Geology of the Tepee Creek Quadrangle Montana-Wyoming

GEOLOGICAL SURVEY PROFESSIONAL PAPER 609

Geology of the Tepee Creek Quadrangle Montana-Wyoming

By IRVING J. WITKIND

GEOLOGICAL SURVEY PROFESSIONAL PAPER 609

A description of the stratigraphy, structure, and surficial deposits in a 15-minute quadrangle that includes the meizoseismal area of the Hebgen Lake earthquake of August 17, 1959

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1969

UNITED STATES DEPARTMENT OF THE INTERIOR

WALTER J. HICKEL, *Secretary*

GEOLOGICAL SURVEY

William T. Pecora, *Director*

Library of Congress catalog-card No. GS 68-317

For sale by the Superintendent of Documents, U.S. Government Printing Office
Washington, D.C. 20402

CONTENTS

ILLUSTRATIONS

TABLES

GEOLOGY OF THE TEPEE CREEK QUADRANGLE, MONTANA–WYOMING

By Irving J. Witkind

ABSTRACT

The Tepee Creek quadrangle, in Gallatin County, Mont., and Yellowstone National Park, Wyo., is part of the Northern Rocky Mountains physiographic province.

The rocks range in age from Precambrian to Recent and have been divided into 28 mapped units. The basement complex consists of pre-Belt crystalline metamorphic rocks, including granite gneiss, dolomite, tremolite marble, amphibolite, mica schist, and quartzite, that are provisionally correlated with rocks exposed near Ennis, Mont. Five and possibly six of the seven periods of the Paleozoic and all three periods of the Mesozoic are represented by the sedimentary strata which rest on the basement complex; only Silurian and possibly Ordovician rocks are missing. The exposed strata have a maximum thickness of about 6,200 feet and range in lithology from coarse-grained conglomeratic sandstone to microcrystalline marine limestone. In all, 19 sedimentary units have been mapped: of these, four are Cambrian, one is Cambrian and Ordovician, one is Devonian, two are Mississippian, two are Pennsylvanian, one is Permian, two are Triassic, two are Jurassic, and four are Cretaceous.

This sedimentary sequence is broken by eight unconformities, of which three represent widespread major erosional episodes. The first major interruption occurred after Late Cambrian but before Late Devonian time; the second occurred during Late Mississippian time; and the third occurred after Early Triassic but before Middle Jurassic time. The remaining five breaks in the sedimentary record are probably local and presumably reflect minor temporary oscillations of the Wyoming shelf.

Locally the sedimentary units are either deformed by intrusive dacite porphyry dikes, sills, and a large semiconcordant mass known as the Gallatin River laccolith or are buried by such extrusive rocks as andesite breccia, shoshonite, or rhyolite welded tuff, known as the Yellowstone Tuff.

Two basic tectonic elements, long since destroyed, determined the nature of the sediments deposited in the quadrangle: the Cordilleran geosyncline which lay to the west, and the Wyoming shelf, part of which underlay the quadrangle but most of which was to the east. Marine conditions generally prevailed, but minor oscillations of the Wyoming shelf determined whether the quadrangle was below sea level, receiving deposits of shallow marine sediments from seas which spread eastward from the geosyncline, or above sea level exposed to subaerial erosion.

The area has undergone at least two major episodes of structural deformation since the Precambrian. The first occurred during the Laramide deformation (Late Cretaceous and early Paleocene), when lateral compressive forces from the southwest folded, overturned, and locally shoved the country rocks northeastward on extensive northwest-trending thrust faults. Three structural elements attributable to this episode of orogeny can be delineated in or near this area: (1) the Madison thrust block (upper plate) southwest of the mapped area, formed by Precambrian crystallines, capped here and there by Paleozoic rocks, that was thrust northeastward onto (2) the Cabin Creek zone (lower plate), part of which is in the mapped area, and which includes overturned and thrust-faulted strata formed in front of the Madison thrust block; and (3) the Pika Point zone, marked by broad symmetric warps which reflect the great distance between this zone and the Madison thrust block.

The second major deformational episode, which probably began during the middle Tertiary (Miocene?) and is still active, may be the result of relaxation following renewed regional arching and uplift. As the elevated region subsided, it broke along high-angle normal faults to form tilted fault blocks. These normal faults, which closely parallel the thrust faults, were probably guided by zones of weakness formed during Laramide deformation. Four of these blocks are in or near this area. Southernmost, beyond the mapped area, is the Hebgen block (including Hebgen Lake), whose north flank is determined in part by the Hebgen normal fault. Next is the Red Canyon block, whose north flank is defined in part by the Red Canyon normal fault. Next is the Kirkwood block, which is bounded along part of its northeast edge by the Upper Tepee normal(?) fault. The fourth tilted(?) fault block is the Monument Mountain block, whose northern edge lies beyond the quadrangle.

Reactivation of the Hebgen and Red Canyon normal faults on August 17, 1959, resulted in an earthquake which was felt over 600,000 square miles. As a result of differential movement on these faults, the Hebgen and Red Canyon fault blocks were tilted northeastward following a pattern of subsidence which seems to have recurred many times in the past.

Before Pleistocene ice spread across the quadrangle, the following major events had probably occurred: (1) Folding followed by thrust faulting which probably began in this general area during the very Late Cretaceous, persisted through the Paleocene, and may have continued through all or part of early Eocene time; (2) emplacement of intrusives shortly thereafter, most likely before early middle Eocene time; (3) partial burial of the eastern part of the quadrangle by andesite breccia and shoshonite flows during the early middle Eocene; (4) epeirogenic uplift accompanied by normal faulting during the middle Tertiary, most likely in Miocene time; and (5) pyroclastic eruptions of rhyolite tuff possibly during either the late Pliocene or the early Pleistocene.

The area was glaciated at least three times. One advance, possibly during the pre-Bull Lake time of the Pleistocene, may have reached from the Madison Range in Montana to the Absaroka Range in Wyoming and buried the entire quadrangle. An ice advance during the Bull Lake Glaciation is represented by morainic deposits of two ice lobes which advanced westward into the eastern reaches of the quadrangle. Drift of younger glaciers, presumably of Pinedale age, clogs valley floors and is plastered on valley walls; these deposits are the residue of alpine glaciers

which spread radially from the high mountains in the western extremities of the quadrangle.

The oil and gas potentialities of the Tepee Creek quadrangle are still unknown, although four wells have tested areas adjacent to the quadrangle. Three tests were in the West Yellowstone basin, but none passed completely through the basin fill of outwash sand and gravel and volcanic rocks. The fourth well tested the broad doubly plunging Carrot Basin anticline which is about 3 miles west of the northwest edge of the quadrangle. This test was abandoned after penetrating the upper part of the Madison Group of Mississippian age. No shows of oil or gas were reported from the wells in the West Yellowstone basin, but two oil shows were noted in the Carrot Basin test.

Although no mineral deposits are known in the Tepee Creek area, a small contact-metamorphic deposit of magnetite and hematite occurs about 1 mile north of the quadrangle along the roof of the Gallatin River laccolith.

INTRODUCTION

The Tepee Creek quadrangle is one of several quadrangles mapped and being mapped by geologists of the U.S. Geological Survey in southwestern Montana. The geologic work in this quadrangle began in June 1958, and the southern third had been mapped by August 1959.

Near midnight on August 17, 1959, a major earthquake was felt over about 600,000 square miles throughout the Northwestern United States and adjacent parts of Canada. This earthquake,[1] centered in the southern third of the area described in this report (fig. 1), created great public interest, for Yellowstone National Park is directly to the east and was crowded with tourists at the time. The extensive damage in the epicentral area and in Yellowstone National Park and the resultant rescue efforts were described in the Nation's newspapers and magazines. The U.S. Forest Service has since set aside the area of maximum destruction (the meizoseismal area) as the "Madison River Canyon Earthquake Area," and the interested tourist, guided by signs and display exhibits, can examine at first hand the fault scarps, landslides, and other features formed during the earthquake.

Two field parties of the Geological Survey were in or near the epicentral area at the time of the earthquake. J. B. Hadley, assisted by W. D. Long, was at Ennis, Mont. (fig. 2), and J. B. Epstein (my assistant), and I were camped on a small knoll overlooking Hebgen Lake. Hadley and I immediately began to study the geologic effects of the earthquake. Additional personnel—including geologists, geophysicists, hydrologists, and engineers—from the Geological Survey and cooperating Government agencies were furnished, and the work was carried on to the end of September 1959, when inclement weather stopped all further fieldwork. Detailed results of this study are in U.S. Geological Survey Professional Paper 435, "Geology of the Hebgen Lake, Montana, Earthquake of August 17, 1959."

[1] Hereafter referred to as the Hebgen Lake earthquake.

The remainder of the quadrangle was mapped during 1960, 1961, and 1962. Major objectives were to study the folded and faulted Paleozoic and Mesozoic rocks near the south end of the Madison Range, to study the Precambrian crystalline rocks in this sector of Montana, to study the surficial deposits and their correlation with established type localities, and to prepare a detailed geologic map of the quadrangle as an addition to the geologic map of the United States.

FIELD METHODS

The geologic contacts were plotted in the field on topographic base maps at a scale of 1:24,000 and a contour interval of 80 feet. Subsequently the completed geologic map was reduced photographically to a scale of 1:48,000 and issued as Miscellaneous Geologic Investigations Map I–417 (Witkind, 1964a). The same map, modified slightly and in multicolor format, is plate 1 of this report.

As much of the quadrangle is distant from roads, most of the work was done from temporary field camps which were moved whenever the work in an area was completed. Pack strings were used exclusively to transport supplies.

PREVIOUS WORK

F. V. Hayden (1873) visted the Gallatin Canyon–Yellowstone National Park area briefly in 1871 and at length in 1872. During the second visit, his party camped for several days in the West Yellowstone basin, then a broad grass-covered valley containing isolated clusters of lodgepole pine, while Hayden crossed Targhee Pass and examined the country near Henrys Lake. A. C. Peale, a member of Hayden's Survey, explored one of the branches of the Madison River and, to see the country, climbed a high hill north of the West Yellowstone basin. Impressed by the scenery, he (1873, p. 160) wrote, "The view from the point was one of the fairest that I have ever gazed upon."

The eastern third of the quadrangle—that part within Yellowstone National Park—was visited by Iddings and Weed at some time between 1883 and 1891, during the course of Arnold Hague's broad survey of the park (Hague and others, 1896, 1899). As part of their work they studied and described the Gallatin River laccolith exposed at the junction of Snowslide Creek with the Gallatin River (pl. 1, and p. 39) (Iddings and Weed, 1899, p. 57–59, 84).

This sector of Montana has been visited briefly by various stratigraphers interested in specific problems. Condit (1919), studying the pre-Jurassic unconformity and the distribution of the Phosphoria Formation (p.

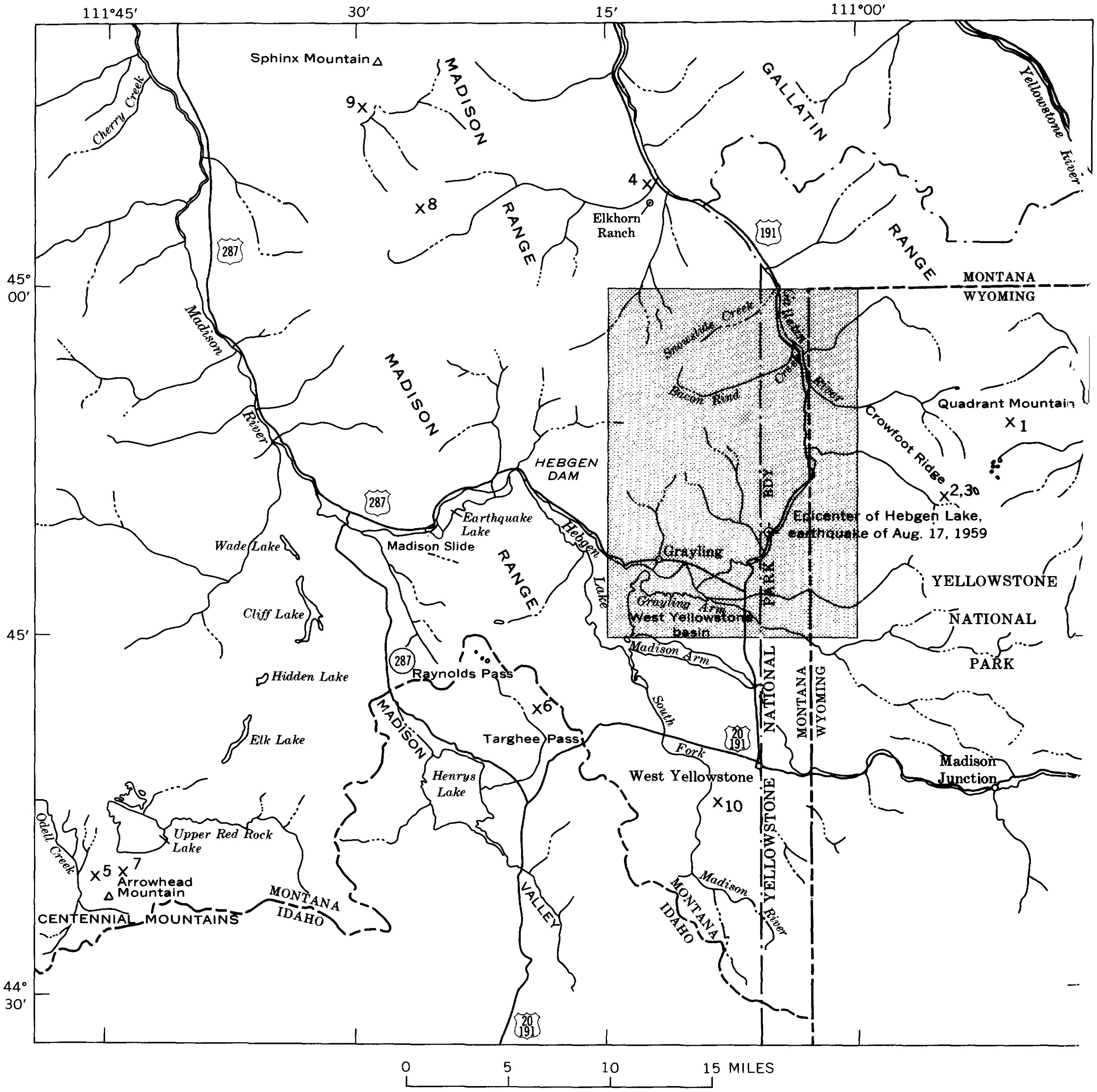


FIGURE 1.—Area near the Tepee Creek quadrangle (stippled), showing location of detailed measured sections (×). Measured sections are identified as follows: (1) Scott (1935); (2) Deiss (1936); (3) Grant (1965); (4) Gardner, Hendricks, Hadley, and Rogers (1945); (5) Arrowhead Mountain section, Sloss and Moritz (1951); (6) Targhee Creek section, Sloss and Moritz (1951); (7) Arrowhead Mountain section, Hanson (1952); (8) Taylor Peaks section, Hanson (1952); (9) Cressman and Swanson (1960); (10) Richmond and Hamilton (1960).

31), passed through the area in 1916 during a reconnaissance trip from Helena, Mont., to Yellowstone National Park. Scott (1935), interested in the "Quadrant-Amsden" relations, measured a section on Quadrant Mountain in the northwest corner of Yellowstone National Park, about 8 miles to the east. Deiss (1936) measured detailed sections of the Cambrian strata exposed on Crowfoot Ridge, about 3 miles to the east in Yellowstone Park, and the same area has been visited recently by Grant (1965), who studied the Snowy Range Formation of Late Cambrian age.

During the early 1940's, as interest in the petroleum potential of upper Paleozoic and lower Mesozoic rocks increased, a program involving stratigraphic measurements at selected localities throughout central and western Montana was completed by members of the U.S. Geological Survey (Gardner and others, 1945; Rogers and others, 1945). As part of this work several stratigraphic sections were measured near the Tepee Creek quadrangle; the nearest is in sec. 2, T. 9 S., R. 4 E., at the Elkhorn Ranch.

Moritz (1951) and Sloss and Moritz (1951), as part of a stratigraphic study of southwestern Montana, also measured sections of Paleozoic and Mesozoic rocks southwest of the quadrangle. A. M. Hanson (1952), while with the Montana Bureau of Mines and Geology, measured several stratigraphic sections of Cambrian strata exposed both southwest and north of the quadrangle. More recently, this sector of southwestern Montana was visited by E. V. Cressman and R. W. Swanson (1956, p. 2852) as part of a broad study of the Permian rocks in the western phosphate fields of the United States. In the period 1951–54, the welded tuffs and flows of the Yellowstone rhyolite plateau were mapped and classified by F. R. Boyd (1961). The northwest corner of his map adjoins the east edge of this quadrangle.

Graduate students from Princeton University, the University of Wyoming, and the University of Michigan have mapped in and near this area as integral parts of their master or doctoral dissertations. During the field seasons of 1946, 1947, and 1948, James A. Wilsey, then a graduate student at Princeton University, studied the geology of the upper Gallatin River area. He was assisted in 1948 by W. B. Hall who, after Wilsey's early death in 1949, took over the work. Hall, by then a graduate student of the University of Wyoming, expanded the area of investigation to include much of the southern parts of the Gallatin and Madison Ranges (Hall, 1961).

In 1950, graduate students from the University of Michigan mapped several areas in and near the quadrangle, and the results of their work are now available as unpublished M. S. theses (Freeman and others, 1949; Jacques, 1949; Leeder and others, in Hume and Leeder, 1950; LeVan and McLean, 1951; Liddicoat and May, 1950; Reiter, 1950).

