER MOUNTAINS,
BEAVERHEAD COUNTY, MONTANA



J. J. STONER Madison Wis.

BECK & PAULI, Litho. Milwaukee, Wis.

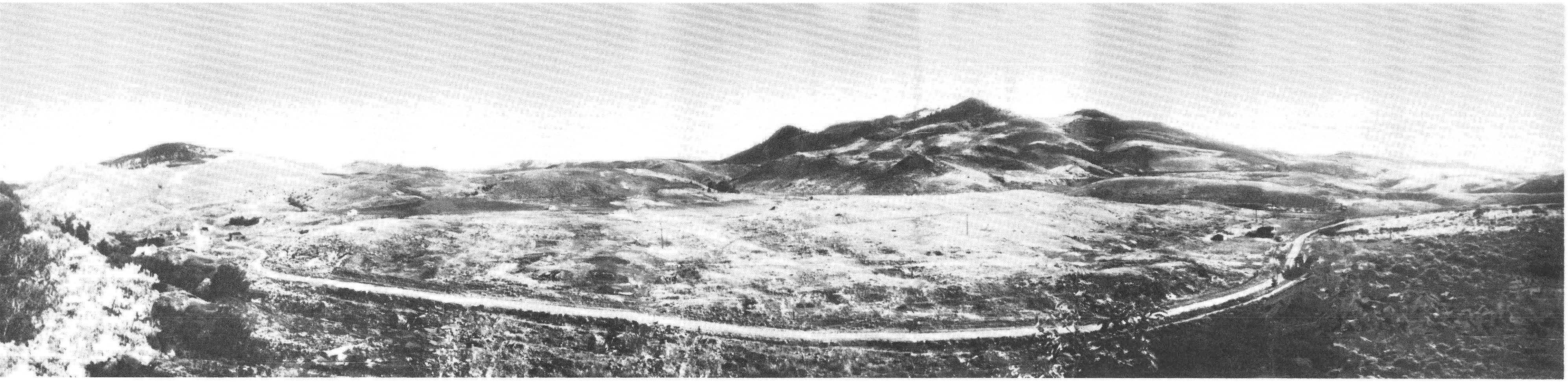
BIRD'S EYE VIEW OF

GLENDALE, MON.

BEAVER HEAD CO.

1883.

1 Methodist Church.
2 Baptist Church.
3 School House.
4 Skating Rink, Mahan & Howe.
5 Hecla Consolidated Mining Company's Smelting Works, H. Knippenberg, General Manager.
6 Glendale Brewery, L. Heinboekel.
7 Bank, N. Armstrong & Co.
8 Avery House, H. H. Avery, Prop.
9 Armstrong & Losee, Gen'l Mdse.
10 H. Schmalhausen, Physician and Surgeon.
10 A. G. Noble, " " "
12 J. C. Keppler, Jewelry and Silverware.
13 Wilson Rote & Co., General Merchandise.
14 Wm. J. Parkinson, Mint Saloon.
15 M. Goldberg, Harness, Boots and Shoes.
16 Narcisse Ledoux, Saloon.
17 F. Murray, Livery and Feed Stable.
18 Ponet & Generaux, General Merchandise.
19 E. F. Alward, Drugs, Books, Stationary, &c.
20 Bateman & Wells, Saloon.
21 J. C. Welch, Books, Notions, &c.
22 Stager House, E. O. Hulsizer, Prop.
23 Peter Wagner, Saloon and Billiard Hall.
24 R. Z. Thomas, Notary Public and Justice of Peace.
25 J. R. Reynolds, Livery, Feed and Sale Stable.
26 Avare & French, Blacksmith and Wagon Shop.
27 H. Stuart, Furniture, &c.



Glendale, Mont., in 1883 (top). This old drawing of Glendale, viewed from the southeast, shows the town in more thriving days. Center of activity then was the smelter (left center) that processed the ore material mined at Hecla and vicinity. Only the chimney, a few shells, and cellar holes of buildings remain of the town. Trapper Creek valley extends from upper left to lower right in the perspective. The foreground is underlain by Cretaceous sedimentary rocks of the Colorado Group. The hills in the upper right part of the picture are south of Dry Hollow Gulch; from left to right the peaks against the skyline are at 6,500 ft, 6,700 ft, and 6,628 ft elevation on the Vipond Park topographic quadrangle map. These hills are underlain by the Upper Paleozoic Amsden, Quadrant, and Phosphoria Formations in fault-bounded blocks. The hills to the left of town are part of the complex anticline cored by the Amsden Formation; the four readily identified peaks, from left to right, are the 6,484-, 7,493-, 6,674-, and 6,760-foot peaks on the Vipond Park topographic map. The valley between the two groups of hills is the Trusty Gulch syncline; the sharp knobs in the center of the valley are on either side of Canyon Creek valley and are underlain by the Quadrant Quartzite. The original photograph of this drawing is in the Beaverhead Museum in Dillon, Mont. Photograph shown at bottom is of the same area, taken in 1983.

Bedrock Geology of the Vipond Park 15-Minute, Stine Mountain 7½-Minute, and Maurice Mountain 7½-Minute Quadrangles, Pioneer Mountains, Beaverhead County, Montana

By E-AN ZEN

A description of the stratigraphy, structure, and igneous rocks of a three-quadrangle area at the north end of the Pioneer Mountains

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DEPARTMENT OF THE INTERIOR
DONALD PAUL HODEL, Secretary

U.S. GEOLOGICAL SURVEY
Dallas L. Peck, Director

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Bedrock Geology of the Vipond Park 15-Minute, Stine Mountain 7 1/2-Minute, and Maurice Mountain 7 1/2-Minute Quadrangles, Pioneer Mountains, Beaverhead County, Montana

By E-an Zen

Abstract

The Vipond Park, Stine Mountain, and Maurice Mountain quadrangles, which occupy an area of about 810 km^2, are located in the northern part of the Pioneer Mountains in Beaverhead County, Mont. Rocks underlying the area include Early Proterozoic gneiss and amphibolite; Middle Proterozoic sedimentary sequences; Cambrian(?) clastic rocks; Cambrian clastic and shelf-facies marine carbonate rocks; Devonian, Mississippian, Pennsylvanian, and Permian marine rocks of the shelf facies; Triassic marine rocks; and Cretaceous rocks that pass from deltaic beds into fluviatile beds deposited in a braided river system. Tertiary rocks are mainly volcanic and subaerial. The Cambrian(?) rocks are conglomerate and coarse quartzite deposited in local troughs and aggregate at least 500 m thick; these grade upward into more mature, finer grained, and more calcareous sediments that were deposited as the troughs filled. By late Middle to Late Cambrian time, a carbonate shelf was well established on which the Hasmark Dolomite was deposited. Ordovician through Middle Devonian rocks are not present in the area. Likewise, Late Triassic through Late Jurassic (pre-Kootenai) rocks are absent. The Amsden (Upper Mississippian) and Kootenai (Cretaceous) Formations both reflect in their lithology the deposition of deeply weathered material. By later Cretaceous (Colorado Group) time, immature sediments were supplied by a landmass formed by Sevier or early Laramide orogeny; the Colorado Group is a synorogenic sedimentary unit. Tertiary volcanic rocks are Eocene and Oligocene and are subaerial flows of calc-alkalic lavas and minor ash. Upper Tertiary river gravels, Quaternary glacial deposits, and the Holocene Mazama ash complete the sedimentary record.

The rocks were faulted, tilted, and probably locally folded in Late Cretaceous time, during and shortly after the deposition of the Colorado Group. This tectonic event was followed in Campanian time by the arrival from the west of thrust sheets of Middle Proterozoic sedimentary rocks that overrode the earlier structures. Erosion of the thrust sheets produced sediment which became part of the Beaverhead Formation preserved east of the map area. Arrival of the thrust sheets terminated the sedimentary cycle by covering up the source area and filling the basins. Early Proterozoic gneiss and amphibolite may be part of the thrust complex, but there are not enough data to prove such relations.

The area is cut up by systems of high-angle faults. One large group of faults trends west-northwest and parallels a series of truncated folds that trend in the same direction; these include the major Cherry Creek syncline that contains Middle Proterozoic rocks in the Morrison Hill klippe. Strike-slip movements, which caused some of the observed juxtaposition of Phanerozoic sedimentary rocks, are important on one or more of these faults. A second group of high-angle faults trends north-northeast; these are most important in the western part of the area. One fault of this group, the Fourth of July fault, has a west-side down-movement sense. To the west of the fault, allochthonous rocks of Middle Proterozoic age dominate, whereas the same units to the east are only locally preserved in klippen. The Wise River valley is a graben in this system.

Intrusive rocks of the Pioneer batholith dominate the southern and western parts of the area. This composite batholith consists of many plutons of different ages. The early phases (Late Cretaceous) are calc-alkalic and range from quartz diorite, tonalite, and granodiorite, to granite. The later phases (earliest Paleocene) are two-mica granite and leucogranite; associated with this latter rock are quartz-sericite veins that locally contain molybdenite of economic value. The time span of intrusions included the time between the deposition of the Colorado Group directly above their roof, and during and after the emplacement of the thrust sheets. Intrusion was at shallow levels of a few kilometers and was at least locally (outside of the map area) accompanied by volcanism.

The Pioneer batholithic rocks are highly radiogenic in terms of the initial strontium ratios; two separate magma series among the various plutons can be recognized by this ratio. The association of molybdenum mineralization with the Paleocene intrusions probably reflects a later stage of magma differentiation rather than a difference in source magma. The highly radiogenic nature of the rocks distinguishes them from the Boulder batholith, which has the same age range and whose sodic series is otherwise chemically similar to the calc-alkalic rocks of the Pioneer batholith. Generation of the magma is probably related to Laramide tectonic activity at the western margin of the North American craton. The hypothesis here advanced calls for partial subduction of sialic crust, already homogenized in lead isotopes but differentiated in its Rb-Sr system through partial fusion, under the margin of the North American craton; partial fusion of the subducted crust led to the formation of the Cretaceous-Paleocene magmas. Subduction, however, was probably inefficient because the crust was sialic; instead, the bulk of the crust was transported northward by transform faults prior to the Cretaceous accretion of the terranes now found west of the Idaho batholith. The Middle Proterozoic sedimentary rocks that now occur in thrust sheets in western Montana and adjacent Idaho may have been transported here on this postulated removed (decreted) terrane as its sedimentary cover rock.

The Early Proterozoic gneiss and amphibolite exposed in the Vipond Park quadrangle could be a record of this terrane preserved by oblique obduction onto the craton.

Economic mineral deposits in unmetamorphosed sedimentary sequences include phosphate ore of the Phosphoria Formation and minor manganese deposits, preserved at the base of the Kootenai Formation as pre-Kootenai weathering residue. Major economic resources are base and noble metals in quartz veins associated with intrusive rocks (molybdenum mostly), as contact skarn (wolfram) of the Amsden Formation, and as skarn and hydrothermal vein deposits in Paleozoic carbonate rocks, especially the Cambrian and Devonian dolostones (Hasmark and Jefferson Dolomites). Metals extracted from the skarn and hydrothermal deposits include gold, silver, lead, zinc, and copper. Lead and sulfur isotopic studies show that the deposits in the carbonate rocks have Mississippi-valley characteristics, whereas those near and within the intrusions have more igneous character. Remobilization of preexisting disseminated metals in the sedimentary units, as a result of hydrothermal circulation initiated by the intrusion, is a possible explanation of the observed relations. If this remobilization occurred, the circulation mixed metals of different sources very inefficiently, and a short distance from the intrusive contacts the main effect of the intrusion was thermal.

INTRODUCTION

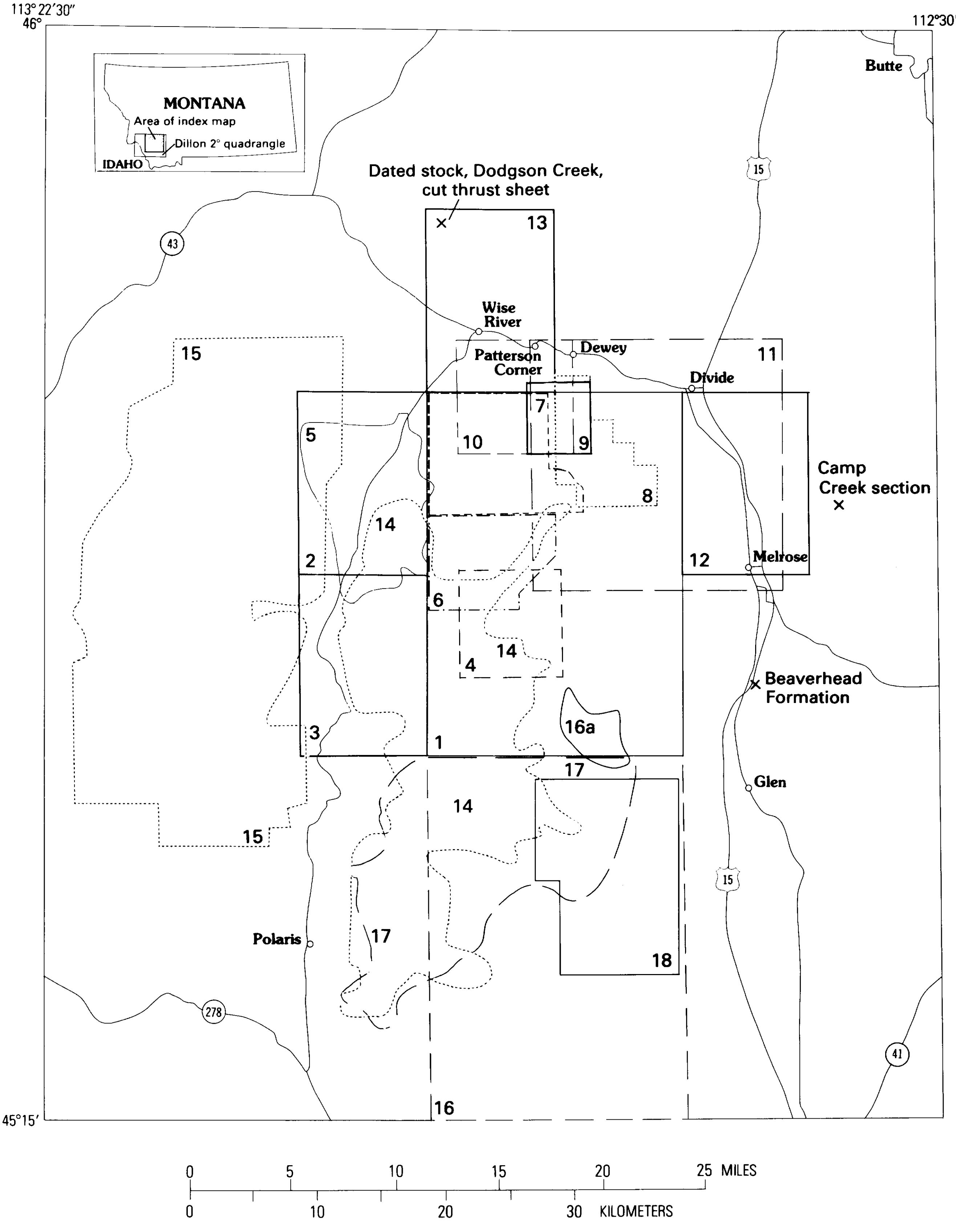
The Vipond Park 15-minute quadrangle, the Stine Mountain 7½-minute quadrangle, and the Maurice Mountain 7½-minute quadrangle occupy contiguous areas in the northern half of the east Pioneer Mountains and the northeastern part of the west Pioneer Mountains in Beaverhead County, southwestern Montana (fig. 1). Most of the area is in the Beaverhead National Forest. Some of the area is administered by the U.S. Bureau of Land Management, and small parts are State and private land.

Total relief in the area is about 1.6 km, or a mile, the highest point being the summit of Granite Mountain (10,633 ft, or 3,240 m above sea level) and the lowest point being the Big Hole River (about 5,400 ft, or 1,650 m). Local relief is considerable in the glaciated part of the high range underlain by intrusive rocks. A diagonal line drawn from the northwest to the southeast corners of the Vipond Park quadrangle divides the unforested areas into a part to the northeast that is mainly open rangeland and a part to the southwest that is mainly above timberline. Access to the area by car is generally good, especially with a small four-wheel-drive vehicle. Access to the main part of the range in the south-central part of the area is more limited, and foot trails are the main avenues. The jeep and pack trails shown on the Vipond Park topographic sheet, edition of 1958, contain numerous minor to serious inaccuracies and should not be relied upon. Jeep trails made after 1958 provide easy access, especially in the Cannivan Gulch to Gray Jockey Peak area. The arduous access to the higher parts of the area is more than compensated for by the geologic challenges, magnificent scenery, wildflowers, wildlife, and presence of untrampled land.

Geologic mapping in the Vipond Park quadrangle was much more detailed than in the Stine Mountain and Maurice Mountain quadrangles. Those two areas were mapped primarily for a RARE II (Second Roadless Area Review and Evaluation) report (Pearson and Zen, 1985) and a wilderness study area report (L.W. Snee and E-an Zen, unpublished data). Because regional interpretations of tectonic relations and construction of structural cross sections would be greatly aided, these two quadrangles are reported with the Vipond Park quadrangle even though the mapping in these areas was considerably more sketchy. Where geologic relations of the Maurice Mountain quadrangle depicted on the geologic map differ from those in Pearson and Zen's (1985) report, they reflect different interpretations of ambiguous field relations.

Rock units in the map area range from Early Proterozoic to Quaternary. Previous mapping in the area (fig. 1) was largely related to economic mineral deposits. In the Vipond Park quadrangle, earlier mapping included the work of Richards and Pardee (1926), Theodosis (1956), and Fowler (1955) in the northeastern part, chiefly concerned with phosphate deposits; Karlstrom (1948) in the Hecla mining area and environs; Myers (1952), Hutchinson (1956), and Collins (1975, 1977) on the tungsten deposits of Brownes Lake area; and Guttormsen (1952) and Goudarzi

Figure 1 (right). Dillon 1°×2° quadrangle, showing location of study area and other mapped areas cited in the text. Insert shows location of the 2° area in southwestern Montana and adjacent Idaho. 1, Vipond Park quadrangle. 2, Stine Mountain quadrangle. 3, Maurice Mountain quadrangle. 4, Hecla area, Karlstrom, 1948. 5, central Wise River valley area, Calbeck, 1975. 6, Black Lion Mountain area, Mero, 1962. 7, Sheep Mountain area, Obert, 1962. 8, Trusty Lake-Quartz Hill Gulch area, Fowler, 1955. 9, Quartz Hill area, Goudarzi, 1941. 10, Swamp Creek-Triangle Gulch area, Guttormsen, 1952. 11, Melrose phosphate field studies: Richards and Pardee, 1926; Theodosis, 1956. 12, Melrose quadrangle, H.W. Smedes and G.D. Fraser, unpublished map. 13, Wise River quadrangle, Fraser and Waldrop, 1972. 14, East Pioneer Mountains, Pearson and Zen, 1985. 15, West Pioneer Wilderness Study area, L.W. Snee and E-an Zen, unpublished data. 16, Northwest quarter, Willis 30-minute quadrangle (this quarter quadrangle is now the Twin Adams, Torrey Mountain, Argenta, and Ermont 7½-minute quadrangles), Myers, 1952; 16a, Brownes Lake area, Myers, 1952. 17, southern part of Pioneer batholith, Snee, 1978. 18, Greenstone Mountain area, Sharp, 1970. Other localities noted on the figure are as follows: Camp Creek section of Precambrian and Paleozoic rocks, east of Melrose. Beaverhead Conglomerate outcrops between Melrose and Glen that contain clasts of Proterozoic Y quartzite interpreted as derived from Late Cretaceous thrust sheet of Morrison Hill klippe. A small tonalitic stock in Dodgson Creek area in northwest part of Wise River quadrangle, that cuts a thrust sheet of Proterozoic Y quartzite over Colorado Group sediments and that provides a minimum date of thrusting at this place.


113° 22′30″
46°
112°30′
MONTANA
Area of index map
Dillon 2° quadrangle
IDAHO
Butte
15
43
Dated stock, Dodgson Creek, cut thrust sheet
13
Wise River
Patterson Corner
Dewey
11
Divide
15
7
5
10
9
8
Camp Creek section
14
2
12
Melrose
6
14
4
Beaverhead Formation
3
16a
1
17
Glen
14
15
15
17
Polaris
18
278
41
16
45°15′
0
5
10
15
20
25 MILES
0
10
20
30 KILOMETERS

(1941) in the northwestern part of the quadrangle. Obert (1962) mapped in the Sheep Mountain area, and Mero (1962) mapped in the Black Lion Mountain area. These two studies extended into the Stine Mountain quadrangle. Calbeck (1975) mapped an area in the Wise River valley that straddles the Stine Mountain and Maurice Mountain quadrangles. Mapping efforts immediately surrounding the area include (fig. 1) Myers (1952) in the "Northwest 1/4 of the Willis Quadrangle," which now consists of the Twin Adams, Torrey Mountain, Argenta, and Ermont 7 1/2-minute quadrangles; Sharp (1970) in the eastern half of the Twin Adams quadrangle; Fraser and Waldrop (1972) in the Wise River 7 1/2-minute quadrangle; H.W. Smedes and G.D. Fraser (U.S. Geological Survey, unpublished data) in the Melrose 7 1/2-minute quadrangle; and Snee (1978) in the batholithic rocks to the south. Berger, Breit, and others (1979) and Berger, Van der Voort, and others (1979) provided data on the geochemical patterns in parts of the area. Some preliminary data on the rocks have been published (Zen, 1977; Zen, Marvin, and Mehnert, 1975; Zen and Dutro, 1975; Zen and Hammarstrom, 1979; Zen, Taylor, and Wilson, 1979; and Zen and others, 1980; Dutro and others, 1975; Hammarstrom, 1979, 1982; and Marvin and others, 1982, 1983).

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This report is designed as a summary of the geology to accompany the geologic map and to provide background for future topical reports.

ROCK UNITS AND STRATIGRAPHY

Detailed descriptions of the rock units are given on the geologic map of the Vipond Park, Stine Mountain, and Maurice Mountain quadrangles (pl. 1) and will not be repeated. Instead, this section of the study will deal with a summary discussion of the geologic relations and implications of the rock units, as well as their age and regional correlations.

Early Proterozoic Gneiss and Amphibolite

Winchell (1914, p. 79) referred for the first time to gneiss on the northern slope and southern base of Sheep Mountain. This terrane is dominated by microcline-plagioclase gneiss (fig. 2*A*) and amphibolite; of these, gneiss is far more abundant. Migmatitic gneiss and augen gneiss are minor; a few outcrops exist of slightly calcareous garnet-bearing schistose gneiss. On the west wall of the cirque north of Sheep Mountain, tight, decimeter-scale, ptygmatic folding of amphibolite and feldspar gneiss can be seen. Most of the gneiss exhibits at least two foliations; the oldest, most prominent one is defined by the alinement of platy minerals and the long dimensions of feldspar. One outcrop (fig. 2*B*) shows the amphibolite and gneiss in a tight, nearly isoclinal and overturned fold passing into a thrust fault, and the oldest foliation that contains mineral alinement is parallel to axial surface of the fold. Because outcrops of the gneiss are confined largely to the ridge extending west from Sheep Mountain and to only a few scattered outcrops on lower slopes, it was not possible to disentangle age and stratigraphic relations of the rock units. Some amphibolites clearly intrude the gneiss, but it is not clear that all of them are intrusive. The calcareous gneiss probably had a sedimentary protolith. A greenish-gray phyllitic rock found at the Queen of the Hills Mine southeast of Sheep Mountain is near a major high-angle fault and is interpreted as tectonically milled gneiss.

Age of the gneiss and amphibolite is based entirely on isotopic geochronology. A four-point Rb-Sr whole-rock isochron including both rock types gave an age of 1,630±38 Ma with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7040±0.0003 (Arth and others, 1986). Uranium-lead ages, from a single zircon separated from foliated plagioclase gneiss, are discordant and range from 1.1 to 1.8 Ga (T.W. Stern, 1975, written commun.); this range suggests that these are all modified metamorphic ages and that the true age of the rock is at least 1.8 Ga, or possibly older. The Rb-Sr isochron date presumably records the last episode of isotopic homogenization.

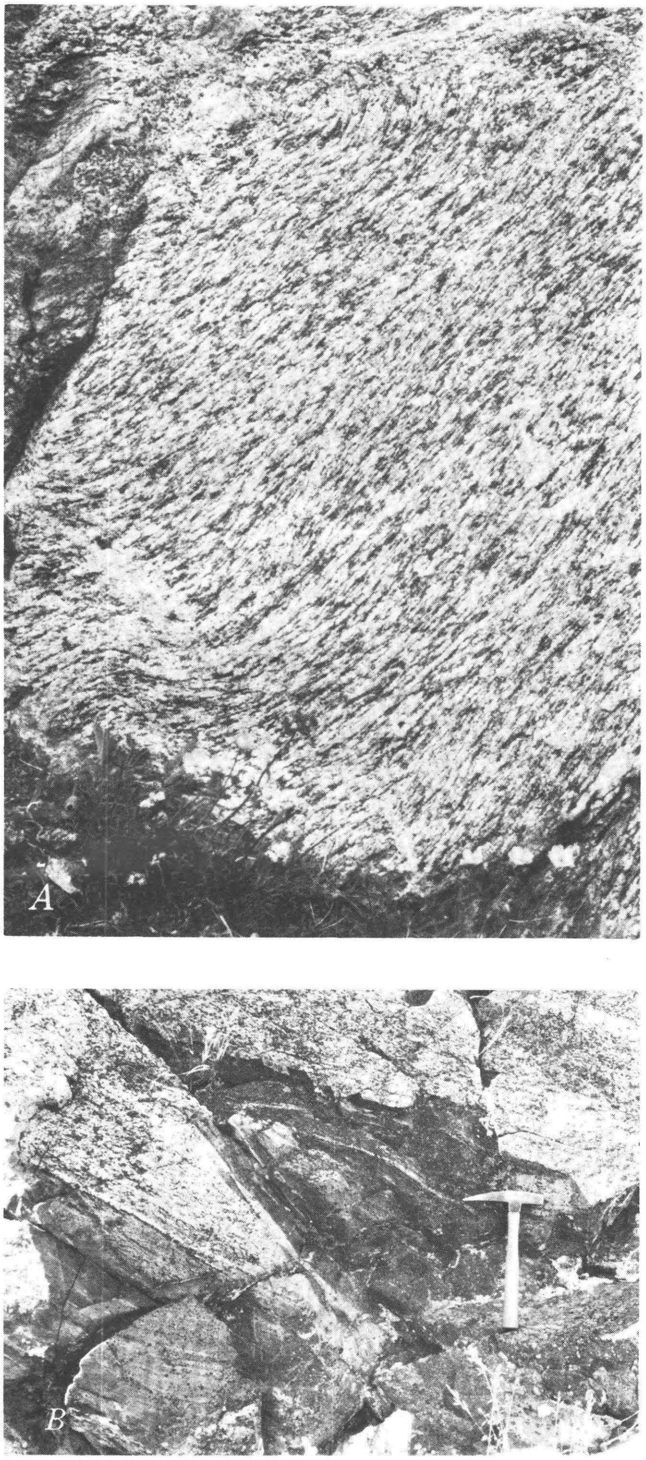


Figure 2. *A*, Proterozoic X gneiss, rim of the west sidewall, cirque north of Sheep Mountain. Notice at least three sets of planar features. *B*, Recumbent and nearly isoclinal fold involving Proterozoic X gneiss and amphibolite; axial surface is parallel to foliation caused by mineral alignment in the rocks. North of 8,860-foot peak northwest of Sheep Mountain.

The nearly total absence of evidence for sedimentary protolith of these gneisses and amphibolites suggests a plutonic or volcanic and (or) volcanogenic origin; the bulk chemistry and rare earth element patterns support this idea as does the initial $^{87}Sr/^{86}Sr$ ratio. Arc volcanism of an andesitic suite seems possible. If the rocks are autochthonous, their present position is not inconsistent with a continental-margin environment of original rock deposition, but such an interpretation is more readily made if the rocks have been transported from farther west. Knowledge of the relation of these rocks to the augen gneiss reported from the Shoup, Idaho, area (Armstrong, 1975; Evans and Zartman, 1981) would be important in regional interpretation. The rocks do not closely match any other Early Proterozoic rocks found elsewhere in southwestern Montana, although different deformational and metamorphic histories could have masked original similarities.

Middle Proterozoic Sedimentary Rocks

Four distinct Precambrian lithostratigraphic sequences, presumably of Middle Proterozoic age, are recognized in the map area. These sequences are separately identified on the map and are informally referred to as the sequence at Maurice Mountain (Ymm; fig. 3*A*), the sequence at Big Point (Ybp; fig. 3*B*), the sequence at Boner Knob (Ybk; fig. 3*C*), and the sequence at Swamp Creek (Ysc). Of these, the first three named sequences are allochthonous relative to the Phanerozoic rocks exposed in the map area and occur in two thrust sheets. The sequence at Swamp Creek is autochthonous with respect to the overlying Phanerozoic sedimentary rocks that constitute the bedrock of most of the area east of the Fourth of July fault, but there is no information to prove that it is truly autochthonous. The klippen east of the Fourth of July fault (west of Grace Lake, at Morrison Hill, on Vipond Park, and on Ponsonby Peak) are correlated with the sequences of Boner Knob and of Maurice Mountain. However, that patch east of Gray Jockey Peak may be an integral part of the associated unit.