During the comprehensive examination of the epicentral area of the Hebgen Lake earthquake, some of the volcanic, surficial, and glacial deposits as well as the structural relations were studied by geologists of the Geological Survey. These workers subsequently published the results of their findings in scientific journals, as preliminary articles in various of the Annual Reviews of the Geological Survey, and as chapters of Professional Paper 435 (Myers and Hamilton, 1961, 1964; Witkind, 1961; Witkind and others, 1962).

Warren Hamilton (1964) examined some of the volcanic rocks exposed in the West Yellowstone and Madison Junction quadrangles during a reconnaissance of an area near the West Yellowstone basin, and compared these rocks with samples of volcanic rock cored from the Morgan-Martzell well (p. 84), which is in the basin. Pollens associated with both the exposed rocks and the cored volcanic rocks were studied by Estella B. Leopold and a paper by Hamilton and Leopold (1962) discusses the significance of the pollen and the Oligocene age assigned to the volcanic rocks (p. 51).

G. M. Richmond (1964) studied various glacial deposits in and near the mapped area. During this work, large lateral moraines of the Quarternary Bull Lake Glaciation were found partly buried by an obsidian-rhyolite flow, and this exposure was summarily described by Richmond and Hamilton (1960).

ACKNOWLEDGMENTS

Many people and organizations furnished information and assistance during the course of this project and it is not possible to list them all. I thank particularly F. A. Dorrell, L. O. Peck, C. W. Silvernale, and E. C. Slusher and their staffs, of the U.S. Forest Service, and J. R. Fraser and J. W. Hodges, of the National Park Service, for information about trails, permitting use of their agencies' facilities, including patrol cabins, and at times lending me special field equipment.

The work was aided by the kind and generous help of many area residents, including the late Mr. F. W. Kersenmacher of the Diamond K ranch, who first guided me through the southern part of the report area. Mr. Martin B. Portmann of the Diamond P ranch suggested access trails and suitable camp sites. Mr. R. B. Whitman of West Yellowstone, Mont., and Messrs. Bob Miller and Ron Hymus of the Elkhorn Ranch packed in supplies to my field camp and time after time moved my camp to new localities.

Mr. William Martzell of West Yellowstone furnished the core of much of the Morgan-Martzell well (p. 84):

a detailed log of the core was prepared by W. B. Hamilton (1964).

I thank the field assistants who worked with me during the course of the project. In 1958, C. E. Harris helped map the southwest corner of the quadrangle. In 1959, J. B. Epstein worked with me in the Kirkwood Ridge area and assisted in the earthquake study. In 1960, C. B. Mason assisted in the mapping of both the southeast corner of the quadrangle and the Cabin Creek–Sage Peak–White Peak area. In 1961, J. C. Davis assisted in the mapping of the Red Mountain–Snowslide Mountain–Monument Mountain area; and in 1962, N. F. Davis assisted in the mapping of the northeastern and northwestern corners of the quadrangle.

GEOGRAPHY

LOCATION

The Tepee Creek 15-minute quadrangle, directly west of and including part of the northwest corner of Yellowstone National Park (fig. 1), consists of about 218 square miles of forested semiwilderness. It extends from long 111°00′ W. to 111°15′ W., a distance of about 12½ miles, and from lat 44°45′ N. to 45°00′ N., about 17½ miles. The quadrangle is partly in southwestern Montana and partly in northwestern Wyoming.

Deeply dissected rugged mountains are in the central, western, and northwestern parts of the quadrangle (pl. 2). Along the west edge, parallel ranges enclose broad spacious basins. The eastern part of the area (in Yellowstone Park) is a primitive wilderness underlain chiefly by volcanic rocks. The southern part of the quadrangle includes both the Grayling Arm of Hebgen Lake and the north edge of a broad, almost featureless, outwash-filled basin known as the West Yellowstone basin (fig. 1).

West Yellowstone, Mont., about 8 miles south of the quadrangle, is the nearest community, and most of its residents depend for their livelihood upon the trade of tourists entering Yellowstone National Park through its west entrance. As a result, the town is densely populated during the summer season and almost deserted in the winter months. Bozeman, Mont., a town of about 13,300 population, is 67 miles north of the quadrangle and is one of the richest agricultural communities in southwestern Montana. Ennis, Mont. (population 600), in the Madison Valley, is about 60 miles to the northwest on Montana State Highway 287.

ACCESSIBILITY

Two all-weather highways traverse the quadrangle (fig. 1; pl. 2). U.S. Highway 191, which connects West Yellowstone and Bozeman, follows the valleys of Grayling Creek and the Gallatin River in the eastern part of the quadrangle. Montana State Highway 499 crosses the southern part of the quadrangle parallel to the north shore of Hebgen Lake and joins U.S. Highway 191 at the Duck Creek wye. At the time of the Hebgen Lake earthquake of 1959, several segments of State Highway 499 (then known as Montana State Highway 287) along the northwest arm of Hebgen Lake slumped into the lake, and that part of the highway near the mouth of the Madison Canyon was buried beneath the disastrous Madison Slide (Hadley, 1964). In 1961 a new all-weather road was constructed around the northwest edge of the recently impounded Earthquake Lake (fig. 1) and across the slide to connect with the undamaged part of State Highway 287.

All other roads are short, and some are unusable when wet. Most lead to local ranches or to Forest Service horse trails, which are the trunk routes to the higher and more remote back country. Game trails, locally wide enough for pack transport, extend outward in random fashion from these established trails.

Although West Yellowstone is the terminus of a branch line of the Union Pacific Railroad which extends northeastward from Pocatello, Idaho, the line has carried only freight since 1961.

In 1965 a modern terminal and a landing strip were constructed north of West Yellowstone, and the community is now served during summer months (about June 12 to September 15) by various commercial airlines. The airport is also used during summer months by Forest Service fire-fighting crews (smokejumpers) and by several small-scale commercial operators who offer scenic flights across the earthquake area and Yellowstone and Grand Teton National Parks.

CLIMATE AND ECOLOGY

The climatological data reflect the mountainous nature of this part of southwestern Montana (table 1). The summers are short and cool, and the winters are long and arduous and marked by deep, long-lasting snows. The summers rarely have extremely hot days; mean summer temperatures are in the high 50's (°F), and July and August are the hottest months of the year. The coldest period is during February, when mean minimum temperatures are as low as 1°F. Mean winter temperatures are about 10°F. In general, the first killing frost is in early September, and the last killing frost is in mid-June.

Precipitation is nearly evenly spread throughout the year, the greatest occurring in January and the least in August. Mean annual precipitation is about 21 inches in West Yellowstone and about 26 inches near Hebgen

TABLE 1.—*Climatological data gathered at Hebgen Dam, West Yellowstone, and Grayling, Mont.* [1]

[Tr.=Trace amounts. Data from the U.S. Weather Bureau]

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Annual
Hebgen Dam [1]													
Temperature (° F):													
Mean:													
1906–30	13.4	15.4	23.4	34.2	44.0	51.6	58.9	57.2	48.4	37.8	25.3	13.3	35.2
1931–52	10.5	15.2	22.9	35.4	46.1	53.3	61.5	60.1	51.4	40.9	25.1	16.2	36.6
Mean maximum:													
1906–30	22.1	28.8	36.6	47.0	57.9	67.0	77.0	75.7	65.1	51.0	35.9	23.3	49.0
1931–52	20.5	27.0	35.5	48.1	60.5	68.2	79.0	77.9	67.4	53.2	33.8	24.8	49.7
Mean minimum:													
1906–30	4.6	1.9	10.3	21.4	30.1	36.2	40.8	38.8	31.8	24.5	14.7	3.3	21.5
1931–52	.6	3.3	10.2	22.7	31.6	38.3	43.9	42.3	35.4	28.6	16.3	7.7	23.4
Mean precipitation (in.):													
1906–30	2.50	1.82	1.69	1.44	2.24	2.10	1.80	1.38	1.68	1.71	1.63	1.89	21.88
1931–52	2.73	2.43	2.49	1.59	2.20	2.84	1.59	1.35	1.56	1.88	2.54	2.81	26.01
Mean snowfall (in.):													
1906–30	38.7	25.7	20.8	10.6	3.4	.4	.0	Tr.	1.0	7.0	16.4	27.8	151.8
1931–52	43.7	36.8	27.8	7.6	2.7	Tr.	Tr.	Tr.	1.6	5.7	28.3	41.8	196.0
West Yellowstone													
Temperature (° F):													
Mean:													
1909–30	12.5	17.1	23.6	33.1	42.0	49.1	56.9	54.4	46.3	35.8	23.3	13.2	33.9
1931–52	10.7	16.0	23.0	34.9	44.2	51.4	59.3	56.8	48.0	38.1	22.5	15.2	35.0
Mean maximum:													
1909–30	25.4	31.9	39.3	48.4	57.7	66.4	77.3	75.4	65.0	51.6	37.9	23.9	50.3
1931–52	23.8	31.2	38.9	50.7	60.6	68.8	80.6	78.5	67.8	54.1	35.8	27.0	51.5
Mean minimum:													
1909–30	−.2	2.3	7.9	17.6	26.3	32.0	36.5	33.4	27.6	19.9	8.7	−.5	17.6
1931–52	−2.4	.8	7.1	19.2	27.8	33.9	37.9	35.1	28.2	22.0	9.1	3.4	18.5
Mean precipitation (in.):													
1909–30	2.02	1.47	1.49	1.31	1.65	1.93	1.45	1.28	1.51	1.34	1.67	1.53	18.65
1931–52	2.24	1.81	1.98	1.28	1.94	2.54	1.23	1.23	1.25	1.68	1.76	2.17	21.11
Mean snowfall (in.):													
1909–30	24.5	16.6	16.8	10.1	55.2	1.0	Tr.	Tr.	1.2	5.7	16.0	18.2	115.3
1931–52	31.4	27.2	26.8	8.7	2.7	.6	.0	.1	1.6	6.0	19.8	30.3	155.2

[1] Includes Grayling, Mont., for period 1904–12 inclusive.

Dam. Commonly, 8–10 days of each month are marked by precipitation averaging 0.01 inch or more. Most of the snow falls from November through March, although meteorological records show that some snow falls in the high mountains in every month except July. Mean annual snowfall is about 155 inches in West Yellowstone and about 196 inches near Hebgen Dam.

The prevailing winds are from the south during the summer and from the north during the winter.

Forest, mainly pine, blankets nearly four-fifths of the quadrangle, and four ecologic zones can be recognized. The transition (or foothills) zone is marked by widespread sagebrush, grasses, and many kinds of flowers, plus small stands of quaking aspen, chokecherry, and willow. This zone, chiefly along the north shore of Hebgen Lake, probably extends up to an altitude of about 7,000 feet.

The montane zone, above 7,000 feet, contains extensive forests of lodgepole pine, limber pine, whitebark pine, and Douglas-fir. Dense, almost impenetrable stands of lodgepole flourish in the thin soil developed on the rhyolite welded tuffs which underlie the eastern part of the quadrangle. Unfortunately, the superficial root system developed by these trees is inadequate to support them, and every windstorm topples hundreds. This zone also contains many meadows which furnish grazing for big-game animals; the chief grass is Idaho fescue.

Alpine fir, normally found in the next higher ecologic zone, locally grows in small stands along north slopes and valley walls in deep, narrow canyons intersected by the montane zone. The montane zone blends imperceptibly into the subalpine zone near the 8,800-foot contour.

The subalpine zone, restricted to the upper flanks of the highest ranges in the quadrangle, includes stands of Engelmann spruce, alpine fir, whitebark pine, and some lodgepole pine. With increasing altitude, spruce and fir replace the lodgepole pine. The rich verdancy of the forests typical of the lower montane zone is broken in the subalpine zone by patches of bare rock or by small alpine meadows supporting a sparse growth of Idaho fescue and dwarf flowers. The subalpine zone ends at an altitude of about 9,600 feet (timberline), where it passes into the uppermost ecologic zone, the alpine zone—a striking treeless area of grass-covered meadows, colorful thin crustlike lichens and mosses, ground-hugging shrubs, and dwarf flowers inured to wintry blasts.

PHYSICAL GEOGRAPHY

Three wide valleys, or basins, separated by towering mountains dominate the physical geography of this part of southwestern Montana and reflect the basic struc-

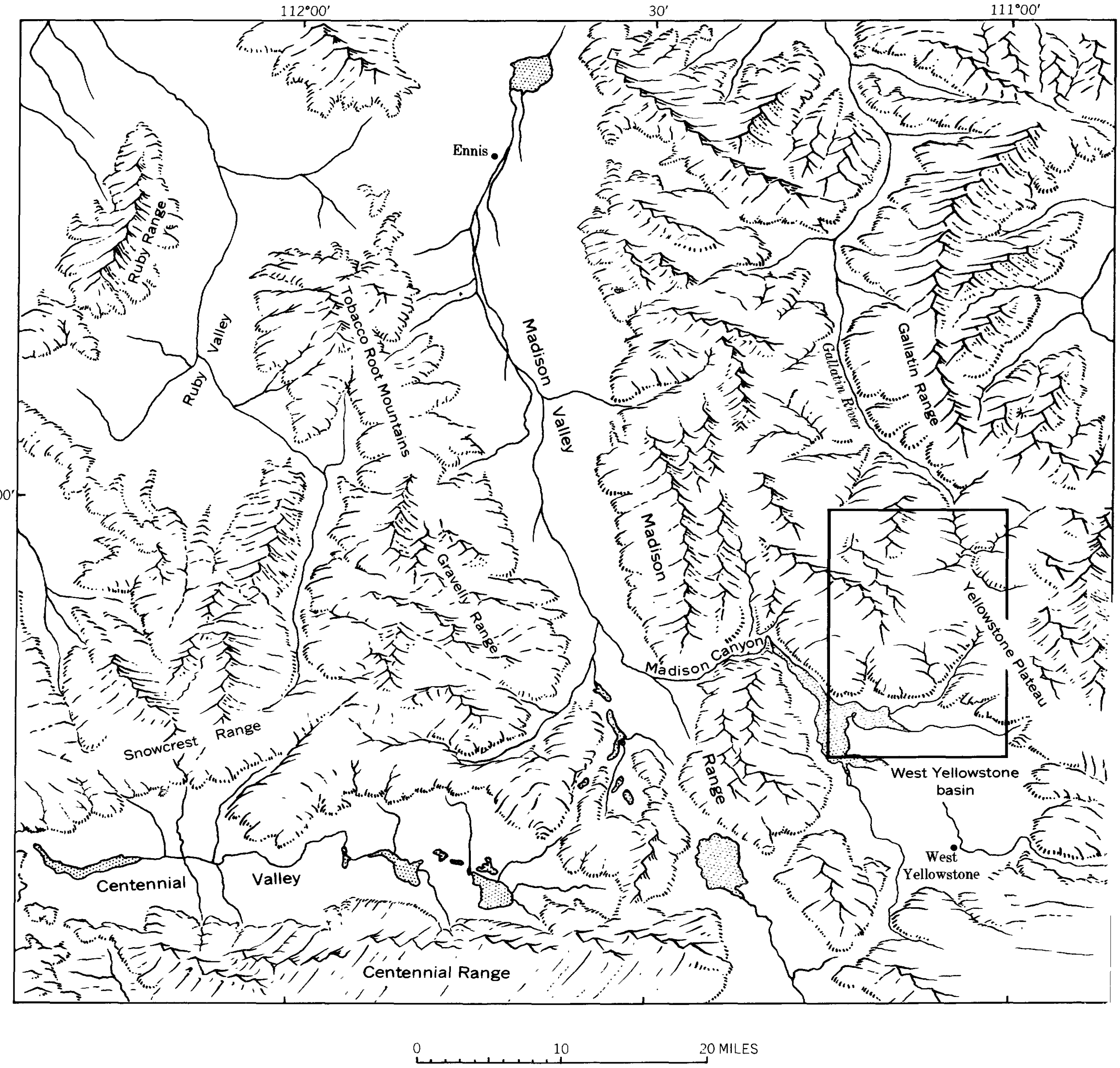


FIGURE 2.—Physiographic diagram of the Tepee Creek area (outlined by the oblong).

tural framework of the region. The largest of the three is the Madison Valley, which is about 55 miles long and about 12 miles across at its widest part. The Centennial Valley, about 47 miles long and 6 miles wide, trends eastward, seemingly cutting across the northwestward-trending structural grain of the region. The West Yellowstone basin, the smallest of the three valleys, is oval. Its long axis, which trends northwestward, is about 18 miles long, and its short axis is about 12 miles long.

The Tepee Creek quadrangle, part of the Northern Rocky Mountains province, is east of the Continental Divide and includes parts of four physiographic units: the Madison Range, the Gallatin Range, the Yellowstone Plateau, and the West Yellowstone basin (fig. 2). The quadrangle, in gross aspect, is divided unequally by a continuous valley, followed by U.S. Highway 191, which extends northward through the eastern part of the quadrangle (pl. 2). The northern part of this valley, about half a mile wide and known as the Gallatin Canyon, is followed by the Gallatin River, which flows westward from the park and bends sharply to the north where it enters the valley. By contrast, the southern part of the valley, only about an eighth of a mile wide, is used by Grayling Creek, which also flows westward from the park but turns sharply to the south where it enters the valley. Between these abrupt bends, the valley, about half a mile wide, appears as a low, broad divide choked with glacial debris and occupied by Divide Lake and small inconsequential streams (fig. 16*D*).

The western two-thirds of the quadrangle is underlain by the folded and faulted Paleozoic and Mesozoic rocks which form the lesser mountains of the Madison Range (pl. 1). The eastern third of the quadrangle includes, in the north, the folded sedimentary rocks which form the south tip of the Gallatin Range and, in the south, the well-dissected volcanic rocks which form the west edge of the Yellowstone Plateau. The southernmost sector of the quadrangle includes the north flank of the West Yellowstone basin, which is filled with obsidian sand and gravel and welded rhyolite tuffs.

Topographic relief is about 4,100 feet; the lowest point in the quadrangle is Hebgen Lake, at an altitude (when filled) of about 6,550 feet. The highest point is Sage Peak, altitude 10,664 feet, on the west edge of the quadrangle (pl. 2).

Streams that drain the area discharge into the Gallatin and Madison Rivers, two of the three parent streams of the Missouri River. In the northern half of the area, Sage, Little Sage, Monument, Snowslide, and Bacon Rind Creeks flow northward or eastward to empty into the north-flowing Gallatin River. Fan Creek, also tributary to the Gallatin River, winds southwestward through the uplands in the northwest corner of Yellowstone National Park. In the southern half of the quadrangle, Red Canyon, Little Tepee, Tepee, Grayling, Campanula, Gneiss, Duck, and Cougar Creeks flow southward or westward to empty into Hebgen Lake, a reservoir formed in 1914 when Hebgen Dam was constructed across the westward-flowing Madison River.

In general, the streams in the western part of the quadrangle rise in the Madison Range, which is composed chiefly of sedimentary rocks. By contrast, streams in the eastern part of the quadrangle rise in the volcanic rocks of Yellowstone National Park. This different source was strikingly apparent after the Hebgen Lake earthquake, when normally crystal-clear streams rising in the volcanic rocks became a murky light brown as a result of immense amounts of colloidal particles discharged into them by adjacent springs. Some of these streams remained murky for almost a year; most were clear by the summer of 1960. By contrast, the streams rising in the sedimentary rocks became muddy after the earthquake but cleared in a week or two.

GEOLOGY

TECTONIC FRAMEWORK

The pattern of sedimentation in southwestern Montana was influenced, directly or indirectly, by two major structural elements long since destroyed: the Cordilleran geosyncline, which had its main axis to the west and southwest of the report area, and the Wyoming shelf, which underlay the area and extended to the east and southeast. These features persisted, in whole or in part, through the Paleozoic and part of the Mesozoic.

The Cordilleran geosyncline, extending along the west flank of the North American continent, was well established long before the Cambrian; its site is now occupied by a thick accumulation of Precambrian Belt metasedimentary rocks which extend through parts of Montana, Idaho, Alberta, and British Columbia. In Early Cambrian time this extensive trough was apparently separated into northern and southern halves by an upland known as Montania (Deiss, 1941, p. 1088) which occupied much of Montana, Idaho, and northern Nevada. The absence of Lower Cambrian strata from Montana, Idaho, and Wyoming reflects this highland (Deiss, 1941, p. 1092).

Montania persisted as a landmass of differing sizes and shapes at least through the early Paleozoic. Deiss (1941, p. 1114) suggested that it disappeared below marine waters in the Late Devonian; if so, from Mississippian time on the Cordilleran geosyncline was again a continuous trough extending along the west border of the North American continent. The eastern part of this trough has been classified as a miogeosyncline, the Millard belt, and the western part as a eugeosyncline, the Fraser belt (Kay, 1947, p. 1290).

The axis of the miogeosyncline extended northwestward through central Idaho (Sloss, 1950, p. 425), and much of southwestern Montana was on the east flank. During late Paleozoic time that part of the eugeosyncline in the central Cordilleran area apparently migrated eastward and merged with the miogeosyncline. The resultant seaway lasted either into Early Triassic time (Kummel, 1954, p. 166) or into the late Mesozoic (Kay, 1947, p. 1291), when it was destroyed during Laramide deformation.

The Wyoming shelf, which occupied much of what is now northwestern Wyoming and southern Montana, sloped gently westward and merged with the east flank of the miogeosyncline. Throughout much of the Paleozoic and Mesozoic Eras this relatively stable platform subsided slowly, and a thin veneer of sediments was laid down across it. The sequence, however, is broken by many uncomformities resulting from short episodes of uplift and consequent subaerial erosion of the shelf area.

Within this tectonic framework, the region, of which the Tepee Creek quadrangle is but a small part, probably acted somewhat as a hinge between the deeply subsiding miogeosyncline to the west and the shelf to the east.

STRATIGRAPHY

The basement complex, here divided into six units, consists of Precambrian metamorphic crystallines that are similar to rocks which crop out near Ennis, Mont. (fig. 2), about 60 miles to the northwest, which were called the "Cherry Creek beds" by Peale (1896, p. 2). These crystallines are here simply called pre-Belt metamorphic rocks to avoid existing uncertainty as to their stratigraphic position and age. (See p. 14–15.) The basement complex is overlain by Paleozoic and Mesozoic rocks, 4,000–6,200 feet thick, which have been divided into 19 mappable sedimentary units (table 2 and pl. 1). Of these, four are Cambrian, one is Cambrian and Ordovician, one is Devonian, four are mostly Carboniferous, one is Permian, two each are Triassic and Jurassic, and four are Cretaceous. Locally, these rocks have been intensely folded and faulted.

TABLE 2.—*Generalized section of pre-Tertiary sedimentary rocks exposed in the Tepee Creek quadrangle*

System	Series	Group and formation		Thickness (feet)	Lithology
Cretaceous	Lower Cretaceous	Thermopolis(?) Shale		?	Includes two units. Upper part consists of dark-gray to black thin-bedded fissile soft unfossiliferous shale; includes dark-brown thin-bedded to platy crossbedded sandstone in lower part. Basal sandstone member is light brown, thin bedded to massive, flaggy, crossbedded, fine grained, and quartzose; forms cliffs.
				UNCONFORMITY	
		Kootenai Formation		300–400	Includes three units. Upper unit consists of two yellowish-brown thin limestone beds separated by claystone; lower bed is composed almost wholly of gastropod molds, and upper bed is composed almost wholly of oolites and pisolites in silica cement. Middle unit is variegated unstable claystone; forms landslide blocks. Basal unit consists of light-gray thick-bedded to massive crossbedded conglomerate lenses of well-rounded pebbles of quartz, quartzite, and chert, and sandstone lenses of medium to coarse angular to subround grains of similar materials; forms cliffs.
		UNCONFORMITY			
Jurassic	Upper Jurassic	Morrison Formation		225–400	Claystone, variegated; contains at least two light-brown massive to thick-bedded crossbedded fine-grained friable sandstone lenses.
		Ellis Group	Swift Formation	40–50	Limestone, olive-gray to brown, sandy, oolitic, thin- to medium-bedded, crossbedded; contains many shell fragments and light-brown well-rounded chert grains.
				UNCONFORMITY	
			Rierdon Formation	40–50	Limestone, light-gray; locally alters to calcareous claystone; weathers to gentle slope.
	Middle Jurassic		Sawtooth Formation	200	Claystone, light-gray; contains intercalated light-gray dense nodular thin limestone beds; fossiliferous; weathers as gentle slopes.
		UNCONFORMITY			
Triassic	Lower Triassic	Thaynes(?) Formation and Woodside Siltstone undivided		400–725	Thaynes(?) Formation consists of about 9 ft of grayish-orange thin-bedded calcareous siltstone. Underlain by sequence dominated by Woodside Siltstone, which is moderate reddish brown, thin to medium bedded, and shaly; contains many ripple marks, mud cracks, and other evidence of shallow-water deposition.
		Dinwoody Formation		70–265±	Limestone, light-brown, thin-bedded, very fine grained; calcareous siltstone beds intercalated locally; contains thin greenish-gray shale partings along bedding planes near base.

TABLE 2.—*Generalized section of pre-Tertiary sedimentary rocks exposed in the Tepee Creek quadrangle*—Continued

System	Series	Group and formation	Thickness (feet)	Lithology
Permian		Shedhorn Sandstone	95–175	Includes four units. Upper unit is sandstone identical with second unit. Third unit is chert and interleaved dark-gray shale layers. Second unit is dark-brown to grayish-brown chert-rich thin- to medium-bedded sandstone. Basal dolomite is light brown and thin bedded; intertongues with Quadrant Sandstone.
Pennsylvanian		Quadrant Sandstone	265–625	Sandstone, yellow to light-brown, thin- to medium-bedded, moderately crossbedded, fine-grained.
Pennsylvanian - - - Mississippian	Middle and Lower Pennsylvanian and Upper Missisippian	Amsden Formation	110–160	Siltstone, red, shaly; locally becomes thin-bedded very fine grained sandstone. Two to three intercalated light-gray dense finely crystalline limestone beds.
		UNCONFORMITY		
Mississippian	Upper and Lower Mississippian	Madison Group: Mission Canyon Limestone	1,000–1,450	Limestone, light-brownish-gray, thick-bedded to massive, dense, saccharoidal, unfossiliferous; forms ridges.
		Madison Group: Lodgepole Limestone		Limestone, light-brownish-gray, thin-bedded, dense, cherty, very fossiliferous; forms slopes.
		UNCONFORMITY		
Mississippian - - - Devonian	Lower Mississippian and Upper Devonian.	Three Forks Formation	100–170	Includes four units. Upper unit is unnamed dark gray to black shale 1–2 ft thick. Third unit, Sappington(?) Member, is yellowish-orange to light-brown thin-bedded calcareous siltstone. Second unit, Trident Member, is dark-green to olive shale. Basal, Logan Gulch Member consists of orange to red thin-bedded limestone and calcalcareous siltstone.
Devonian	Upper Devonian	Jefferson Formation	260–320	Dolomite, light-brown to light-olive-gray, thin- to thick-bedded, dense, medium to coarsely saccharoidal; contains chert; locally has petroliferous odor.
		UNCONFORMITY		
Ordovician - - - Cambrian	Upper Ordovician and Upper Cambrian	Bighorn(?) Dolomite Snowy Range Formation, and Pilgrim Limestone undivided	300–350	Includes three units. Upper unit is light-gray thin-bedded dolomite about 35 ft thick and is tentatively assigned to the Bighorn(?) Dolomite. Middle unit, Snowy Range Formation, consists of light-brown even-bedded thin- to medium-bedded mottled dolomitic limestone and contains many mud-pebble conglomerate beds and much glauconite and is about 100 ft thick. Basal, Pilgrim Limestone consists of yellowish-gray even- and thin-bedded mottled dolomitic limestone and is about 195 ft thick.
		UNCONFORMITY(?)		
Cambrian	Middle Cambrian	Park Shale	12–110	Shale, greenish-gray to grayish-red, fissile; weathers to gentle slope.
		Meagher Limestone	460	Limestone, light-gray to brownish-gray, thin-bedded, finely saccharoidal; marked by distinctive mottles of yellow and orange calcareous siltstone in dense brownish-gray limestone matrix.
		Wolsey Shale	125–225	Shale, greenish-gray to gray; locally contains thin beds of light-brown glauconite-rich fine-grained sandstone.
		Flathead Sandstone	75–125	Sandstone, locally conglomeratic, light-brown, thin- to thick-bedded, crossbedded; matrix consists of fine- to coarse-grained sandstone. Pebbles are angular to well-rounded quartzite and quartz.
		UNCONFORMITY		
Precambrian		Pre-Belt metamorphics		

The igneous rocks, of uncertain age, include both intrusive hypabyssal bodies and extrusive pyroclastic flows. The Gallatin River laccolith, exposed near the junction of Snowslide Creek and the Gallatin River, is the largest intrusive. The extrusive rocks, which have been divided into three units, nearly everywhere mantle the eroded Precambrian crystallines and deformed sedimentary rocks, especially in the eastern third of the quadrangle.

Glacial deposits of at least three ice advances mantle much of the area, either as till or as glaciofluvial sand and gravel. These have been subdivided and mapped as five units. Other surficial deposits mapped include talus piles (mapped as colluvium), earthflows, landslides, rockslides, alluvial fans, and alluvium.

PRE-BELT METAMORPHIC ROCKS

The oldest strata exposed are Precambrian metamorphics whose parental rocks probably were widespread sedimentary units such as impure dolomite, limestone, sandstone, and siltstone. The definitive stratigraphic order is uncertain, but the field relations suggest that the sequence, from oldest to youngest, is granite gneiss, dolomite, amphibolite, mica schist, tremolite marble, and quartzite. All are foliate, having structures ranging from the planar fissility of the mica schist to the gross gneissose structure of the granite gneiss. The total thickness may be as much as 30,000 feet.

These rocks alternate irregularly; locally, they change lithology along strike and pinch out. Some variants of the mica schist pass into amphibolite, and in places the dolomite grades into tremolite marble. The dolomite forms lenses that pinch, swell, and locally disappear.

Thin dolomite beds or attenuated quartz laminae are interleaved with beds of schist or gneiss. I interpret this compositional layering to reflect original sedimentary bedding rather than any foliation resulting from metamorphism. Wherever exposed, these rocks generally dip at steep angles, which suggests that they have been intensely folded, possibly more than once.

GRANITE GNEISS

In general, the granite gneiss is a light-gray coarse-grained foliate rock composed mainly of quartz and feldspar with subordinate biotite so alined as to give the rock its gneissose appearance. Rocks comparable to these have elsewhere been called quartz-feldspar gneiss (Heinrich and Rabbitt, 1960, p. 10) and gneiss (Reid, 1957, p. 3).

Commonly the gneiss is thick bedded, almost massive, and in most outcrops the beds average 10 feet in thickness. In a few places the beds are only 1 inch thick and impart a thin-bedded, almost platy, appearance to the rock. The thickest sequence of granite gneiss mapped is about 4,000 feet.

Dominant minerals are quartz, plagioclase feldspar (oligoclase, An_{17}), microcline, biotite, and the following modal analyses (in percent) are fairly representative. (See fig. 10 for sample locations.)

Field No	Wg-64	Wg-108	Wg-139
Quartz	27.7	34.9	27.5
Oligoclase	58.0	49.8	39.6
Microcline	6.3	8.6	24.4
Biotite	5.1	3.6	7.3
Minor accessories	2.8	3.1	1.2
Total	99.9	100.0	100.0

The predominance of plagioclase feldspar over microcline is striking, especially as Heinrich and Rabbitt (1960, p. 10) noted that for the Cherry Creek area, "microcline is greatly in excess of plagioclase, although in a few places microcline is subordinate."

The quartz is unaltered and the microcline nearly unaltered; by contrast, the plagioclase is marked by small patches and seams of sericite. The biotite, altering to chlorite, forms disseminated blebs and thin, discontinuous laminae.

Minor constituents scattered irregularly through the rock include clinozoisite, garnet, apatite, sphene, and ilmenite.

In several localities (sec. 27, T. 11 S., R. 5 E., and secs. 10 and 15, T. 11 S., R. 5 E.) along the west boundary of Yellowstone Park, the granite gneiss underlies the Precambrian dolomite, which in turn seems to underlie, at least locally, most of the other Precambrian crystallines. On the small hill southwest of Rathbone Lake, however, the sequence is reversed, and the granite gneiss seems to be younger than the other metamorphics. Either these beds are overturned or granite gneiss occurs in more than one zone. If the granite gneiss occurs in more than one zone, then the parental rocks may have been several separated sedimentary beds composed of such arenaceous sediments as arkosic sandstone and conglomerate. Reid (1957, p. 20), however, discussing comparable rocks from the Tobacco Root Mountains, suggested that "the gneisses have probably been derived through metamorphic recrystallization of shale with addition of sodium and potassium, to form the abundant feldspar present."

DOLOMITE

The most extensive exposures of dolomite are at the head and along the north valley wall of Bacon Rind Creek and in small knolls of Precambrian rocks exposed west of Grayling Creek. The dolomite locally becomes a tremolite marble, possibly because the parent rocks contained more silica in one place than in another. Along the southeast flank of Tepee Point (NE¼, T. 11 S., R. 5 E.) the contact between the two is vague, and they seem to merge. In the Cherry Creek area comparable rocks have been called marbles, calc-schists, and lime-silicate gneisses (Heinrich and Rabbitt, 1960, p 4).

Commonly the dolomite weathers light brown, although in places it is light gray or dark bluish gray with prominent dark-gray streaks; its color contrasts with the more somber hues of the adjacent metamorphic rocks. In most exposures it is a dense medium-grained crystalline rock that contains large numbers of crenulated quartz veins and stringers. Most of these quartz veins are 1–2 inches thick and 5–6 feet long, although some are much larger. They commonly are bluish gray, and the weathered surfaces stand as small ridges above the mass of the rock. Many have been tightly deformed into isoclinal ptygmatic folds.

In thin section the dolomite is seen to consist of a tight mesh of intergrown xenoblastic dolomite crystals. A few sections show small blebs of calcite and tremolite; accessory minerals include scattered quartz grains and a colorless mica, possibly phlogopite.

The dolomite is crudely bedded, almost massive, although locally it ranges from thin to thick bedded. Characteristically it breaks into large subangular boulders which mantle the underlying slopes. Where the dolomite forms the tops of hills, the residual soil, resulting from the solution of the carbonate, is dark brown and is rich in manganese and iron.

The main mass of the dolomite underlies several square miles and has an estimated minimum thickness of about 600 feet. Locally, however, it forms beds about 6 inches thick in granite gneiss or mica schist.

The general stratigraphic relations are obscure; the dolomite overlies the granite gneiss in sec. 27, T. 11 S., R. 5 E. (near the Yellowstone Park boundary), and, together with the tremolite marble, the mica schist near Tepee Point. Locally, as in sec. 27 west of Grayling Creek, it underlies the amphibolite.