The age of these three allochthonous sequences is not firmly established. They are assigned to the Middle Proterozoic for two reasons. First, the nature of the rocks, consisting of thick sequences of quartzite, feldspathic sandstone, siltstone, and argillite, shows evidence of shallow-water marine deposition and strongly suggests the Middle Proterozoic rocks of the Missoula Group of the Belt Supergroup of western Montana. In fact, the Maurice Mountain, Big Point, and Boner Knob may be directly correlated by their lithic assemblages to units 1, 2, and 3 of the Mount Shields Formation (C.A. Wallace, 1981, oral commun.). The correlatable sequence of units is so short, however, that prudence suggests the use of informal local names for the beds at this time.

The second basis for the Middle Proterozoic age assignment for the allochthonous sequences is the available minimum isotopic age determinations for two samples. One sample is an altered and highly deformed mafic dike (sheared greenstone and altered diabase (ZYdm)) that cuts

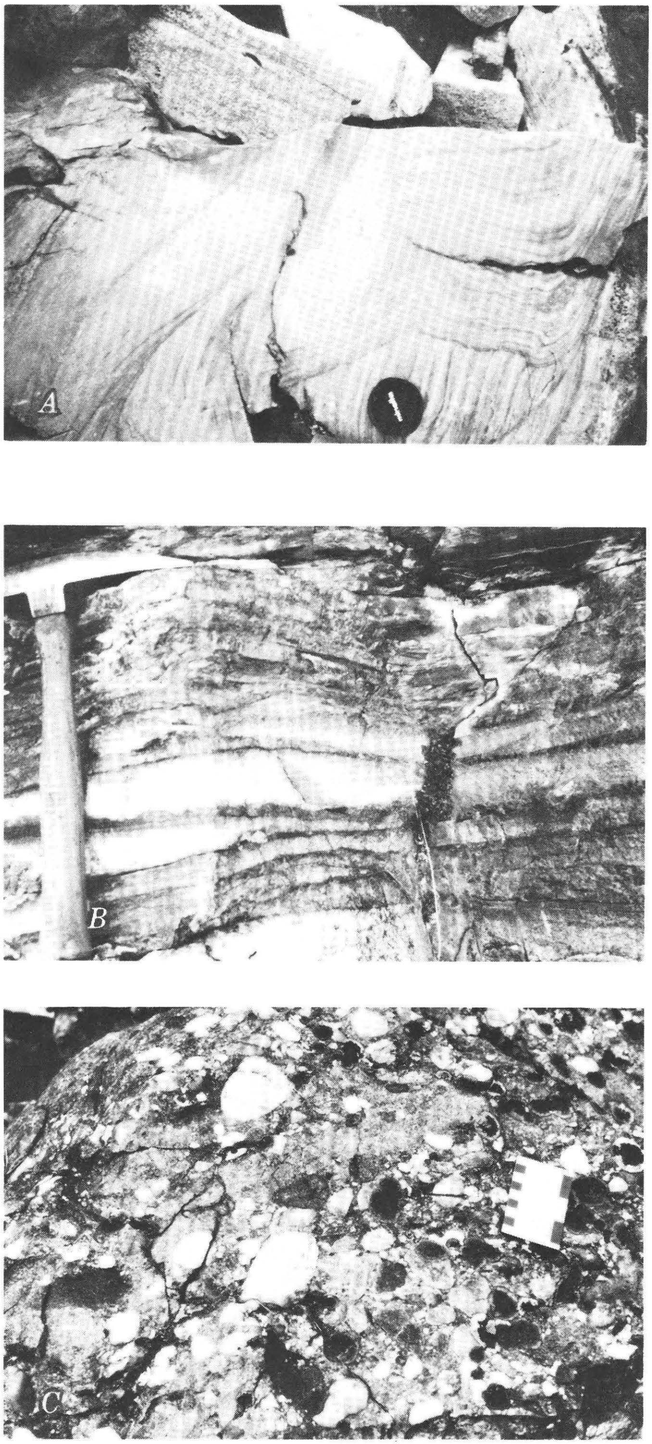


Figure 3. *A*, Thick-bedded quartzite beds of the sequence of Maurice Mountain, showing multiple slumps that involved crossbedded strata. Loose block at base of cliffs west of lowest lake at head of Grouse Creek. *B*, Thin-bedded argillite and siltstone, type locality of the sequence of Big Point. Notice interrupted bedding below and to right of hammer point caused by dewatering of silty beds through argillite; on bedding surfaces these interruptions mimic mudcracks. *C*, Conglomerate in the sequence of Boner Knob, 7,200 ft on ridge between Mammoth Gulch and Adson Creek, just west of Fourth of July fault. Clasts include purple sandstone, white sandstone, vein quartz, jasper, and single-crystal feldspar.

the sedimentary rocks of the Big Point sequence. A sample from north of Lacy Creek gave a disturbed-argon-release pattern that indicates a minimum age of about 660 Ma (Snee, 1982). Second, a single whole-rock Rb-Sr determination on a purple mud chip from unit 2 (Ybk2) of the sequence of Boner Knob on the ridge north of the summit of Morrison Hill (SE1/4 sec. 32, T. 2 S., R. 10 W.) gave an age of 757±15 Ma (C.E. Hedge, 1974, written commun.).

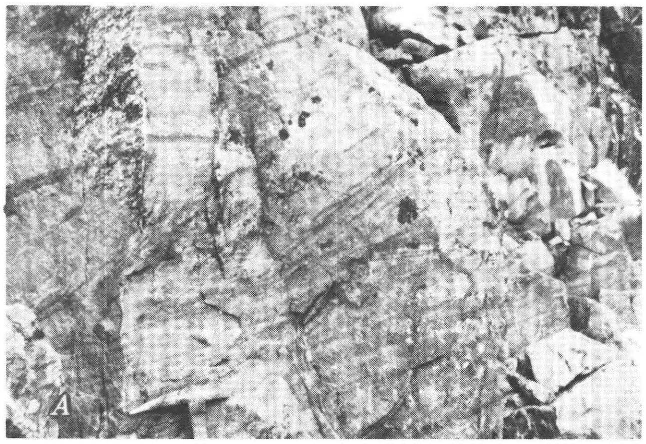
The stratigraphic relations of the three allochthonous sequences cannot be deciphered unambiguously with the information now available. Even their mutual structural relations are not known for certain. Because these two topics are intimately mingled, a fuller discussion of the subject is deferred until the end of the section on regional structural relations.

The "autochthonous" sequence of Swamp Creek underlies Cambrian rocks of the Hasmark Dolomite, a dolostone, or the upper, calcareous member of the Silver Hill Formation in the Swamp Creek and Gray Jockey Peak area. The sequence of Swamp Creek is mapped by Fraser and Waldrop (1972) as the units "white vitreous quartzite" and "dark variegated argillite" (bw and ba). Assignment of this sequence to the Middle Proterozoic is even more tenuous than that of the three allochthonous sequences. The only direct evidence, an argon-release pattern on a mafic dike that cuts the unit on the north side of Swamp Creek immediately north of the Vipond Park quadrangle boundary, gave a disturbed pattern and a minimum age of 400 Ma (L.W. Snee, 1980, written commun.). The sequence does not resemble any of the allochthonous sequences. A minimum thickness of 250 m is demonstrated by a drill-hole record in the Ponsonby Peak area (Guttormsen, 1952, p. 31); Fraser and Waldrop (1972) estimated 0–600 ft for their "dark variegated argillite" unit (ba).

Cambrian(?) and Cambrian Strata

The Cambrian(?) and Cambrian strata in the map area consist of four formations. From oldest to youngest, they are Black Lion Conglomerate, a new unit having its type locality on the cliffs between Black Lion Mountain and Black Lion Lake in the northwest part of the Vipond Park quadrangle; the Silver Hill Formation; the Hasmark Dolomite; and the Red Lion Formation. The lithic character and general contact relations of each unit are given in the map explanation. Special attention is given to the Black Lion Conglomerate because it is nonfossiliferous and its age is limited only by the next younger formation. Field relations show that the Black Lion Conglomerate is in gradational contact with the lower member of the Middle Cambrian Silver Hill Formation; thus, it is assigned to a Cambrian(?) to Cambrian age.

The contact between the Black Lion Conglomerate and the Silver Hill Formation is physically observable (1) on the north slope of Sheep Mountain, especially on the lip and upper cliffs of the cirque; (2) on the east slope of Black Lion Mountain; (3) around Grace Lake; and (4) east of Lion Mountain at Hecla. At Sheep Mountain, a 2-m layer of gray-green silty argillite locally separates typical Black Lion Conglomerate from dark-gray argillite that forms the lower beds of the lower member of the Silver Hill. This gray-green argillite is not a mylonite, as was first suspected, because microscopically the rock shows no deformation features and there is no evidence that it is a blastomylonite. Moreover, intercalated with and above the first few meters of the overlying dark-gray argillite are beds of coarse quartzite, each several decimeters thick, that have the same bedding character and grain size as some finer grained beds in the Black Lion. On the cliffs east of the cirque and above the jeep track, especially adjacent to two prominent snow gullies, crossbedded quartzite (fig. 4*A*) is interbedded with typical conglomeratic strata (fig. 4*B*); this quartzite is similar both to the one above the gray-green argillite bed and to the Flathead Quartzite as exposed at Camp Creek. At least in part, the Black Lion Conglomerate must be homotaxial with the Flathead Quartzite.



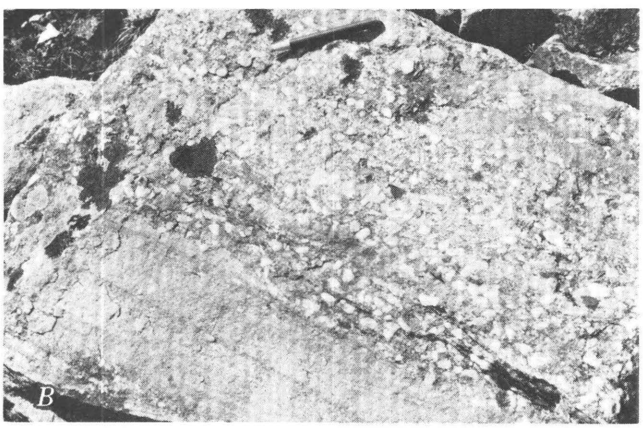


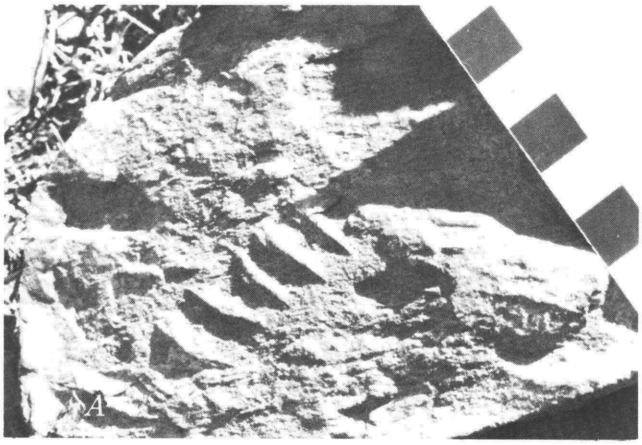
Figure 4. *A*, Crossbedded pale-pinkish-gray feldspathic sandstone within the Black Lion Conglomerate, cliffs next to western couloir, north face of Sheep Mountain. *B*, Conglomerate beds of Black Lion Conglomerate, above quartzite strata of figure 4*A*, same locality.

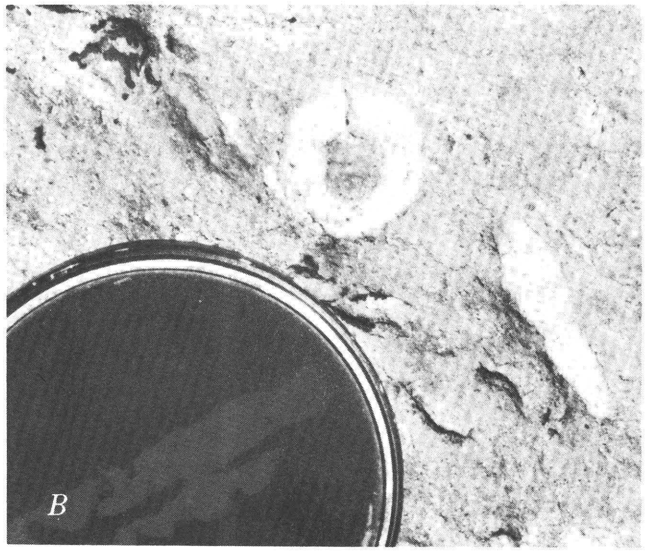
On the open slope east of the summit of Black Lion Mountain, the conglomerate beds pass upsection into interbedded dark-gray argillite and coarse, clean, white-weathering quartzite that does not have the green mica-spangled and argillite-veneered bedding surface characteristic of the conglomerate. The relations here are equivocal but do not contradict a continuous sedimentary section. The same is true at Grace Lake. At both places, the strata above the contact are typical of the lower member of the Silver Hill Formation.

At Hecla, conglomerate identical to the typical Black Lion Conglomerate is found at the south side of the outcrop area. This rock passes upsection into coarse quartzite beds showing the same bedding characteristics (nature and size of crossbedding, argillite veneer on bedding surfaces, and bed thickness) as the conglomerate. The quartzite beds differ from the conglomerate only by the progressive upsection depletion of larger clasts. These upper quartzite beds are also identical with the quartzite interbedded with argillite at Sheep Mountain. The upper quartzite section shows increased interbedding of dark-gray argillite (fig. 5) and siliceous argillite as well as evidence of bioturbation; several bedding surfaces of quartzite display trace fossils and a pelmatozoan stem (fig. 6*A*,*B*,*C*). Thus, the stratigraphic relations indicate a continuous and gradual change of sedimentation, and the contact between the Black Lion Conglomerate and the Silver Hill Formation is arbitrary, placed for cartographic purposes at the first upsection appearance of significant argillite beds.

The nature of the lower contact of the Black Lion Conglomerate is unknown. In the Black Lion Mountain area, the bottom of the unit is cut off by faults and intrusions. In the Hecla area it is not exposed. At Sheep Mountain, it is cut off by faults against the Early Proterozoic gneiss to the north. On the north and west sides of the gneiss, the Black Lion reappears; the contact is either a fault

Figure 5. Interbedded dark-gray argillite and quartzite, lower member of the Silver Hill Formation near Hecla. Note thin seams of argillite in quartzite above the thick argillite bed, and note crossbedding in quartzite below.





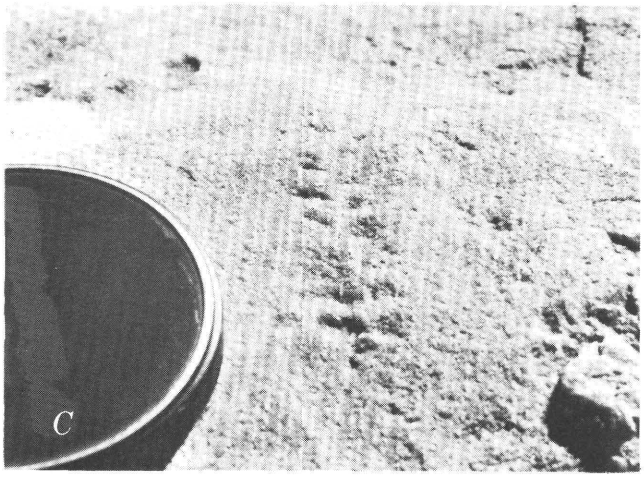


Figure 6. *A*, Trilobite tracks, lower member of Silver Hill Formation, north slope of Sheep Mountain. *B*, Pelmatozoan fragment, bedding surface of a quartzite bed at transition interval between Black Lion Conglomerate and lower member of the Silver Hill Formation, north side of structural dome near Hecla. *C*, Trilobite tracks on quartzite bedding surface, lower member of the Silver Hill Formation, south side of structural dome near Hecla. *D*, Stereoscopic pair showing *Albertella* from lower member of the Silver Hill Formation, north rim of cirque east of Black Lion Mountain, at 9,600 ft elevation.

or an unconformity, but lack of actual exposure of the contact leaves an ambiguity.

The evidence favors assignment of the upper part of the Black Lion Conglomerate to the top of the Lower Cambrian or the bottom of the Middle Cambrian because the best fossil evidence indicates that the lower member of the Silver Hill is early Middle Cambrian (fig. 6*D*; Zen, Taylor, and Wilson, 1979). The bulk of the Black Lion Conglomerate is probably Early Cambrian, but it may range down into latest Precambrian. This age is consistent with the finding of occasional red silty quartzite pebbles, which could have a Middle Proterozoic origin, in the conglomerate.

Regionally, the Black Lion Conglomerate may correlate (1) with the autochthonous and parautochthonous Tintic Quartzite of Utah (Morris and Lovering, 1961), which contains similar conglomerate and crossbedded quartzite in its lower half; (2) with the Cash Creek Quartzite of the Clayton area, central Idaho (Hobbs and others, 1975); and (3) with the fossiliferous Lower Cambrian rocks of the Lemhi Range (Derstler and McCandless, 1981; A.R. Palmer, 1981, oral commun.).

The Black Lion Conglomerate aggregates at least 500 m thick, as calculated from the topographic relief of the gently dipping beds between Black Lion Lake and just east of the summit of Black Lion Mountain; the section is everywhere right side up, and throughout the continuous exposure on the cirque headwall it shows no evidence of repetition by faulting. This cliff section between the summit and the top of talus above Black Lion Lake, at about 9,000 ft elevation, is here designated the type section for the formation. At Sheep Mountain, a little over 100 m of the right-side-up section is exposed, mainly in cliffs. At Hecla, only about 50 m of the section is exposed.

Trough crossbeds, best seen at Hecla (fig. 7), are observed in the Black Lion Conglomerate. This crossbedding, characteristic of the unit, cuts into underlying crossbed sets. Orientations of the troughs at Hecla consistently yielded a south-southeast source of sediment; this finding is confirmed by the orientation of large dune ripples (wavelength about 1 m, amplitude a few centimeters) in the overlying lower member of the Silver Hill Formation.

The Silver Hill Formation in the Vipond Park area is divided into two members. Only the lower member in the area is fossiliferous. The most definitive fossil (fig. 6*D*) is a limonitic impression of the pygidium of an *Albertella* sp. found at 9,600 ft (2,926 m) elevation on the north ridge of a shallow cirque due east of the summit of Black Lion Mountain (Zen, Taylor, and Wilson, 1979). Other fossils include pelmatozoan columnals at the north slope of Sheep Mountain and at Hecla, as well as an actinian coelenterate (sea anemone; possibly related to *Bergaueria*), burrows, and other trace fossils previously reported by Dutro and others (1975) and Zen, Taylor, and Wilson (1979), including *Scolithus*, *Monocraterion*, *Diplichnites* (paired walking tracks), *Cruziana* (grazing burrows), and *Rusophycus* (resting burrows). The lower member thus correlates with the Gordon Shale of the Dearborn valley region (Deiss, 1939, p. 37) and with the lower member of the Silver Hill in the Garnet Range (Kauffman, 1963) and west of Missoula (Wells, 1974) (fig. 8). The lower member of Silver Hill is

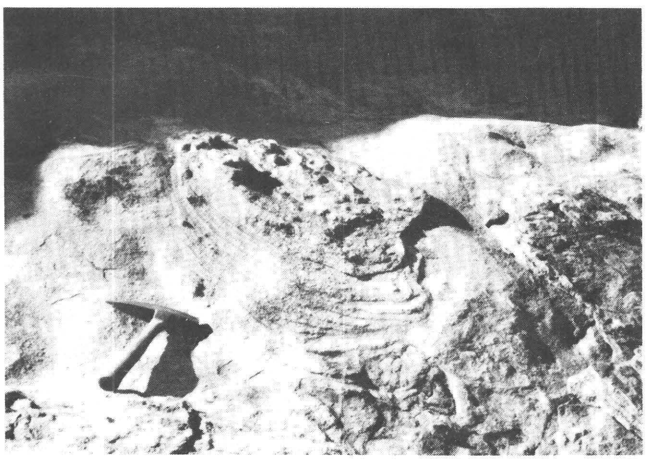


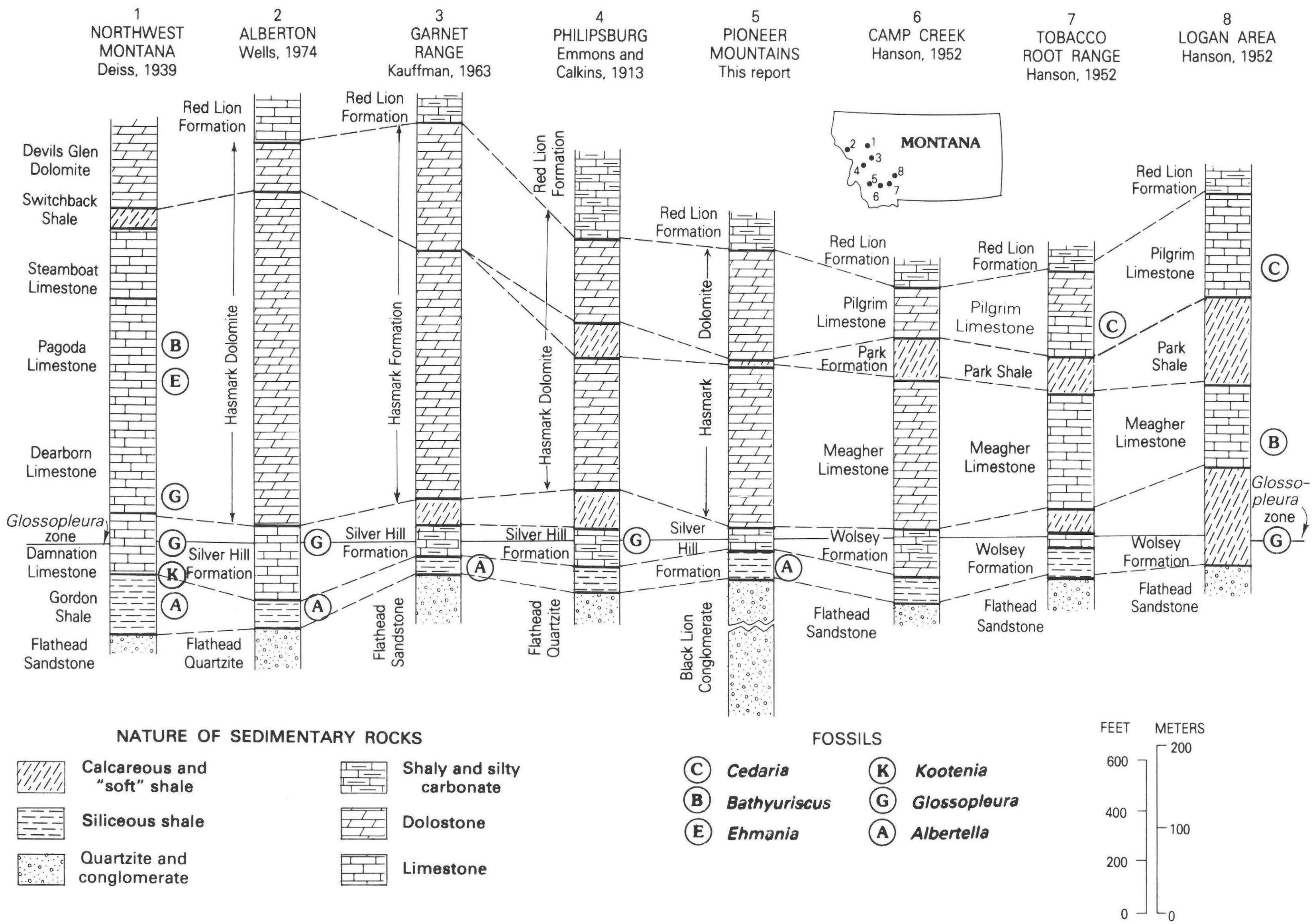
Figure 7. Trough crossbeds in upper beds of Black Lion Conglomerate near Hecla. View is down the trough; in cross section the bedding laminations are part of crossbed sets that cut into underlying crossbeds.

homotaxial, not correlative, with the lower part of the Wolsey Shale of the Three Forks region on the shelf (Hanson, 1952) because the latter is in the *Glossopleura* Zone and thus correlates with the middle member of the type Silver Hill (Emmons and Calkins, 1913; Hanson, 1952, p.14). The Wolsey correlates with the upper member of the Silver Hill in the Vipond Park quadrangle (which includes the middle and upper members of the type Silver Hill). Wolsey Shale is also mapped in the Camp Creek area just east of the quadrangle (Hanson, 1952); here the unit has a lithic division into lower siliceous and upper calcareous beds, similar to the Silver Hill. Lack of fossils from the Camp Creek exposures, however, precludes a more precise correlation; the geographic location of Camp Creek suggests that the Camp Creek section ought to be intermediate between the Silver Hill and the Wolsey of the Three Forks area in its time-facies relationship, for even if the entire section of Phanerozoic rocks in the Vipond Park area should prove allochthonous, it was probably transported from the west.

According to Zen, Taylor, and Wilson (1979), the upper and middle members of the type Silver Hill are not time correlative with the Meagher Limestone and Park Shale of the shelf sequence, as suggested by Emmons and Calkins (1913) and followed by many workers; rather, they are time correlative with the Wolsey (and the lower Silver Hill with pre-Wolsey beds), whereas the Hasmark Dolomite of the Pioneer Mountains and the Philipsburg area correlates with the Meagher, Park, and Pilgrim Formations (fig. 8). The Hasmark, therefore, is Middle and Late Cambrian in age.

Karlstrom (1948, p. 18–19) described a "basal" conglomerate in the Wolsey (= Silver Hill) Formation at Hecla. This conglomerate is probably an intraformational sedimentary lens, reflecting a local and temporary change of hydraulic regime rather than a tectonic break. The conglomerate does contain pebbles of a distinctive green quartzite whose color is imparted by chrome-rich muscovite that chemically resembles the chrome-rich mica of the quartzite of the Early Proterozoic Facer Formation (Crittenden and Sorensen, 1980). The Facer, or a similar formation, might have been the source rock for the pebbles; the locally measured direction of transport, from south-southeast, however, does not point to any source known today, as known rocks of this kind are in the Wyoming-Idaho thrust belt (Crittenden and Sorensen, 1980) and in the Laramie Range (Max Crittenden, written commun., 1980). James (1981) described chrome mica in an Archean quartzite in the Tobacco Root Mountains to the east, but the lithic characteristics of the rocks do not seem to match.

The Black Lion to Silver Hill sequence is interpreted (Zen, 1977) as a trough-filling sequence. In this view, Black Lion deposition originated in a sharply defined trough or troughs, perhaps graben, with surrounding highlands underlain by crystalline gneiss, schist, probably igneous rocks, and some Belt-like red sandstone and argillite. As the highland relief diminished and as the troughs became shallower by infilling, the sediment became finer grained and more


1
NORTHWEST
MONTANA
Deiss, 1939
2
ALBERTON
Wells, 1974
3
GARNET
RANGE
Kauffman, 1963
4
PHILIPSBURG
Emmons and
Calkins, 1913
5
PIONEER
MOUNTAINS
This report
6
CAMP CREEK
Hanson, 1952
7
TOBACCO
ROOT RANGE
Hanson, 1952
8
LOGAN AREA
Hanson, 1952
MONTANA
Devils Glen Dolomite
Switchback Shale
Steamboat Limestone
Pagoda Limestone
Dearborn Limestone
Glossopleura zone
Damnation Limestone
Gordon Shale
Flathead Sandstone
Red Lion Formation
Hasmark Dolomite
Silver Hill Formation
Flathead Quartzite
Red Lion Formation
Hasmark Formation
Silver Hill Formation
Flathead Sandstone
Red Lion Formation
Hasmark Dolomite
Silver Hill Formation
Flathead Quartzite
Red Lion Formation
Hasmark Dolomite
Silver Hill Formation
Black Lion Conglomerate
Red Lion Formation
Pilgrim Limestone
Park Formation
Meagher Limestone
Wolsey Formation
Flathead Sandstone
Red Lion Formation
Pilgrim Limestone
Park Shale
Meagher Limestone
Wolsey Formation
Flathead Sandstone
Red Lion Formation
Pilgrim Limestone
Park Shale
Meagher Limestone
Glosso-pleura zone
Wolsey Formation
Flathead Sandstone
NATURE OF SEDIMENTARY ROCKS
Calcareous and "soft" shale
Siliceous shale
Quartzite and conglomerate
Shaly and silty carbonate
Dolostone
Limestone
FOSSILS
C Cedaria
B Bathyuriscus
E Ehmania
K Kootenia
G Glossopleura
A Albertella
FEET
METERS
600
400
200
0
200
100
0

calcareous, leading to the passage from the Black Lion upward through the lower member to the upper member of the Silver Hill Formation. Silver Hill deposition was in a normal marine environment; the Black Lion may also be marine. Toward the end of Silver Hill time, the troughs became completely filled, permitting the development of the widespread and uniform deposition of the shelf-type carbonate sediments of the Hasmark Dolomite. Whether the trough formation was part of the postulated rifting event for western North American craton during the late Precambrian (Stewart, 1972) is unknown.

At least locally in the Vipond Park area, the basal strata of the Hasmark are peloidal limestone, grading upward into pisolitic and oolitic carbonate, algal-mat carbonate, intraformational conglomerate, and other rock types that suggest a continuing shoaling from subtidal to intertidal or possibly even to supratidal environments. This condition persisted through the remainder of the Cambrian. Many of the primary sedimentary features are lost in the metamorphosed parts of the formation, except for the pisolites that had previously been replaced by chert (fig. 9). The metamorphosed Hasmark in isolated outcrops may be difficult to distinguish from the Jefferson Dolomite. Useful clues to identifying the Hasmark are a more blocky, massive weathering, which contrasts with the thin-layered weathering nature of the Jefferson; a bluish-gray instead of yellowish-buff weathered hue; the presence of beds of chert; and less tendency than the Jefferson to weather into carbonate sand.

The youngest Cambrian unit in the Vipond Park quadrangle is the Red Lion Formation (fig. 10). Because fossils are scarce and the next younger unit is a dolostone, the Devonian Jefferson Dolomite, the age assignment is not firm. The only fossils found in the area are in the Red Lion, a dominant red dolostone, at 7,960 ft elevation on top of cliffs above Adson Creek and north of the Adson Creek fault. According to M.E. Taylor (1979, written commun.) the collection consists of poorly preserved molds and casts of ptychoparioid trilobites and fragments of phosphatic linguloid brachiopods, the former being similar to known Upper Cambrian forms. The Cambrian assignment is supported by the fact that beds typical of the Red Lion interbed with dolostone typical of the Hasmark Dolomite on the cliffs immediately south of the fossil locality. Some of the rocks assigned to the Red Lion in the map area, nevertheless, may be part of the Devonian Maywood Formation, but no separation on a lithic basis is possible, and there is no fossil evidence to suggest the presence of any Maywood in the area. The contact between the mapped Red Lion and the Jefferson is always sharp; Ordovician and Silurian rocks are altogether absent.

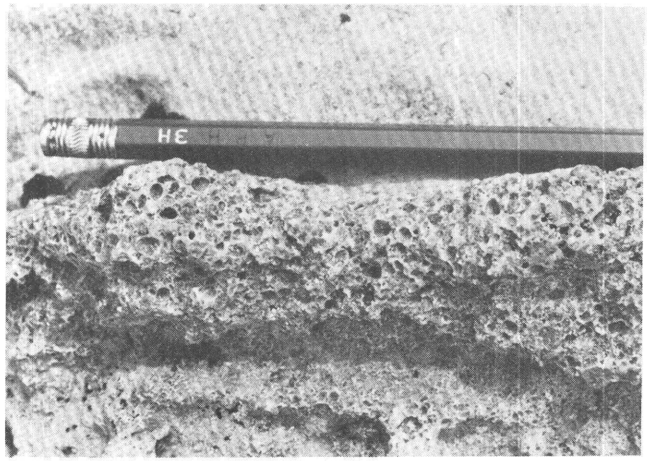


Figure 9. Pisolitic dolostone beds in the Hasmark Dolostone replaced by massive chert. Location is southwest slope of Sheep Mountain.

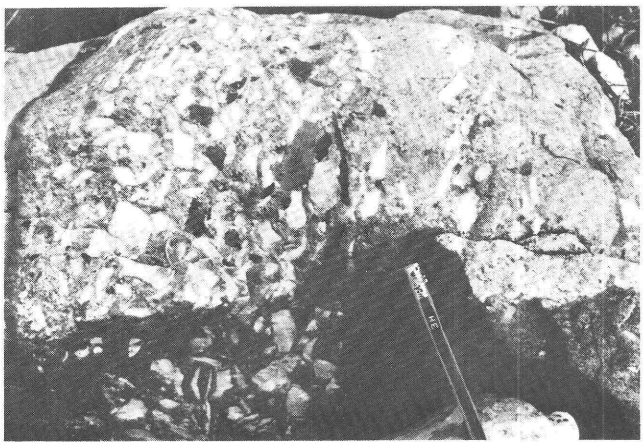


Figure 10. Polymictic conglomerate in the Red Lion Formation, south slope of Sheep Mountain. The angular clasts are tan, pink, and purple calcareous argillite in a matrix of sandy carbonate that is slightly more resistant to weathering.

A small area of green argillite interbedded with thin-bedded pale dolostone occurs at the Blue Bell Mine and on the slopes to the north in the Sheep Creek valley. This area

Figure 8 (left). Correlation of Cambrian strata of the Pioneer Mountains with those in nearby areas of western Montana. The *Glossopleura* Zone is arbitrarily shown as a horizontal line, and other trilobite faunae are shown relative to this horizon. Various lithic facies are shown by symbols, and suggested lithic correlations are shown by dashed lines. The suggested scheme correlates the lower member of the Silver Hill Formation in the map area with the lower member of the type Silver Hill, and correlates the upper member of the Silver Hill in the map area with the combined middle and upper members of type Silver Hill. Unlike the scheme advanced by Emmons and Calkins (1913), I do not correlate the Meagher Formation and Park Shale of the Logan area with the middle and upper members of the type Silver Hill, but include the Meagher, Park, and Pilgrim in the Hasmark Dolostone both in the map area and in the Philipsburg area. In this scheme, the shaly, calcareous beds in the middle of the Hasmark at both places are correlated with the Park Shale. Note that the scheme removes a need for drastic changes of thicknesses of correlative strata implied by the scheme of Emmons and Calkins.

is tentatively assigned to the Red Lion Formation on the geologic map.

The Cambrian carbonate rocks in the southeastern part of the Pioneer Mountains were mapped in the past using the stratigraphic nomenclature of the Three Forks region and of the Camp Creek section (Myers, 1952; see, however, Zimbelman, 1981); the Black Lion, at least in the same facies, is absent in all these regions. The difference in lithofacies between the northern and southern Pioneer area is noticeable. It is not clear, however, that the differences are regionally significant or that they merit conclusion of major geographic separation at the time of deposition. Finding of Phanerozoic trace fossils, very similar to those of the Silver Hill at Sheep Mountain, in rocks also identical to those found at the latter locality, on the hill south at Cat Creek near the south end of the east Pioneer Mountains (R.C. Pearson, M.E. Taylor, W.B. Myers, and E-an Zen, unpublished data), greatly reduces the perceived difference between the lithostratigraphy of the two regions. Perhaps most if not all of the interval recorded by the Black Lion Conglomerate is present and recorded in areas to the south by brown to maroon shaly sandstone, fissile shale, and shaly carbonate (Zimbelman, 1981); *Scolithus* occurs near the middle of this unit, indicating Phanerozoic age for this part of the section. These differences in pre-Silver Hill rocks may reflect local sediment sources.

Upper Paleozoic Strata

The Upper Devonian Jefferson Dolomite rests unconformably on Cambrian strata. Rocks mapped as the Red Lion are not everywhere recognized in the field, but the angularity of any pre-Jefferson truncation must have been slight as the Hasmark everywhere underlies the Jefferson-Red Lion in pseudoconformity. The Jefferson is a dark, fetid, sugar-textured dolostone in the Quartz Hill area, but it is dominantly a creamy white dolostone with interbedded, 1-m-thick, gray, faintly fetid dolostone farther south in the map area (figs. 11*A*,*B*). The white-and-gray sequence closely resembles rocks mapped as Jefferson in the Georgetown area to the northwest (Winston, 1965, p. 147); both the Cambrian and Devonian rocks of much of the map area thus resemble the rocks of the Philipsburg quadrangle as described by Emmons and Calkins (1913) and later workers.

The white-and-gray dolostone sequence locally helps in distinguishing the Jefferson from the Hasmark. Because the Jefferson is in part a two-carbonate rock (dolomite rhombs cemented by calcite), it tends to weather into carbonate sand; this feature, as mentioned previously, helps to distinguish these two units. No fossil has been recovered from the Jefferson of the map area. Two fossil collections outside of the quadrangle were studied by A.G. Harris: (1) Just north of the map boundary in the Wise River quadrangle (NW1/4, sec. 14, T. 1 S., R. 11 W.), 1 m above the base of the Jefferson, conodonts (*Polygnathus* sp. indet., ozarkodiniform elements, hindeodellan elements, and trichonodelliform elements) indicative of Middle to Late Devonian age were found. (2) An *Icriodus expansus*-bearing conodont fauna was recovered in the Jefferson, recognized by its dolostone lithology, in the southern part of the east Pioneer

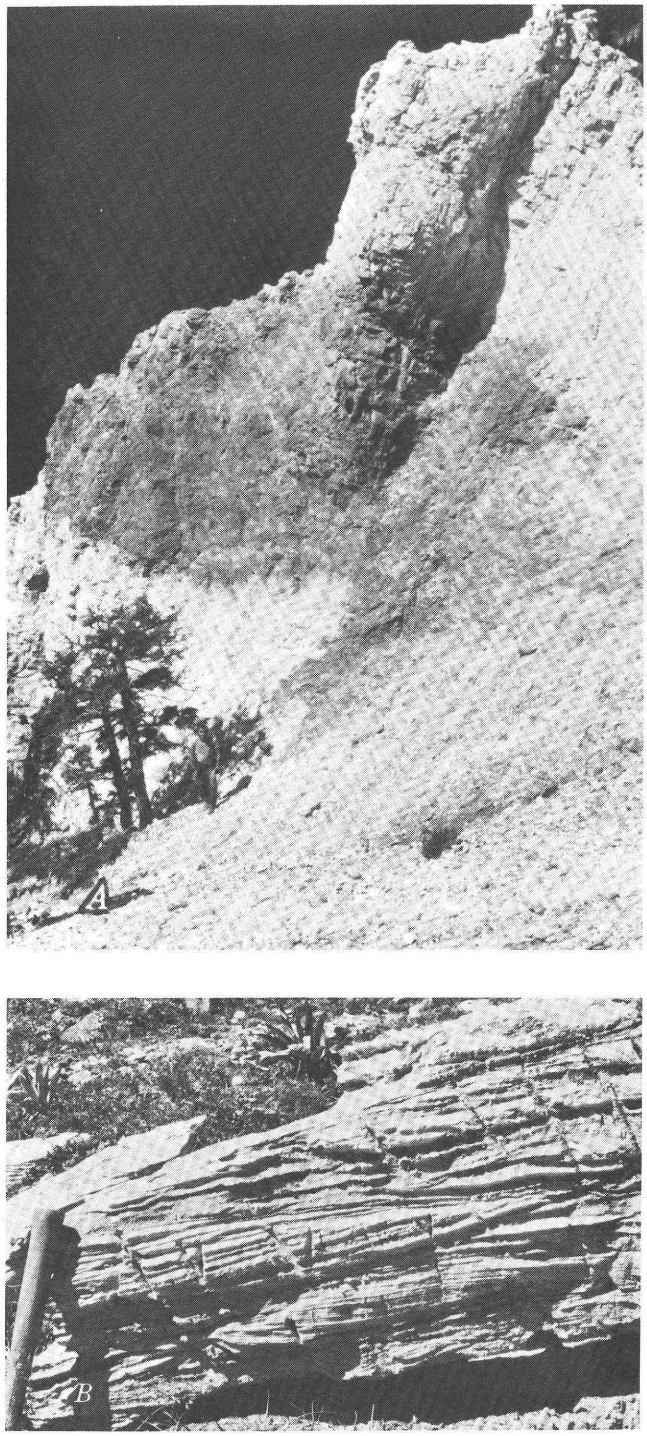


Figure 11. *A*, Jefferson Dolostone; interbedded light creamy white and steel-gray dolostone. Location is south slope of 8,938-foot peak west of Lion Mountain. *B*, Crossbedded Jefferson Dolostone, north of jeep trail northeast of Keokirk Mountain.

Mountains (NE1/4 sec. 22, T. 6 S., R. 12 W.), indicating the Givetian to Frasnian part of the Devonian (late Middle to early Late). These findings agree with the established age for this formation.

The Three Forks Formation is discontinuous in the map area, and no fossils have been recovered to confirm its age. Thus, the assignment of the rocks to this unit is tentative, and its patchy nature may reflect incomplete deposition, faulty field detection and identification, or post-Three Forks erosion before the deposition of the Mississippian Madison Group.

Although rocks of the Madison Group are extensively exposed and several fossil collections have been made in the map area, no detailed study of the stratigraphy has been made. Two general comments are in order.

First, the rocks capping Lion Mountain near Hecla are the Lodgepole Limestone, not the Jefferson as mapped by Karlstrom (1948); this reassignment is supported by the identification of Mississippian corals in the rock and by the presence of rocks similar to the Trident Member of the Three Forks Formation between it and the underlying Jefferson. Reassignment of the rocks that cap Lion Mountain removes the need for the fault between Lion Mountain and Sheriff Mountain shown by Karlstrom; the saddle is now interpreted as the location of an anticlinal crest. The Lodgepole outcrops on the two peaks look somewhat different, but such lateral changes over short distances are not exceptional, and the original nature of the rocks is partly obscured by extensive solution of carbonate as shown by solution breccias and pockets of sooty black, carbonaceous, and insoluble residues.

Second, extensive development of solution breccias in both the Lodgepole and Mission Canyon Limestones of the Madison Group, locally even involving the Jefferson Dolomite between Queens Gulch and the hills north of Canyon Creek Ranch, obscures the internal stratigraphy of this group here and makes interpretation of structural relations between the western and central parts of the quadrangle difficult. Extensive development of breccia and karst topography is also present at Cattle Gulch Ridge and can be seen along side canyons of the lower Canyon Creek (W1/2 sec. 6, T. 2 S., R. 9 W.); the breccia and karst topography is spectacularly evident in a less accessible ravine west of the 8,576-foot peak of Cattle Gulch Ridge (secs. 21 and 28, T. 1 S., R. 10 W.) as well as on the spires and cliffs of Canyon Creek opposite Vipond Park (secs. 7 and 18, T. 2 S., R. 10 W.). Whether there are penecontemporaneous depositional conglomerates among the breccias is not certain. Locally, there are conglomerates that lack a terra rosa matrix but include older rocks as blocks; these blocks could point toward a sedimentary origin.

Fossils are abundant in many outcrops of the Madison Group except where the rock has been severely deformed, metamorphosed, or brecciated through dissolution. The fossils recovered from rock mapped as Lodgepole Limestone and as Mission Canyon Limestone are tabulated in table 1. Small areas of rocks that lithically are typical of either the Lodgepole or the Mission Canyon Limestone, but without fossil control, are identified on the map by their lithic affinities.

No unconformity exists above the Madison Group in the area. The next higher stratigraphic unit is the Amsden Formation. Recent studies of its conodont content (B.R. Wardlaw, 1983, written commun.) show that rocks called the Amsden in the Vipond Park area are entirely Mississippian and thus time-equivalent to the Big Snowy Group or Formation. However, the lithologic units of the Big Snowy are not present, and so for the purpose of this report I will continue to use the name Amsden Formation in the sense used by Myers (1952), Theodosis (1956), and Fraser and Waldrop (1972). The Amsden is heterogeneous lithically; approximately, it consists of 50 percent marl, 30 percent carbonate, and 20 percent sandstone. In the southern part of the quadrangle there is more carbonate and less marl and sandstone, and the basis for separating the Amsden from the Mission Canyon Limestone in the area around Brownes Lake is the first upsection appearance of a marly bed. The abundance of red marl in the Amsden points to possible regolithic source material that resulted from the weathering and erosion of underlying carbonate rocks in the source area of the Amsden sediments. The thickness of the unit, about 100 m, is highly variable owing to the ease with which the Amsden was deformed and locally faulted.

The Amsden Formation has yielded numerous fossil collections from the Vipond Park quadrangle. These collections are listed in table 2. The fauna indicate a latest Mississippian age for this formation, although it may be in part Early Pennsylvanian.

At contacts against granite and granodiorite, the marl of the Amsden Formation typically becomes garnet-diopside-epidote-idocrase skarn, and at such places it commonly shows a tungsten anomaly. All the tungsten (scheelite) deposits commercially exploited in the Pioneer Mountains are in the garnet-rich skarn of the Amsden. At the Ivanhoe Pit above Brownes Lake, the marl beds have been metasomatized into garnet rocks and diopside-epidote rocks; the shales are tactites, and the carbonate beds are white marbles. Determination for tungsten using the instrumental neutron activation analysis (INAA) method, however, failed to show detectable tungsten in samples of unaltered Amsden marl beds collected away from contacts with intrusive rocks; and tungsten was also not detected in the granodiorite near the Ivanhoe deposit.

The contact between the Amsden Formation and the Quadrant Quartzite is cartographically easily located because of the contrast between the cliff-forming Quadrant and the grassy slopes of Amsden. Where outcrops are exceptional, as at several places on Cattle Gulch Ridge, the contact is seen to be gradational, consisting of a zone a few meters thick of interbedded carbonate rock and quartzite.

Table 1. Fossils from rocks mapped as the Mississippian Madison Group

Locality, elevation, and field station	Forms identified	Zone[1] [2]	Paleontologist[1]
Lodgepole Limestone			
E1/2 sec. 36, T. 1 S., R. 10 W.; 7,300 ft (sta. 901-1, 2, 3; USGS locs. 26563–5–PC).	*Homalophyllites*? sp. *Vesiculophyllum*? sp. *Lophophyllum*? sp.	C1	W.J. Sando.
NW1/4 SW1/4 sec. 36, T. 1 S., R. 10 W.; 7,170 ft (sta. 1014; USGS locs. 26566–8–PC).	*Vesiculophyllum* sp. *Homalophyllites* sp. ?*Straparollus (Euomphalus)* sp.	C1	W.J. Sando and E.L. Yochelson.
NW1/4 SW1/4 sec. 36, T. 1 S., R. 10 W.; 7,560 ft (sta. 1015; USGS locs. 26569–70–PC).	*Vesiculophyllum* sp. *Homalophyllites* sp. ?*Loxonema* sp. indet.	C1	W.J. Sando and E.L. Yochelson.
SE1/4 SW1/4 sec. 36, T. 1 S., R. 10 W.; 7,560 ft (sta. 1016; USGS loc. 26571–PC).	*Vesiculophyllum* sp.	C1	W.J. Sando.
Sec. 3, T. 3 S., R. 11 W. (unsurveyed); 9,644 ft top Lion Mountain (sta. 266).	*Homalophyllites* sp.	C	J.T. Dutro, Jr.
Sec. 3, T. 3 S., R. 11 W. (unsurveyed); 9,600 ft, east face Lion Mountain (sta. 266; USGS loc. 25392–PC).	*Ptylopora*? sp. Brachiopod, possibly *Composita* sp. Horn coral, indet.	C	W.J. Sando and J.T. Dutro, Jr.
NW1/4 sec. 6, T. 2 S., R. 9 W.; 6,450 ft (sta. 1005; USGS loc. 26355–PC).	*Spirifer* cf. *S. grimesi-logani* complex.	C	J.T. Dutro, Jr.
SE1/4 sec. 36, T. 1 S., R. 10 W.; 7,364 ft (sta. 899; no USGS number).	*Composita*?	C	J.T. Dutro, Jr.
NE1/4 sec. 26, T. 1 S., R. 10 W.; 7,050 ft, in Dry Gulch (sta. 924; no USGS number).	Gastropod: ?*Straparollus (Euomphalus)* sp.		E.L. Yochelson.
Mission Canyon Limestone			
NW1/4 SE1/4 sec. 8, T. 2 S., R. 10 W.; 6,900 ft (sta. 148; USGS loc. 25394–PC).	*Syringopora* aff. *S. surcularia* Girty. *Homalophyllites* sp. *Vesiculophyllum* sp. *Rylstonia*? sp. *Unispirifer* sp.	C2	W.J. Sando and J.T. Dutro, Jr.
SE1/4 NW1/4 sec. 18, T. 2 S., R. 10 W.; 7,920 ft (sta. 207; USGS loc. 25395–PC).	*Syringopora* aff. *S. surcularia* Girty. *Rylstonia* sp. *Vesiculophyllum* sp. Large spiriferoid brachiopod, indet. Horn coral, indet.	C2	W.J. Sando and J.T. Dutro, Jr.

[1]Each entry refers to entire locality.

[2]Sando, W. J., Mamet, B.L., and Dutro, J. T., Jr., 1969, Carboniferous megafaunal and microfaunal zonation in the northern Cordillera of the United States: U.S. Geological Survey Professional Paper 613–E, p. E1–29.

Table 2. Fossils from rocks mapped as the Mississippian and Early Pennsylvanian(?) Amsden Formation

Locality, elevation, and field station	Forms identified	Age[1]	Paleontologist[1]
SE1/4 SW1/4 sec. 18, T. 2 S., R. 10 W.; 7,950 ft on ridge crest (sta. 208-2; USGS loc. 26862–PC).	Echinoderm debris, indet. *Derbyia* sp. *Antiquatonia*? sp. *Linoproductus* sp. *Anthracospirifer* sp. *Composita* cf. *C. subtilita* (Hall) *Punctospirifer* sp. Pelecypod, indet.	Possibly late Mississippian or very early Pennsylvanian.	J.T. Dutro, Jr.
Same locality as above (sta. 208–4; USGS loc. 26863–PC).	Fenestrate and ramose bryozoans, indet. Orthotetid brachiopod, indet. Linoproductid, indet. *Composita* sp. Punctate spiriferoid, indet. *Eumetria*? sp. (large)	Late Mississippian; possibly very early Pennsylvanian.	J.T. Dutro, Jr.
Same locality as above, 75 ft below top of formation (measured from contact with the Quadrant Quartzite). Red marl (sta. 208A; USGS loc. 25401–PC).	Echinoderm debris, indet. Conularid fragment, indet. Fenestrate bryozoan fragments, indet. *Orthotetes* sp. (same as in 25400–PC). Linoproductid fragments, indet. *Composita* sp. *Hustedia*? sp.	Late Mississippian; possibly very early Pennsylvanian.	J.T. Dutro, Jr.
Same locality as above, 90 ft below top as measured above, red marl (sta. 208B, USGS 25400–PC).	Echinoderm debris, indet. *Orthotetes* sp. (abundant) Spiriferid fragments, indet. (possibly *Anthracospirifer*). *Composita*? sp. Proetid trilobite pygidium, indet.	Late Mississippian; possibly very early Pennsylvanian.	J.T. Dutro, Jr.
Same locality as above, 25 ft above base, tan silt (sta. 208C; USGS loc. 25399–PC).	*Lingula* sp. *Orbiculoidea* sp.	Unknown	J.T. Dutro, Jr.
Same locality as above, 15 ft above base (sta. 208D; USGS loc. 25398–PC).	*Tabulipora* sp. Fenestratid fragments, indet.; several species. Fish tooth, indet.	Unknown	O.L. Karklins and J.T. Dutro, Jr.
Near center sec. 1, T. 3 S., R. 11 W.; 2,000 ft ENE of Cleve Mine, elev. 8,270 ft (sta. 464; USGS loc. 25397–PC).	Endothyrid foraminifer, indet. Horn coral, indet. Echinoderm debris, indet. Bryozoans: Fenestratid fragments, indet.; several species. Stenoporid, possibly a tabuliporid Rhabdomesid, indet. *Rhombopora*? (or a similar genus)	Unknown Unknown Unknown Unknown	O.L. Karklins. W.J. Sando. J.T. Dutro, Jr. O.L. Karklins.

Table 2. Fossils from rocks mapped as the Mississippian and Early Pennsylvanian(?) Amsden Formation—*Continued*

Locality, elevation, and field station	Forms identified	Age[1]	Paleontologist[1]
	Bryozoans—Continued *Tabulipora*? sp. (same as in loc. 25398–PC).	Unknown	O.L. Karklins.
	Orthotetid fragment, indet. *Anthracospirifer* sp. *Composita* sp.	Unknown	J.T. Dutro, Jr.
Near center sec. 35, T. 2 S., R. 11 W.; 8,560 ft (sta. 562).	Fenestrate bryozoans, indet. Orthotetid brachiopods, indet.	Unknown	J.T. Dutro, Jr.
NE1/4 sec. 7, T. 2 S., R. 9 W.; 6,500 ft (sta. 889, USGS loc. 26356–PC).	Fenestrate and ramose bryozoans, indet. Orthotetid brachiopod fragment, indet. *Eolissochonetes*? sp. *Linoproductus*? sp. *Diaphragmus* cf. *D. nivosus* Gordon. *Inflatia*? sp. *Composita*? sp. *Reticulariina* sp. Pelecypod fragments, indet.	Late Mississippian	J.T. Dutro, Jr.

[1]Each entry refers to the entire locality, unless otherwise indicated.

Bedding is rare in the Quadrant, but faint bedding laminations that can be recognized after some practice at many places conform to the "grain" of the outcrops, even in seemingly massive ledges. Generally silica-cemented, the Quadrant locally does have a sparse carbonate cement, especially toward the top in the transition to the Phosphoria; this carbonate-cemented Quadrant is best seen on cliffs north of Canyon Creek Road east of Devils Hole (west center, sec. 10., T. 2 S., R. 10 W.).

Two lithic members of the Permian Phosphoria Formation have been mapped in the Vipond Park area. The lower member is dominated by dolostone and subsidiary phosphatic sandstone; the upper member is dominated by massive chert and has subsidiary phosphatic sandstone. The phosphate-ore zone is between these members, but the western feather edge of the ore beds is in the map area. The lower contact of the formation against the Quadrant Quartzite is at least locally gradational, effected by the local presence of carbonate cement in the Quadrant near its upper contact and by interbedding of quartzite and dolostone on a decimeter scale. This contact is best studied on the cliffs near Devils Hole, where a nearly complete section of the Phosphoria is exposed. Another feature visible here concerns the origin of the tan-colored nodular and tubular chert. At least some nodular chert in the lower member of the Phosphoria is here seen to be of replacement origin, as original millimeter-size laminations in the dolostone cross the boundary of the chert nodules (fig. 12).

Two collections of fossils from the Phosphoria of the Vipond Park quadrangle were made and studied by J.T. Dutro, Jr. (1974, written commun.). For station 78, NW1/4SW1/4 sec. 10. T. 2 S., R. 10 W., 7,000 ft elevation on cliffs, just below oölitic phosphate zone, he reported *Phestia, Astartella, Nuculopsis*?, an indeterminate bellerophontacean, and *Plagioglypta* sp. indet. According to Dutro, the fauna indicates correlation with "some part of the Word [Formation] of west Texas," probably late Early Permian. This collection is nearly identical with that reported by Cressman and Swanson (1964, p. 542) for the lower Canyon Creek area in the northeast part of the Vipond Park quadrangle. For station 51, NE1/4SE1/4 sec. 22,

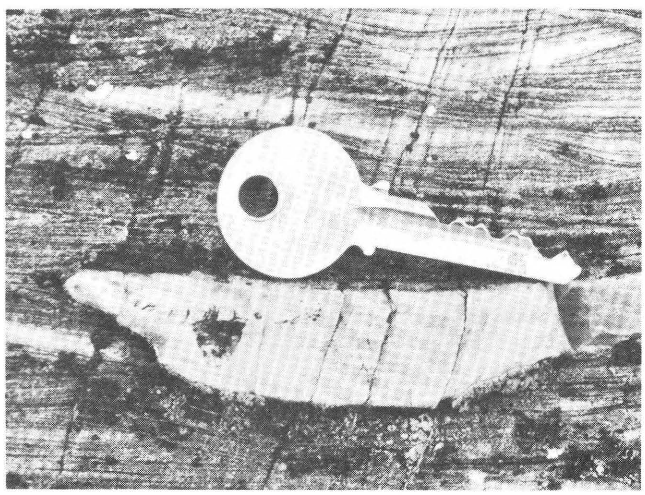


Figure 12. Chert in lower member of the Phosphoria Formation, cliffs in sec. 10, T. 2 S., R. 10 W., east of Devils Hole north of Canyon Creek. Note lamination in dolostone, caused by quartz sand in crossbeds, that passes as a ghost lamination into the chert.

T. 2 S., R. 10 W., elevation 6,630 ft on north slope of Trapper Creek, Dutro reported silicified *Bathymyonia nevadensis* (Meek), late Early Permian, probably Word equivalent; this is a first reported occurrence of this brachiopod species in Montana.