The parental rocks probably were discrete and interleaved beds of dolomite which contained little or no quartz in most places. Where quartz was abundant, the dolomite apparently reacted with the silica to form tremolite marble.

AMPHIBOLITE

Dark-gray foliate rocks, classed as amphibolite, are widespread. They commonly form long, prominent ridges whose flanks are mantled by a scree of angular tabular fragments which range in size from a quarter of an inch to several inches on a side. The amphibolite, rich in hornblende, ranges from dark gray to black; the lighter colors are the result of thin randomly oriented laminae of plagioclase feldspar (oligoclase, An_{25}) and quartz. Locally, myriad veins of these light minerals both follow the lineation and cut across it. In places, exposures of the amphibolite are dark greenish gray, a color which suggests that some chlorite may be present.

Rocks comparable to these have been called hornblende gneiss in the Cherry Creek area (Heinrich and Rabbitt, 1960, p. 8), and amphibolite in the Tobacco Root Mountains (Reid, 1957, p. 5).

The amphibolite is extremely thinly laminated. The planes of foliation commonly are smooth and flat. Here and there the laminated beds form persistent ledges 2–3 feet thick which stand as prominent steps on the outcrop.

In thin section the rock is seen to consist of alternating compositional laminae of bluish-green hornblende and lighter minerals, chiefly plagioclase feldspar (oligoclase, An_{25}) and quartz. The following modal analyses (in percent) are representative. (See fig. 10 for sample locations.) Possibly some alkalic feldspar

Field No	Wg-88	Wg-137	Wg-239
Hornblende	63. 2	65. 9	73. 4
Plagioclase feldspar	31. 5	27. 2	17. 5
Quartz	1. 1	5. 9	6. 0
Minor accessories	4. 2	1. 1	3. 1
Total	100. 0	100. 1	100. 0

is present, but if so it does not exceed 2 percent of the total volume of the rock. Some ilmenite grains (about 2 percent) are scattered irregularly through the matrix, and many are surrounded by a reaction rim of sphene. Other minor accessory minerals include clinozoisite, apatite, sericite (from the alteration of feldspar), leucoxene (from the alteration of ilmenite), and chlorite (from the alteration of hornblende). The hornblende grains are xenoblastic and generally unaltered. The plagioclase grains similarly lack crystal faces; some are thoroughly altered, but most are merely encircled by a thin reaction rim of alteration products. The quartz is unaltered and appears both as discrete grains scattered through the matrix and as mosaic clusters; it appears to be unstrained. Clinozoisite and sphene both commonly show good crystal outlines.

Although the amphibolite commonly occurs as uninterrupted even beds of dark-gray foliate rocks, locally this monotonous sequence is broken by light-gray thin-bedded rocks provisionally identified as quartzite. The quartzites are bimodal and consist of about equal parts of quartz and tremolite. They form beds which range from a few feet to as much as 200 feet in thickness.

These rocks may be similar to the silicified marbles of Runner and Thomas (1928). The presence of tremolite strongly suggests that the parent rock of the quartzite was a dolomitic sandstone and thus strengthens the concept that these pre-Belt metamorphic rocks were derived from sedimentary units.

The relative age of the amphibolite is uncertain. In some places the amphibolite overlies the dolomite and underlies the mica schist in what appears to be conformable contact. No evidence of interfingering was noted between the amphibolite and the mica schist. Elsewhere, however, the amphibolite grades along strike into a crenulated mica schist rich in almandine(?) garnet.

The parental rocks of the amphibolite are uncertain. The mineral assemblage implies that the amphibolite was derived from impure mixed calcareous sediments. The hornblende is twice as common as plagioclase; quartz and biotite, although not common, make up about 5 percent of the volume of the rock; epidote (clinozoisite) is conspicuous; and almandine garnet is absent. Other workers, however, studying comparable rocks in southwestern Montana, have suggested that the amphibolites are derived from mafic and semimafic igneous rocks, chiefly metamorphosed mafic sills and flows (Heinrich, 1953).

MICA SCHIST

In a typical outcrop the mica schist ranges in color from reddish brown to gray and dark greenish gray. Where deeply weathered, as near U.S. Highway 191 mile post 270 (on the Montana-Wyoming border) along the west valley wall of Grayling Creek (pl. 1), it is very dark reddish brown. There the schist is overlain by a pyroclastic flow, and the weathered zone, formed prior to deposition of the volcanic rocks, is as much as 25 feet thick (p. 48 and fig. 12).

The mica schist has a typical crenulate, locally ptygmatic, foliate, banded appearance owing to interleaved veins, stringers, and alined nodules of bluish-gray quartz. The quartz veins, locally as much as 1 foot thick, nearly parallel the foliation for 6–8 feet. The quartz nodules, about 2 feet long and 2–6 inches in diameter, are oval; their long axes are subparallel to the foliation. Similar rocks have been called mica schist in both the Cherry Creek area (Heinrich and Rabbitt, 1960, p. 6) and the Tobacco Root Mountains (Reid, 1957, p. 4, 5).

In thin section, the mica schist is seen to consist dominantly of quartz, biotite, and plagioclase feldspar (oligoclase, An_{28}); accessories such as clinozoisite, chlorite, garnet, tourmaline, and ilmenite are scattered throughout the matrix. The following modal analyses (in percent) are representative. (See fig. 10 for sample locations.) Commonly the biotite grains are elongate

Field No.	Wg-56	Wg-106	Wg-238
Quartz	35.8	43.4	40.4
Plagioclase feldspar	40.2	34.5	16.7
Biotite	22.4	19.2	36.0
Minor accessories	1.6	2.9	6.8
Total	100.0	100.0	99.9

and strikingly alined; most are about 0.52 mm long and about 0.10 mm wide; a few have altered along their edges to chlorite (penninite). The quartz grains are moderately strained, xenoblastic, and unaltered, and they also are alined but much less so than the mica. The plagioclase grains, which also lack crystal faces, are slightly to moderately altered to sericite(?). Both the quartz and the plagioclase grains average 0.15 mm in diameter. The garnet porphyroblasts, about 0.60 mm in diameter, are rounded and seemingly irregularly dispersed through the rock.

Near the mouth of Upper Tepee Basin (N½ sec. 7, T. 11 S., R. 4 E.) several thin bands of black shiny garnetiferous mica schist are interleaved with thin-bedded amphibolite. These schists, which contain many garnet porphyroblasts, are composed chiefly of quartz and biotite. The following modal analysis is representative: Percent quartz=39.6; percent biotite=55.2.

The mica schist is estimated to be about 5,000 feet thick. It overlies the amphibolite and underlies the tremolite marble.

TREMOLITE MARBLE

One of the more unusual Precambrian metamorphic rocks is an extremely resistant medium-bedded to massive tremolite marble which forms prominent topographic ridges. It ranges in color from light gray to gray and is composed almost wholly of radiating needles and interwoven masses of bladed tremolite crystals. These crystals range in width from 1/32 inch to 1 inch; in places they are as much as 5 inches long. Locally the rock is tightly folded.

These rocks have been described as "lime-silicate gneiss of tremolite and calcite with crenulated quartz bands, and bedded marble" (May, 1950, p. 7). In the Cherry Creek area, Heinrich and Rabbitt (1960, p. 5) referred to comparable rocks as tremolite marble, and in the Tobacco Root Mountains, Reid (1957, p. 5) referred to them simply as marble.

In thin section the rocks are seen to consist of tremolite (57.4 percent), quartz (29.7 percent), and calcite (12.8 percent). The tremolite, as elongate crystals about 1.50 mm long and 0.35 mm wide, is moderately to intensely altered to talc(?) and calcite(?). The quartz is unaltered and commonly forms a mosaic of subangular to angular grains about 0.07 mm across. The calcite, as angular grains about 0.24 mm wide, fills interstices.

I estimate that the tremolite marble is about 3,500 feet thick. Locally, it seems to grade imperceptibly into the dolomite. At Tepee Point the tremolite marble apparently overlies mica schist and, together with the dolomite, probably underlies the amphibolite.

The parental rocks may have been impure carbonate rocks rich in silica.

QUARTZITE

Beds of quartzite, which crop out as either discrete stratigraphic units or light-gray lenses in darker metamorphic rocks, may be the youngest pre-Belt rocks exposed. Outcrops are limited; one of the largest is on Horse Butte, where the quartzite apparently overlies granite gneiss and unconformably underlies volcanic rocks of Tertiary age. Other rock units provisionally called quartzite (p. 12) are interbedded with the other metamorphic rocks, chiefly the amphibolite, but most are so thin that their outcrops cannot be mapped separately.

Comparable rocks, also classed as quartzite, have been found in the Cherry Creek area (Heinrich and Rabbitt, 1960, p. 5), in the Tobacco Root Mountains (Reid, 1957, p. 4), and in the Ruby Mountains (Heinrich, 1960, p. 12).

The quartzite weathers dark gray but on fresh surfaces is either light gray or bluish gray. Locally the rock has a very distinctive emerald-green hue which Heinrich (1960, p. 22) suggested is the result of concentrations of slightly chromiferous green muscovite.

In many places the quartzite crops out as short stubby ledges in which the beds range in thickness from 6 inches to 20 feet. In places, these beds are intensely folded and faintly marked by pytgmatic folds.

The quartzite is bimodal and consists chiefly of quartz (80 percent) and muscovite (20 percent). The quartz grains range in size from about 0.02 mm to about 1.40 mm and average 0.18 mm. Most are intensely strained and xenoblastic. They commonly form a tight mosaic which is broken here and there by small muscovite flakes. A few quartz grains are rounded and seem to be relicts of a water-laid sandstone. The muscovite grains, generally about 0.07 mm long and 0.04 mm wide, are alined to give the rock a crudely banded appearance. Minor constituents include tremolite, microcline, calcite, sphene, possibly a little clinozoisite, and some opaque mineral grains (probably ilmenite).

The quartzite is estimated to be about 600 feet thick. Where present in other units, however, individual beds have a maximum thickness of about 20 feet.

CLASSIFICATION OF THE PRE-BELT ROCKS

The pre-Belt metamorphic rocks have undergone at least one episode of regional metamorphism which probably ranged in intensity from moderate to high. They are grouped with the amphibolite facies on the basis of such characteristic rocks as amphibolite, mica schist, and granite gneiss.

The parental rocks of all the metamorphic rocks probably were sedimentary, although some question exists concerning the parental rocks of the amphibolite. The quartzite clearly was derived from an impure sandstone; the dolomite from a dolomite; the tremolite marble from an impure dolomite rich in silica; the mica schist from a mudstone; and the granite gneiss from an arkosic sandstone. The amphibolite, on the basis of its mineral assemblage, may have been derived from impure calcareous sediments, although mafic sills and lava flows could have been the parental rocks (p. 13).

AGE OF THE PRE-BELT ROCKS

The age of the metamorphic rocks in the Madison Valley–Madison Range area is uncertain. Originally, the metamorphics exposed in the Madison Valley and near Virginia City, Mont., were divided into unnamed Archean gneisses and younger rocks called the "Cherry Creek beds" by Peale (1896, p. 2) who described these younger rocks as "a series of marbles, or crystalline limestones, and interlaminated mica-schists, quartzites, and gneisses * * *."

As the result of a geologic study of the Tobacco Root Mountains, north of Virginia City, Tansley, Schafer, and Hart (1933, p. 9) divided the metamorphic rocks of that region into, in ascending order, the Pony Series, composed of rocks of both sedimentary and igneous origin; the Cherry Creek Series (retaining the name proposed by Peale), of sedimentary origin; and the Belt Series, youngest of all and not as intensely metamorphosed as the Pony or Cherry Creek Series.

Heinrich (1953) subsequently reaffirmed the concept that the rocks of the Pony Group are older and underlie the crystallines of the Cherry Creek Group. But Reid (1957), after studying the north end of the Tobacco Root Mountains, part of the area discussed by Tansley, Schafer, and Hart, suggested that the long-accepted stratigraphic sequence was in error and that the Pony metamorphics overlies, and thus is younger than the Cherry Creek metamorphics. He stated (1957, p. 21): "the conclusion seems inescapable that in the Tobacco Root Mountains, the type locality, the Pony metamorphics overlie and are younger than the Cherry Creek metamorphics. However, * * * they have undergone the same metamorphic history. Thus, they are broadly of the same age; * * *."

Since then, Heinrich (1960), as a result of his work in the Ruby Mountains west of Virginia City, suggested that Cherry Creek rocks are underlain by older metamorphic rocks, which he (1960, p. 16) called pre-Cherry Creek rocks. These older metamorphics apparently include the Pony rocks as well as other similar metamorphic crystallines (Scholten, 1955).

In view of these uncertain stratigraphic relations, it would seem wise to forego use of either name until the problem is resolved. The metamorphic rocks in this area, therefore, are classed simply as pre-Belt metamorphic rocks.

PRECAMBRIAN UNCONFORMITY

A distinct angular unconformity marks the contact between the pre-Belt metamorphic rocks and the overlying Flathead Sandstone of Middle Cambrian age, the oldest sedimentary rock exposed. In general, the steeply dipping matamorphic rocks were beveled to a near plane before the deposition of the Flathead. Relief rarely exceeds 10 feet, although locally the uppermost pre-Belt rocks are dissected by small channels, some of which are 20 feet deep and 50–75 feet wide.

By Middle Cambrian time the island of Montania (p. 8) was vastly reduced in size as a result of erosion and moderate subsidence; it probably occupied part of northern Idaho and northwestern Montana (Deiss, 1941, p. 1101–1102). At this time the Wyoming shelf formed an uneven surface which sloped gently westward to the miogeosyncline. Transgressive seas rapidly spread eastward from the miogeosyncline (Hanson, 1952, p. 20), at first depositing coarse clastics (Flathead Sandstone), then siltstone (Wolsey Shale), and finally carbonate (Meagher Limestone), as the waters deepened.

SEDIMENTARY ROCKS

CAMBRIAN

FLATHEAD SANDSTONE

The Flathead Sandstone is a brown thin- to thick-bedded crossbedded quartzose sandstone; it is fine to coarse grained, in part poorly sorted, and locally is conglomeratic. Commonly it forms rounded ledges or broad benches from which younger, softer strata have been removed (fig. 3).

Thin conglomerate lenses, chiefly in the basal part, normally consist of well-rounded pebbles of quartzite and quartz, which range from 1/10 inch to 3 inches in diameter and average ½ inch, in a matrix of fine- to coarse-grained quartzose sandstone. Comparable pebbles are scattered throughout the sandstone. Locally the basal part of the Flathead contains cobbles and angular to subround boulders, as much as 3 feet on a side, of the pre-Belt crystalline rocks.

Scattered through the quartzose sandstone are grains of microcline and plagioclase feldspar and a few thin grains of muscovite. The grains are angular to subround and range in size from 0.08 mm (very fine) to 1.20 mm (very coarse), and average about 0.32 mm (fine). Secondary overgrowths of quartz add to the angularity of the quartz grains. Small drusy vugs lined with quartz crystals occur locally. Silica is the dominant cement—so much so that locally the unit becomes a quartzite.

The upper part of the Flathead Sandstone, which interfingers with the Wolsey Shale, contains many thin, platy beds of very fine grained sandstone which is extremely rich in glauconite. These alternate irregularly with sandstone beds typical of the Flathead.

The Flathead ranges in thickness from 75 to 125 feet; a thickness of about 115 feet is most common.

WOLSEY SHALE

The Wolsey Shale, of Middle Cambrian age, is transitional between the underlying Flathead Sandstone and the overlying Meagher Limestone (fig. 3*A*). As a result, the Wolsey can be divided into a basal part that contains intercalated brown sandstone beds typical of the underlying Flathead Sandstone, a middle part of greenish-gray shaly siltstone and sandstone beds rich in glauconite which typify the Wolsey, and an upper part that contains intercalated gray thin dense crystalline limestone beds characteristic of the overlying Meagher.

Commonly, outcrops of the nonresistant Wolsey form broad benches or gentle grass-covered slopes between the ledges of the underlying Flathead and the cliffs of the overlying Meagher (fig. 3*A*). The Wolsey is well exposed at the mouth of Bacon Rind cirque, and a section measured there is considered typical (p. 87; fig. 3*A*).

The sandstone beds of the basal part of the Wolsey are brown and thin to platy. The grains are moderately well sorted and range in size from 0.04 mm (silt) to 0.40 mm (medium sand); most are about 0.30 mm in diameter. Grains of glauconite are common.

The middle part of the Wolsey consists chiefly of light-greenish-gray shaly siltstone beds that locally are fissile and paper thin and break into angular fragments, ½–1 inch on a side on weathered slopes. In a few places the siltstone beds grade laterally into a dark-greenish-gray thin-bedded fine-grained friable "quartzitic" sandstone in which quartz (61 percent) and glauconite (27 percent) are the dominant constituents. The quartz grains range in shape from angular to subround; the glauconite grains are generally subround to round.

Minor constituents include opaque minerals (chiefly "limonite"), microcline, plagioclase feldspar (oligoclase), and a few grains of muscovite.

The upper part of the Wolsey is formed of interbedded greenish-gray shaly siltstone and gray thin-bedded crystalline limestone (fig. 3*B*). The limestone beds, which contain scattered grains of quartz and glauconite, increase in abundance toward the top of the formation.

The Wolsey ranges in thickness from 125 to 225 feet and most commonly is about 150 feet thick. It thickens northward (Hanson, 1952, pl. 6), although this is not apparent within the report area.

MEAGHER LIMESTONE

The Meagher Limestone, of Middle Cambrian age, is the older of the two carbonate sequences in Cambrian strata. It crops out as a steep slope or cliff between a moderate slope formed on the underlying Wolsey Shale and a broad bench formed on the overlying Park Shale (fig. 3*C*). A section measured in Bacon Rind cirque is considered characteristic (p. 87, fig. 3*B*).