At the lower Canyon Creek gorge, several phosphate mines had been active in the Phosphoria Formation. In the Devils Hole and Cleve Mountain areas, only a few decimeters of phosphate ore can be found. The phosphate mineral in the ore is francolite, or hydroxy-apatite, except for one outcrop east of Cleve Mountain (trench shown on the topographic map), where similar-looking oölitic phosphate proves to be a mixture of the iron phosphate minerals strengite and metastrengite (identified by H.C. Skinner, Yale University), presumably formed by metasomatic reaction of francolite as a result of emplacement of the Pioneer batholith.

Mesozoic Strata

In the field, no angular discordance can be recognized between the Phosphoria and Dinwoody Formations. However, the data do not rule out a hiatus in sedimentation. The Dinwoody contains sedimentary features indicative of shallow-water deposition. It is overlain in most places by the Cretaceous Kootenai Formation, and the hiatus there suggests a major erosional interval, an interpretation supported by the presence of erosion-concentrated manganese deposits along the contact at many places. At one place (N1/2 sec. 15, T. 2 S., R. 10 W.), however, the Dinwoody is overlain by a red stylolitic limestone. This limestone contains the conodonts *Ellisonia* sp. and *Furnishius* sp. and is Smithian in age. This assemblage is typical of a shallow-water facies of the Thaynes Formation (B.R. Wardlaw, 1983, written commun.). As the Dinwoody that underlies the red limestone has its usual thickness, the relation suggests that the thickness of the Dinwoody in most of the area is the full original thickness. The Trapper Canyon section of the Dinwoody is the northernmost section measured by Newell and Kummel (1942).

Fossils recovered from the Dinwoody, in the Canyon Creek area (NW1/4SW1/4 sec. 10, T. 2 S., R. 10 W.), include *Lingula borealis* Bittner, *Eumorphotis* cf. *E. multiformis* (Bittner), and *Unionites*? sp., suggesting a probably Scythian age for the middle part of the formation (J.T. Dutro, Jr., 1974, written commun.). In addition, cf. *Anodontophora* sp. and *Myalina*? were recovered from the Dinwoody from Quartz Hill Gulch area (NE1/4 sec. 20, T. 1 S., R. 10 W.) (B.R. Wardlaw, 1977, written commun.).

Myers (1952) mapped a narrow band of Triassic Woodside(?) Formation and of Jurassic Morrison(?) Formation in the eastern part of the present Twin Adams 7 1/2-minute quadrangle, and Sharp (1970) also recognized a narrow band of the Triassic Woodside Formation in the same area. The pre-Kootenai unconformity is easily mapped over much of the area because the basal Kootenai beds are resistant, ledge-forming conglomerate and sandstone unlike any of the older sequence. Just north of the Trapper Creek Road (SW1/4SW1/4 sec. 23, T. 2 S., R. 10 W.), and also on the north slope of the 8,910-foot peak of Sugarloaf Mountain, the unconformity between the Dinwoody and the Kootenai is marvelously exposed. The Trapper Creek locality (fig. 13) shows a small angular discordance, and local depressions in the erosion surface are filled with pebbles of the basal conglomerate bed of the Kootenai Formation. The Sugarloaf locality was mapped by Myers (1952) as a slumped block of Permian-Triassic beds; as the angular discordance is actually exposed, I interpret the basal Kootenai to rest on beveled Dinwoody unconformably with a restored angular discordance of 40°.

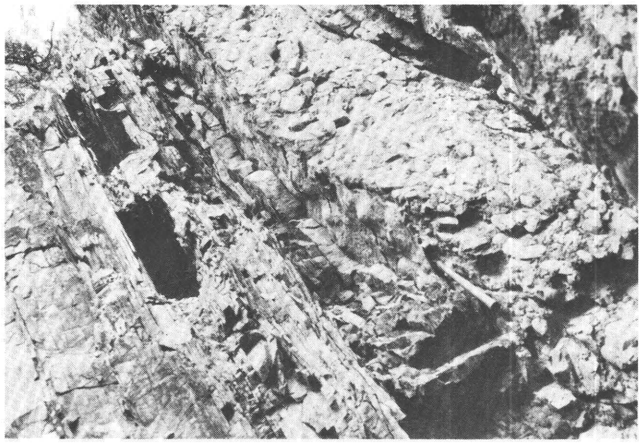


Figure 13. Unconformity between the Triassic Dinwoody Formation and basal conglomerate of the Lower Cretaceous Kootenai Formation. Location is north of Trapper Creek Road just west of confluence of Sucker Creek, 20 m above road level. Hammer is on unconformity, and plane of hammer is parallel to contact. Angular discordance is about 15°; pebbles of Kootenai conglomerate fill local depressions of the unconformable surface developed in the Dinwoody.

Elsewhere in the area, the Kootenai rests on the Phosphoria and possibly even on the Quadrant. (See also Myers, 1952.) Although the relations are conceivably due to faulting rather than unconformity, the fact that the beds above the contact are basal Kootenai lends credence to the unconformity interpretation and indicates that considerable erosion preceded Kootenai deposition.

The upper contact of the Kootenai with the Colorado Group is generally easily placed, even in the absence of the distinctive dark-gray gastropod (*Reesidella montanensis*) limestone[1] used to define the top of the Kootenai, because the argillites of the Kootenai are varicolored whereas those of the Colorado are drab olive, and the basal Colorado is

[1]In addition to these gastropods, a fish tooth identified as ?*Xiphactinus* sp. (F.C. Whitmore, Jr., and J.T. Dutro, Jr., 1974, written commun.) was found in the same bed near the head of Dry Hollow Gulch.

commonly, though not invariably, graywacke and sandstone. This shift in the nature of the fine-grained beds indicates a change of source area; the change may be related to the emplacement of thrust sheets of upper Paleozoic beds in the area to the west that supplied the bulk of the Colorado sediments, discussed below. At a few places, for instance along Trapper Creek Road just west of Glendale, there are local varicolored argillites in typical Colorado strata; how these beds originated remains a problem, but lack of structural disturbance suggests that they are not in-slumped Kootenai; rather, local persistence of a different source of sediments is indicated. Volcanic ash beds are common in the Colorado, but so far only one outcrop of ash has been found in the Kootenai; it is on the north side of Cherry Creek Road opposite Gross Ranch.

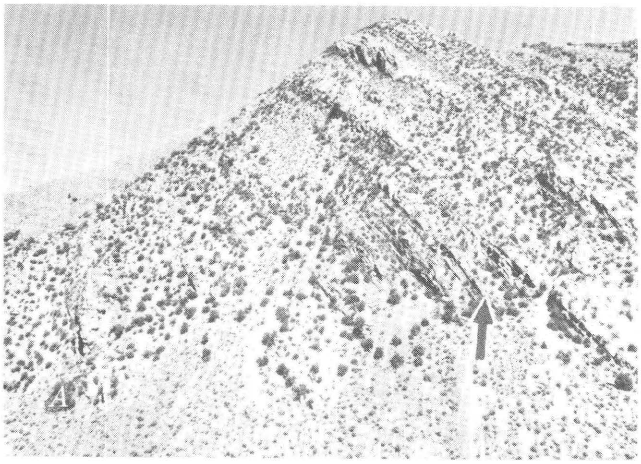
The volcanic and volcanogenic components of the Colorado Group include sandstones and white-weathering indurated ash beds approaching porcellanite; these ash beds failed to provide meaningful K-Ar dates presumably due to cation exchange since deposition; no zircon has been seen. The ash beds generally do not exceed a few decimeters in thickness. They are very fine grained and are presumably a distal deposit. Ash beds that show close-packed, ellipsoidal shapes about 5 mm across (for example, at center sec. 36, T. 2 S., R. 10 W.) may be indurated accretional lapilli. Source of the ash is not known, but the general age relations suggest possibly the Elkhorn Mountains Volcanics and (or) the 73-Ma volcanic rocks in the Argenta area (mapped by Myers, 1952, as Tertiary andesite; dated by Snee and Sutter, 1979).

Impure, massive feldspathic and lithic sandstone and conglomerate in the Colorado are interpreted as deposits in braided stream channels or possibly channel levees. Fine-grained, carbonate-bearing mudstone and olive-drab calcareous argillite beds, locally containing light-gray micritic limestone, are lateral equivalents of the coarser grained strata (for instance, basal Colorado at Storm Peak, below the ridge-forming massive sandstone; compare with basal beds of impure sandstone along lower Trapper Creek (fig. 14*A*) and along Canyon Creek road at the Canyon Creek-Trapper Creek divide). These are interpreted to be backwater facies and flood-plain facies; the carbonates might have been formed in local ponds. Rare outcrops of detrital carbonate, such as those south of Trapper Creek road near Glendale (NW1/4 sec. 25, T. 2 S., R. 10 W.), might have had local supplies of carbonate from restricted drainage subsystems.

The enormous thickness of the Colorado Group, compared to all preceding stratigraphic units in the area, reflects the tectonic activities that began in early Late Cretaceous time. The general nature of the coarse clastic components of the material in the Colorado is not very different from that in the underlying Kootenai. Much more coarse clastic material is present in the Colorado, however, and the clastic component is less mature and less diluted by carbonate than that of the Kootenai. I interpret the source area for much of the Colorado as being to the west, as suggested by scanty transport direction indicated by the imbricated cobbles of the conglomerate beds; source direction that might be implied by crossbedding is being studied by R.K. Schwartz of Allegheny College. The conglomerate lenses (fig. 14*B*) may have been gravel bars in a braided stream system; sandstone lenses in the conglomerate are apparently foreset beds in the downstream sides of gravel banks. On the basis of the nature of clasts in the sandstone and conglomerate, I interpret the source terrane as imbricately thrusted sheets of upper Paleozoic rocks emplaced during the Sevier orogeny, possibly as part of the northward extension of the Idaho-Wyoming thrust belt. These rocks are known to underlie thrust sheets of Middle Proterozoic rocks in the west Pioneer Mountains and may have created highlands there as well as over the present Big Hole Valley. This interpretation of the nature of the source terrane for the Colorado is supported by findings of fossil-bearing cobbles, which are composed of rocks ranging from the Madison Group to the Phosphoria Formation, in the conglomerates of the Colorado.

Deposition of the Colorado Group in the Vipond Park area overlapped in time with the evolution of the Mesozoic phase of the Pioneer batholith; the younger Cretaceous plutons approximately coincide in age with the deposition of the upper beds of the Colorado. The batholith thus invaded an area of active stream development and subsidence. It is possible that the stream system was in a graben, analogous to the modern relations in the Rio Grande valley of New Mexico where subsurface magma chambers have been detected by vibroseismic methods. (See, for example, Brown and others, 1979.)

Ryder and Ames (1970) described microfossils as old as the Aptian (early Late Cretaceous) in rocks assigned to the Beaverhead Formation from the Lima region in Montana, south of the Pioneer Mountains. The sequence is lithically closely comparable to rocks mapped in the Vipond Park area as the Colorado Group (Robert Scholten, 1979, oral commun. in the field; see also Nichols and others, 1985). In the map area, the Colorado Group is sparsely fossiliferous. Aside from some unidentifiable carbon films of plant stems, one collection of palynomorphs from a dark-gray silty shale in the upper member of the formation was found just south of Brownes Creek (fig. 14*C*, which also shows a large flame structure), east of the Beaverhead National Forest boundary (6,200 ft south of 6,280-foot knoll, NW1/4 sec. 19, T. 3 S., R. 9 W.). D.J. Nichols identified these forms, which date the bed as mid-Campanian to Maestrichtian: *Aquilapollenites* sp., *Balmeistporites minor*, *Cicatrocosisporites* sp., *Proteacidites thalmannii*, *Quadripollis krempii*, *Taxodiaceaepollenites hiatus*, and *Tricolpites interangulus*. This age agrees with the previous correlation of the Colorado Group in the area to the south (Myers, 1952, unit Ku_4), based on leaf impressions, with the Judith River Formation.



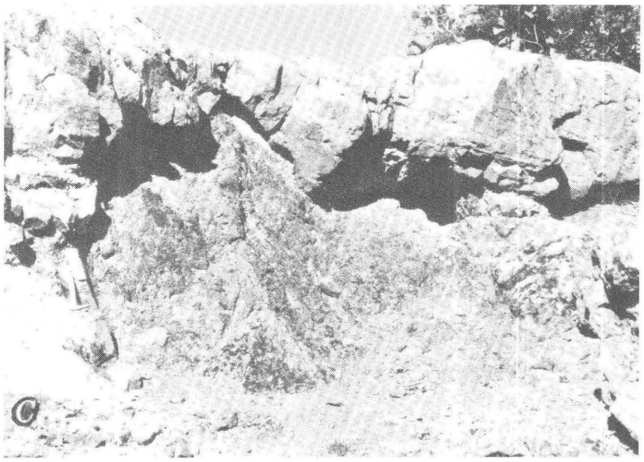


Figure 14. *A*, Contact between Kootenai Formation and Colorado Group, on north side of Trapper Creek, south slope of 6,484-foot peak. Visible relief is about 500 ft (150 m). Top of the Kootenai is a gray gastropod limestone, succeeded by a massive (5–10 m) graywacke bed of the Colorado. Contact is just above right-hand talus slide marked by arrow. *B*, Conglomerate in the Colorado Group, interpreted as gravel bar; pen just above middle of picture is shown for scale. Location is near 6,690-foot knob, Lelo (Louie Lowe) Basin. Large elongate cobble in lower right corner is a sheared clast of volcanic ash typically found as beds in the formation. Most clasts are chert and quartzite typical of the upper Paleozoic formations of the area. *C*, Dark-gray fissile shale in the Colorado Group forming a large flame structure; note that adjacent sandstone beds have been turned to upright position but that the overlying sandstone bed continues without interruption across the structure. Palynomorphs described in text are found in shale in foreground. Location is ridge south of Browns Creek, just east of National Forest boundary.

Tertiary Strata

Eocene volcanic rocks in the map area consist largely of flows and flow breccias having compositions ranging from basaltic andesite through andesite and dacite to rhyodacite. (SiO_2 content, on volatile-free basis, ranges from 52.9 to 64.5 percent.) These rocks are here named the Trusty Gulch Volcanics, which have their type section on the slopes immediately north of Trusty Gulch in the northeast part of the Vipond Park quadrangle. Two separate flows here have whole-rock K-Ar ages of 47.0±1.6 and 47.3±1.6 Ma (Marvin and others, 1982). Other flows in the Vipond Park area and nearby having similar petrographic character and age are assigned to the same unit and are so depicted on the geologic map. An area of outcrop near Dry Hollow Gulch (SE1/4 sec. 13, T. 2 S., R. 10 W.) of volcanic or volcanic neck rock (Tlc), however, is petrographically and geochemically identical to the Lowland Creek Volcanics (Smedes, 1962); it has a K-Ar age of 50.4±1.7 Ma on a biotite separate and is not included in the Trusty Gulch Volcanics. Likewise, pumice and ash deposits (Tlca) that bear strong petrographic and chemical resemblance to the Lowland Creek Volcanics, intercalated with the flow units of the Trusty Gulch Volcanics, are assigned to the Lowland Creek. Detailed discussion of the petrogenesis of the rock units is deferred to a later report.

A typical flow of the Trusty Gulch Volcanics has a thin chilled zone at the base, followed upward by a zone showing flow-shear strain (phenocryst orientation and incipient fracture); this zone is succeeded, for the thicker flows, by a massive zone. The next zone above is amygdaloidal, and the amygdules may be flow-stretched. Finally, the typical flow has an upper carapace of highly oxidized, brick-red volcanic rubble. These features, as well as the abundance of columnar joints and the total absence of features indicative of subaqueous flow, show that the Trusty Gulch Volcanics were erupted on land. The open space between blocks of the rubble is locally filled with crossbedded volcanogenic sandstone; thus, at least some of the oxidized rubble is probably tops rather than bases of flows.

An interval of erosion apparently preceded Trusty Gulch volcanism, because the volcanic rocks rest on fresh or nearly fresh substrate and occur without any paleosoil and with little or no indication of a weathering interval. On the open ridge east-southeast of the 8,344-foot peak (secs. 2 and 3, T. 3 S., R. 10 W.), the volcanic rocks rest on Middle Proterozoic rocks of the Morrison Hill klippe. The volcanic rocks locally baked the clay galls into tactite. The pink Precambrian quartzite itself is baked to a chocolate-brown color. The open ridge is thinly veneered with remnants of volcanic rocks and is an exhumed Eocene erosion surface.

As mentioned, rocks identified with the Eocene Lowland Creek Volcanics are found in the Vipond Park quadrangle, and pumiceous ash beds, some deposited by air fall but others apparently reworked by running water, are assigned to it; these rocks are labeled "Tlca" on the geologic map. The synchronous evolution of two chemically and petrographically distinct volcanic series at the same place requires intricate, nonintersecting plumbing systems that are not understood. The field relations are similar to those reported in the area south of Anaconda (Iagmin, 1972) and may well be of regional scope.

Immediately east of the Wise River valley, in both the Stine Mountain and Maurice Mountain quadrangles, there are several areas of pumiceous ash and ash-matrix conglomerate strata (fig. 15). These strata typically occur on ridgetops and are made conspicuous because of the distinct yellow-ochre color of the ash. In areas of good exposure, such as the north side of the ridge north of Fourth of July Creek, the strata consist of alternating layers of nearly clast-free, well-sorted pumiceous ash, a few centimeters to several decimeters thick, and of boulder beds, several meters thick each, that contain poorly sorted and poorly rounded clasts as much as 2 m across supported in a pumiceous ash matrix.

The clasts are from the Middle Proterozoic quartzites and from the Cretaceous and early Tertiary intrusive rocks of the Pioneer batholith. All the clasts are lithically those of bedrock that crops out within a few kilometers of the clasts, indicating a local origin of the boulder beds. The clast-free layers are interpreted to be the original ash beds, with or without reworking by water. The boulder beds are interpreted to be locally derived paleolandslide material associated with the explosive volcanism, the clasts being the material entrained during mass movement. The nearly horizontal layering suggests catastrophic events that left deposits on nearly flat topography. Some of the deposits are superficially like moraine material because of the nature of

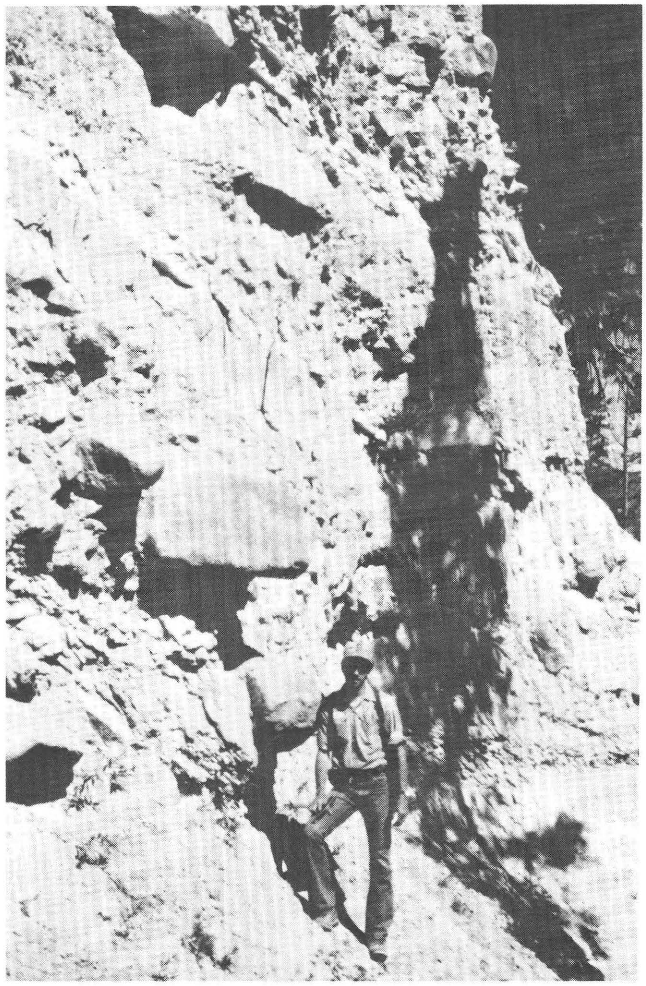


Figure 15. Oligocene pumiceous ash interbedded with paleolandslide material that has ash matrix and clasts of quartzite and granite. Exposure on north side of ridge north of Fourth of July Creek (NE1/4 sec. 14, T. 2 S., R. 12 W.). A clast-free layer occurs above large boulder next to man's head; another occurs just to left of tip of tree's shadow below protruding block near top of picture.

the clasts; this problem is compounded by the numerous postglacial landslides of the ash-rich deposits. Careful examination of the matrix helps to identify the origin of the rock, but locally, for instance on the bold bluffs east of the mouth of Elk Creek, distinguishing a pumice-matrixed landslide from a glacial moraine remains a problem. The landslide beds are the first direct indication of unroofing of the batholith. Unfortunately, although field relations indicate proximity of the source areas, no eruptive center that could have produced such volumes of siliceous ash beds has been identified.

The age of the ash-conglomerate sequence is given by a single fission track date of 27.0±1.3 Ma on zircon separated from the ash east of Happy Creek (Pearson and Zen, 1985). This is the same age as a rhyodacite (SiO_2 content, 56.8 percent on a volatile-free basis) that occurs northeast of Morrison Hill in the Vipond Park quadrangle (SW1/4 sec.

27, T. 2 S., R. 10 W.) and that gave an age of 27.3±1.0 Ma (Marvin and others, 1982). Although pumice of this age has not been identified east of the east Pioneer range divide, the possibility of its occurrence cannot be excluded. The probable presence of volcanic soil on Vipond Park, best indicated in wet weather (the road has now been covered by road metal), suggests the presence of ash beds here.

Stream gravel (fig.16; shown on pl. 1 as QTg) and landslide deposits of possible late Tertiary age have not been studied.

Quaternary Deposits and Other Geological Records

Quaternary deposits include gravel, talus, landslide deposits, some travertine and hillside limestone conglomerate (cemented slope wash), slope wash, alluvium, and various forms of glacial deposits (moraine and lake beds). These deposits have not been studied. Extensive lateral moraines noted in the field are shown on the map; lodgment tills are especially notable at Yanks Pasture (fig. 17) and near Siria Ranch. Cirques, arêtes in the higher elevations (fig. 18), ice-margin channels, and U-shaped valleys record the valley glaciation; some outcrops on steep north-facing slopes, notably along upper Rock Creek, still preserve specular polish. Recessional terminal moraines are best seen in the lower Rock Creek gorge; Brownes Lake owes its existence to one of these. I have made no effort in this study, however, to distinguish ice-contact gravel deposits from lodgment and ablation tills.

Preceding the valley glaciation, the Pioneer Mountains apparently went through a stage of local icefield development, as already noted by Alden (1953, pl. 1). In the east

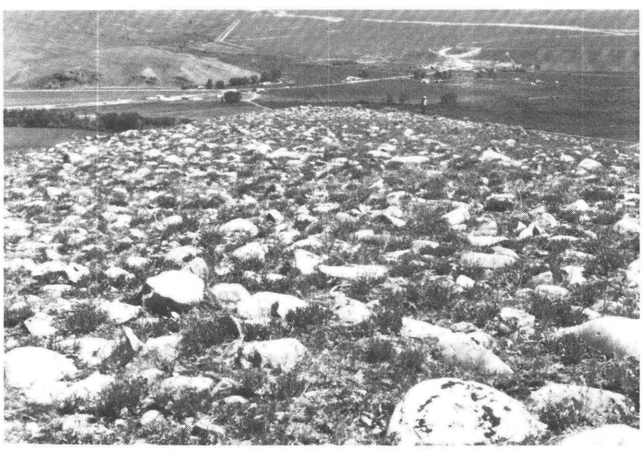


Figure 16. Large rounded boulders on ridge top, southwest of Divide (visible in background), interpreted as Tertiary stream gravel. Boulders are nearly exclusively Proterozoic Y quartzite of types not found east of the Fourth of July fault. Man in distance indicates scale.

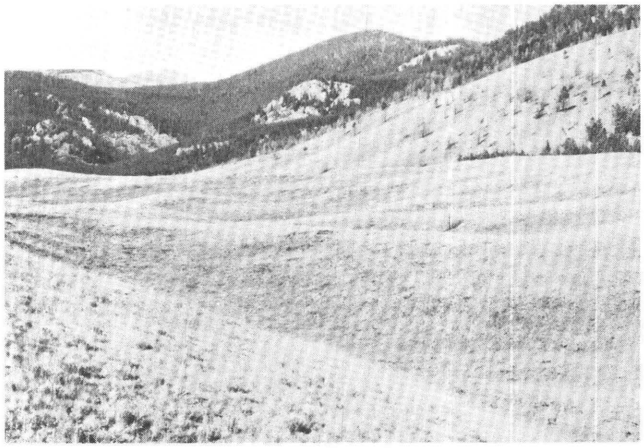


Figure 17. Lodgment till, Yanks Pasture along Trapper Creek. Slope in lower left corner is part of lateral moraine that borders the stream course and is younger than the lodgment till. View is to southwest.

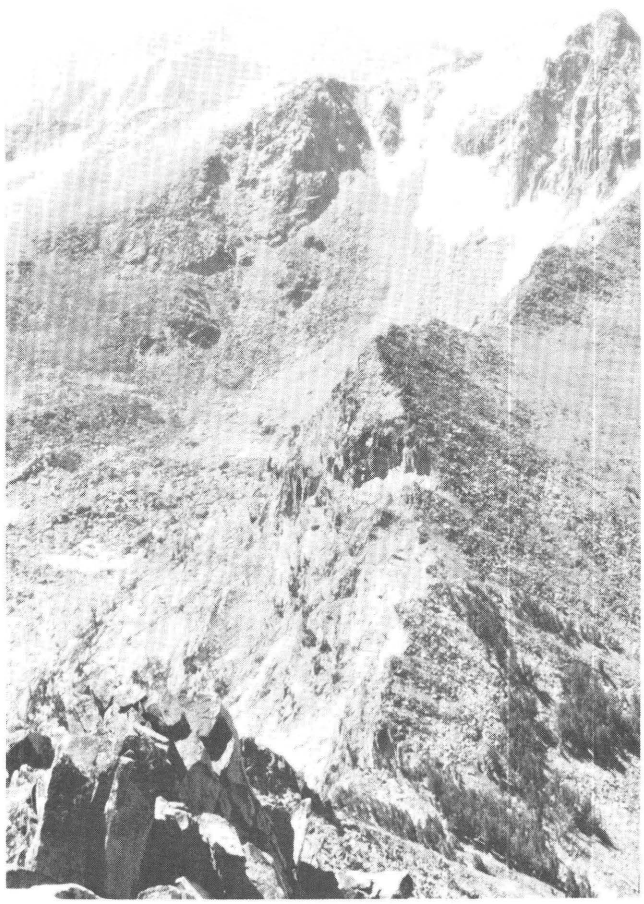


Figure 18. The high country of the east Pioneer Mountains that is underlain by the batholithic rocks. View is south from summit of 10,352-foot peak. Mount Tahepia (10,473 ft) is in right background; northeast part of peak was destroyed by prehistorical landslide (slide material visible in left center of picture). The two high peaks on the horizon are Tweedy Mountain (11,154 ft) on the left and Torrey Mountain (11,147 ft) on the right, the highest and second highest peaks in the Pioneer Mountains.

Pioneers, evidence for this development comes from the distribution of exotic erratics in the glacial deposits, not only in valleys but along ridges. These erratics include bedrock types not found in the present valley; they must have crossed ridge lines. As a rule of thumb, the lowest level of moraine deposits in the valleys is about 6,000 ft, whereas on ridges they are at least as low as about 8,000 ft. Along the main mountain divide such as behind Tendoy Lake, the uppermost few meters or tens of meters of the divide are held up by fragile pinnacles and teetering blocks of granitic rocks (fig. 19*A*); these rocks are commonly deeply weathered and rest on a fresh, solid, and smooth granite surface (fig. 19*B*). I interpret them to have stuck above the ice surface at least during the terminal stages of icefield glacial erosion, when at the highest reaches of the mountain the ice and rock divides coincided. As support for this reconstruction, I note that no material unequivocally from the west Pioneer Mountains has been found in the east Pioneer glacial deposits.

Major tributaries of the Wise River, notably Pattengail Creek, Lacy Creek, and David Creek, show exceptional development of U-shaped valley profiles. On the south shoulder of Stine Mountain (elev. 9,490 ft) at about the 9,250-foot level, the bedrock shows glacial striae that trend east-west and end abruptly at the edge of the cirque headwall (fig. 20). An episode of glaciation that included the higher ridge crests must have occurred. The location of the grooves, just below the summit of the highest peak of the west Pioneer Mountains, suggests that the west Pioneers were inundated by ice at one time.

Isolated higher peaks in the east Pioneer Mountains, such as Storm Peak (9,492 ft), Sugarloaf Mountain (8,912 ft), and Sheep Mountain (9,578 ft), show scant evidence of glacial erosion even though they are at elevations that in the main range supported icefield and valley glaciations. Sheep Mountain has one notable cirque; the others have none. These mountain peaks all show patterned ground near their summits (fig. 21). There is no evidence that the patterned ground is active today.