The Meagher consists of thin dense limestone beds which range in color from light gray to brownish gray (about 5*YR* 5/1)[2] and in texture from microcrystalline to finely crystalline. Widespread irregular-shaped bright "gold" mottles of calcareous siltstone, about 1 inch long and half an inch wide, contrast with the gray matrix of the limestone. Most of the mottles are yellow (about 5*Y* 7/2) and orange, although some are red or light brown. The Meagher is sometimes known as the "black and gold marble," a provincial term that was used when the rock was quarried near Radersburg for decorative stone (Hanson, 1952, p. 14).

In most exposures the Meagher is thin bedded; many of the beds are faintly crenulated. The beds range in thickness from about ¼ inch to 3 feet, although most are about ½ inch thick. Commonly they are separated

[2] Numerical designation is from the "Rock Color Chart" (Goddard and others, 1948); however, colors are described as they appeared to me, and the nomenclature may differ from that used in the chart.

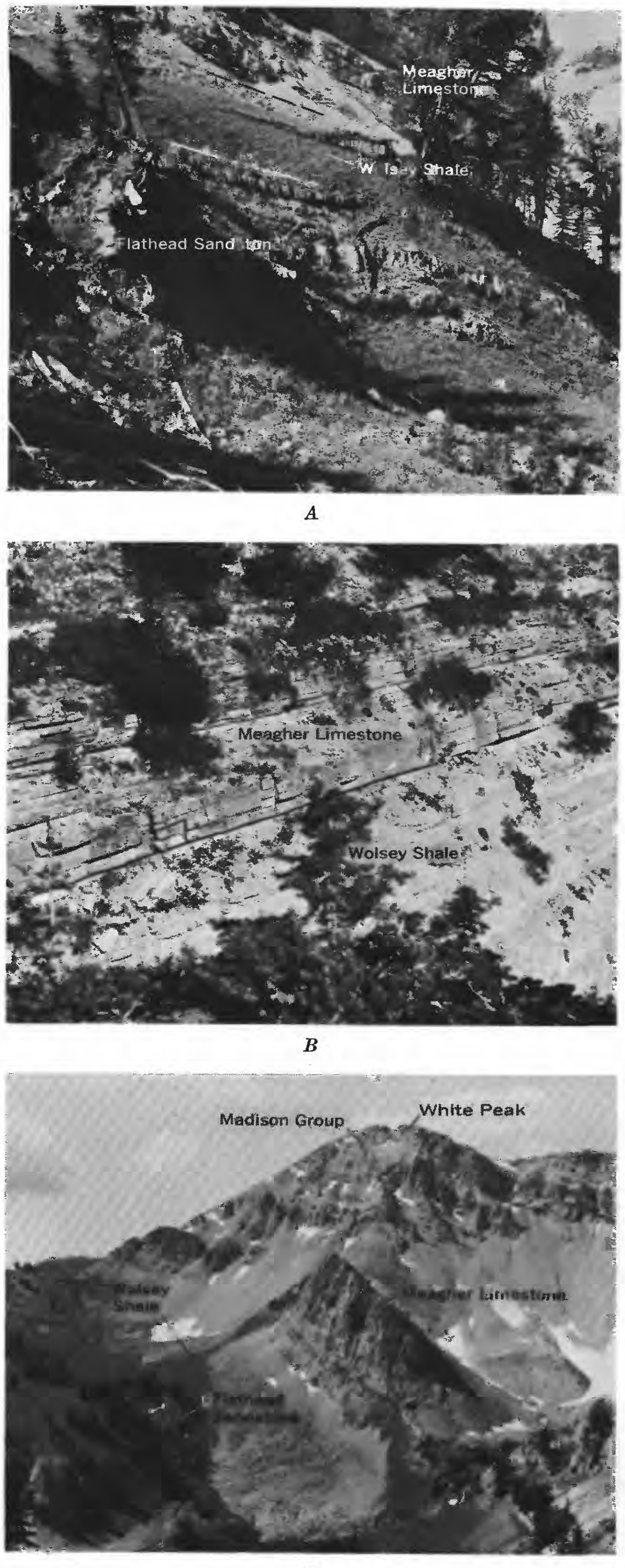


FIGURE 3.—Cambrian strata exposed near Bacon Rind cirque. *A*, The Wolsey Shale forms a moderate grass-covered slope between the steplike ledges formed by the underlying Flathead Sandstone and the cliffs formed by the overlying Meagher Limestone. *B*, Contact between the thin-bedded crystalline limestones of the Meagher and the upper part of the Wolsey Shale, which here consists of interbedded beds of limestone and shaly siltstone. The Wolsey invariably disintegrates into fragments which mantle its slopes. *C*, South flank of Bacon Rind cirque, showing the general mode of outcrop of the Flathead Sandstone, the Wolsey Shale, and the Meagher Limestone. Rockfalls at the base of the Meagher have nearly covered the Wolsey Shale. View is southward.

by paper-thin fissile shale laminae, and in those outcrops from which the shale has been eroded the thin-bedded aspect of the Meagher is accentuated (fig. 3*B*). Where the shale laminae are missing, the bedding planes of the Meagher are vague and indistinct and the unit is thick bedded, almost massive. Locally the Meagher has a distinctive nodular appearance, possibly because the mottles are alined nearly parallel to the bedding plane.

Dolomite is a minor component of the Meagher, as shown by the following chemical analyses[3] (in percent). (See fig. 10 for sample locations.) The average

Lab. No. / Field No.	289640 Wg-a-109	289641 Wg-111	289642 Wg-112
Calcium (Ca)[1]	37. 83	37. 60	37. 62
Magnesium (Mg)[2]	0. 26	0. 25	0. 40
Molal ratios	88. 31	91. 28	57. 07

[1] $CaCO_3$ percent=Ca percent × 2.4973.
[2] $MgCO_3$ percent=Mg percent × 3.4675.

molal ratio of calcium to magnesium is about 80; therefore, the Meagher is a limestone. Calcite seams, which both follow and cut across bedding planes, are widespread.

The contact between the Meagher and the overlying Park Shale is not clearly exposed in the quadrangle. Limited exposures suggest that it is gradational and that the two units intertongue.

The Meagher is about 460 feet thick in this area.

Few fossils were found in the Meagher during this study; those that were found are from the basal part and include the following trilobites and brachiopods identified by A. R. Palmer (written commun., 1962), of the U.S. Geological Survey, and considered by him to be a characteristic Middle Cambrian fauna for southern Montana (fossil-collection locality 415 (3827–CO[4])). (See fig. 10 for fossil-collection localities.)

Trilobites:
- *Bathyuriscus*? *powersi* Walcott
- *Ehmania weedi* Resser
- *Peronopsis bidens* (Meek)

Brachiopods:
- *Homotreta* sp.
- *Pegmatreta*? sp.
- *Micromitra sculptilis* (Meek)

PARK SHALE

The Park Shale, of Middle Cambrian age, is a greenish-gray (about 5*G* 6/1) to grayish red ("purplish," about 5*R* 4/2) even-bedded uniformly fissile noncalcareous shale which commonly forms broad grass-covered benches or strike valleys between the steep slopes formed on both the underlying Meagher and the overlying undifferentiated Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite. As the Park is easily weathered, good exposures are rare, but minute shale fragments 1/8–1/4 inch in size are commonly discernible in the residual soils. In a few places, gray thin dense crystalline limestone beds composed of an irregular mixture of fossil fragments, oolites, and quartz sand grains, all tightly cemented by calcite, are intercalated. These normally break into angular fragments several inches on a side which mantle the lower shale slopes.

The contact with the overlying Pilgrim Limestone is not exposed in the quadrangle; Sloss and Moritz (1951, p. 2145) suggested that a widespread unconformity separates the Park from the Pilgrim in southwestern Montana and that the thickness and distribution of the Park reflects the intensity and extent of this episode of erosion. Although no localities were found where the Park has been completely removed, the Park is only 12 feet thick where exposed in the Bacon Rind cirque (pl. 1). It thickens away from this locality, however, and is about 65 feet thick west of Cone Peak and about 100 feet thick north of that point. This abrupt variation in the thickness of the Park may reflect erosional beveling at the top. Because of poor exposures, the thickness of the formation is uncertain; I estimate it to range from a few feet (in the Bacon Rind cirque) to about 110 feet—averaging 100 feet across most of the quadrangle.

Fossils are not common in the Park in the quadrangle, although the thin intercalated limestone beds contain bits and fragments of indeterminate trilobites and brachiopods. A. R. Palmer noted (written commun., 1962) that one brachiopod from the Park "had the distinctive ornamentation of forms assigned to *Westonia*, a subgenus of *Obolus* that has been found in other collections from the Park Shale."

CAMBRIAN AND ORDOVICIAN

PILGRIM LIMESTONE, SNOWY RANGE FORMATION, AND BIGHORN(?) DOLOMITE UNDIVIDED

The Park Shale is overlain by 300–350 feet of yellowish-gray to pale-yellowish-brown even-bedded thin-to thick-bedded limestone and dolomite which have been mapped as the Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite undivided. Of the total thickness, the lower 195 feet is probably best correlated with the Pilgrim Limestone of Late Cambrian age. The next higher 100 feet probably represents part of the Snowy Range Formation, also of Late Cambrian age. The uppermost 35 feet is provisionally assigned to

[3] Analyst, Dorothy Ferguson, U.S. Geol. Survey. Ca determined volumetrically; Mg, gravimetrically.
[4] Number in parentheses is U.S. Geol. Survey fossil collection number.

the Bighorn Dolomite of Late Ordovician age, although Grant (1965, p. 63) assigned comparable beds exposed on Crowfoot Ridge, 3 miles east of the quadrangle, to the Jefferson Formation of Late Devonian age (p. 19). A regionally significant disconformity (A. R. Palmer, written commun., 1962) unrecognizable in the report area, separates the two Cambrian units throughout the Montana-Wyoming region.

In the undivided sequence there seems to be a gradual increase in dolomite content from oldest to youngest. Thus, the beds in the lower part of the sequence are almost wholly limestone; these are overlain by thin limestone beds, each of which is encircled by a thin rind of dolomite. The highest beds are almost pure dolomite.

PILGRIM LIMESTONE

The Pilgrim Limestone ranges in color on weathered surfaces from yellowish gray (5*Y* 8/1) to light gray (*N* 7); on fresh surfaces it is pale yellowish brown (10*YR* 6/2). Mottles, chiefly dolomitic siltstone in varied shades of yellow and brown, nearly parallel the bedding planes but are not so prominent as those in the Meagher.

The Pilgrim is even bedded and ranges from thin bedded to massive. The thin limestone beds range in thickness from ½ inch to about 8 inches; most are crenulated and have a distinctive nodular or crinkly appearance. They are dense, microcrystalline, and only locally finely saccharoidal. Thicker beds are from 2 to 4 feet thick and more coarsely crystalline.

Edgewise (mud-pebble) conglomerate beds are common; each consists of oblong angular to subangular fragments of light-gray dense microcrystalline limestone in a matrix of light-brown saccharoidal dolomite. The fragments, subparallel to the bedding planes, commonly are about 2 inches long and ¼ inch thick, although some are as much as 7 inches long and 1 inch thick. Locally, limestone fragments have been leached from a bed of edgewise conglomerate, imparting a distinctive vuggy appearance to the bed.

Glauconite, both as discrete grains and as thin seams, is characteristic of these beds. Most of the seams range in thickness from ⅒ inch to about 8 inches and persist laterally for hundreds of feet. Locally, there is an almost regular alternation of glauconite seams and limestone beds. The glauconite seams contain many oolites, and in places they become glauconite-rich oolitic limestone. The quartz grains that form the nuclei of the oolites are angular; by contrast, almost all the glauconite grains are subround to round.

Although the Pilgrim is similar to the Meagher Limestone, it differs in three main respects: it contains many edgewise conglomerate beds; it contains widespread glauconite, both as disseminated grains and as thin seams and lenses; and it is rich in dolomite, as shown in the following chemical analyses [5] (in percent). (See p. 17 for analyses of the Meagher Limestone, and fig. 10 for sample locations.)

Lab. No.	289643	289644
Field No.	Wg-110A	Wg-113
Calcium (Ca) [1]	21.93	17.66
Magnesium (Mg) [2]	12.20	10.04
Molal ratio	1.09	1.07

[1] $CaCO_3$=Ca percent × 2.4973.
[2] $MgCO_3$=Mg percent × 3.4675.

The average molal ratio of calcium to magnesium is about 1.08; therefore the selected beds are dolomite.

The following fossils from the Pilgrim Limestone were identified by A. R. Palmer (written commun., 1962). (Fossil-collection localities are shown in fig. 10.)

Fossil-collection locality 412 (3824–CO):
- Trilobites:
  - *Arapahoia* sp.
  - *Kormagnostus* sp.
- Brachiopod:
  - *Opisthotreta*
- Mollusks:
  - *Hyolithes* sp.
- Sponges:
  - *Chancelloria* sp.
- Conodonts:
  - *Westergardodina* sp.

Fossil-collection locality 391 (3825–CO):
- Trilobites:
  - *Cedarina cordillerae* (Howell and Duncan)
  - "*Homagnostus*" *lochmanae*? Howell and Duncan
  - *Nixonella montanensis* Lochman
  - *Semnocephalus centralis* (Whitfield)
  - *Syspacheilus dunoirensis* (Miller)
- Mollusks:
  - *Hyolithes* sp.
- Sponges:
  - *Chancelloria* sp.

Fossil-collection locality 413 (3826–CO):
- Trilobites:
  - *Bolaspidella* sp.
- Mollusks:
  - *Hyolithes* sp.

These fossils, wholly from the lower 195 feet of the undivided sequence, are all older than middle Dres-

[5] Analyst, Dorothy Ferguson, U.S. Geol. Survey. Ca determined volumetrically; Mg, gravimetrically.

bachian age and are from the *Cedaria* zone (the oldest zone of the Upper Cambrian), which represents the lower beds of the Pilgrim.

SNOWY RANGE FORMATION

Throughout this part of southwestern Montana three lithologic units of Late Cambrian age locally overlie the Pilgrim Limestone: a basal shale unit; a middle unit of interbedded shale, glauconitic limestone, and flat-pebble limestone conglomerate; and an upper unit chiefly of conglomeratic limestone and shale. The lower unit is correlative with Lochman's (1950, p. 2212) Dry Creek Shale Member of the Snowy Range Formation, and the medial member is correlative with her upper Member, the Sage Pebble-Conglomerate Member of the Snowy Range. Commonly the two units are not separated and are mapped together as the Snowy Range Formation (Richards, 1957; Roberts, 1964). The upper unit is equivalent to Dorf and Lochman's (1940) Grove Creek Formation, and many workers have mapped the two formations together as the Grove Creek and Snowy Range Formations undifferentiated, either because the contact between the two is almost impossible to trace in the field (A. E. Roberts, oral commun., 1964) or because the outcrop of either formation is too narrow to show on a published map (Richards, 1957, p. 400).

Grant (1965) suggested, as a result of a study of Late Cambrian invertebrate faunas collected in southwestern Montana and northwestern Wyoming (fig. 1), that the three lithologic units be considered members of the Snowy Range Formation. He proposed that the basal shale be known as the Dry Creek Shale Member, the medial unit as the Sage Member, and the upper unit as the Grove Creek Member. Of these three members, only the lower two are present in this area; of these, the Dry Creek Shale Member is present only locally.

The Dry Creek Shale Member is light-green thin-bedded to fissile shale which forms a moderate slope between the steep cliffs and slopes formed on both the underlying Pilgrim Limestone and the overlying Sage Member. Commonly it is concealed beneath debris shed by the overlying Sage Member, and exposures are rare. It is well exposed, however, on a steep slope at the junction of Migration Creek with Bacon Rind Creek (fig. 4). There the shale is about 30 feet thick, but it thins to 15 feet on Crowfoot Ridge, 5 miles to the east (Grant, 1965, p. 64), and is missing in the Bacon Rind cirque.

The Sage Member is much like the Pilgrim Limestone, and where the Dry Creek Shale Member is absent the two form a steep slope or cliff and are differentiated only with great difficulty. Commonly the Sage Member is light-brown to light-grayish-brown even-bedded thin- to medium-bedded limestone and dolomitic limestone. It contains some edgewise conglomerate beds, which differ from those in the Pilgrim chiefly in containing less glauconite. Many of the limestone beds have a rind of dolomite, and the upper part of the member consists of light-tan to light-gray thin dolomite beds wholly devoid of fossils and glauconite. The beds in the Sage Member range in thickness from ¾ inch to 6 inches, and most are about 3 inches thick. I estimate the Sage Member in this area to be about 100 feet thick; Grant (1965, p. 64) gave its thickness as 65 feet on Crowfoot Ridge.

The Snowy Range fauna characteristically includes the silicified brachiopod *Billingsella* at many localities. The best assemblage of Snowy Range fossils is listed below. Identifications are by A. R. Palmer (written commun., 1961, 1962). (See fig. 10 for fossil-collection locality.)

Fossil-collection locality 390 (3829–CO):
- Trilobites:
  - *Maustonia nasuta* (Hall)
  - *Taenicephalus shumardi* (Hall)
- Brachiopods:
  - *Angulotreta tetonensis*? (Walcott)
  - *Billingsella perfecta* Ulrich and Cooper

BIGHORN (?) DOLOMITE

The uppermost 35 feet of the undivided sequence consists of light-gray thin dense cryptocrystalline dolomite beds devoid of edgewise conglomerates, glauconite grains, or fossils. The age of these strata is uncertain; they are grouped with the Bighorn Dolomite on the basis of their lithology and stratigraphic position.

Grant (1965, p. 64), however, referred to 23 feet of comparable light-gray to yellow thin dolomite beds exposed on Crowfoot Ridge as the "Maywood unit," and assigned these beds to the overlying Jefferson Formation. He stated, referring to the exposure on Crowfoot Ridge (1965, p. 63): "Fossiliferous Snowy Range Formation is capped by high cliff of hard white dolomite. Weed and Deiss assigned dolomite to Jefferson Formation. Basal 23 feet of Jefferson has gross appearance of underlying fossiliferous limestone but is hard unfossiliferous dolomite * * *. Basal 23 feet probably was part of Snowy Range Formation, later dolomitized along with Jefferson Formation; * * * here it so closely resembles Jefferson except in bedding that it is assigned to the Jefferson."

It seems likely that Grant was in error in assigning these beds to the Jefferson. On Antler Peak, about 4 miles east of Crowfoot Ridge, 55 feet of comparable beds in the equivalent stratigraphic position were assigned to the Bighorn Dolomite by Richards and Nieschmidt (1961). Moreover, recent work by Sandberg and McMannis (1964) has indicated that the southern

FIGURE 4.—Cambrian strata exposed along the north valley wall of Bacon Rind Creek near its junction with Migration Creek.

limit of the Maywood Formation is about 30 miles north of Crowfoot Ridge; so most likely the dolomite beds on Crowfoot Ridge are not Maywood.

In the Tepee Creek area these light-gray thin dolomite beds differ markedly from the overlying brown thick to almost massive irregular dolomite beds assigned to the Jefferson. Jefferson strata in this area contain thin seams of light-gray to dark-gray chert, but chert is absent from the underlying dolomite beds. I suggest, therefore, that these thin-bedded dolomites represent the feather edge of the Bighorn Dolomite of Late Ordovician age. C. A. Sandberg (oral commun., 1963) stated that dolomite beds identical in lithology and stratigraphic position with those described here are exposed to the north near the Squaw Creek Ranger Station and that available evidence indicates that the beds in that locality belong in the Bighorn.

The Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite undivided ranges in thickness from 300 to 350 feet, although about 325 feet seems most common.

One and possibly two widespread paraconformities separate these strata, but they are inconspicuous in this area. The first may separate the Pilgrim Limestone from the Snowy Range Formation. If it does extend throughout the Montana-Wyoming region, as suggested by A. R. Palmer (written commun., 1962), it probably represents an episode of erosion at the end of early Late Cambrian (Dresbachian) time, when a broad uplift in central Idaho forced a brief withdrawal of the seas from the general area near southeastern Idaho (Hanson, 1953, p. 19).

The second paraconformity, between Snowy Range–Bighorn(?) strata and the Jefferson Formation, probably represents at least two episodes of erosion. The first episode occurred at the end of Early Ordovician time when Lower Ordovician carbonates and some Upper Cambrian strata were probably removed; the second occurred at some time after the close of the Ordovician but before the Late Devonian. During this second episode most of the Bighorn Dolomite (Late Ordovician) that may have been deposited in this area was removed, and locally all of the Bighorn and Grove Creek and possibly some of the Sage Member of the Snowy Range Formation were removed.

DEVONIAN

JEFFERSON FORMATION

The steep slopes formed on the Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite undivided continue unbroken across the overlying dolomite beds here assigned to the Jefferson Formation of Late Devonian age. In most places these beds are obscured beneath foliage and debris, and good exposures are rare. A characteristic section is along the north wall of Bacon Rind cirque (p. 88 and locality *D* of fig. 10).

Although Sloss and Laird (1947) suggested that the Jefferson, throughout much of central and southwestern Montana, consists of a lower, limestone member and an upper, dolomite member, the two units cannot be distinguished everywhere. Sloss and Laird's proposed division of the Jefferson seems to be valid only locally (Robinson, 1963, p. 28), and in this area the dominant lithology is dolomite; if present, the limestone member is not distinguishable.

On weathered exposures the Jefferson of this area is light brown and light olive gray (about 5*Y* 6/1); fresh exposures are yellowish gray (about 5*Y* 7/4).

In general, the Jefferson is crudely bedded, in sharp contrast to the even, thin dolomite beds which form the upper part (Bighorn(?) Dolomite) of the underlying strata. Locally the Jefferson is thick bedded to massive; here and there, however, it is flaggy and formed of beds 2–6 inches thick. These generally are finely crystalline, almost cryptocrystalline, and in places each bed consists of extremely fine wavy laminae. The thicker beds, generally 5–6 feet thick, are slightly coarser, and a few are coarsely crystalline; they tend to form massive ledges.

A rubbly, faintly vuggy, pitted surface is characteristic. Most of the beds are broken by irregular fractures which have been filled by calcite. In fact, calcite seams and vugs lined with calcite crystals are widespread.

Chert, both as irregularly scattered small angular fragments and as thin seams along bedding planes, is indigenous to the Jefferson but is not in older Paleozoic rocks. This relationship also seems to be true in lower Paleozoic rocks exposed near Ennis, Mont. (J. B. Hadley, oral commun., 1958). The chert seams, commonly light gray or dark gray, range in thickness from 1/4 inch to 1/2 inch and locally thicken to as much as 2 inches. Most are 1–2 feet long, although some form discontinuous lenses as much as 10 feet long.

The Jefferson contains a few limestone beds but is chiefly dolomitic. The following chemical analyses [6] (in percent) are of typical Jefferson strata. (See fig. 10 for sample locations.) The average molal ratio of calcium to magnesium is about 1.20; therefore the Jefferson is dolomite to calcitic dolomite.

Lab. No.	D112766	D112767
Field No.	Wg-450	Wg-453
Calcium (Ca) [1]	13. 40	16. 29
Magnesium (Mg) [2]	6. 59	8. 44
Molal ratio	1. 23	1. 17

[1] $CaCO_3$=Ca percent × 2.4973.
[2] $MgCO_3$=Mg percent × 3.4675.

The fetid odor commonly associated with the Jefferson (Sloss and Moritz, 1951, p. 2151) is faint and elusive.

The upper contact of the Jefferson with the Three Forks Formation is exposed along the north wall of the Bacon Rind cirque (pl. 1); there, the brown and olive-gray thick to massive dolomite beds of the Jefferson appear to grade into the red calcareous shaly siltstone beds of basal Three Forks strata.

As the upper and lower contacts of the Jefferson commonly are concealed, the thickness of the units is problematical; I estimate it to range from about 260 to about 320 feet.

Originally the beds here assigned to the Jefferson Formation were grouped with undivided Cambrian and Devonian strata (Witkind, 1962, p. B6). Since then, corals (listed below) collected from these beds have been examined by W. A. Oliver, of the U.S. Geological Survey (written commun., 1962). Although the fossils are very poorly preserved, Oliver indicates that most "are pre-Mississippian and probably Devonian, and it is probable that all are correctly assigned to the Jefferson." He suggests that their age is probably Middle or early Late Devonian. One or more of the following fossils were found at fossil-collection localities 402, 423, 450, and 453 (fig. 10).

Stromatoporoids:
- *Amphipora*? sp.
- Stromatoporoids, undet.

Corals:
- *Thamnopora* sp.
- *Pachyphyllum* sp.
- *Neostringophyllum* sp.
- Rugose corals, undet.

[6] Analyst, E. J. Fennelly, U.S. Geol. Survey. Ca determined volumetrically; Mg gravimetrically.

DEVONIAN AND MISSISSIPPIAN

THREE FORKS FORMATION

Commonly a gentle slope, partly covered by talus and vegetation, is formed across rocks here assigned to the Three Forks Formation. In most places the nature of the concealed rocks is suggested by thin small chips and fragments of yellow dolomitic and red calcareous siltstone.

Near Three Forks, Mont., 70 miles to the north, the Three Forks Formation can be divided into three units (Robinson, 1963, p. 29): a basal orange thin-bedded limestone and siltstone (Logan Gulch Member, Sandberg, 1965); a medial dark-green and olive shale (Trident Member, Sandberg, 1965); and an upper orange siltstone, shale, and limestone (Sappington Member, Sandberg, 1965). In the Tepee Creek quadrangle, however, only the basal, Logan Gulch Member is widely recognized, although all three units are poorly exposed on the bluff northeast of the junction of Migration Creek with Bacon Rind Creek. Elsewhere, segments of the formation are preserved chiefly in fault blocks. For example, part of the Sappington(?) Member (measured section *F*, p. 89) abuts the Upper Tepee fault in Upper Tepee Basin (secs. 1 and 2, T. 11 S., R. 4 E.) and the White Peak fault in Tepee Basin (SW¼ sec. 7, T. 11 S., R. 5 E.) (pl. 1).

Where the Logan Gulch Member is well exposed it forms a very striking red and yellow band. The basal part of the member is red (about 5*R* 4/4) calcareous shaly siltstone about 35 feet thick. This is overlain conformably by a sequence of yellow (about 10*YR* 7/2) limestone and dolomitic siltstone beds about 95 feet thick. These strata are thin bedded, ranging in thickness from ½ inch to 1 inch, and break into angular chips and fragments. Intercalated in the upper part of these strata are a few thin beds of greenish-gray shale, commonly 2–3 feet thick, and several thin beds of gray finely crystalline dense limestone each 6 inches to 1 foot thick.

The next higher unit, the Trident Member, seems to be missing over most of this area, suggesting either nondeposition (Rau, 1962, p. 58–59), regional erosion (Sandberg, 1965, p. N3), or possibly removal as a result of detachment faults (Witkind, 1962). Where present, the member is represented by a bed of fossiliferous greenish-gray shale and siltstone 2–15 feet thick. This bed is best exposed along a northeast-trending ridge—opposite the Ernest Miller Ridge—on Red Mountain, where it forms a bench between the moderate slopes formed on both the overlying and underlying members of the Three Forks.

The following faunal assemblage, collected from the shale bed of the Trident Member (fossil-collection locality 427(7594–SD) of fig. 10), was identified by J. T. Dutro, Jr., of the U.S. Geological Survey.

Gastropods:
- Platycerated gastropod?, undet.

Brachiopods:
- "*Productus*" *arcuatus* (Hall)
- *Rhipidomella* sp.
- *Syringothyris* sp.

Echinoderms:
- Echinoderm debris, undet.

The Trident Member is overlain by 15 feet of light-grayish-brown (about 10YR 8/2) to light brown thin-bedded calcareous siltstone here tentatively correlated with the Sappington Member. The beds range in thickness from about ½ inch to 1 foot and are marked by extremely fine laminae, so that the member has a somewhat banded appearance. The Sappington(?) consists chiefly of angular to subround quartz grains tightly cemented by calcite. A few grains of sodic plagioclase feldspar (albite-oligoclase) and an opaque mineral, possibly ilmenite, are included. Most of the grains are about 0.07 mm in diameter, although the range seems to be from about 0.02 mm to about 0.1 mm. Rounded specks and blebs of dark-brown limonite ⅛–¼ inch in diameter are disseminated throughout the siltstone, imparting a speckled appearance to it.

The following fossiles were collected and identified by J. T. Dutro, Jr., from that part of the Sappington (?) Member exposed in Upper Tepee Basin along the Upper Tepee fault (p. 75, and measured section F, p. 89; 7595–SD, W½ sec. 1, T. 11 S., R. 4 E.).

Brachiopods:
- Productellid brachiopod
- *Rhipidomella* sp.
- *Syringothyris* sp.

Bryozoa:
- Bryozoan fragments undet.

Echinoderms:
- Echinoderm debris, undet.

Both fossil collections contain Late Devonian Brachiopods, and the host beds are correlated wtih the lower part of the Sappington Member near Logan, Mont. (J. T. Dutro, Jr., written commun., 1966).

The fossil assemblage, ripple marks, and small worm tracks(?) suggest shallow-water depositional conditions.

On the bluff near the junction of Migration Creek with Bacon Rind Creek (pl. 1) the Sappington(?) is overlain, apparently conformably, by dark-gray to black fissile shale 1–2 feet thick. This black shale, exposed throughout southwestern Montana and northwestern Wyoming, has been included in the Three Forks by Sando and Dutro (1960, p. 122) and in the basal Lodgepole Limestone by Sloss and Laird (1947, p. 1411) and McMannis (1962, p. 5, 10).

A comparable black shale exposed near the Little Rocky Mountains was originally called the Little Chief Canyon Member of the Lodgepole by Knechtel, Smedley, and Ross (1954). Sandberg (1963, p. C18), however, believes that the southern limit of the Little Chief Canyon Member is near the Little Rocky Mountains and that it is not continuous with the black shale exposed to the south in southwestern Montana and northern Wyoming. Pending further work, Sandberg (1963, p. C17) suggested that the black shale of southwestern Montana be known informally as "a dark shale unit of Devonian and Mississippian age."

The black shale seems to be absent over much of the Tepee Creek area, and in most places the yellowish-brown Sappington(?) is overlain directly by bluish-gray fossiliferous limestone of basal Lodgepole Limestone. The contact is clear and definite; no interfingering was noted. The absence of the shale may be due to any one of a series of causes, but on the basis of exposures I suggest that an unconformity exists in this area between the Three Forks and the overlying Madison Group.

The Three Forks Formation ranges in thickness from 100 to 170 feet. A section measured in the Bacon Rind cirque is considered characteristic (p. 89 and locality *E* on fig. 10).

In several widely separated localities—along Red Canyon Creek, Bacon Rind Creek near its mouth, and the north flank of Red Mountain—the top of the Three Forks has been brecciated and consists of boulders and fragments of silicified siltstone. The thickness of the breccia is uncertain because of poor exposures, but the breccia may include both the upper part of the Three Forks and basal Lodgepole strata. The breccia may be a solution phenomenon, or it may be a result of faulting (p. 28).

MISSISSIPPIAN

MADISON GROUP

A thick sequence of thin, thick, and massive limestone beds constitutes the Madison Group of Mississippian age. The lower two-thirds of the group, here correlated with the Lodgepole Limestone, is thin to medium bedded, locally massive, and forms moderately steep slopes. By contrast, the upper third of the group, here correlated with the Mission Canyon Limestone, consists of thick to massive beds which normally stand as steep cliffs or imposing blocky ledges (fig. 5). Because of inadequate exposures, the two formations have not been mapped separately, and on plate 1 the Madison Group is shown as a single, undivided unit.

LODGEPOLE LIMESTONE

The Lodgepole Limestone is mainly light brownish gray (about 5*YR* 6/1) to medium brownish gray (5*YR* 3/1) but is in part pale red (5*R* 6/2) and light brown (about 10*YR* 8/2). It is characterized chiefly by its thin even beds, its interleaved thin chert beds, and its rich fossil content.

Near the base of the Lodgepole, individual beds range in thickness from ½ inch to about 3 inches, near the middle they are 3–5 feet thick, and near the top they are as much as 40 feet thick; those near the top are, however, faintly divided by imperfect bedding planes into units about 8 feet thick. Thin gray shale seams separate the thin limestone beds, and where the shale has weathered out, the beds form thin parallel ledges. Commonly the thin-bedded lower part of the Lodgepole is fine grained; only a few beds are medium or coarse grained. As the beds thicken higher in the section the grain size coarsens, and the upper part of the Lodgepole sequence is dominated by both coarse-grained beds and beds formed almost wholly of fossil fragments.

Beds of brownish-gray chert ½–3 inches thick are common in the lower part of the Lodgepole (mainly in the basal 155 ft), but chert, either as seams or as fragments, is sparse to absent in the upper part of the Lodgepole as well as in the overlying Mission Canyon.

Two units in the Lodgepole are distinctive but are not persistent throughout the area. The first, about 150 feet above the base, is a pale-red limestone with some intercalated beds of dolomitic limestone and is 50–60 feet thick. Most of the strata in this unit generally range in thickness from ¼ inch to 4 inches, and these become extremely rich in shell fragments toward the top of the unit. The second distinctive unit is an intraformational conglomerate about 200 feet above the base of the Lodgepole. The conglomerate, ranging in thickness from several inches to several feet, consists of angular tabular fragments of finely crystalline light-gray limestone in a silty matrix. Most of the fragments are about ½ inch thick and 2–3 inches long.

The Lodgepole, for the most part, is so fossiliferous that in places the beds are a hash of fossil fragments—chiefly brachiopod, coral, and crinoid remains. Of these, brachiopods and corals are the most important fossils used in stratigraphic correlation (Sando and Dutro, 1960, p. 119). Well-formed discrete brachiopod and coral fossils are plentiful in the lower 150 feet of the Lodgepole but become less plentiful higher in the section where many of the beds are composed of a fossil hash. The basal 30 feet of the Lodgepole is marked by a profusion of silicified crinoidal fragments which weather to form minute protuberances.

FIGURE 5.—Bacon Rind cirque viewed from the southeast. Strata exposed in the cirque range from Precambrian crystallines (not shown) to the thin-bedded sandstones of the Quadrant and form part of the nose and southwest limb of the Skyline anticline.