At several places, for instance in the Lelo (Louie Lowe) basin and immediately to the east, a thin (a few centimeters), conspicuous layer of white ash occurs about 1 m below the present surface of the olive-drab colluvium. The ash particles have textures (including bubble shapes), glass chemistry, mineralogy, and a glass refractive index all supporting correlation with the Mazama ash deposit (6,845±50 B.P., Bacon, 1983). Alden (1953) described ash in similar positions elsewhere in southwestern Montana, but the localities he mentioned were not visited during this study. Several road cuts along the old alinement of U.S. Route 91 between Divide and Butte showed similar ash deposits; samples were collected from some of these deposits, but most of the cuts were destroyed during the con-

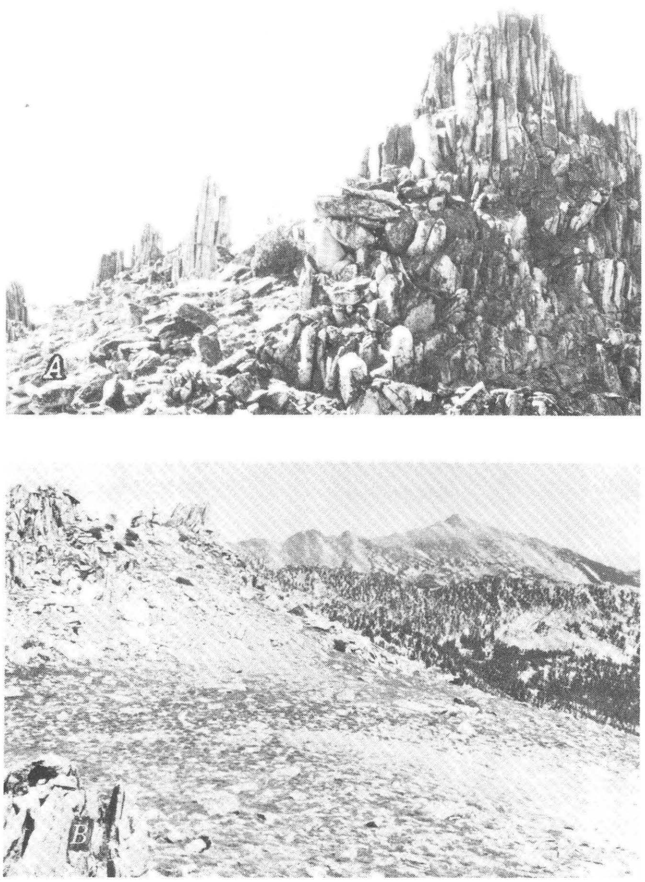


Figure 19. *A*, Rock pinnacles of deeply weathered piles of teetering rock, interpreted as not to have been covered by the icefield glaciation, rising 10 to 20 m above the smooth glaciated surface. Divide is above rock basin south of Tendoy Lake, looking south. *B*, View northeast from same spot as figure 19*A*, with Granite Mountain in background. In contrast to the pinnacles, the smooth surface of most of the divide area has been glacially polished.

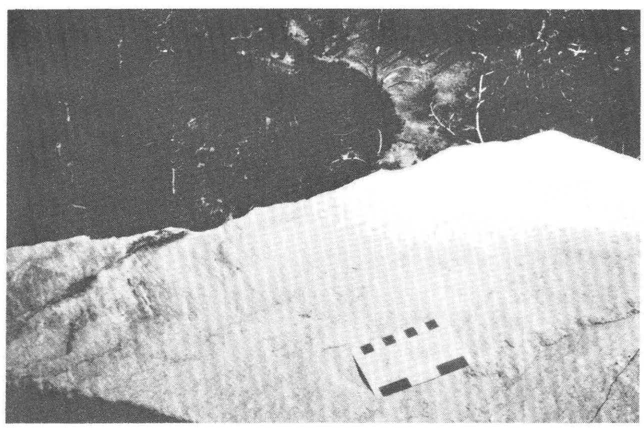


Figure 20. Glacial striae on flat-lying quartzite, 9,230 ft on south shoulder of Stine Mountain. The striae stop abruptly at brink of headwall of later cirque glaciation; the uppermost of the Grouse Lakes, visible in background, is nestled in the cirque. Formation of striae probably required the ice to be at least 100 m thick. Because the top of Stine Mountain, the highest peak of the west Pioneer Mountains, is at 9,490-foot elevation, the implication is that the peak was also under ice.

struction of Route I–15. An excellent deposit remains visible, however, in a borrow pit west of I–15 at mileage 118.9 south of Butte; this deposit is 10 cm thick, 1.5 m below the surface, and is clearly visible from the highway (fig. 22).

STRUCTURE

Folds

The east-central part of the Vipond Park quadrangle is dominated by a large syncline that trends and plunges southeastward. This syncline, here designated the Cherry Creek syncline, controls the outcrop pattern of most of the Colorado Group (pl. 2). The center and thus structurally highest part of the syncline is occupied by the Morrison Hill klippe and many of the Tertiary volcanic rocks.

Isoclinal, recumbent folds in Paleozoic and Mesozoic rocks below the Morrison Hill klippe in the area just south of Canyon Creek are interpreted as local rather than regional in scope; they reflect the movement of the superjacent thrust sheet. The folds can be seen best in the contact between the Lodgepole and Mission Canyon Limestones at the base of the cliffs northwest of the 7,779-foot peak near Canyon Creek Guard Station; a good place to view the structure is on the road to Vipond Park at the top of the steep grade up the face of the moraine. The keel of the recumbent fold can be followed on the steep, heavily wooded slope south of Canyon Creek due south of Devils Hole (N$^{1}/_{2}$ sec. 16, T. 2 S., R. 10 W.), where right-side-up beds of the Dinwoody Formation at stream level can be traced up slope into nearly vertical beds. At the top of the steep slope (open pasture, extreme west center of sec. 15, T. 2 S., R. 10 W.), upside down and gently south-dipping basal conglomerate of the Kootenai Formation occurs; this conglomerate is covered to the south successively by the Dinwoody, Phosphoria, and Quadrant Formations, which together constitute the overturned limb of the recumbent fold.

Two long and narrow southeast-plunging synclines, recognized by Richards and Pardee (1926) and Theodosis (1956), flank the east side of the Cherry Creek syncline. These two synclines are separated from the principal syncline by a partly recumbent anticline. The two subsidiary synclines are mutually separated by a high-angle fault, the Trusty Gulch fault, which locally brings the Colorado Group against the Madison Group; this contact is readily seen along the jeep trail southeast of Short Slope. Eocene volcanic rocks lap across the fault. The fault is no older than the synclines, which are related to the emplacement of the Morrison Hill klippe which in turn rests on rocks of the Colorado Group. Therefore, the folding is probably latest Cretaceous to Paleocene and thus is a Laramide structure.

The remnant of an important compound anticline occurs south of the Cherry Creek syncline; the oldest rocks exposed are the Black Lion Conglomerate, visible at Hecla and above Lion Creek northwest of Barbour Hill. Most of the anticline, including all of its southwest limb, has been removed by the intrusion of the plutonic rocks. Though the intrusions locally distorted the geometry of the fold, the overall relations show that the folds predate the intrusions and that the various plutons plucked away the structure as if they were cookie cutters. A remnant of the south limb of the anticline is preserved, partly as roof pendants, on Keokirk Mountain. The Black Lion Conglomerate of the Grace Lake area may be part of the same structure; the structurally complex south-facing section north of upper Gold Creek in the Maurice Mountain quadrangle (Pearson and Zen, 1985) is also part of the same structure.

Thrust Faults

The important thrust faults mapped in the northern Pioneer Mountains are flat lying or nearly flat lying, and

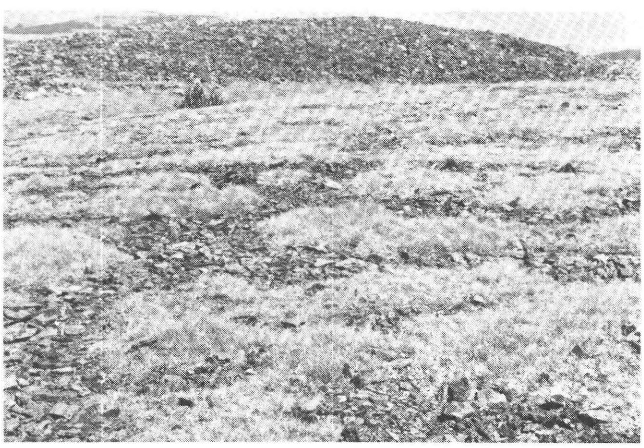


Figure 21. Frost polygons, north slope of Sheep Mountain, looking northeast toward rampart of conglomerate talus above the cliffs. The polygons are about 3 m across.

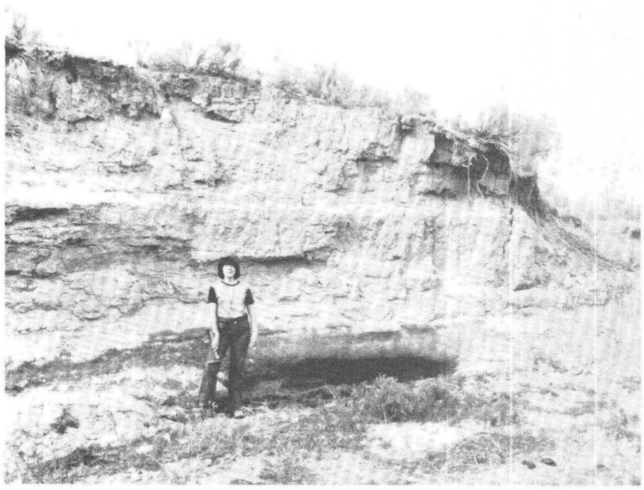


Figure 22. Mazama ash in colluvium, borrow pit west of Interstate 15 south of Butte, at mileage 118.9. The light ash layer is about 10 cm thick.

bring Middle Proterozoic sedimentary rocks and possibly Early Proterozoic gneiss over rocks that range from other Middle Proterozoic rocks and the Cambrian(?) and Cambrian Black Lion Conglomerate to the Cretaceous Colorado Group. In the Stine Mountain and Maurice Mountain quadrangles, mapping of structural features in the Middle Proterozoic rocks was hampered by the extensive development of talus slopes and the paucity of outcrops. At least two thrust sheets are involved, as determined by slivers of Paleozoic rocks between them. This relation is the basis for a conceptual model of stratigraphic-sedimentary relations, to be discussed in a later section. The two thrusts are designated the Wise River thrust (lower) and the Pattengail thrust (upper).

The slivers of Paleozoic carbonate rocks occur near and on both sides of the mouth of Ross Gulch (Stine Mountain quadrangle); the more northerly of the two tracts was first recognized by Calbeck (1975), who interpreted it as sheared carbonate of unknown age along the high-angle Ross Gulch fault. The rocks consist of assorted calcareous rocks (gray limestone and dolostone, yellow calcareous siltstone, yellow- and orange-weathering marl) indicative of the Amsden Formation. The tract south of Ross Gulch is poorly exposed, but the abundant Amsden float on a grassy terrace amidst talus of Proterozoic quartzite and the projected geometric relations suggest a sliver at the base of a thrust fault. This interpretation is consonant with the behavior of the Amsden at the base of known Middle Proterozoic thrust sheets such as at Morrison Hill or at the base of the Johnson thrust in the Wise River quadrangle (Fraser and Waldrop, 1972). This sliver of the Amsden is at the base of the upper thrust slice.

The Black Lion Conglomerate occurs at the base of the eastern cliff face of the 8,920-foot ridge north of Maurice Mountain and is succeeded upward by the sequence of Maurice Mountain. This contact is interpreted as the base of the lower, Wise River thrust that cuts down into the Paleozoic substrate. However, about 50 m below the ridge line, a zone several meters thick of conglomerate lithically identical to the Black Lion recurs, bounded both above and below by the Maurice Mountain. This conglomerate could be a lithic variation in the Maurice Mountain, but it is not found elsewhere in that sequence. My favored interpretation is that this is a sliver of the Black Lion. If this is true, then other splays of the thrust sheets might remain undetected.

At Grace Lake, the same thrust contact reappears and again brings quartzite of the Maurice Mountain sequence onto Black Lion Conglomerate. The rocks on both sides are right side up and show no evidence of local deformation; a prominent white band on the cliffs is a weathering phenomenon, not reflecting the rock lithology. The thrust sheet at one time must have extended at least another 1.5 km farther east, because along the trail north of Grayling Lake, a large xenolith of the Middle Proterozoic quartzite of the Maurice Mountain sequence occurs in the Cretaceous Grayling Lake Granite. The thrust fault and its attitude can be readily seen from the cirque of Black Lion Lake and along either side of the upper Gold Creek valley.

Within the same thrust sheet, the Maurice Mountain sequence is succeeded northward and eastward by the younger Boner Knob sequence in the area of Boner Knob, in the Wise River quadrangle, and in the central part of the Vipond Park quadrangle (Morrison Hill klippe). All known exposures of the base of the lower thrust are east of the Fourth of July fault, however; west of this fault, the thrust surface has been down dropped below the present level of erosion.

At Morrison Hill, the klippe was cut by north-south-trending high-angle faults that are younger than the Eocene Trusty Gulch Volcanics. The klippe is underlain on the north side by imbricated slivers of the Amsden, Quadrant, and Phosphoria Formations; outliers of the Quadrant form prominent free-standing panellike cliffs visible from Melrose. The Middle Proterozoic rocks of the Morrison Hill klippe are nearly everywhere badly shattered; few large outcrops exist. Thrusting is inferred to have occurred near the land surface, in part because of the occurrence of a conglomerate south of Melrose in the area southeast of the present klippe (fig. 23). Most of the clasts of the conglomerate are red quartzite similar to rocks in the klippe—they are subrounded to rounded, are as much as 30 cm across, and have minor components of clasts of Paleozoic and Mesozoic rocks. The conglomerate is probably to be correlated with the Beaverhead Conglomerate (Lowell and Klepper, 1953). It is underlain by the Colorado Group and is overlain by Eocene volcanic rocks (47.1±1.6 Ma based on whole rock K-Ar); presence of the conglomerate indicates the availability of the klippe rocks to erosion at the time of or shortly after their emplacement.

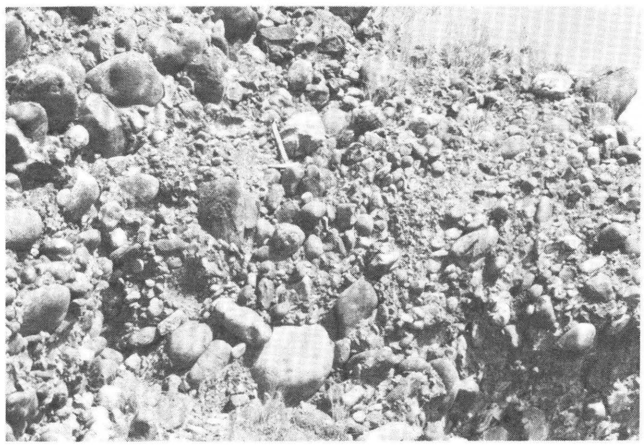


Figure 23. Conglomerate, correlated with the Beaverhead Conglomerate, east of Interstate 15 south of Melrose, center sec. 23, T. 3 S., R. 9 W., Earls Gulch quadrangle. Nearly all the cobbles are pink and red feldspathic quartzite typical of that of the Proterozoic Y sequence of Morrison Hill klippe. A few Paleozoic limestone cobbles (one is next to the hammer) and rare pale-green volcanic ash pebbles (not visible on photograph) exist.

The base of the Morrison Hill klippe is not well exposed, but its location can be fixed quite closely. In addition to use of floats of the Colorado Group, a line of seepage springs occurs in the contact zone; presumably, the springs owe their existence to meteoric water which moves down through the shattered rocks of the klippe and emerges upon encountering the relatively impermeable argillite of the Colorado Group.

The klippe is not found south of Cherry Creek, even though projection of attitudes indicates that it should be on the north slope of Storm Peak. A pre-Eocene high-angle fault or flexure (sections J, L) in the southwest limb of the syncline could explain the relations; the minimum Eocene age is based on concordant altitudes of Eocene ash beds across Cherry Creek valley. Elsewhere, as on Vipond Park, isolated outcrops of Middle Proterozoic rocks surrounded by much younger rocks are interpreted as partly buried erosional remnants of the same klippe. The present upland surface may coincide to a significant degree with the Late Cretaceous surface over which the klippe was emplaced.

Another low-angle fault in the eastern part of the Vipond Park quadrangle is within the upper Colorado Group. This contact is exposed mainly south of Cherry Creek near the Beaverhead National Forest boundary and is manifested by the abrupt juxtaposition of moderately to steeply dipping Colorado beds against nearly flat-lying strata. (A good viewpoint is along Rock Creek Road near the National Forest boundary.) At a few places, outcrops from the two sides are within a few meters of each other and persist in their attitudes; thus, a structural discordance must be present. The collection of palynomorphs was made in the flat-lying sequence. An unconformity is ruled out, as it would require a deformational episode not otherwise recorded in the area within the time span of the upper Colorado. Where the upper, nearly flat-lying sequence came from, and when and how, are unresolved problems; the sequence may be a gravity slide that came off the rising land above the developing batholith, or it may have been shoved off the front of the advancing thrust sheets.

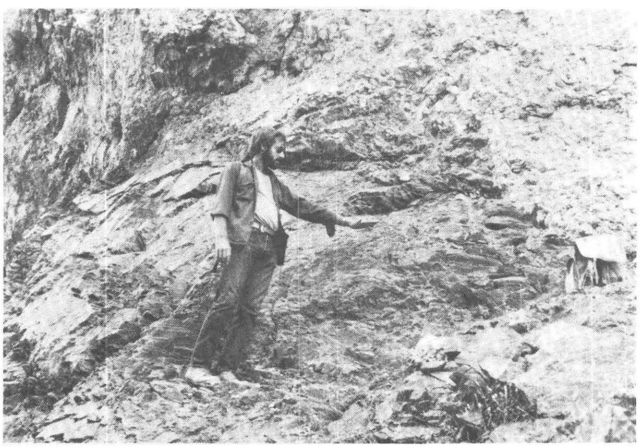


Figure 24. Devonian Jefferson Dolostone thrust upon the Mississippian Lodgepole Limestone. Location is west slope of 8,938-foot peak west of Lion Mountain. The Jefferson is highly brecciated and yellow, typical of that below the main thrust of Proterozoic quartzite at head of Gold Creek; its presence here suggests that this thrust extended this far east. Man's hand is parallel to bedding in the Lodgepole.

Minor thrust faults include one that is well exposed on the east face of Lion Mountain in the Hecla basin. The Red Lion Formation, here a calcareous quartzite, is repeated. This fault may be related to one or more faults encountered underground (Karlstrom, 1948), but it is not the "Great Fault" postulated by Karlstrom; as indicated previously, such a fault is not needed to explain the field relations. A small area of Devonian Jefferson Dolomite is in fault contact against the underlying Lodgepole Limestone on the west slope of the 8,938-foot peak west of Lion Mountain. Here, the rocks are brecciated and altered, identical to those exposed at the head of Gold Creek, and they may be a sliver that once underlay the extension of the Wise River thrust fault east of Grace Lake (fig. 24).

High-Angle Faults

Two major systems of high-angle faults occur in the map area; one system trends generally north, the other west-northwest (pl. 1). These two systems intersect to produce an intricate pattern of fault-bounded blocks.

West-Northwest Faults

The major faults in this group can be described as (1) the Sawmill Gulch fault, (2) the Trusty Gulch fault, (3) the group of faults in the Gray Jockey Peak-Swamp Creek area, and (4) the fault at Grace Lake and Moose Creek.

The Sawmill Gulch fault brings the Colorado Group against the Madison Group. The Colorado Group locally occupies lower slopes, and the fault might be interpreted as a low-angle thrust. However, the Colorado Group is in apparent stratigraphic continuity with rocks that flank it on the west-northwest side. As the trace of the fault is independent of the topography, it must have a fairly steep dip. The trend is parallel to other faults in the area that can be shown to be high-angle faults. At the ridge northeast of Lime Kiln Gulch and west of the road, on the projected trend of the fault, the Mission Canyon Limestone is altered to massive jasper, and the Quadrant Quartzite is intensely brecciated. These deformed and altered rocks are assigned to the fault zone. The fault is probably the northern boundary of a prominent topographic feature in the adjacent Melrose quadrangle near Maiden Rock, where the mountain front is offset eastward several miles and where the ridge trends are west-northwest.

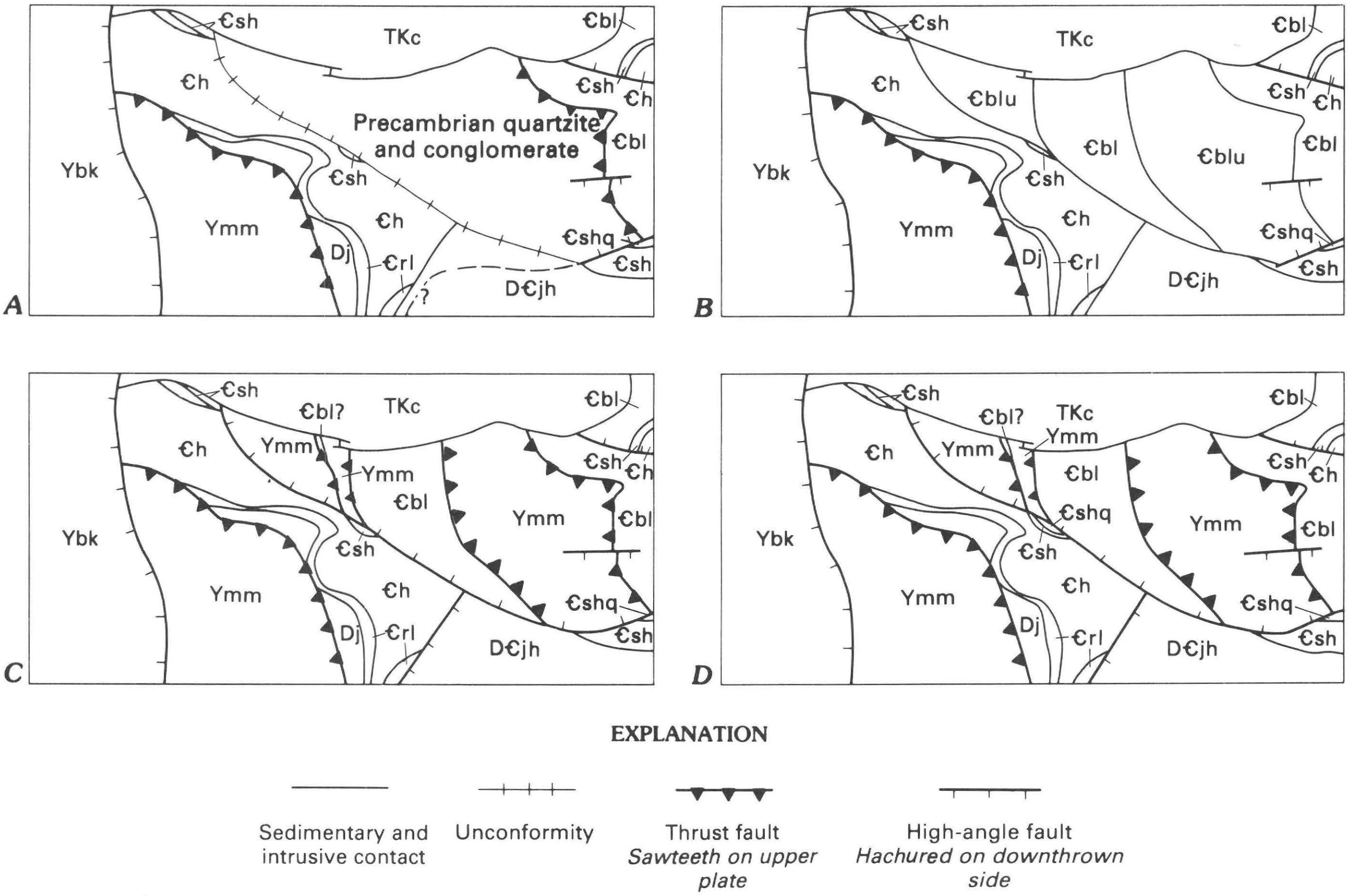
The southern boundary of the same topographic feature is south of Shepherd Mountain in the Melrose quadrangle and coincides with the northwest-trending Trusty Gulch fault. The latter fault appears to converge with the Sawmill Gulch fault in the area of Lime Kiln Gulch, despite the complications due to several splays, so that the area around Maiden Rock is a fault-bounded block. The Trusty Gulch fault separates two large synclines and shows stratigraphic throw of at least 700 m (upper Mission Canyon to lower Colorado). Eocene volcanic rocks lap across it with little or no perturbation, whereas the underlying rocks are folded and tilted. The dog leg of the fault at Dry Hollow Gulch may be caused by a later, left-lateral movement on a northeast-trending fault, but the relations cannot be proven in the field. A hillock of Quadrant Quartzite, at the dog leg just south of the jeep track and north of the Lowland Creek Volcanics, is badly shattered and partly jasperized. The proposed intersections of faults may account for the location of the volcanic neck as well as explain the intricate map pattern, which consists of fault blocks of the Quadrant, Amsden, and Phosphoria Formations, in the hills nearby.

A whole group of west-northwest-trending high-angle faults in the area of Swamp Creek and Gray Jockey Peak is probably a continuation of the fault system just described. This group of faults locally offsets the north-trending Fourth of July fault, though I shall show in a later section that the west-northwest fault *system* must have begun at a much earlier date. A major fault of the Gray Jockey Peak area is one that now occupies part of the Adson Creek valley. North of this fault, here designated the Adson Creek fault, the Hasmark Dolomite and upper member of the Silver Hill Formation rest on the Proterozoic sequence of Swamp Creek; south of the fault, both members of the Silver Hill Formation rest on the Black Lion Conglomerate. The Early Proterozoic gneiss-and-amphibolite is restricted to the area south of the fault; the sequence of Swamp Creek is restricted to the north. The Hasmark is little deformed or metamorphosed north of the fault but is more deformed and has lost many of its primary sedimentary structures to the south. The Jefferson Dolomite is a dark-gray, sugary, fetid rock to the north, but it is a cream-white dolostone with subsidiary dark-gray dolostone to the south. The Adson Creek fault thus appears to be an important tectonic-stratigraphic boundary that either was subjected to large lateral displacement or had a long history dating back at least to Silver Hill time. Unfortunately, its trace is lost in the covered area of Vipond Park; beyond the Park and south of Canyon Creek valley, the Adson Creek fault has not been recognized.

The Early Proterozoic terrane is intersected toward its south side by another high-angle fault. This fault is exposed on the headwall of the cirque north of Sheep Mountain, and it can be followed westward along a shallow gully. Both the Black Lion Conglomerate and the gneiss occur on both sides of this fault, however, so it is of local importance. The relation between the two rock sequences will be discussed in a later paragraph.

Figure 25 (right). Alternative explanations of relations north of Maurice Mountain. Heavy lines denote faults; thrust faults have teeth on the upper slice. Unconformities are signified by railroad track pattern. Formation symbols are used as on the geologic map (pl. 1) except as noted below. *A*, The Cambrian Silver Hill Formation unconformably rests on allochthonous Precambrian quartzite and is, therefore, itself allochthonous. Some of the younger Cambrian formations conformably above the Silver Hill must also be allochthonous; the geometric problem could be solved by extending the Grace Lake fault southwestward as schematically indicated. This is alternative (1) of the text. *B*, The quartzite above Grace Lake and underlying the 8,920-foot peak north of Maurice Mountain is not Proterozoic despite the lithic identity; instead this quartzite is a post-Black Lion sequence of Cambrian age, conformably between the typical Black Lion and the Silver Hill. This is alternative (2) of the text. *C*, The quartzite above Grace Lake and on the 8,920-foot peak is part of the Maurice Mountain sequence that is thrust over the Black Lion Conglomerate, but a high-angle fault separates this rock from the Silver Hill Formation on the east side of the saddle south of the 8,920-foot peak. This is alternative (3) of the text. *D*, The quartzite immediately underlying the Silver Hill Formation on the east side of the saddle south of the 8,920-foot peak is part of the Cambrian sequence comparable to similar rocks south of Grace Lake; a high-angle fault separates this limited area of quartzite and younger rocks from the allochthonous Proterozoic quartzite of the 8,920-foot peak and above Grace Lake. This is alternative (4) of the text. Geologic relations of all four diagrams simplified from the geologic map to eliminate irrelevant information.

Finally, the Grace Lake fault juxtaposes the lower member of the Silver Hill and the sequence of Maurice Mountain, and it must have a northside downthrow. Although the trend of the fault on the ridge west of Grace Lake is just north of due east, I interpret this as a local deflection of a dominant west-northwest trend; the sequences on both sides of the fault can be followed westward (for the northern block, to be sure, this continuation requires projection across an interpreted structural window at the south branch of Boulder Creek; see below) to the 8,920-foot peak and the saddle between it and Maurice Mountain, thus defining the overall trend. On the east side of the saddle south of the 8,920-foot peak, quartzite that is lithically identical to the Maurice Mountain sequence dips at a moderate angle southward under the right-side-up lower member of the Silver Hill Formation, which here consists of interbedded quartzite and black argillite. The lower member of the Silver Hill in turn passes southward and stratigraphically upward into an attenuated upper member (estimated at 10 m), followed by typical but attenuated Hasmark Dolomite, Red Lion Formation, and finally Devonian Jefferson Dolomite. This set of relations is much like that seen on the ridge west of Grace Lake, but the field relation at the saddle invites an obvious interpretation of the Silver Hill as resting unconformably on the Maurice Mountain-like quartzite of the 8,920-foot peak. Four alternative explanations are possible (fig. 25) and will be briefly reviewed:



(1) The lower Paleozoic rocks east of the saddle, including the Silver Hill Formation, are indeed in depositional contact on the allochthonous quartzite, possibly of the Maurice Mountain sequence. Despite the identical lithofacies and detailed lithic features of these lower Paleozoic rocks with those in the same formations of the Vipond Park quadrangle, they have been transported on a low-angle thrust fault for an unknown but probably great distance. The relation with the "autochthonous" rocks of the same formations south of Grace Lake must be a fault, possibly the westward extension of the Grace Lake fault, now hidden in the carbonate rocks on the ridge south of the south branch of Boulder Creek. This alternative does not exclude a window of Black Lion Conglomerate in the valley of the south branch of Boulder Creek, but if the window is admitted, then the unconformity would have to be locally utilized by a thrust fault. In my opinion, the need for fortuitous juxtaposition of identical Paleozoic carbonate rocks of once great remove renders this alternative unacceptable.

(2) The quartzite underlying the 8,920-foot peak, and by extension that on the cliffs above Grace Lake and underlying the 10,428-foot peak north of the lake, is not part of the Maurice Mountain sequence despite lithic similarities. Rather, it is an upper member of the Black Lion in normal sedimentary contact. The Silver Hill is conformably above this supposed upper Black Lion quartzite (labeled in fig. 25*B* as Єblu). That such a unit, several hundred meters thick, should be stratigraphically intermediate between the typical Black Lion and the Silver Hill here and nowhere else, seems implausible. A particularly strong argument against this interpretation is the absence of such a rock unit east of Black Lion Mountain, barely 2 km away. In addition, to accept this hypothesis, the lithic similarity between this unit and the Maurice Mountain sequence must be ignored, as must the appearance of the thrust sheet of the Maurice Mountain sequence on the south side of Gold Creek, an area that bears the same general geometric relation to the underlying Paleozoic rocks.

(3) The Maurice Mountain-like rocks of the 8,920-foot peak and above Grace Lake are indeed part of the allochthonous Maurice Mountain sequence that has ridden over the Black Lion Conglomerate. The Black Lion Conglomerate, moreover, reappears in the south branch of Boulder Creek as a structural window. This interpretation assumes that the contact between the Silver Hill and the quartzite on the east side of the saddle south of the 8,920-foot peak is a fault despite lack of deformation at many exposed contacts. For this reason the interpretation is improbable.

(4) The argillite of the Silver Hill Formation east of the saddle is immediately underlain by a quartzite that is part of the Silver Hill Formation. The quartzite here is clean,

lacking the conglomerate pebbles of the Black Lion and the characteristic green lamination of the Maurice Mountain, but is not unlike the quartzite associated with the Silver Hill at Sheep Mountain, the quartzite south of Grace Lake, or the quartzite at Hecla. The interpretation postulates that a high-angle fault separates this quartzite and the other Cambrian units from the quartzite of 8,920-foot peak, thus allowing the latter quartzite to be assigned to the allochthonous Middle Proterozoic sequence. This hypothesis interprets the contact of the quartzite and the Silver Hill as a normal sedimentary contact, without incurring impossible corollaries as under (1), and draws comparison with similar stratigraphic and structural relations in the nearby Grace Lake area within the same tectonic blocks. Finally, on the saddle south of the 8,920-foot peak, as noted above, the Hasmark and upper Silver Hill are exceptionally thin; on the crest of the saddle the Silver Hill is not exposed and no float exists to suggest its presence. These features could be readily explained by the truncation of the Hasmark and Silver Hill along the saddle: Because the Cambrian beds dip moderately south, a vertical or near-vertical fault should cut higher up the stratigraphic section at higher elevations.

Weighing the various factors and hypotheses, I favor interpretation (4) as doing least violence to local field evidence or to regional stratigraphic relations; thus, the main contact is shown as a high-angle fault that is a continuation of the Grace Lake fault, but the local conformable contact between the argillite of the Silver Hill Formation and the quartzite is interpreted as an intra-Silver Hill bedding contact.

Earlier, I mentioned that a pre-Eocene flexure on the south limb of Cherry Creek syncline could account for the lack of the Morrison Hill klippe south of Cherry Creek; the flexure would have to be pre-Eocene because a distinctive Eocene pumiceous ash bed occurs on the two sides of the valley at the same elevation. Alternatively, the deformation could be the result of a small down-to-north movement on a fault along the valley, alined with the Grace Lake fault.

Many plutons of the Late Cretaceous to early Tertiary Pioneer batholith are elongated in the west-northwest direction; examples are the Grayling Lake Granite, the granite of Brownes Lake, and the granite of Stine Creek. The elongated shapes of these plutons may have been controlled by the pre-existing west-northwest-trending faults. The southern contact of the Clifford Creek pluton can be demonstrated to be an intrusive contact, but it is straight and is alined with the high-angle fault that crosses the high ridge between Black Lion Mountain and the 10,428-foot peak to the southwest. Very likely the pluton stoped out all the rocks on one side of the fault.

The valley of Big Hole River, approximately between Seymour Creek and the hamlet of Dewey north of the map area, is straight and wide; the rocks on the two sides do not match. I suggest that this anomalously wide west-northwest-trending valley is the locus of a major west-northwest fault; the valley thus has a physiographic history similar to that of the north-trending Tertiary basins. East of Dewey, the fault would have to be south of the valley. Paleozoic rocks in the area of Patterson Corner (locally known as "Watercress Springs") in the Wise River quadrangle are intensely deformed and faulted; formations are juxtaposed out of their normal sequence. This is where the wide valley ends abruptly and becomes a narrow gorge. I suggest that Patterson Corner harbors a splay of the main fault and that the Sawmill Gulch fault is part of this group. The abrupt northward termination of the fault-bounded Wise River valley against the Big Hole valley could be explained by the postulated fault.

North-South Faults

The most important single fault of this family is the Fourth of July fault (pl. 1), first mapped and named by Calbeck (1975). The fault enters the map area from the south on the west slope of the high ridge east of Elkhorn Creek, generally parallels the course of the Wise River valley and, with the valley, bends north-northeast opposite the confluence of Wise River and Pattengail Creek. A short distance to the north of this point, the fault is offset to the east by a right-lateral fault between Butler Creek and Swamp Creek and enters the Wise River quandrangle north of a left-lateral offset. Movements on this fault and on the west-northwest fault system must have overlapped in time.

The Fourth of July fault has a west-down sense of throw; Calbeck (1975) drew the contrary conclusion presumably because he did not realize that the Middle Proterozoic rocks west of the fault are in thrust sheets that overrode the Paleozoic rocks. Displacement on the fault is not known as it depends on the thickness of the Middle Proterozoic sequences; the cross sections show 1 to 2 km of dip slip, which is probably a conservative estimate. Presence along the fault of rocks not immediately related to either wall (such as the Jefferson Dolomite along Clifford Creek) also suggests fairly large displacements. As inferred from the orientations of fractures in the zone of sheared rocks found along the fault, the fault surface is steep to vertical; however, though the fault at places can be located to within a few meters, its surface is nowhere exposed.

The lower reaches of the Wise River valley have straight range fronts, offset here and there. I interpret the relations to mean that the Wise River valley is a graben; the present physiography of the range front may reflect exhumed fault-line scarps. The Fourth of July fault may be part of the Tertiary basin-and-range fault system. Age of development of this system in southwestern Montana is commonly considered to be late Tertiary (Reynolds, 1979; Pardee, 1950).

The Fourth of July fault was not recognized by Fraser and Waldrop (1972) in the Wise River quadrangle. The fault projects into the center of section 16, T. 1 S., R. 11 W., where these authors show the Cambrian Hasmark Dolomite as a klippe on Middle Proterozoic feldspathic quartzite (bq) on their map. I correlate this feldspathic quartzite with the sequence of Boner Knob and correlate their white vitreous quartzite (bw) and dark, variegated argillite (ba) with the sequence at Swamp Creek. Thus, I interpret the feldspathic quartzite as allochthonous and as overriding the Paleozoic sedimentary rocks where it appears east of the Fourth of July fault. I also interpret it as having been downdropped west of the fault. With Fraser and Waldrop, I interpret the dark argillite and the white quartzite as autochthonous relative to the Paleozoic rocks; these rocks are now exposed only east of the Fourth of July fault. The hillslope northeast of the Hasmark outcrops in section 16 is shown by Fraser and Waldrop as underlain by the feldspathic quartzite, but I could find no outcrop on the hillslope. Thus the field relations permit extending the Fourth of July fault into the area just west of the outcrops of the Hasmark. The continuation of the fault farther north is not determined, but it may follow the lower reaches of Wise River itself which, because of a recess of the mountain front east of the valley, has taken a dog leg to the east just north of the Vipond Park-Wise River quadrangle boundary.

With this revised interpretation, it would be simple to correlate the Johnson thrust mapped by Fraser and Waldrop (1972) with the Wise River thrust in the map area. The rocks on the Johnson thrust have the same structural relation as do rocks of the Morrison Hill klippe; interestingly, both thrusts have slivers of upper Paleozoic rocks along them. I agree with Fraser and Waldrop (1972) that Granulated Mountain and Little Granulated Mountain, in the northern part of the Wise River quadrangle, constitute another group of klippen resting on the Colorado Group as do rocks of the Morrison Hill klippe. These klippen must be east of the projected northward extension of the Fourth of July fault. At one place, the thrust contact is intruded by a small tonalite stock ("Cretaceous granodiorite," Kg, of Fraser and Waldrop, 1972) which baked the rocks on both sides of the thrust fault. Biotite from the stock has yielded a K-Ar age of 76 Ma. This age is probably close to a real intrusive age rather than a cooling age because the stock is barely 0.1 km across at the present level of exposure.

The above interpretation requires that the Johnson thrust predate the Fourth of July fault, so its trace must be offset by the latter rather than continue without a break northwest of the village of Wise River along the Big Hole valley, as shown by Fraser and Waldrop (1972). My explanation, which requires a third, west-northwest-trending fault, was given at the end of the section on west-northwest faults.

Minor Planar Features

Minor, though numerous, northeast-trending faults displace strata by amounts ranging from a few meters to a few hundred meters, but where they displace resistant strata such as the Quadrant Quartzite, notable topographic features

Figure 26. Swarm of mafic dikes in the Grayling Lake Granite, south wall of cirque above Green Lake. Dikes occupy north-south fracture system. Man (just to right of center foreground) is standing upright.

are developed. The ages of these movements are unknown but appear to postdate all other structures in the quadrangle. Whenever possible, these faults have been mapped.

A persistent north-northeast-trending fracture cleavage exists in the sedimentary rocks and locally is the most readily observed minor structure in a given outcrop. This system of fracture cleavage coincides with the axial-surface orientations of some later folds, for instance the conspicuous fold in the Mission Canyon Limestone displayed on the cliffs on the south bank of Canyon Creek upstream from Devils Hole and the fold in the south-plunging anticline defined by the Phosphoria Formation on the north side of Trapper Creek. This may be the same fracture system that pervasively cut up rocks of the Pioneer batholith and that provided conduits for mafic dikes (fig. 26). Presence of fractures of the same orientation in the Eocene flows as, for instance, in the Trusty Gulch area shows that the fractures and Eocene volcanism were synchronous.

Joints are locally spectacular in the intrusive rocks, and their orientation affects the morphology of the mountain slopes. Excellent places to examine the joints include the south slope of the upper reaches of Rock Creek and the northwest cliff face of Granite Mountain. The joints are commonly spaced a few meters apart, appear perfectly parallel to one another, and some show mineral growth such as epidote in the spaces. Tourmaline, however, occurs only in joints in sedimentary rocks near intrusive bodies. At many places, close examination shows that the joint orientation is but slightly different from the faint mineral foliation in the rocks as seen, for example, in habit orientation of hornblende and biotite; possibly the slight anisotropy imparted by the mineral orientation helped determine the orientation of the joints. Possible correlation of joint character and spacing to rock type and grain size is being studied (Judy Ehlen and E-an Zen, 1983, unpublished data).

Tectonic Relation of Early Proterozoic Gneiss Terrane

The Early Proterozoic gneiss terrane bears an uncertain structural relation to the immediately surrounding Paleozoic rocks, mainly the Black Lion Conglomerate. The south boundary of the main mass is clearly a fault, well exposed on the headwall of the cirque north of Sheep Mountain and in the straight, shallow ravine leading west from the cirque. The western and northern boundaries, however, cannot be proved to be faults; field relations could be interpreted as an unconformity: At both places, a narrow grass-covered zone separates the gneiss from right-side-up Black Lion Conglomerate, which dips away from the gneiss. The relation is ambiguous. Regardless of whether the zone is a fault or a sedimentary contact, however, there remains the question of how the gneissic rocks fit into the overall structural jigsaw puzzle of the Pioneer Mountains. Is the gneiss terrane autochthonous? If so, how does the sequence relate to the autochthonous Swamp Creek sequence north of Adson Creek fault? Why are the Cambrian and Cambrian(?) sequences so different above, respectively, the gneiss and the Swamp Creek sequence? If the gneiss terrane is allochthonous, is the movement surface above or below the Black Lion Conglomerate? If below, how does the presence of the Black Lion fit the relations of rocks north of Adson Creek fault? If above, can the observed relations be reconciled with a thrust fault?

The Early Proterozoic gneiss is shown in the cross sections to underlie the Black Lion Conglomerate. Absence of both units north of Adson Creek is explained by a combination of circumstances: (1) Prior to the deposition of the Black Lion Conglomerate, significant erosion locally removed the sequence of Swamp Creek down to the level of the gneissic "basement"; (2) troughs formed in the same areas and were filled with sediments that became the Black Lion Conglomerate and the lower member of the Silver Hill Formation; (3) the upper Silver Hill was deposited across the entire area after the troughs were filled, and a level surface was reestablished. Such a sequence of complex events confined to a restricted area seems improbable.

A second explanation is to regard the gneiss terrane as allochthonous and postulate that it represents part of the basement to Middle Proterozoic sedimentary rocks thrust in from the west. The relations would be broadly analogous to the relations suggested for the Early Proterozoic gneiss and Belt Supergroup sedimentary rock just east of Spokane, Wash., briefly mentioned by Harrison and others (1980; see especially their cross section). Such a hypothesis would remove the local problem of explaining away the presence of an Early Proterozoic "basement" south of Adson Creek, though it would not explain the presence of the Black Lion Conglomerate in the same area. However, it introduces the embarrassment of having the Early Proterozoic gneiss and the Black Lion in geometrical relations not obviously attributable to thrust faulting; subsequent local downdropping of the Early Proterozoic rocks, everywhere against the Black Lion, would be required. Because the gneiss terrane bears geochemical affinity to the intrusive rocks of the Pioneer batholith, its origin and structural position in the area have direct and important bearing on the genesis of the batholith.

Yet a third possible explanation, alluded to earlier, is that the Early Proterozoic gneiss terrane and the Black Lion Conglomerate are rocks of the same tectonic block, though they may now be in fault contact at the present level of exposure. The Adson Creek fault, however, may have a large strike-slip component, so that the rock sequences on the two sides of the fault may have had considerable horizontal separation. This strike-slip component would explain the sharp stratigraphic contrast across the fault without appealing to improbably large local topographic and sedimentological differences necessary for either of the other two explanations. This solution begs the question of whether the Early Proterozoic rocks are allochthonous relative to the Black Lion Conglomerate, but it does make an autochthonous hypothesis more palatable.

Stratigraphic and Structural Relations of Middle Proterozoic Sedimentary Rocks in Thrust Sheets

The problem of the age and facies relations of the Middle Proterozoic rocks that now occur in thrust sheets is entwined with the problem of the areal structural interpretations of the rocks and the assignment of individual outcrops to structural blocks. The rocks belong to three lithic sequences, informally referred to as the sequence at Boner Knob, the sequence at Big Point, and the sequence at Maurice Mountain. At their named localities, the sequences are distinctive. However, away from the Boner Knob, Big Point, and Maurice Mountain localities, the affinities of many outcrops are less clear. It is also an assumption, not a proven fact, that each lithic sequence occurs uniquely in the stratigraphic reconstruction. The problem is compounded by the exceptionally poor outcrops in large parts of the area underlain by these rocks. The interpretation presented in this section is, as a result, at places necessarily rather speculative.

The reconstructed stratigraphy of the sequences and their lithology indicate that the rocks are part of the Missoula Group of the Belt Supergroup. The nature of the rocks of the individual sequences suggests lithic correlation of the Maurice Mountain sequence and the Big Point sequence, respectively, with members 1 and 2 of the Mount Shields Formation of the Missoula Group (Wallace and others, 1981); the Boner Knob sequence is correlated either with member 3 of the Mount Shields or possibly with the overlying Boner Quartzite (C.A. Wallace and Don Winston, 1981, oral commun.). Preliminary paleomagnetic data (D.P. Elston, 1982, oral commun.) on the three sequences indicate that their more likely correlations are entirely with the Mount Shields. The combined thickness of the three sequences in the map area, however, is apparently greater than that of Mount Shields in the area of the Butte 1°×2° quadrangle sheet (Wallace and others, 1981). If the thickness increase is real, the three sequences in the map area might represent more proximal deposits and could include some pre-Mount Shields rocks such as the Shepard Formation. I have thus chosen to avoid direct use of the established stratigraphic names of the Missoula Group, but on the map explanation I have indicated plausible correlations with these established units.

One of my basic working assumptions is the uniqueness of each lithic sequence. A second basic assumption is that the sequences appear in only two thrust sheets, one resting above the other at least in part of the area; these are called the Pattengail thrust sheet (upper) and the Wise River thrust sheet (lower). This assumption of two thrust sheets is justified by the presence in two areas of Paleozoic carbonate rocks, probably the Mississippian Amsden Formation, around the mouth of Ross Gulch west of Wise River, interpreted as slivers caught between thrust sheets. An immediate conclusion is that the Maurice Mountain sequence and the Boner Knob sequence both appear in the Pattengail thrust sheet which rests on rocks mapped as the Big Point sequence. The Maurice Mountain sequence is interpreted to grade upward into the Boner Knob sequence in the Pattengail thrust sheet for the following reasons (fig. 27, loc. a): On the steep cirque sidewall above the Grouse Lakes, thick and characteristically slumped beds of quartzite of the Maurice Mountain sequence appear (fig. 3*A*). They pass upward into pink, feldspar-poor, massive quartzite beds that contain thin argillite interbeds and are transitional into the Boner Knob sequence. The higher parts of the ridge crest of Stine Mountain, above the Grouse Lakes, consist of this intermediate type; the northern cirque sidewalls of the Grouse Lakes basin show actual interbedding of typical Maurice Mountain-type quartzite and these pink intermediate types.

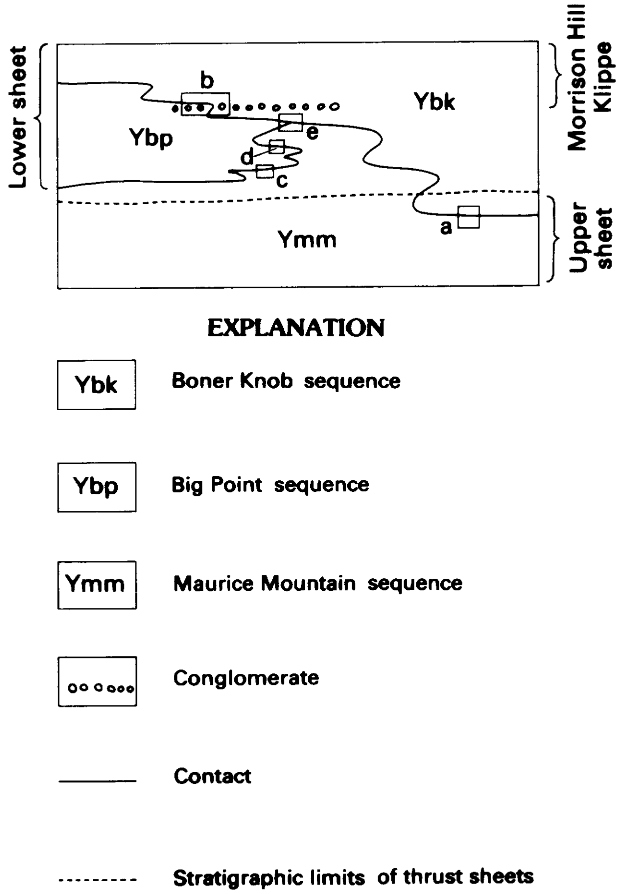


Figure 27. Schematic restoration of relations of allochthonous sedimentary sequences of Proterozoic Y age. Ymm, sequence of Maurice Mountain. Ybp, sequence of Big Point. Ybk, sequence of Boner Knob. The conglomerates are assumed to have a unique stratigraphic position as indicated, and their presence is used to restore the section as discussed in the text. Letter symbols refer to specific localities (see text): a, ridges of Stine Mountain and the Grouse Lakes area. b, Eagle Rock and slope southwest of Pattengail Creek-Wise River confluence. c, upper slopes and summit of Maurice Mountain. d, ledges above (east of) Little Joe Campground of the U.S. Forest Service and on the north side of Happy Creek. e, relations near the 8,980 ft peak and the mouth of Moose Creek. The general sections encompassed by major thrust sheets are indicated.

The next significant observation (fig. 27, loc. b) is that typical pink feldspathic quartzite and argillite of the lower part of the Boner Knob sequence are interbedded with massive polymictic conglomerate beds at Eagle Rock (Calbeck, 1975, gave measured section at this place); these beds are distinctive and readily recognized. The conglomerate here rests on massive greenish-gray quartzite; these quartzite beds crop out at the lower ledges along the west side of the Wise River Road from Eagle Rock to near the confluence of Wise River and Pattengail Creek and are identified with the quartzite of the Maurice Mountain sequence. Southwest of the confluence, the quartzite and distinctive conglomerate reappear on the steep slopes; apparently on strike and at the same level above the stream (with bedding strike nearly parallel to the stream course), these quartzite beds pass laterally into rocks that lithically resemble beds at Big Point; small conglomerate beds are found in this area, and I interpret the relations to mean that the two sequences interdigitate, though on the geologic map (pl. 1) the rocks are shown as the Boner Knob sequence. The beds occurring on the hillslope pass upward abruptly on the ridge crest into quartzite beds of the Maurice Mountain sequence, which has a different structural attitude. I interpret these beds to be part of the upper thrust sheet, but the leading or eastern edge was truncated by a later high-angle fault whose southern extension within the lower thrust slice separates dominant Boner Knob sequence from dominant Big Point sequence. By this token, the conglomerate at Eagle Rock and southwest of the Pattengail Creek-Wise River confluence is considered to be in the lower, Wise River thrust sheet. Inasmuch as identical conglomerate occurs on the steep slope immediately across the Wise River from Eagle Rock, this part of the Boner Knob sequence is included in the lower thrust sheet as well.

On the west ridge of Maurice Mountain, the Fourth of July fault separates pink feldspathic quartzite and maroon argillite assigned to the Boner Knob sequence on the west from massive, clean, white-to-gray, striped quartzite of the Maurice Mountain sequence on the east. However, the uppermost 100 m or so of the summit of Maurice Mountain itself is underlain not by the typical Maurice Mountain sequence but by a sequence of steel-gray, silty, poorly sorted, feldspathic quartzite, generally less than 1 m thick, and by impure thin-bedded argillite and siltstones having the bedding characteristics of the Big Point sequence. I interpret these beds to be transitional toward the Big Point sequence, and the upward change forms the basis of my conclusion that the Big Point on the whole is younger than the Maurice Mountain (fig. 27, loc. c). Thus, the Big Point is at least in part a lateral facies equivalent of the Boner Knob, as already inferred from the relations of the conglomerate beds from Eagle Rock southwestward.

The argillite interbeds in the Boner Knob and the Big Point sequences are very similar in bedding characteristics, lithology, and thickness; the main difference is that the argillite at Big Point has more mudcracks and other desiccation features as well as beheaded ripples in the siltstone beds, all of which indicate relatively shallower water of deposition. The quartzite of the Big Point is less thick-bedded, finer grained, and more silty, and it occupies a smaller proportion of the section than does the quartzite in the Boner Knob sequence. Again, these characteristics could be attributed to a less well washed and a shallower environment of deposition.

Southeast of Big Point on the slopes above Happy Creek, outcrops containing typical thin-bedded argillite and siltstone interbed with clean quartzite closely resembling that of the Maurice Mountain sequence (fig. 27, loc. d). These outcrops apparently are part of a coherent structural block that is the upward continuation of and thus younger than the section at Big Point itself. The balance of the evidence for relative ages of the two sequences, however, seems in favor of the Big Point being younger, so the significance I attach to the outcrop above Happy Creek is to demonstrate that the two sequences do intergrade. The same can be said of some outcrops on the roadside just south of Willow Campground and north of the confluence of Moose Creek and Wise River: The outcrops show interbedding of quartzite typical of the Maurice Mountain sequence and tan siltstone and maroon argillite of the Big Point and (or) the Boner Knob sequence, on a scale of individual beds.

Above Happy Creek on the slope toward the 8,899-foot peak to the east, there is virtually no outcrop. Scattered loose blocks of quartzite and conglomerate on the land surface are lithically assignable to the Boner Knob, but their origin is wholly unclear; they may well be material that once formed blocks in a former cover of Oligocene ash-matrixed landslide material, and so I exclude them in my effort at interpretation.

Rocks on and around the 8,940-foot peak north of Little Joe Creek and north of the 8,899-foot peak are a peculiar flesh pink, sparsely feldspathic quartzite interbedded with some maroon argillite (best seen northeast of the peak). These rocks find their nearest lithic comparison with the pink, sparsely feldspathic quartzite on the Stine Mountain ridge, interpreted earlier as transitional between the Maurice Mountain sequence below and the Boner Knob sequence above. The occurrence on the 8,940-foot peak may occupy an analogous stratigraphic position, here above the level of the Maurice Mountain-Big Point transition (fig. 27, loc. e). An outcrop on a knob, approximately at 7,930 ft, and north of the 8,940-foot peak, is a green-stripe quartzite strongly resembling the Maurice Mountain but containing much more feldspar than normal for the sequence. This occurrence tends to support the assignment of rocks. Because the rocks occur in the vicinity of the Fourth of July fault, some of the juxtaposition of rock types could be the result of tectonic mixing, but an interpretation of straightforward facies changes seems compatible with available evidence.

North of the 8,940-foot peak, scattered outcrops occur of a mafic dike having a west-northwest trend. Though the lower, northwestern slopes of the hill are without outcrop and the continuity of the dike cannot be proven, projection of this trend, across Wise River, leads to outcrops of similar dikes north of the Lacy Creek-Wise River confluence that have the same trend. I interpret these observations to mean that the rocks on both sides of Wise River in this area belong to the same coherent structural block and probably stratigraphic level and that all are part of the lower thrust slice.

East of the Fourth of July fault and thus at a deeper structural level, the only allochthonous Middle Proterozoic rocks (relative to the Phanerozoic sedimentary rocks) are those of the Morrison Hill klippe, scattered klippe rocks in the Vipond Park and Ponsonby Peak area, and the rocks in an area that extends from Grace Lake and the 10,428-foot peak at the northeast to the area southwest of Crescent Lake and west to Maurice Mountain. These areas are underlain by rocks definitely assignable to the Boner Knob sequence and the Maurice Mountain sequence. No rock assignable to the Big Point sequence has been discovered except, as noted, for the transitional beds on the summit of Maurice Mountain. Because both the Boner Knob and the Maurice Mountain sequences are also found in the upper thrust slice, one could interpret the rocks east of the Fourth of July fault as belonging only to the upper slice. This interpretation has serious deficiencies. First, it is not obvious why the sole thrust east of the fault should be the higher thrust of the west side, as there is no indication that the lower slice is thinning toward the fault. Second, the conglomerate identical to that found at Eagle Rock and other localities west of the Fourth of July fault is found east of the fault (see pl. 1), albeit in a white quartzite matrix, in the Morrison Hill klippe. Conglomerate such as this has not yet been found west of the Fourth of July fault anywhere in rocks demonstrably in the upper slice. Thus, this point could be used to argue that rocks east of the fault are within the lower slice. One might argue that if rocks east of the Fourth of July fault are in the lower slice, then some Big Point sequence ought to occur; however, there is no occurrence of Big Point rock next to the fault on the west side. The transitional beds on the summit of Maurice Mountain, in fact, might be a measure of the stratigraphic throw on the Fourth of July fault.