Some indication of the profusion of fossil remains in the Lodgepole is given by Table 3, which lists the invertebrates collected by J. T. Dutro, Jr., when he measured a section of the Madison Group along the southeast flank of White Peak (measured section *G*, fig. 10 and p. 89). The fossils were identified by Dutro

TABLE 3.—*Invertebrate fossils collected from Madison Group strata exposed along the north flank of White Peak (NE¼ sec. 2, T. 11 S., R. 4 E.)*

[Fossils collected by J. T. Dutro, Jr.; identified by Dutro and W. J. Sando (corals). Collections are arranged with lowest on left and highest on right]

Taxonomy	USGS Upper Paleozoic locality (-PC) 20162	20163	20164	20165	20166	20167	20168	20169	20170	20171	20172	20173	20174	20175	20176	20177	20178	20179	20180	20181	20182	20183	20184	20185	20186	20187	20188	20189	20190	20191	20192	20193	20194	20195	20196	20197	20198	20199	20326	20327
	Zone B			Zone C																																		Zone D?	
Echinoderms:																																								
pelmatozoan debris, indet	X	X	X						X	X		X	X	X	X		X		X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X		X
Cryptoblastus cf. *C. pisum* (Meek & Worthen)		X																																						
Cryptoblastus sp			X																																					
Platycrinites sp. (columnals)									?	X					X						X	X																		
actinocrinitid columnals, indet															X			X																						
Sponges:																																								
hexactinellid spicules, indet		X	X																																					
Corals:																																								
Aulopora sp		X	X						X																															
Syringopora surcularia Girty										aff																								X					aff	
Syringopora sp			X																	X								?												X
"*Roemeria*" sp											X																													
Cleistopora sp									X																															
Michelinia sp												?													X															
Amplexus sp		X	X																																					
Cyathaxonia sp			X																																					
Rylstonia sp	?								X	?																														
Homalophyllites sp									X	X		X								X													X	X		X	X	X	:	
Vesiculophyllum sp					?				X	X		?												X				X				X	X	X				X		
Zaphrentites sp	X	X							X				X			X	X							?																?
horn coral, indet																													X											X
horn coral, indet		X		X				X																				X		X							X			
Lithostrotionella sp																																						X		
Bryozoans:																																								
fenestrate bryozoans, indet	X	X	X						X													X	X	X									X	X						
ramose bryozoans, indet			X						X													X																		
Cystodictya sp									X				X	X	X							X	X										X							
Brachiopods:																																								
Rhipidomella sp												X	X				X				X																			
Leptagonia analoga (Phillips)																										X														
Schellwienella sp									X												X				X	X	?		?	X	?									
Schuchertella sp													X	X																										
orthotetid fragments, indet													X	X					X		X		X					X								X				
"*Camarotoechia*" sp			X															X	X		X		X																	
rhynchonellid fragments, indet									X																						X									
Reticulonetes logani (Norwood & Pratten)	?		X						X						X																									
Rugosochonetes loganensis (Hall & Whitfield)													X	X	X							X	X		X	X	X				X					?				
Ovatia cf. *O. laevicosta* (White)						X																			?															
Rhytiophora? sp																									X	X					X									
productoid fragments, indet																	X																			X				
Unispirifer aff. *U. albertensis*									X				X																											
Unispirifer sp									X			X	X	X	X	X							X			X		X	X		X					X				
"*Spirifer*" sp. (*centronatus*-type)	?	X	X		X													?	?																					
"*Spirifer*" aff. "*S.*" *biplicoides* (Hall)														X	?	?		?			?	?	?	?																
Spirifer rowleyi Weller		aff																																						
Imbrexia sp																X	?	?	X						X										?					
Brachythyrina? sp																									X															
Anthracospirifer? sp																									X										?					
Nucleospira cf. *N. obesa* Rowley													X	X	X	?																								
Martinia? sp									X					X	X			X	X																					
Composita humilis Girty									X																															
Composita sp																X			X																X	X				
Cleiothyridina cf. *C. tenuilineata* (Rowley)		X	X						X			X	X	X		X	X				X																			
Cleiothyridina cf. *C. obmaxima* (McChesney)							X		?																															
Torynifer cf. *T. cooperensis* (Swallow)																		X																						
Punctospirifer solidirostris (White)													X	X	X	X	X	X	X							cf														
Punctospirifer sp		X	X															X							X			?												
Cyrtina cf. *C. burlingtonensis* Rowley		X	X									?	?																											
Eumetria sp		?														?					?									X										
Beecheria sp									?								?																							
brachiopod fragments, indet			X		X					X										X							X					X	X	X						
Gastropods:																																								
Platyceras sp		X	X										X	X		X	X	X	X		X																			
Straparollus sp																												X		X										
Trilobites:																																								
proetid fragments, indet		X	X																																					
Worms and trace fossils:																																								
spirorbid tubes, indet		X	X																																					
Taonurus sp																															X									
trails and burrows, indet																															X									

and his colleague W. J. Sando who stated the following concerning the age, correlation, and coral zonation[7] of the fossils (written commun., 1966):

The fossil assemblages compare well with those from other measured sequences in southwestern Montana and northwestern Wyoming. All appear to be of Early Mississippian age; no definite Late Mississippian forms were identified. Correlation with the zones established by Sando and Dutro (1960) is as follows: Zone B includes collections 20162–PC through 20164–PC and perhaps extends as high as 20169–PC; Zone C includes the remaining collections in the section starting with 20170–PC, with the possible exception of the higher two. No definite Zone A assemblages were collected, but this is not unusual. Zone A, where it has been recognized, usually occurs in the lowest few feet of the Lodgepole. No definite Zone D was identified, although collections 20326 and 20327 may represent this zone.

A subdivision of Zone C is difficult because of a lack of distinctive C_2 species. It generally falls at about the Lodgepole–Mission Canyon boundary, however, and there is a suggestion in the fossil distribution that the base of Zone C_2 could be placed just below collection 20193–PC.

In terms of the Mississippi Valley sequence, the Madison in this section is Kinderhook and Osage in age.

It seems likely that measured section G (p. 89), from which these fossils came, is not a complete sequence of the Madison Group, although it was once so considered (p. 75). Part of the sequence may have been removed by one or more bedding plane faults; so although all zones seem to be present, they may not exhibit their full thicknesses. The positions of the collections in the measured sequence suggest that Zone A is about 5 feet thick and that Zone B at least 52 and quite likely as much as 104 feet thick. The thickness of Zone C is unknown, for at least one bedding-plane fault is suspected (p. 92, measured section *G*). The strata here assigned to Zone C total 512 feet. Zone D may be as much as 317 feet thick.

MISSION CANYON LIMESTONE

The generally thin beds of the Lodgepole give way gradually to the thicker and more massive beds of the Mission Canyon. These beds are comparable in color to those of the underlying Lodgepole and differ from them chiefly in three major aspects: they are thicker, lack chert, and are much less fossiliferous.

The Mission Canyon consists of coarse-grained limestone. Commonly it forms rimrocks, bluffs, and ridges (fig. 8) and weathers to a rough, angular, blocky surface. The beds range in thickness from 15 to 50 feet and are separated by thin lenses of greenish-gray shale. The bedding planes are not persistent and die out along strike, only to begin again at a slightly higher or lower horizon a short distance away. Overall, the formation appears as a poorly bedded cliff former (fig. 8).

One of the most distinctive units in the Mission Canyon is a pale-red coarse limestone breccia 15–60 feet thick which is about 100 feet below the top of the formation. The breccia, traceable for miles as a continuous unit, consists of a chaotic mass of angular to subangular fragments of light-gray limestone in a matrix of pale-red to red siltstone. Most of the fragments range in size from about ½ inch to 5 inches. Locally, the fragments have been cemented to form discrete cobbles as much as 8 inches in diameter. Much larger material is also included; scattered through the breccia are angular fragments and boulders of limestone which are as much as 2 feet on a side. Also included are fragments of angular very fine grained sandstone and siltstone.

Limestone breccias are widespread throughout western and southwestern Montana; breccias much like those exposed in this area occur in the Gravelly Range (J. B. Hadley, oral commun., 1958), in Milligan Canyon in the Three Forks area (Robinson, 1963, p. 42), in the Bridger Range (McMannis, 1955, p. 1400–1401), in the Livingston area (Roberts, 1961), and in the Elkhorn Mountains (Klepper and others, 1957, p. 19–20). Comparable breccias occur also in central Montana (Nordquist, 1953) and in western Wyoming (Sando and Dutro, 1960).

The general appearance and the continuity of the breccia led Thom, Hall, Wegemann, and Moulton (1935, p. 35) to suggest that it represents "collapse features and cave fillings related to an extensive development of karst topography on the top of the Madison." A similar interpretation was suggested by Klepper, Weeks, and Ruppel (1957, p. 20), who explained the breccia as the result of extensive ground-water solution while a pre-Amsden water table maintained a constant position beneath an erosion surface.

Strickland (1956, p. 51), however, suggested that the breccia represents an intraformational disconformity. But Norton (1956, p. 54), reviewing much the same evidence as Strickland, thought that the persistent breccia is a collapse feature rather than an unconformity. Sando and Dutro (1960, p. 119) also were not convinced that a major unconformity is present in the Madison.

[7] As part of a comprehensive study of Mississippian rocks in the western interior of the United States undertaken by J. T. Dutro, Jr., and W. J. Sando, of the U.S. Geological Survey, the corals in the Madison Group have been divided into five zones which are remarkably persistent throughout much of southwestern Wyoming, northeastern Utah, and western Wyoming (Sando and Dutro, 1960). The lowest coral zone, Zone A, includes the basal 10–50 feet of the Lodgepole. Zone B, generally poorly fossiliferous, is also in the Lodgepole; although Sando and Dutro failed to cite any specific interval occupied by this zone, they implied in their diagrams that it is restricted to be the lower part of the Lodgepole. Zone C contains an abundant fauna of both solitary and colonial corals, and Sando and Dutro arbitrarily divided it into two parts, which are, in ascending order, Zones C–1 and C–2. The boundary between these two zones was arbitrarily placed at the top of the Lodgepole Limestone. Zone D, the highest coral zone in the Madison Group, includes the uppermost Mission Canyon Limestone and, locally, later Mississippian carbonates.

Another interpretation of considerable merit offered by various authors suggests that the breccia developed from the solution of evaporites and the resultant collapse of the bedded carbonates. Nordquist (1953, p. 80) noted that the subsurface Charles Formation consists of carbonates interbedded with evaporites, chiefly anhydrite and gypsum. He pointed out that where these rocks are exposed in the mountains of central Montana they are characterized by brecciated zones, the result of the leaching of the evaporites and the resultant collapse of the carbonates. In south-central Montana three distinct solution breccias were apparently derived from the solution of three evaporite zones (Andrichuk, 1955, p. 2179–2180); similarly in western Wyoming, the upper part of the Mission Canyon contains an evaporitic sequence in which solution breccias are common (Sando and Dutro, 1960, p. 124). Roberts (1961; 1966) also suggested that the laterally persistent breccias in the Mission Canyon of the Livingston area are the result of leaching of evaporites to form solution breccias.

The age of the breccia in the Tepee Creek area is unknown, for no faunal evidence exists. If the breccia formed much as suggested by Klepper, Weeks, and Ruppel (1957, p. 20), its age is restricted to the rather broad post-Mission Canyon pre-Amsden time interval. But in both the Three Forks (Robinson, 1963, p. 44) and the Bridger Range areas (McMannis, 1955, p. 1401), Big Snowy rocks overlie the Madison, and this relation implies that the breccia formed during the post-Mission Canyon pre-Big Snowy time interval. The amount of time available for the breccia to have formed, thus, was short. Severson (1952, p. 39–45), however, suggested that these solution breccias may have been formed much earlier, possibly in post-Colorado (Late Cretaceous) time. He based this belief on the observation reported to him by O. T. Hayward that near the Smith River in central Montana large areas of Colorado Shale of Cretaceous age are in depressions or sinks which presumably formed over major solution zones. He concluded from this that the solution breccias probably formed either during periods of mountain building in that area or during the period of erosion which followed.

In addition to this solution breccia, which is near the top of the Madison, two breccias occur locally near the base of the formation. The lower one seems to involve the uppermost beds of the Three Forks Formation as well as basal Madison strata (p. 23). The thickness of this basal breccia is unknown, for its debris mantles the lower contact; I estimate the breccia to be about 100 feet thick. The second breccia is 300–400 feet above the basal breccia and is confined wholly to Madison strata; it may also be about 100 feet thick.

In the southern part of the quadrangle, as along the west fork of Red Canyon and in adjacent tributary valleys (secs. 26 and 27, T. 11 S., R. 4 E.; fig. 6), the two breccias consist chiefly of angular to rounded fragments and boulders of limestone in a matrix of small limestone fragments firmly cemeted by carbonate (fig. 7). By contrast, in the northern part of the area the two breccias are silicified; the limestone fragments have been thoroughly impregnated by silica and now appear as gray crystocrystalline fragments tightly held in a reddish-brown siliceous matrix.

The silicified breccias crop out along both valley walls near the mouth of Bacon Rind Creek; there they form thick, prominent, jutting ledges nearly masked by dense forest cover. They also crop out along the west valley wall of the Gallatin River between Bacon Rind and Snowslide Creeks and along the east flank of the hill directly west of Divide Lake. A small patch of the basal silicified breccia is preserved on the crest of Red Mountain.

In both breccias, cobbles are widespread (fig. 6*A*) and boulders as much as 2 feet on a side are common (fig. 6*B*). The cobbles and boulders all contain the typical fossil hash so characteristic of the Madison. Here and there the basal breccia contains many chert fragments as much as 3 feet long and 2–3 inches thick. These apparently are disjointed remnants of once-continuous chert beds. The basal siliceous breccia yielded the following silicified forms identified by W. J. Sando of the U.S. Geological Survey (fossil-collection locality 387 (21403–PC) of fig. 10).

Corals:
- *Amplexus*? sp.
- *Zaphrentites*? sp.

Bryozoans:
- *Fenestella* sp., undet.
- Rhomboporoid bryozoan, undet.

Brachiopods:
- Brachiopods, undet.

Gastropods:
- Gastropods, undet.

Sando (written commun., 1963) reported that although this faunal assemblage does not contain fossils diagnostic of the Madison, it does seem to represent a Mississippian faunule.

The two siliceous breccias end abruptly against the near-vertical Bacon Rind fault (p. 79), which cuts the rocks along the north valley wall of Bacon Rind Creek near the mouth of Migration Creek (pl. 1). On the west side of the fault, which trends about N. 25° W., is a normal stratigraphic sequence—the Madison (lacking the basal breccia) rests on a thin layer of black shale which in turn overlies the yellowish-brown calcareous

A B

FIGURE 6.—The basal breccia of the Madison Group in the west fork of Red Canyon. *A*, Steep embankment formed by the basal breccia. Note the faint trace of bedding planes which here dip northward (toward left side of photograph). *B*, Subangular to subround limestone boulders in the basal breccia.

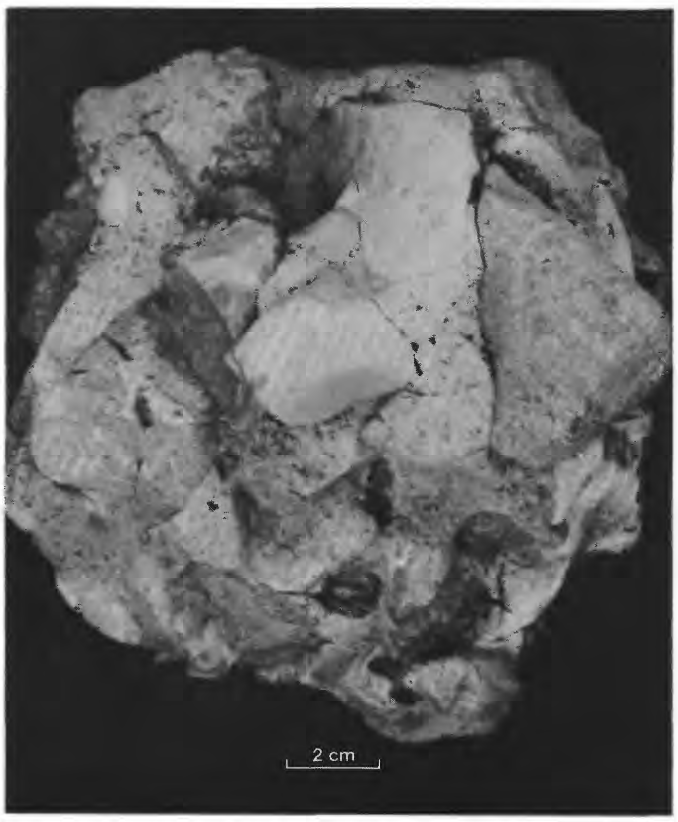


FIGURE 7.—Boulder of the basal breccia of the Madison Group, showing the degree of rounding of the enclosed limestone fragments. Photograph by E. P. Krier, U.S. Geological Survey.

siltstone beds of the Sappington(?) Member of the Three Forks Formation. East of the fault, however, the two breccias are well developed and can be traced eastward as continuous units.

Although no localities are known where the silicified breccias pass into the unsilicified ones, the two types of breccia occupy similar stratigraphic positions and seem to differ only in cementation. The reason for this difference is not known; it may be significant, however, that the silicified breccias lie closer to the pyroclastic flows which cover the eastern part of the quadrangle. Possibly when these flows were erupted either in very late Tertiary or in early Pleistocene time (p. 53) silica-rich waters entered the brecciated zones of the Madison and deposited the silica cement. The Bacon Rind fault could have prevented westward migration of these waters.

The breccias may be solution phenomena comparable to the breccia near the top of the Mission Canyon, or they may be tectonic breccias and reflect bedding-plane or detachment faults (p. 76). Some fossil evidence is available which suggests that, locally at least, part of the Madison Group may have been duplicated as a result of several detachment faults. This evidence is discussed in a later part of this report (p. 75).