In the absence of definitive information to settle these questions, I use the following working hypotheses:

1. Three lithostratigraphic sequences obtain; the Maurice Mountain sequence is the oldest at any given area, grading upward into the Boner Knob sequence or the Big Point sequence. The Big Point is laterally equivalent to or slightly older than the Boner Knob (fig. 27).
2. These sequences occur in two thrust slices. The rocks east of the Fourth of July fault are entirely in the lower, Wise River slice. Within the map area, the Pattengail slice is preserved only west of Wise River and north of Lacy Creek.
3. Only the Maurice Mountain sequence and the Boner Knob sequence occur in the Pattengail slice; all three sequences occur in the Wise River slice. All polymictic conglomerates such as those found at Eagle Rock are in the lower slice.

The original relative dispositions of the three sequences are difficult to reconstruct. It seems clear that the Big Point lithology is confined to the more southerly parts of the map area; to the north the Boner Knob lithology takes over. If rocks east of the Fourth of July fault are in the lower slice, as postulated, then the Maurice Mountain lithology is areally extensive and may underlie the other two rock types everywhere. If the upper slice was originally more westerly than the lower slice, as seems reasonable since the slices moved west to east, then the finer grained variety of the Boner Knob was more westerly. A source area of sediments to the south and east, in general, would be indicated, though it is certainly not proven.

IGNEOUS AND CONTACT-METAMORPHIC ROCKS

The petrogenesis of the Pioneer batholith is a large and complicated subject; its full discussion is beyond the scope of this report. Only a summary of the information, particularly that dealing with field relations and geologic history, is given here.

The Pioneer batholith, as the term is used here, includes all the Cretaceous and early Tertiary instrusive rocks in the area, except some younger dikes, and spans the Late Cretaceous and early Paleocene. The rocks range from quartz diorite through granite in a calc-alkalic suite, which comprises the dominant rocks of the batholith as it is exposed; later plutons are mainly two-mica granite and related minor intrusions of leucogranite. Isotopic evidence indicates that these later granitic rocks are probably genetically related to the bulk of the calc-alkalic rocks, whereas a second contemporary group of calc-alkalic granitic to granodioritic rocks are from a distinct batch of magma (Zen and others, 1980). Because the strontium in the rocks is highly radiogenic, the magmas had crustal sources. Rocks similar in chemistry to the Early Proterozoic gneiss and amphibolite terrane are possible candidates, but their identity, origin, location, and relation to the larger Idaho and Boulder batholiths remain unknown. The strontium isotopic geochemistry shows that despite similarity in age and bulk chemistry with the Boulder batholith, especially its sodic series (Tilling, 1973; Doe and others, 1968), the Pioneer batholith is distinct. The Bitterroot lobe of the Idaho batholith was probably emplaced under different tectonic conditions (Hyndman, 1977; Chase, 1977) and cannot be directly compared (Arth and others, 1986).

Five plutonic units of the Pioneer batholith are formally named in this report. They are, from the youngest to the oldest, the following:

- The Clifford Creek Granite. Type locality is upper part of Clifford Creek in northeast part of the Stine Mountain quadrangle and northwest of Black Lion Mountain. Upper Cretaceous to Paleocene.
- The Grayling Lake Granite. Type locality is vicinity of Grayling Lake in the southwest part of the Vipond Park quadrangle. A granodiorite (Kgp), a tonalite (Ktcl), and two granites (Kgb and Kgm) are spatially and (or) petrographically and chemically closely related to the main body of Kgl and may be part of the same pluton. Upper Cretaceous.
- The Uphill Creek Granodiorite. Type locality is the drainage area of Uphill Creek at the south edge of the Vipond Park quadrangle in the south-central part of the map area. Upper Cretaceous.
- The Trapper Tonalite. Type locality is the south slope of Trapper Creek in the area north of Granite Lake in the Vipond Park quadrangle. Upper Cretaceous.
- The Keokirk Quartz Diorite. Type locality is Keokirk Mountain, but the main body is exposed on the high ridge west of Granite Lake in the Vipond Park quadrangle. Upper Cretaceous.

The detailed descriptions of the rocks are given in the map explanation. The ages of the rocks are based on isotopic dating; data are given in table 3 as well as in Marvin and others (1983). All rocks names used are those of the International Union of Geological Sciences system (Streckeisen, 1973).

Field evidence suggests that the level of emplacement of the batholith in the east Pioneer Mountains was fairly shallow, though quantitative assessment is difficult. Miarolitic cavities in rocks of the Uphill Creek Granodiorite (fig. 28) show the separation of a fluid phase during crystallization. The typical contact-metamorphic assemblages in pelitic rocks include andalusite-biotite-cordierite. Stratigraphic reconstruction of the overburden can be made at some places. For instance, for the Brownes Lake pluton of granite (Kgb on map), whose roof is well exposed on the north wall of Rock Creek and whose age is known, a reconstruction gives about 2 km (section J) of roof rock including the Colorado Group. To this roof rock, however, may have to be added the unknown thickness of the rocks of the Morrison Hill klippe which might have extended to this point. Lack of strange rock types in conglomerate of the Beaverhead Formation south of Melrose, already described, suggests that at the time of its emplacement the klippe contained only rocks of the Boner Knob sequence; this suggestion agrees with the relations found now at Morrison Hill. Although the original stratigraphic thickness of the Boner Knob is not known, another 1 to 2 km of section seems a reasonable upper bound, and a stratigraphic overburden of some 3 to 4 km at the time of batholithic emplacement seems acceptable. (Approximately 600 m of section is preserved in the klippe; relation of the klippe to the Trusty Gulch Volcanics suggests that the section was not much thicker in Eocene time.) This thickness corresponds to a pressure of about 1 to 1.5 kbar and is consistent with the formation of the miarolitic cavities and the andalusite-cordierite-biotite assemblages. Eventually, another estimate on the thickness of overburden during intrusion can probably be obtained by calculating the rate of cooling of the intrusive rock (initially at nearly solidus temperatures) calibrated on the basis of a combination of blocking temperature and fission track dating of the igneous rocks near a contact, as a function of the thickness of overburden using observed and assumed geometry. (See Snee, 1982.)

The two-mica granites have a mineralogy indicating that part of their crystallization occurred at considerably greater depth than the few kilometers indicated by the field relations (Zen and others, 1980). Much of the crystallization process might well have occurred at depth of nearly 10 km, an appreciable depth into the crust. At the time of final emplacement, the magma-crystal ensemble probably had rather high bulk viscosity and little heat content to give to the country rock, thus explaining the near absence of contact aureoles.

Are the batholithic rocks classically "bottomless," or are they shallow sheets as envisioned by Hamilton and Myers (1967; 1974)? Field evidence does not elucidate this point. Wherever observed, the contacts are steep. Gravity maps do not help much because of uncertain density contrasts for rocks buried at depth. Aeromagnetic maps show strong pattern correlation with the occurrence of the batholith, because the latter is rich in magnetite and the country rock is not. The aeromagnetic data are not readily

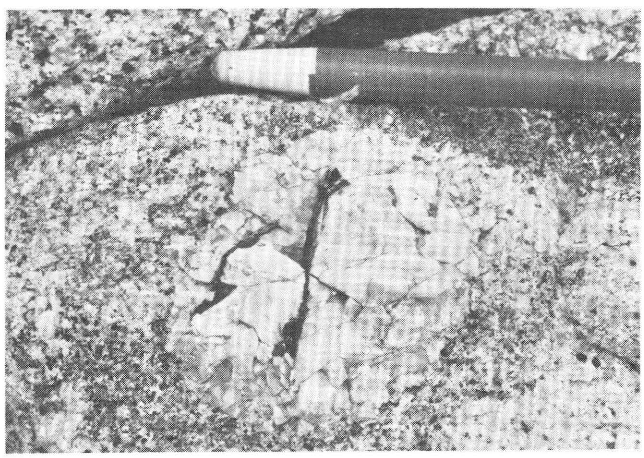


Figure 28. Miarolite cavity filled with quartz, two feldspars, and biotite. Sample is loose block south of Tendoy Lake in the Uphill Creek Granodiorite.

interpreted in terms of thickness of the batholith. If parts of the plutons were crystallized at depths of about 10 km, and if these are not diapiric intrusions, then the depths are comparable to the lateral dimensions of the plutons, and the three-dimensional shapes would be more like cubes or spheres than sheets and sills.

The idea that the magma was already laden with crystals during the final emplacement of the plutons is consistent with the field observation that at least the largest pluton, the Uphill Creek Granodiorite, contains compositional layers defined by biotite concentration that decreases upward in each cycle (fig. 29). Whether the layering is caused by gravitational settling (graded bedding), by viscous flow, or by penecontemporaneous multiple intrusion, the crystals must already have existed in abundance. The mineralogy in the layers—biotite, plagioclase, quartz—suggests a rather advanced stage of crystallization. On the other hand, field evidence shows that the typical texture and mineralogy of the intruding rock commonly extends right up to the intrusive contacts, with little or no chilled zone. In addition, the plutons by and large are remarkably free of xenoliths of the country rock despite abundant evidence of local crosscutting relations. Where did the country rock go? A simple answer might be that the rocks were removed within the magma by convection vigorous enough to move them and to ensure compositional and textural homogeneity up to the contacts. But could a magma consisting of much solid undergo such convection? Until modeling and calculations on the rheological and thermal history of the rocks have been carried out and until the rocks near the contact have been studied in detail to determine the chemical evolution of the contact rocks, this question remains moot.

The palynomorphs in the upper Colorado Group gave an age of mid-Campanian to Maestrichtian, indistinguishable from or slightly younger than the K-Ar ages on the major Cretaceous plutons of the Pioneer batholith (Sohl and Wright, 1980; Armstrong, 1968). Just north of the map area in the gorge of the Big Hole River, plutonic rocks dated at about 80 Ma (Snee, 1982) directly intrude the Colorado Group. Therefore, the Cretaceous plutons were emplaced in an area where a large, braided, and aggrading stream system draining from the west was developing at the surface. We are left speculating as to whether there were geomorphic indications of the intrusive process at depth and whether there were geothermal anomalies. One might expect to find evidence of contemporaneous and related volcanism. In fact, such volcanic rocks are found near Argenta (Snee and Sutter, 1979) but not within the area of this report. Because a period of active erosion preceded Eocene volcanism in the Pioneer Mountains, however, any volcanic rock of Late Cretaceous age that was present might have been removed.

The rocks of the Pioneer batholith crystallized under moderately oxidizing conditions. Ilmenite is unknown except as rare inclusions of a manganiferous variety (Hammarstrom, 1982) in muscovite of the two-mica granite. The other plutonic rocks contain magnetite and sphene, thus they were more oxidizing than the ilmenite-hedenbergite-sphene-magnetite-quartz buffer (Wones, 1966). The Grayling Lake Granite has magnetite as well as pink K-feldspar due to hematite inclusions, so these latter rocks probably were on the hematite-magnetite buffer at least during the last stages of crystallization. If at least some of the epidote in the rocks were late magmatic (Zen and others, 1980), the pistacite content of about 30 percent suggests that the oxygen fugacity was between the nickel-bunsenite and the magnetite-hematite buffers (Liou, 1973); this conclusion is borne out by a study of the composition of the biotite (Hammarstrom, 1982). Emplacement temperatures have not been independently estimated. Based on the data on minimum melting of haplogranite (Tuttle and Bowen, 1958) and of two-mica granite (Wyllie, 1977), the temperature, which is slightly pressure dependent, is $(700\pm100)+273$ K.

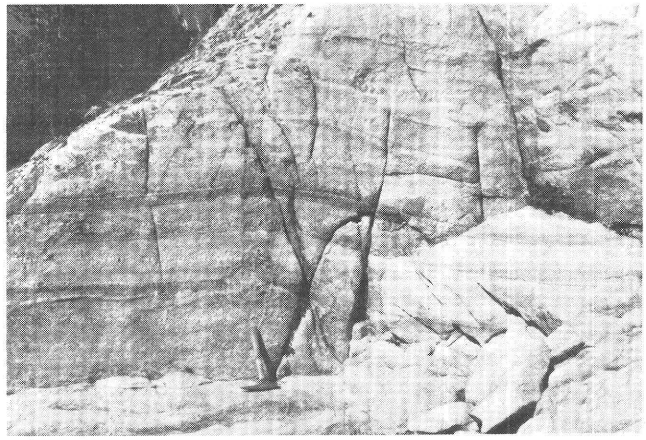


Figure 29. Biotite banding in Uphill Creek Granodiorite, rock basin south of Tendoy Lake. There is no size gradation in biotite from mafic to leucocratic layers.

The S-type vs. I-type classification of granite as used by White and Chappell (1977) cannot be directly applied to the rocks of the Pioneer batholith. The calc-alkalic rock suite contains hornblende and sphene, the inclusions are hornblende-biotite-feldspar-rich schlieren or discoid bodies (fig. 30), and the contacts are cleancut intrusive. Aluminous inclusions or minerals are totally lacking, the δ-^{18}O value is low (+6.8 to +8.6); the rocks have a molar alkali+lime/aluminum index of near unity. These are all hallmarks of "I-type" granite. Yet the initial $^{87}Sr/^{86}Sr$ ratio ranges from 0.7113 to 0.7160, distinctly crustal. The Paleocene two-mica granite has initial ratios of 0.7116 and 0.7123, a mildly peraluminous character (as much as 3 percent normative corundum), magmatic muscovite, no hornblende and also no aluminum silicate minerals; moreover, the contact relations are crosscutting. Both sets of peculiarities are compatible with a hypothesis of magma generation from a crustal rock, perhaps volcanic in origin, deep within the crust and intruded at a much higher elevation. Anatexis of a volcanic pile or volcanogenic sedimentary pile could produce a

Figure 30. Discoid mafic inclusions paralleling contact between the Grayling Lake Granite (matrix of the discoids) and its porphyritic border phase. Location is base of cliffs west of Keokirk Mountain.

magma of the requisite composition, isotopic nature, and the intrusive relations as observed.

Upon emplacement, the plutons produced contact-thermal metamorphism of the host rocks. The calcareous rocks produced assemblages of diopside-epidote-actinolite-plagioclase rocks, grossular-diopside-epidote rocks, and marbles. Very coarse marbles that developed in the Mission Canyon Limestone may be seen along contacts of the Trapper Tonalite and of the porphyritic granodiorite (Kgp), at Keokirk Mountain, and of Uphill Creek Granodiorite south of Sugarloaf Mountain. Idocrase and wollastonite are found associated with diopside and grossular near Brownes Lake in metasomatic rocks. (See Collins, 1975, 1977.) Massive chert of the Phosphoria Formation becomes snow-white massive quartz. Marls and calcareous shales of various formations become purple-to-green, massive, fine-grained biotite-chlorite-actinolite-quartz-feldspar hornfels. Impure limestones such as the Lodgepole have their clay-rich layers altered nearly isochemically to grossular rocks; other lime-

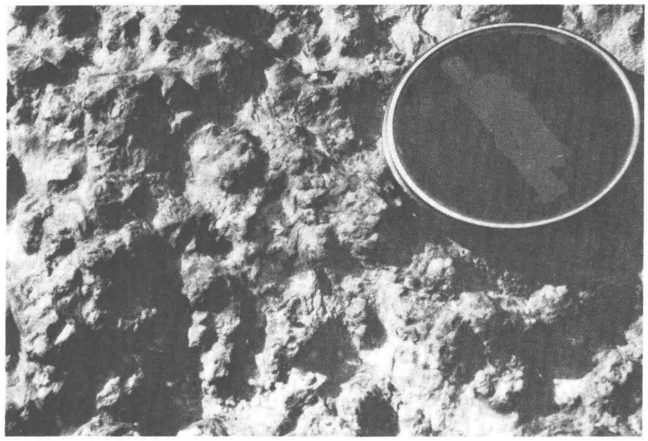


Figure 31. Pseudomorphs of garnet or staurolite in a micaceous matrix, lower member of the Silver Hill Formation, east side of structural dome at Hecla, above swamps of Sappington Creek.

stone layers have mizzonitic scapolite. Dolostones such as those in the Hasmark and Jefferson produce sunbursts of white tremolite at quartz-rich sites such as chert beds.

Contact metamorphism of noncalcareous pelites is best studied at the intrusive contacts with the lower member of the Silver Hill Formation at Hecla. The rock has undergone more than one episode of development of mineral assemblages related to one or more distinct thermal events. Most of the rocks on the east side of the dome above the swamp of Sappington Creek have recrystallized to produce a silky muscovite-rich schistose texture; many equant or nearly equant, subrounded to rectilinear pseudomorphs ranging from a few millimeters to more than 1 cm across (fig. 31) project through this matrix. These pseudomorphs have altered to quartz-chlorite-muscovite mixtures; examination of many failed to yield any original material. The shapes of the pseudomorphs strongly suggest original garnet or possibly staurolite. (See also Karlstrom, 1948, p. 18.) These rocks now contain andalusite, biotite, quartz, tourmaline, muscovite, chlorite, magnetite, and rare plagioclase. Cordierite has not been found unless it be part of the large pseudomorphs. (Entirely fresh cordierite is found with the same mineral assemblage, however, in the basal Kootenai Formation along the trail northwest of the 9,092-foot peak south of Storm Peak and at the summit of the 8,892-foot south peak of Sugarloaf Mountain.) The andalusite commonly has red-pleochroic cores and colorless rims. Clearly, two distinct stages of metamorphic recrystallization occurred. This recrystallization could be explained by a single stage of waning thermal metamorphism. However, some evidence suggests multiple thermal events. At several places in Hecla, rocks of the lower member of the Silver Hill contain a strange reddish-orange material in thin section. This material is isotropic, occupies interstices of minerals or similar narrow spaces, and at first sight seemed to be discolored plastic used to prepare the thin sections. However, the material is different; it also contains inclusions of magnetite. My next thought was that it might be due to quenched minimum melt of a granitic material from a later intrusion or from lightning strikes. Similar material was deliberately collected under overhangs of cliffs to eliminate the latter possibility, yet the material persists. It has a refractive index of 1.532±0.004, is X-ray amorphous, and has a specific gravity of between 2.51 and 2.58. Microprobe study shows that it contains just silica, alumina, and a small amount of iron; if computed as ferric iron, the atomic ratio of Si/Fe+Al is nearly 1:1 (1.02:1). I interpret the material as "metakaolinite" caused by the heating of pre-existing kaolinite to about 500 °C (Richardson, 1961). I conclude that the earlier mineral assemblage containing the porphyroblasts (now altered) was followed by minor argillic alteration producing kaolinite. A second thermal event, under conditions of considerably higher H_2O fugacity, heated the rocks to about 500 °C, destroyed the kaolinite, and hydrated the porphyroblasts to pinnitic material. The area is known to have been intruded by the earlier quartz diorite and the later, more hydrous granite; biotite from two samples of the contact rock gave K-Ar ages of 71.6±2.5 and 72.2±2.5 Ma that agree with the age of the granite and with the reset age of biotite in the quartz diorite. Thus, I associate the first metamorphic event with the quartz diorite, and the second event with the granite. Whether the two events occurred at different depths is unknown; because during the intervening years the Colorado Group continued to be deposited, it is unlikely that the older intrusion was at a deeper level.

The exceedingly limited extent of contact metamorphism by the two-mica Clifford Creek Granite attests to its lack of thermal capacity and probably to a high degree of consolidation at the time of intrusion. Local contact relations, however, seen well at the head of the cirque north of Black Lion Mountain, for instance, leave no doubt that the contact is a crosscutting intrusion rather than a fault. Calbeck (1975) reported a fault contact between the southern boundary of the Clifford Creek pluton and the Middle Proterozoic quartzite. In the Stine Mountain quadrangle, along the ridge just north of the top of the 8,920-foot peak (center sec. 36, T. 2 S., R. 12 W.), apophyses of muscovite granite intrude along bedding surfaces immediately adjacent to the main contact. A welded intrusive contact is also found on the steep eastern face of this ridge. Thus, the main contact is not a fault, even though some contacts might have been located at preexisting faults.

The two-mica granites are cut by one or more generations of smoky quartz-eyed leucogranite; the large body of leucogranite at Bobs Lake area (included in Tqp on the map) is probably one of these. The thin dikes of aplite and leucogranite seem to have no particular orientation; their relations in any given outcrop suggest the possibility that their positions were determined by hydrofracting. Successive intrusions of leucogranite and aplite (±pegmatite) seem to culminate in quartz-sericite veins that are locally the main bearers of molybdenite. Berger (1979), Berger and others (1981), and Siems and others (1979) explored the possible geochemical evolution of intrusive rocks leading to regional geochemical anomalies (see also Pearson and Berger, 1980) and to economic mineral deposits.

ISOTOPIC DATA AND AGE OF ROCKS

Ages of the stratigraphic units are determined by fossils and by isotopic dating methods. The paleontological data have been presented in connection with individual stratigraphic units; localities where fossils have been found and studied, as well as localities where fossils have been observed but not yet properly collected and studied, are marked on the geologic map (pl. 1).

The geochronologic data for the rocks pertain mostly to the igneous rocks, both intrusive and extrusive; a few data points were on contact-metamorphic rocks, and one was on a sedimentary rock. The data were obtained by a variety of methods: Potassium-argon method on mineral separates and on fine-grained volcanic rocks; $^{40}Ar/^{39}Ar$ on mineral separates; rubidium-strontium on whole rock and on rock isochron; uranium-lead on zircon; and fission track on sphene. The results are summarized in table 3; however, a full discussion of the results and their implications is beyond the scope of this report and will be taken up in topical studies to come.

In general the K-Ar determinations, especially on intrusive rocks, must be considered minimum ages because they reflect cooling of the pluton to the temperatures where individual mineral systems closed to argon diffusion (Snee, 1982). In a few favorable situations, such as when samples are collected at contacts of small projections of plutons into country rocks that have not been heated by a previous intrusion, the cooling age may closely approximate the intrusive age. In the majority of cases, however, the cooling age may be as much as a few million years younger than the real age of intrusion. Snee's (1982) detailed study of this problem in the areas to the south and west of the study area suggests that the problem can be overcome and that the real intrusive ages can be determined by the $^{40}Ar/^{39}Ar$ spectra; this study is underway.

Detailed reports of the chemistry for both major and minor elements will be made in topical studies to come. In the "Description of Rocks" part of the geologic map (pl. 1), the major element chemistry of each major plutonic unit is summarized. The initial $^{87}Sr/^{86}Sr$ ratios of many plutons and several volcanic rocks were determined by Arth and others (1986), and the results are summarized in table 4. These data directly bear on the interpretation of the origin of the batholith.

Table 3. Geochronological ages from volcanic, metamorphic, and intrusive rocks
[For intrusive rocks, cooling age is given unless otherwise specified. (See Snee, 1982.) WR, whole-rock]

Sample no.	Rock unit (pl. 1)	Age (Ma)	Method	Reference
Volcanic and metamorphic rocks				
612–1	Oligocene volcanics (Tov).	27.3±1.0	WR K-Ar	Marvin and others, 1982.
836–2	Trusty Gulch Volcanics (Ttg).	45.6±1.6	WR K-Ar	Do.
627–1	do	46.1±1.6	WR K-Ar	Do.
621–1	do	46.7±1.6	WR K-Ar	Do.
114–1	do	47.0±1.6	WR K-Ar	Do.
I–15	do	47.1±1.6	WR K-Ar	Do.
115–2	do	47.3±1.6	WR K-Ar	Do.
914–2	do	47.6±1.7	WR K-Ar	Marvin and others, 1983.
913–1	do	47.8±1.7	WR K-Ar	Do.
507–2	do	47.9±1.6	WR K-Ar	Marvin and others, 1982.
429–1–2	do	49.8±1.9	WR K-Ar	Do.
648–1	do	49.9±1.7	WR K-Ar	Do.
395–1	Lowland Creek Volcanics (Tlc).	50.4±1.7	Biotite K-Ar	Do.
287–1	Lower member, Silver Hill Formation (Gshl).	[a]71.6±2.5	Biotite K-Ar	R.F. Marvin, written commun., 1980.
287–2	do	[a]72.2±2.5	Biotite K-Ar	Do.
199–2	Middle Proterozoic sequence of Boner Knob at Morrison Hill (Ybk3).	757±15	WR (clay gall) Rb/Sr (minimum).	C.E. Hedge, written commun., 1974.
Intrusive rocks				
41–3–1	Mafic dike (Tdm)	48.8±1.8	WR K-Ar	Marvin and others, 1983.
516–1	Leucogranite (Tqp)	66.8±2.4	Biotite K-Ar	Do.
741–2	Silicic dike (TKds)	64.4±2.2	WR K-Ar	Do.
500–1[b]	Clifford Creek Granite (TKc).	64.6±2.1	Biotite K-Ar	Do.
		[c]65.6±1.4	Biotite $^{40}Ar/^{39}Ar$.	J.F. Sutter, written commun., 1982.

Table 3. Geochronological ages from volcanic, metamorphic, and intrusive rocks—*Continued*

Sample no.	Rock unit (pl. 1)	Age (Ma)	Method	Reference
32–1–4[b]	Clifford Creek Granite porphyritic phase (TKcp).	64.9±2.2	Biotite K-Ar	Marvin and others, 1983.
BHS	Grayling Lake Granite (Kgl).	72.3±2.5	Biotite K-Ar	Zen and others, 1975[d].
		72.1±1.6	Biotite $^{40}Ar/^{39}Ar$	J.F. Sutter, written commun., 1982.
1293–1	Grayling Lake Granite (Kgl).	74.1±2.7	Biotite K-Ar	Marvin and others, 1983.
744–2[b]	Granite of Brownes lake (Kgb).	71.3±2.4	Biotite K-Ar	Do.
		[e]72.7±2.7	Biotite K-Ar	Kruger Laboratories; Dan Powell, written commun., 1979.
1298–2	Granite of Mono Park (Kgm).	67.3±0.9	Biotite K-Ar	Snee, 1978.
		72.6±1.2	Hornblende K-Ar	Do.
BH9850[b]	Granodiorite, porphyritic border phase of Kgl (Kgp).	71.5±2.5	Biotite K-Ar	Zen and others, 1975[d].
		70.0±2.0	Hornblende K-Ar	Do.
IVP	Uphill Creek Granodiorite (Kuc).	72.6±2.8	Biotite K-Ar	Do.
		69.5±2.0	Hornblende K-Ar	Do.
		[e]72.0±2.0	Biotite K-Ar	Kruger Laboratories; Dan Powell, written commun., 1979.
1228–1	do[f]	71.7±2.6	Biotite K-Ar	Marvin and others, 1983.
1272–2[b]	do[f]	72.2±2.6	Biotite K-Ar	Do.
		68.8±4.2	Hornblende K-Ar	Do.
		70.4±1.8	Biotite $^{40}Ar/^{39}Ar$	J.F. Sutter, written commun., 1982.
881–1[b]	do[f]	72.0±2.5	Biotite K-Ar	Marvin and others, 1983.
		72.0±1.6	Hornblende $^{40}Ar/^{39}Ar$.	J.F. Sutter, written commun., 1982.

Table 3. Geochronological ages from volcanic, metamorphic, and intrusive rocks—*Continued*

Sample no.	Rock unit (pl. 1)	Age (Ma)	Method	Reference
121–1 -----------	--------do[f]--------	71.3±2.4	Biotite K-Ar	Marvin and others, 1983.
		71.3±4.5	Hornblende K-Ar	Do.
547–1–78[b] -------	Trapper Tonalite (Kt)	74.3±2.7	Biotite K-Ar	Do.
		71.9±4.3	Hornblende K-Ar	Do.
		73.5±1.0	Biotite $^{40}Ar/^{39}Ar$	J.F. Sutter, written commun., 1982.
Cliff above Cherry Lake.	---------do[f]--------	72.5±1.1	Biotite K-Ar	Snee, 1978.
1162–1 ----------	Granite of Stine Creek (Ksc).	72.3±2.6	Biotite K-Ar	Marvin and others, 1983.
		74.9±1.6	Biotite $^{40}Ar/^{39}Ar$	J.F. Sutter, written commun., 1982.
313–1 -----------	Keokirk Quartz Diorite (Kke).	72.6±2.5	Biotite K-Ar[g]	Marvin and others, 1983.
		79.8±2.2	Hornblende K-Ar	Do.
		80.3±2.3	Hornblende K-Ar	Do.
		76.5±7.5	Sphene fission track.	C.H. Naeser, written commun., 1976.
984–2–78 --------	Tonalite of Lime Kiln Gulch (Klk).	77.1±2.8	Biotite K-Ar	Marvin and others, 1983.
		77.3±1.5	Biotite $^{40}Ar/^{39}Ar$	J.F. Sutter, written commun., 1982.
		81.0±2.5	Hornblende $^{40}Ar/^{39}Ar$[h].	Do.
	Early Proterozoic rocks (Xga):		Rb-Sr WR isochron.	Arth and others, 1986.
RS–1 ------------	Amphibolite -----	} 1,630±38		
RS–6 ------------	Gneiss ----------			
RS–7 ------------	Augen gneiss ----			
RS–12 -----------	Banded gneiss----			
Dodgson Creek in Wise River quadrangle.	Tonalite (Granodiorite, Kg, of Fraser and Waldrop, 1972).	[c]76.4±2.6	Biotite K-Ar	Marvin and others, 1983.
		95.9±2.8	Hornblende K-Ar[i]	Do.
		98.8±2.8	Hornblende K-Ar[i]	Do.

[a]Metamorphic cooling age.

[b]Sample collected from loose block either at base of cliffs or in cirques having a single rock type that exactly corresponds to the block.

[c]Probably closely approximates intrusive age.

[d]Slight difference in reported ages reflects different constants used; see Marvin and others, 1983.

[e]Same general outcrop area but not same sample.

[f]Sample from zone where magma mixing between Kuc and Kgl is probable.

[g]Textural evidence indicates recrystallization due to intrusion of Kgl nearby.

[h]Minimum on a saddle-shaped release pattern; may correspond to intrusive age.

[i]Problem of excess radiogenic argon in material.

Table 4. Initial $^{87}Sr/^{86}Sr$ ratios of volcanic and intrusive rocks
[Rock units are defined in table 3. For intrusive rocks, cooling ages are given. Data are from Arth and others, 1986]

Sample no.	Rock units (pl. 1)	Age (Ma)	Initial $^{87}Sr/^{86}Sr$ ratio
836–2	Ttg	45.6±1.6	0.7058±.0002
114–1	Ttg	47.0±1.6	.7082±.0002
648–1	Ttg	49.9±1.7	.7064±.0002
395–1	Tlc	50.4±1.7	.7094±.0025
500–1	TKc	65.6±1.4	.7123±.0002
32–1–4	TKcp	64.9±2.2	.7116±.0002
BHS	Kgl	⩾72.3±2.5	.7137±.0002
1293–1	Kgl	74.1±2.7	.7129±.0002
744–2	Kgb	⩾72.7±2.7	.7131±.0002
BH 9850	Kgp	71.5±2.5	.7138±.0002
IVP	Kuc	72.6±2.8	.7113±.0002
BC	Tonalite of Torrey Mountain.	[1]70.4±2.4	.7118±.0002
1272–2	Kuc/Kgl mixing zone.	70.4±1.8	.7121±.0002
1298–2	Kgm	72.6±1.2	.7126±.0002
1228–1	Kuc/Kgl mixing zone.	71.7±2.6	.7123±.0002
121–1	Kuc/Kgl mixing zone.	71.3±2.4	.7114±.0002
881–1	Kuc/Kgl mixing zone.	72.0±2.5	.7119±.0002
547–1	Kt	74.3±2.7	.7160±.0002
1162–1	Ksc	74.9±1.6	.7139±.0002
313–1	Kke	80.3±2.3	.7113±.0002
984–2–78	Klk	81.0±2.5	.7069±.0002

[1]Zen, Marvin, and Mehnert (1975); age revised to reflect new constants.

ECONOMIC MINERAL DEPOSITS

Among the nonmetal resources in the map area, the most important is phosphate in the Permian Phosphoria Formation. Several theses (Theodosis, 1956; Fowler, 1955) and reports (Richards and Pardee, 1926; Cressman and Swanson, 1964) have been written on the geology and economic potential of the phosphate deposits of the area. Within the Vipond Park quadrangle, the principal mines are in the lower part of the Canyon Creek drainage, including the West Limb Mine and West Central Mine of the Melrose-Divide area; prospect trenches are widely scattered throughout the quadrangle where the Phosphoria Formation crops out. The minable phosphate beds thin rapidly westward and southward from the northeast part of the quadrangle; I know of no past or present operating mine southwest of the Trusty Gulch area.

The principal economic mineral deposits in the area are metal deposits. Noble and base metal deposits dot the area, particularly surrounding the plutons of the Pioneer batholith. The mines and prospects have been listed by Geach (1972), who gave, in addition to the location and brief description of the local geology of each mine and prospect, a brief summary of the metals produced and their production history. The principal districts are the Bryant District (Hecla and environs), the Sheep Mountain-Vipond Park-Quartz Hill district (which is not obviously related to plutons except for the developing prospect at Cannivan Gulch), and the Rock Creek district. The report by Geach does not give the ore mineralogy or petrology; for Hecla such data were supplied in part by Karlstrom (1948). A survey of the economic mineral deposits of the Pioneer Mountains is part of the U.S. Geological Survey's Dillon 1°×2° quadrangle CUSMAP (Conterminous United States Mineral Assessment Program) project; parts of the map area are also the subjects of studies of the mineral resources of the east Pioneer Mountains (Pearson and Zen, 1985) and of the West Pioneer Wilderness Study Area (L.W. Snee and E-an Zen, unpublished data). Accordingly, no specific description of mining districts will be attempted here; a few general remarks will suffice.

The deposits related to the intrusive rocks can be broadly grouped into three types: skarn deposits at the immediate contacts between country rock and intrusive bodies; hydrothermal metal deposits in country rocks not immediately associated with igneous contacts; and deposits within plutons.

Skarn deposits are presumably mineralized by metasomatic reactions between the host rock and the intrusive rocks and their fluids. Especially notable among these are the tungsten deposits, best exemplified by the Ivanhoe Mine (Myers, 1952; Geach, 1972; Pattee, 1960; Collins, 1975) at the contacts between the Amsden Formation and both the Uphill Creek Granodiorite and the granite of Brownes Lake. Scheelite occurs as mineral inclusions in grossular crystals in layers resulting from metasomatism of the marly beds. All the tungsten deposits and prospects known to me are along the contact of intrusive bodies with the Amsden, with the single exception of the Sheep Mountain tungsten prospect (Geach, 1972, p. 141). Instrumental neutron activation analyses of tungsten in both representative lithic types of the Amsden away from skarn localities and in the granodiorite near the Ivanhoe deposit, however, failed to show tungsten above background level, so the source of the tungsten remains unknown.

The gold, silver, and base metal deposits of the Hecla area are not in immediate contact with intrusive rocks at the present level of erosion, and the mine maps of Karlstrom (1948) do not show underground intersection of the ore bodies with intrusive rocks, except for minor dikes. The ages of formation of the Hecla ore bodies may or may not be as old as the age of intrusion. Interestingly enough, at Hecla and elsewhere, the hydrothermal deposits are in dolostones wherever both dolostone and limestone exist. At Hecla, for instance, all the significant prospects and workable deposits occur in the Hasmark or the Jefferson

Dolomite. The rocks south of the saddle between Lion Mountain and Keokirk Mountain were mapped by Karlstrom (1948) as Cambrian Pilgrim, Park, and Meagher Formations (equivalent to my Hasmark Dolomite) which are the host rocks below Lion Mountain. Karlstrom mapped the unit capping Lion Mountain as Jefferson. Neither the rocks south of the saddle nor the caprock of Lion Mountain has a mineral deposit. We now know that these barren strata are Mississippian limestones rather than pre-Carboniferous dolostones. Some chemical control, possibly pH buffering, caused sulfide precipitation in dolostone. On the other hand, along Birch Creek south of the quadrangle, the Indian Queen Mine is a small copper deposit in skarn at the contact between the granodiorite and the Mission Canyon Limestone; here no dolostone is present.

The Vipond Park-Quartz Hill area has many mines but none, except for the newly developing Cannivan Gulch molybdenite deposit, are associated with igneous rocks at the present level of erosion. McClernan (1977) proposed, on the basis of his interpretation of field relations and of unpublished lead and sulfur isotope data, that Hecla is a deformed Mississippi-valley-type deposit in terms of form, composition, and origin. Unpublished sulfur and lead isotope data of the U.S. Geological Survey (B.R. Doe, written commun., 1980) also show that samples from the Hecla area, the Sheep Mountain area, and the Quartz Hill area, as well as from other mines associated with carbonate host rocks, have lead-isotope abundances approaching those of Mississippi-valley-type lead deposits. These deposits might be the result of remobilization and concentration of disseminated metals in the carbonate rocks (a refinement of McClernan's suggestion) by hydrothermal circulation systems set up within the carbonate rocks near the intrusions. Because Doe's lead isotope data for the carbonate-host deposits and for the igneous rock-related bodies show a crude correlation with the mapped distance of the body from igneous rocks, local mixing, by coupled circulation, of lead from the igneous and sedimentary sources seems reasonable. Study to discover the lead isotope characteristics of galena in quartz veins in different host rocks of various ages, and of trace lead in skarn deposits adjacent to intrusive rocks, is underway. Data available suggest that hydrothermal circulation systems were initiated by intrusions but that systems in sedimentary rocks remote from intrusive bodies were largely independent, except for thermal effects, from circulations in the igneous rocks.

Economic mineral deposits found within intrusive rocks are uncommon in the map area. The molybdenite prospect at Cannivan Gulch (Schmidt and Worthington, 1977; Schmidt and others, 1979) is one example. K-Ar dating provided a range of apparent ages (Schmidt and others, 1979; Armstrong and others, 1978) suggesting a complex themal and (or) alteration history that has not been satisfactorily elucidated. The small pluton at Cannivan Gulch (Kcg on map) has a K-Ar biotite age of 68 Ma, but 68 Ma may not be the age of principal mineralization. Various K-Ar and $^{40}Ar/^{39}Ar$ release ages suggest that the quartz-molybdenite veins in the Pioneer Mountains may be closely related to the 65-Ma intrusive event (Snee, 1982; Pearson and Berger, 1980) and that the veins are spatially associated with the leucogranite such as that of the Bobs Lake area. This late episode of intrusion of more silica-rich magma produced more and larger scaled hydrothermal circulations leading to mineral deposits than did the earlier, larger, and less silica-rich intrusions of the Pioneer batholith. Interestingly enough, 65 Ma is also the age of several silicic dikes (TKds on map) that petrographically closely resemble the pre-main stage quartz-porphyry dikes of about the same age in the Butte district (Brimhall, 1979).

Finally, small-scale mining operations for manganese are scattered along the unconformity between the Triassic Dinwoody and the Cretaceous Kootenai Formations. The mines are clearly for the residual manganese concentrations on the pre-Kootenai erosion surface. No such mines were in operation in the area during the 1970's.

GEOLOGICAL HISTORY

Even though important pieces of the jigsaw puzzle that depicts the geologic history of the Pioneer Mountains are still missing, enough is known of the areal geology to place useful constraints on possible hypotheses. This section describes and recapitulates some of these constraints; the next section proposes some possible schemes as a basis for further discussion.

The oldest rocks known, the Early Proterozoic gneiss and amphibolite, are in an uncertain structural position in the tectonic framework of the Pioneer Mountains. If they are autochthonous, they may be the basement on which the Cambrian(?) and Cambrian sediments were laid down. The gneiss and amphibolite terrane shows virtually no rock of sedimentary origin; its protolith is most likely a volcanic and volcanogenic-sedimentary mass that had been repeatedly deformed and metamorphosed. The last major episode that resulted in elemental redistribution on outcrop scale was at about 1,600 Ma. This suite of rocks of calc-alkalic and intermediate chemical affinities might have been part of an island arc suite that was added to the Proterozoic craton. Alternatively, if the rocks were allochthonous, they might have been accreted somewhere else and moved to the present position through much later processes of transportation in response to events on the cratonal margin farther to the west. (See Coney and others, 1980.) Of particular interest is a report by Harrison and others (1980) which discusses gneissic rocks 2.2 Ga in age that occur near Spokane as part of the thrust assemblage of Belt Supergroup rocks. Brown (1980) and Read (1980) described overthrusting of Precambrian gneiss ("core rock") about 2 Ga in age (Wanless and Reesor, 1975) over cratonal basement rocks in the Shuswap terrane. The overthrust rocks came in from an area to the west now occupied by younger rocks of an accreted terrane (Monger and others, 1972). The age of the overthrust rocks, based on discordant lead ages, is similar to that of the Early Proterozoic terrane in the Vipond Park quadrangle, and this superficial analogy deserves further study.

The presence of the Cambrian(?) and Cambrian Black Lion Conglomerate and the lower member of the Silver Hill Formation at Sheep Mountain and their absence across the Adson Creek fault in the Gray Jockey area only a few kilometers away is difficult to understand. To recapitulate briefly: Rocks immediately above the Proterozoic sedimentary rocks at Gray Jockey Peak (the sequence of Swamp Creek) are either the upper, limy member of the Silver Hill Formation or the next younger Hasmark Dolomite; the Black Lion and the lower, nonlimy member of the Silver Hill Formation are absent. Thus, a section several hundred meters thick representing the lower Paleozoic clastic sediments is missing. Rapid lateral changes in conditions of erosion (removal of the sequence of Swamp Creek in the Sheep Mountain area), tectonism (formation of the troughs that received Black Lion clastic material and the maintenance of the trough), and sedimentation may have occurred. The formation of the troughs may be related to graben formation, perhaps peripheral to an Early Proterozoic highland which supplied the clastic and coarsely crystalline feldspar, mica, and rare clasts of gneiss and schist. This highland exerted little effect outside of the trough area. This event may be part of the late Precambrian rifting postulated by Stewart (1972) and could be related to the deposition of a similar rock sequence near Clayton, Idaho, reported by Hobbs and others (1975). (See also Zen, 1977.) In this view, the troughs became progressively filled, the sediments became finer grained, better sorted, and richer in lime up section, until the rocks became largely carbonate in late Silver Hill time. At that time, also, the troughs were insignificant topographic features, and the surrounding terrane not previously included in the troughs and still underlain by the sequence of Swamp Creek became inundated. By Hasmark time, the entire area was a subtidal to intertidal carbonate bank.

This hypothesis would not be drastically modified if the Early Proterozoic gneiss terrane was allochthonous rather than autochthonous and was emplaced as part of the Laramide thrust system. The problem of sedimentation of the Black Lion Conglomerate and the lower, quartzose and noncalcareous member of the Silver Hill Formation in the Sheep Mountain area and farther south, as well as the problem of their absence north of Adson Creek, remains; nevertheless, the pre-Black Lion period of erosion would not be required. Instead, a system of troughs could be developed on a preexisting Middle Proterozoic sedimentary base, and the Black Lion would be conformable above the unexposed sequence of Swamp Creek. The source of the crystalline material in the Black Lion is a separate problem; this source must now be hidden under younger rocks.

Yet a third possibility is that the course of Adson Creek defines a major high-angle fault. Large horizontal movement along this fault could have brought the rocks on either side in juxtaposition, and the original site of deposition of Black Lion Conglomerate may have been far from that of the sequence of Swamp Creek. The relation of the Early Proterozoic gneiss to Black Lion Conglomerate remains a separate problem, and both the autochthonous and thrust hypotheses are permissible.

The presence of chrome-mica-bearing green quartzite pebbles in the conglomerate lens of the lower member of the Silver Hill at Hecla could bear on this subject. The size of the pebbles (1–3 cm) suggests that the source was not far away, and the nature of the clasts suggests a source not unlike that for the Black Lion Conglomerate. The chrome-mica-bearing quartzite lithically matches the Early Proterozoic Facer Formation of the Willard thrust area of northern Utah, described by Crittenden and Sorensen (1980). That source seems too far away and, upon restoration of the thrust, would be from a wrong direction since at Hecla the source direction is consistently south-southeast. James (1981) reported a chrome-mica-bearing quartzite in the Archean of the Tobacco Root Mountains; a more satisfactory source rock in the Ruby Mountain-Dillon region remains to be discovered. Conversely, the availability of chrome-mica-bearing rocks as source terrane would constrain the paleogeographic reconstructions.

Other pre-Cretaceous tectonic events and changes in sedimentary patterns and sources are implied by the nature of the sedimentary rocks (Hasmark through the Red Lion, the Upper Devonian Jefferson and Three Forks, the Madison, the Amsden through the Phosphoria, and the Dinwoody) and gaps in the record (absence of Ordovician through Middle Devonian rocks; absence of post-Dinwoody and post-the unnamed Triassic red limestone (₸r) unit and all Jurassic rocks). The preservation of rocks attributed to the Jurassic in the areas just to the south (Myers, 1952) and of the Triassic red limestone apparently continuous with the Dinwoody sedimentation suggests that their absence was at least in part due to pre-Kootenai erosion rather than solely to absence of sedimentation.

The Kootenai Formation records a new cycle of sedimentation that clearly shows a drastically different tectonic regime compared to previous times. Source terranes include sedimentary rocks of identifiable units of the known Phanerozoic section. From Kootenai through the Colorado time, the sedimentary material generally became less well weathered and less mature, and rock fragments became more important components. The environment became progressively more continental, changing from deltaic for the lower Kootenai, through fresh, standing-water bodies suggested by the gastropod limestone of the uppermost Kootenai, to the fluviatile environments of the Colorado. Volcanic ash became a more significant component in the younger rocks, reflecting the inception of volcanism and plutonism that were penecontemporaneous with Sevier and Laramide orogenic events. These events reflect a plate tectonic regime that predated the sequence of events discussed by Atwater (1970).

The Cretaceous sedimentary cycle was terminated in Late Cretaceous time by two sequential events. Deformation of earlier rocks was followed by the emplacement of at least two thrust sheets of Middle Proterozoic sedimentary rocks. Evidence for the deformation is seen today in the fact that the allochthonous Middle Proterozoic rocks rest on Paleozoic and Mesozoic rocks that show folding and faulting independent of the geometry of the thrust sheets. For instance, at Morrison Hill the thrust is on the Colorado Group; above Grace Lake it is on the Black Lion Conglomerate; along Gold Creek, it is on various Paleozoic carbonate units, and the rocks below the thrust are intricately folded and faulted; at Calvert Hill in the Foolhen Mountain quadrangle at the north end of the west Pioneer Mountains, a rock sequence ranging from Mission Canyon Limestone to Colorado Group appears in a window below Middle Proterozoic rocks. These thrusts were all shallow-level, as rocks derived from the thrust sheet make up the bulk of the clasts of conglomerate assigned to the slightly younger Beaverhead Formation in the area south of Melrose, east of the map area. Thus a period of active deformation and erosion preceded the arrival of the thrust and was the first major event of the Laramide orogeny in the area. The thrusting covered up both the depositional site and the source area for the Colorado Group, respectively, in the east and west Pioneer Mountains.

Dating of the thrust depends on dating the crosscutting igneous rocks. The Keokirk Quartz Diorite with a minimum age of 80 Ma was probably concomitant with the deposition of the upper part of the fluviatile Colorado Group, including volcanogenic, clastic, and ash beds; intrusion of the Keokirk Quartz Diorite, however, did not significantly affect the environment and the nature of the sediments. A stock that intrudes the thrust contact at Dodgson Creek in the Wise River quadrangle has a 76-Ma biotite age. This age may be slightly diachronous with the time of arrival of the rocks at the site of Morrison Hill, but the differences cannot be very great. In the immediate vicinity of Vipond Park quadrangle, the youngest time constraint for the arrival of the thrust sheet is given by the crosscutting relation of the Grayling Lake Granite with the thrust rocks (sequence of Maurice Mountain) in the area along the west shore of Crescent Lake and by the xenolith of the Maurice Mountain northeast of Grayling Lake. The Grayling Lake Granite has an age of about 72 Ma; this may be a cooling age, slightly younger than the intrusive age. Thus, the period between about 76 and 72 Ma was a very active time. It involved (1) sedimentation reflecting the early events of the Laramide orogeny (or the final phase of the Sevier orogeny, which probably earlier had created the terrane in the west Pioneer Mountains that supplied the sediments), (2) the beginning of batholithic intrusions below the sediment cover, (3) folding, faulting, and local erosion, (4) thrust faulting and emplacement of at least two nested thrust sheets, and (5) culmination of the intrusion of calc-alkalic rocks of the Pioneer batholith at a shallow depth beneath an aggrading braided river system.

The intrusive activities were accompanied at least locally by volcanism, for extrusive rocks in the Argenta-Ermont area to the south, mapped by Myers (1952) as Tertiary andesite, were recently shown by Snee and Sutter (1979) to be latest Cretaceous in age. Whether these volcanic rocks were comagmatic with the Pioneer batholith remains to be shown. The preservation of the volcanic rocks is consistent with the conclusion that the later phases of calc-alkalic intrusions were at very shallow depth. We have no record of this volcanism in the map area, but the period of pre-middle Eocene erosion may explain this lack.

The latest Cretaceous and earliest Tertiary intrusive activities must have effectively altered the topography of the area and changed the drainage system. The slide of Colorado Group rocks in the southeast part of the quadrangle may also record this event and the thrusting that preceded it.

The Eocene Trusty Gulch Volcanics were extruded on land, as indicated by the abundant cooling columns, oxidized rubble, and deposits of volcanogenic sand in the rubble beds. No volcanic center is known, and probably none existed; the mode of eruption seems to have been through fissures and, more rarely, pipes. Each individual eruption was of short duration and probably of local extent; thicknesses of individual observed flows are no more than a few tens of meters. Therefore, no individual flow can be used as a time reference. The Eocene mafic dikes follow fractures, and similar fractures also occur in the Trusty Gulch Volcanics; thus, the north-northeast-trending fracture systems are probably Eocene.

The Lowland Creek volcanic activities, most abundantly preserved in the area near Butte, largely thin out toward the Pioneer Mountains. A shallow perlitic intrusion having highly variable flow directions at the head of Dry Hollow Gulch could have been a volcanic neck; equivalent erupted pumiceous rocks are today intercalated with Trusty Gulch Volcanics in the Cherry Creek valley. However, although the two groups of volcanic rocks are synchronous and are found in the same areas, there is neither petrographic nor isotopic evidence that they mixed; distinct plumbing systems must have existed.

A volcanic flow of middle Oligocene age (27.3±1.0 Ma) shows that igneous activity was present during this time, but the sparse data do not permit further conclusion. West of the Fourth of July fault, areas of ash deposits interbedded with ash-matrixed mudflow deposits containing large clasts of local sedimentary and igneous bedrocks are preserved; a fission track age of 27.0±1.3 Ma dates the deposits also as middle Oligocene (Pearson and Zen, 1985). By that time, erosion had clearly unroofed the Pioneer batholith. The centers of these clearly explosive eruptions have not been found.

Some of the large-scale high-angle faults are older than Eocene volcanism and are probably pre-batholith; the ages of most of the faults remain problematic. Although the Oligocene ash beds are bounded locally by the Fourth of July fault, the age relations remain undefined, and the contact could be a fault-line scarp or a later reactivation of a preexisting fault. The major episode of mountain-front faulting that defines, for instance, the lower Wise River valley, may be as young as the latest Tertiary as suggested by Pardee (1950; see also Reynolds, 1979), but local evidence does not permit an assignment of time of faulting.

The formation of the Pleistocene icecaps in the mountains was followed by valley and cirque glaciation; the last stages were clearly Pinedale or younger, as on canyon walls the rock surfaces locally retain their specular polish. The chronology, however, largely remains to be established. (See Alden, 1953.)

The origin of the flat area known as Vipond Park has caused much speculation. (See Alden, 1953; Pardee, 1950; Perry, 1934.) A conspicuous lateral moraine from the Canyon Creek system of valley glaciers defines the southeastern border of the park. It is not certain that the park itself is the remnant of an old peneplain (Pardee, 1950, p. 368); Vipond Park is not developed on granite as Pardee suggested. There is one outcrop of flat-lying limestone pebble, tufa-cemented conglomerate (Qcg on map) exposed in a small stream cut under the surficial material. In parts of the area, the road is underlain by material that behaves like volcanic ash in wet weather; thus, Oligocene volcanic ash might underlie part of the park. In addition, the origin of the flat park seems to differ from that of the flat-to-tilted, smooth erosion surface that cuts across hard rocks in the higher ranges; these latter surfaces are clearly products of an earlier episode of erosion.

The final record in the geologic history is deposition of the Mazama ash beds, 6845±50 years ago, when land surface development was essentially as it is today. Earthquakes in southwestern Montana are frequent, and there is every reason to assume that the evolution of the tectonic and topographic features in the Montana part of the basin and range province is still proceeding.

TECTONIC SPECULATIONS

Final remarks are in order. We need to know whither came the thrust sheets of Middle Proterozoic sedimentary rocks, and what is the nature of the source rock from which the batholithic magmas were derived. I believe the two questions are related and tentatively propose an answer to both. I have little factual evidence to compel the answer; it is stated here in hope of stimulating discussions and further search.

Middle Proterozoic sedimentary rocks and rocks that bear strong lithic and tectonic affinity to them occur widely in western Montana and in Idaho; with minor exceptions, they seem to be allochthonous. Harrison and others (1980) suggest that all Belt rocks north of the Lewis and Clark Line and west of the Disturbed Belt, as far west as Spokane, Wash., are allochthonous. Harrison and others also suggest that a piece of older crystalline rock apparently underlying the Middle Proterozoic sedimentary rocks was attached to the Belt rocks in the allochthon, thus providing a rare glimpse of the substrate on which the Belt and other Middle Proterozoic rocks were laid down. Restoration of these allochthons probably would lead to a pre-Belt basement many kilometers wide approximately west of the present Idaho-Washington and Idaho-Oregon line; this terrane is nowhere in sight as rocks west of this line are of much younger age and belong to terranes interpreted to have been accreted during and since Mesozoic time (Coney and others, 1980).

I suggest that the terrane on which the Middle Proterozoic rocks were deposited was a part of the North American craton that had been removed (decreted) prior to the accretion of the terranes that now underlie much of Oregon, Washington, northern California, and northwestern Nevada. This basement was largely sialic rather than oceanic in composition and might have been, in part, old accreted arc terranes. During the Mesozoic plate-margin orogeny, this terrane did not subduct readily because of its buoyancy; rather, it moved along right-lateral transform faults comparable to the modern San Andreas and Tintina fault systems. Prior to the accretion of the present terranes of the Pacific coastal United States, the decreted terrane had moved out of the way in an earlier cycle of plate motion. For some reason, rocks of Middle Proterozoic age originally deposited on the now-decreted terrane to the south and west were thrusted onto the remaining craton in this area as the terrane moved past.[2] This unloading must have preceded the arrival of the new accreted terrane and took place sometime in the middle of late Mesozoic time (Jones and others, 1977; Churkin and others, 1980; Coney and others, 1980; Monger and Irving, 1980; Monger and others, 1972). The age of about 75 Ma determined on the younger limit of the allochthon in the Pioneer Mountain region is the time of final arrival, which must be somewhat younger than the age of initiation of the odyssey of the thrust sheets; the age relations as known are at least acceptable. The distance between the Pioneer Mountains and the Idaho-Oregon line is 300 km; if an allochthon moved at an average rate of 1/3 to 1/2 cm per year, then 60 to 90 Ma would be needed, making the initiation of the odyssey an event at 130 to 160 Ma. Both the

[2]In this context, the paper by Aleinikoff and others (1981) of Early Proterozoic age zircon from augen gneiss immediately southwest of the Tintina fault in Alaska especially invites speculation of a former connection.

ages of unloading of different thrust masses and ages of movement of a given mass from the point of unloading to the point of final rest must have been diachronous.

I further propose that the Early Proterozoic gneiss and amphibolite of the Vipond Park quadrangle might provide a glimpse of the nature of the basement now lost to view. Especially if this terrane is itself allochthonous, as indicated under the second or third alternatives in the section on structural geology, it could in fact be a piece that arrived as part of the thrust package of Middle Proterozoic rocks. The Early Proterozoic terrane in the area is a geochemically admissible source rock for the Pioneer batholith in terms of its major, trace, and rare earth elements, as well as isotopic chemistry (Zen and others, 1980; Snee, 1981, written commun.). I interpret it as an accidentally preserved piece of the lost basement that once occupied the margin of the Mesozoic North American craton, while the bulk of the lost terrane was obliquely subducted to a depth of a few tens of kilometers or moved laterally out of the way through strike-slip motion on a transform fault. Subduction of this and similar terranes to a depth of about 30 km would, upon thermal equilibration, provide the temperature for a hydrous gneiss-and-amphibolite suite to begin extensive partial melting, leading to the generation of magmas for not only the Pioneer batholith but for other batholiths of the region.

The Pioneer batholith is, by the calculated initial $^{87}Sr/^{86}Sr$ ratios, exceptionally radiogenic; the Boulder batholith is much less so, even though its lead isotope signature is like that of the Pioneer batholith (Doe and others, 1968; B.R. Doe, 1980, written commun.). A model that could explain these relations postulates a lead-homogenized old Precambrian terrane. That terrane, sometime during its history, underwent differentiation through melting that did not affect the lead isotopes but concentrated Rb in the melt phase; subsequent segregation and consolidation of the melt caused some portions to become richer in radiogenic strontium than other portions. Subduction and partial melting of portions of this partially differentiated terrane could have given rise to the observed differences in initial strontium ratios that distinguish the Pioneer batholith from the Boulder batholith, batholiths which share the same major chemistry, age, and tectonic relations to the country rocks and to the Laramide deformation.

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