The thickness of the Madison Group ranges from about 1,000 to about 1,450 feet, and this considerable variation is most puzzling. The group is about 1,000 feet thick along the southeast flank of White Peak (NE¼ sec. 2, T. 11 S., R. 4 E.), and a comparable thickness has been measured about 2 miles to the southeast along the north flank of peak 7298 (NW¼ sec. 18, T. 11 S., R. 5 E.). In both localities the top and bottom of the group are well exposed, and, as far as can be determined in the field, the beds are not broken by faults, although a bedding-plane fault is suspected in at least one place (measured section *G*, p. 92). Both localities are on the overturned Skyline anticline, and the strata either dip very steeply, are vertical, or are overturned. Hence, the true stratigraphic thickness may be masked in these areas by some form of structural deformation (p. 76). Elsewhere in the area, however, the Madison is about 1,450 feet thick, and this thickness fits well the regional isopach pattern of the Madison Group (A. E. Roberts, oral commun., 1965). The Madison is about 1,335 feet thick to the north in the Taylors Peak area (Hall, 1961, p. 215).

The Madison is overlain unconformably by the Amsden Formation; the contact between the two is remarkably even and free of any large irregularities. Locally, a 5-foot-thick red breccia cemented by calcite is at the base of the Amsden, and this unit probably formed during a period of erosion prior to deposition of the red siltstone beds of the Amsden (p. 81). It is probably correlative with the karst breccia features exposed near Livington, Mont. (Roberts, 1966, p. B7).

MISSISSIPPIAN AND PENNSYLVANIAN

AMSDEN FORMATION

Although the Amsden Formation is one of the less resistant units exposed in the quadrangle, it crops out chiefly along ridge crests and mountain tops. It occupies this prominent topographic position mainly because it is protected by two firmer units, the Madison Group below and the Quadrant Sandstone above (fig. 8). Between these two, the Amsden is preserved as an elongate vivid-red band. Redstreak Peak, one of the highest peaks in the mapped area, takes its name from the red shaly siltstone beds of the Amsden exposed on its crest (fig. 5).

The Amsden consists chiefly of thin, even siltstone beds which range in color from medium reddish brown (10*YR* 4/4) to dark red (about 10*R* 3/4). A few beds are very firmly cemented by calcite. Throughout most of the area, this siltstone sequence is interrupted by two, in places three, light-gray (about 5*Y* 7/1) to pinkish-gray (5*YR* 8/1) dense finely crystalline limestone beds which range in thickness from 10 to 30 feet. These beds are formed of thin to medium limestone layers, most of which are about 4 feet thick. Here and there, the surface of the limestone is covered by pale-red (5*R* 5/2) irregular-shaped mottles. Locally the limestone beds contain fragments, blebs, and stringers of reddish-orange chert; many of the fossils from these limestone beds, chiefly solitary cup corals, are composed of this chert.

A persistent light-brown (about 5*YR* 8/2) fine-grained quartzose sandstone about 3 feet thick occurs 10–25 feet above the base of the Amsden. This unit may be correlative with the Darwin Sandstone Member of the Amsden of western Wyoming (Blackwelder, 1918, p. 422–423).

Near the top of the Amsden the siltstone grades into a light-brown fine-grained sandstone which is transitional with the overlying Quadrant Sandstone. In a few places, a limestone bed about 6 inches thick is at the top of the Amsden. The following fossils from this bed (fossil-collection localities 71 and 72 (19839–PC) of fig. 10), chiefly brachiopods, suggest a Pennsylvanian (Atoka) age for this part of the Amsden (J. T. Dutro, Jr., written commun., 1961).

Brachiopods:
- *Antiquatonia*? aff. *A*.? *coloradoensis* (Girty)
- *Anthracospirifer* cf. *A. occidentalis* (Girty)
- *Composita* cf. *C. subtilita* (Hall)

The following fossils have been found in a limestone bed about 80 feet above the base of the Amsden (fossil-collection locality 290 (20156–PC) of fig. 10).

Foraminifers:
- Textularians (simple forms)
- *Earlandia* sp. (very minute species)
- *Tuberitina* sp.
- *Endothyra discoidea*? Girty, 1915 (common)
- *E*. aff. *E. phryssa* (D. N. Zeller), 1953 (abundant)
- *Paramillerella* or "*Endothyra*" of Zeller, not Phillips, 1846 (rare)
- *Nankinella* sp. (common)
- *Brunsia* sp.

Corals:
- cf. *Zaphrentites* sp.
- clisiophyllid coral, genus undet.

Bryozoan:
- *Tabulipora*? sp.

Brachiopods:
- *Cleiothyrindina* sp.
- brachiopod, undet.

Gastropod:
- Bellerophontacean gastropod, undet.

A late Mississippian or possibly Early Pennsylvanian age is indicated for this part of the Amsden by these fossils (L. G. Henbest and W. J. Sando, written commun., 1966).

The Amsden is about 160 feet thick throughout most of the area, although locally it thins, possibly as a result of structural attenuation, to about 110 feet. A section was measured on the southeast flank of peak 9830 (p. 93 and locality *H* of fig. 10).

319-733 O—69——3

FIGURE 8.—Peak 9830 from the southeast. The soft vivid-red shaly siltstone beds of the Amsden are preserved between the yellow sandstone beds of the overlying Quadrant Sandstone and the brownish-gray thick to massive limestone beds of the underlying Mission Canyon Limestone of the Madison Group.

PENNSYLVANIAN

QUADRANT SANDSTONE

The Quadrant, which forms many of the high ridges throughout the area (fig. 8), consists of bedded yellow (about 5*Y* 7/4) to light-brown (about *N*7) well-sorted quartzose sandstone which is generally firmly cemented by silica. It contrasts strikingly with the underlying red shaly siltstone beds of the Amsden. Locally a few thin beds of yellowish-gray (about 5*Y* 6/2) dolomite occur about 50 feet above the base of the Quadrant.

On weathered surfaces the Quadrant normally is dull gray or, in places, medium reddish brown. In most outcrops it is even bedded (fig. 5); the beds range in thickness from 2 to 12 feet. Locally it is almost massive. It is moderately to well crossbedded with long sweeping tangential laminae.

Grain size ranges from about 0.02 mm to about 0.20 mm; most grains are about 0.10 mm. Many of the grains are coated by secondary quartz which both cements and adds angularity to once subround grains. The dominant cement is silica, but in places it is ineffective and the sandstone is extremely friable, so much so that the disaggregated grains form large areas of loose, unconsolidated sand. Elsewhere, however, the silica cements the Quadrant into a hard, resistant quartzite.

The Quadrant is conformable with, and grades into, the overlying Shedhorn Sandstone; the upper sandstone beds of the Quadrant pass smoothly into light-brown sandy dolomite.

The Quadrant ranges in thickness from 265 to 315 feet. In some places, however, the formation is much thicker (maximum observed, 625 ft), probably as a result of deformation (p. 74).

No fossils were found in the Quadrant, but it is lithologically equivalent to the Pennsylvanian Tensleep Sandstone of northwestern Wyoming (Sloss and Moritz, 1951, p. 2164).

PERMIAN

SHEDHORN SANDSTONE

For many years, the term "Phosphoria Formation" was used to designate all the rocks of Permian age in southwestern Montana. In 1956, however, McKelvey and others (1956), as the result of a broad regional study of this rock system, proposed a new plan of nomenclature, based chiefly on lithology, for the rocks studied. In essence, the name "Phosphoria" was retained and used for the black shale, phosphorite, and bedded chert exposures in southeastern Idaho. The well-established name "Park City" was restricted to a carbonate facies which rims the black shale–phosphorite sequence, and the name "Shedhorn Sandstone" was coined (Cressman and Swanson, 1956, p. 2852; 1959, p. 31) for the Permian rocks, chiefly sandstone beds, exposed in and near Yellowstone National Park. Accordingly, this name is used in this report.

In most places the Shedhorn is a poorly exposed unit of dark-gray chert-rich sandstone which contrasts markedly with the yellow rocks of the underlying Quadrant Sandstone and the dusky-brown rocks of the overlying Dinwoody Formation. Where well exposed on Skyline Ridge the Shedhorn can be divided into four parts. The lowest part is a thin-bedded sandy dolomite about 40 feet thick which contains many angular and rounded chert fragments. This unit, tentatively correlated with the Grandeur Tongue of the Park City Formation, is overlain conformably by a 25-foot-thick dark-gray to dark-brown fine-grained sandstone, also rich in chert, which probably represents the lower member of the Shedhorn Sandstone. The next higher unit, about 40 feet thick, consists of many thin beds of gray chert and thin interleaved layers of dark-gray fissile shale; each chert bed ranges in thickness from 2 to 4 inches. These bedded cherts are thought to be equivalent to the Tosi Chert Member of the Phosphoria Formation. The uppermost unit is a sandstone identical with the lower member of the Shedhorn and is about 60 feet thick.

The dolomite beds are light brown, slightly sandy and 1½–3 inches thick. The basal beds, which contain the most sand, seem to intertongue with the underlying Quadrant Sandstone, for a few thin dolomite beds occur in the uppermost thick sandstone bed of the Quadrant. Scattered through the dolomite are many light-gray angular to well-rounded fragments of chert, most of which are about 2 inches in diameter. A few irregular-shaped masses of chert about 1 foot long and 2–3 inches thick are crudely alined with the bedding planes.

The sandstone beds, which form the bulk of the Shedhorn, are dark brown (about $5Y$ 6/2) to grayish brown (about $5Y$ 6/1), thin to medium bedded, fine grained, and rich in chert nodules. Although the sandstone is in beds generally only a few inches thick, in places individual beds are as much as 3 feet thick; in many exposures the bedding is accentuated by subparallel chert masses. The sandstones are well sorted; most of the grains are about 0.15 mm in size (fine), although they range from 0.04 mm to about 0.30 mm. The grains range in shape from angular to subround; much of the angularity is due to secondary overgrowths of silica. Quartz and collophane are the major constituents, and quartz is about three times as plentiful as collophane. A few grains of glauconite and some shell and bone fragments are scattered widely through the sandstone. In general, the sandstone is moderately well cemented by silica; locally it is a quartzite.

Chert, as white or light-gray (about $5GY$ 8/1) irregular-shaped nodules and light-gray ribbed cylindrical rods, is abundant in the sandstone members. The nodules, about 3 inches thick and 1 foot long, nearly parallel the bedding. Commonly they occur in dense clusters, and the sandstone appears almost like a bedded chert. The cylindrical rods consist of sandstone and chert, which form a structureless mixture in some rods and concentric rings in others. Generally the rods are perpendicular, or nearly so, to the bedding. The rods are about 2 inches in diameter and range in length from 6 inches to about 2 feet. They have ribbed sides and weather out as light-gray short, somewhat distorted, cylinders or angular light-gray fragments that are easily recognizable in the float.

The thin-bedded cherts, provisionally correlated with the Tosi Chert Member, are light gray to gray and even bedded and weather out as thin angular plates. The beds, which range in thickness from 2 to 4 inches, are separated by thin layers of dark-gray fissile shale.

Each of the four members of the Shedhorn seems to be conformable with, and to grade into, the adjacent members.

As far as can be determined from limited exposures, the Shedhorn is conformably overlain by the Dinwoody Formation. Although comparable relations have been noted elsewhere in southwestern Montana (Kummel, 1954, p. 168; Hadley, 1960, p. 149; Sloss and Moritz, 1951, p. 2167), Cressman and Swanson (1960, p. 231) suggested that a slight disconformity beveled the

Permian rocks throughout a wide area, including the Madison and Centennial Ranges.

The Shedhorn is about 160 feet thick along Skyline Ridge and about 175 feet thick on Kirkwood Ridge. It thins to about 95 feet along the northwest end of Johnson Ridge in Red Canyon, but, because of poor exposures, it is not known whether this is due to the pinchout of one or more of its members or to deformation.

A section measured along the southwest flank of White Peak includes all four members (p. 93, and locality *I* of fig. 10).

No identifiable fossils were found in Shedhorn exposures; the unit, however, was considered to be Permian in age by Cressman and Swanson (1956, p. 2852).

TRIASSIC

DINWOODY FORMATION

The steep slopes formed on the sandstone and chert members of the Shedhorn give way to moderate slopes and elongate strike valleys characteristic of the Dinwoody Formation of Early Triassic age.

Moritz (1951, p. 1788) recognized a twofold division of the Dinwoody throughout southwestern Montana: a lower unit, chiefly shale, which he called the shale member, and an upper unit, chiefly limestone and siltstone, which he called the limestone member. I am uncertain whether both units are present in this area, for exposures of the Dinwoody commonly are poor. Very likely, most Dinwoody exposures are part of the limestone member, although the lowest part of the exposed Dinwoody consists of interbedded calcareous shaly siltstone and dark-gray fissile shale and may be part of the shale member. Both members are thick to the west (in Beaverhead County, Mont.), but both thin rapidly northeastward; in Taylors Fork, Mont., some 7 miles to the northwest, both members are almost pinched out (Moritz, 1951, p. 1788, fig. 6).

In this area Dinwoody strata are light brown (about 5*Y* 7/4) to yellowish gray (about 5*Y* 6/2) and contrast with the somber brownish-gray strata of the underlying Shedhorn and the bright reddish brown strata of the overlying Woodside Siltstone. Near the base of the Dinwoody, thin layers of grayish-green shale intertongue with light-brown thin calcareous siltstone beds. Many of the beds are only a quarter of an inch thick, and each consists of parallel fine laminae. The upper part of the formation consists of light-brown to yellowish-gray calcareous siltstone beds 2–4 feet thick which locally crop out as bluffs or long, partly wooded dip slopes. In places, the siltstone coarsens to form a calcareous sandstone which stands as a rib above the softer subjacent rocks. Ripple marks and other indications of shallow-water deposition are widespread.

The calcareous siltstone beds are composed of angular to subangular quartz grains which are tightly cemented by calcite. The quartz grains range in size from 0.02 mm to about 0.18 mm; most are about 0.08 mm in diameter. Small dark-gray to black manganese dendrites are widespread and seem to be characteristic of the Dinwoody.

The Dinwoody is conformably overlain by, and grades transitionally into, the Woodside.

The Dinwoody ranges in thickness from about 70 to 265 feet. It is thickest on Kirkwood Ridge and thins northeastward to about 100 feet on Little Sage Creek and only 70 feet on Monument Creek.

The following fossils (fossil-collection localities 405, 438, 443 of fig. 10) were provisionally identified by N. J. Silberling of the U.S. Geological Survey (written commun., 1962), who reported: "As a whole, this fauna is like that described from the Dinwoody Formation by Newell and Kummel (1942), and from correlative rocks in Greenland by Spath (1930; 1935)."

Pelecypods:
- *Anodontophora*? sp.
- ?*Claraia* cf. *C. stachei* Bittner
- *Myalina*? sp.
- *Pleurophorus*? sp.

Most of the fossils are poorly preserved and could not be positively identified. In addition to the pelecypod fossils, incomplete specimens of the phosphatic brachiopod *Lingula* were also collected from outcrops near the base of the unit.

The Dinwoody was assigned to the Early Triassic (Scythian Stage) by Newell and Kummel (1942, p. 950–951).

WOODSIDE SILTSTONE AND THAYNES(?) FORMATION UNDIVIDED

Lower Triassic rocks younger than the Dinwoody Formation were mapped as a single unit. The Woodside Siltstone, which ranges in thickness from about 400 to about 725 feet, constitutes the bulk of the sequence. It is overlain by about 9 feet of grayish-orange thin-bedded calcareous siltstone which is tentatively correlated with the Thaynes Formation (fig. 9).

Woodside Siltstone

The Woodside Siltstone, one of the most striking formations in the area, is very well exposed near the head of Red Canyon, where it forms a steep slope in a tributary wash (NW¼ sec. 24, T. 11 S., R. 4 E.) (fig. 9, and pl. 1).

FIGURE 9.—Triassic and Jurassic strata exposed near the head of Red Canyon Creek. About 9 feet of grayish-orange thin-bedded calcareous siltstone, here tentatively correlated with the Thaynes Formation, separates the Woodside Siltstone from the Sawtooth Formation of the Ellis Group. Note the gradational change from the Woodside to the Thaynes(?) and the abrupt lithologic break between the Thaynes(?) and the Sawtooth.

The Woodside is moderate reddish brown (about 10*R* 5/5), and its exposures commonly are visible for miles as either reddish soil or barren outcrops. Its color continuity is broken here and there by a few thin light-yellowish-gray (5*Y* 5/2) beds or by light-gray mottles. These lighter colored units and mottles probably are the result of local reduction of ferric iron (Fe^{+3}) to ferrous iron (Fe^{+2}).

The Woodside is thin bedded (fig. 9), almost fissile; a few beds are as much as 4 feet thick, and these stand as small steplike ledges in the long slopes formed on the softer siltstone beds. Ripple marks, mud cracks, and other indications of shallow-water deposition are widespread.

The Woodside is composed of well-sorted angular quartz grains which range in size from 0.01 mm to 0.05 mm; most are about 0.03 mm in diameter. A thin film of iron oxide coats each grain, so that the formation has a reddish-brown appearance. Calcite and iron oxide are the dominant cements, but they are somewhat ineffective, for the formation on the whole is friable.

The section measured in the wash tributary to Red Canyon Creek is characteristic (p. 93, and locality *J* of fig. 10).

The uppermost beds of the Woodside are transitional with, and grade into, the overlying grayish-orange calcareous siltstone beds of the Thaynes (?) (fig. 9).

The formation is about 725 feet thick in Red Canyon

Creek, but it thins rapidly northward to about 500 feet near Redstreak Peak and about 400 feet on Monument Creek near the Black Butte Ranch (once known as the Dean Ranch), about 2 miles north of the north boundary of the report area (May, 1950, p. 20). Moritz (1951, fig. 6) reported the Woodside Formation to be about 250 feet thick in Taylors Fork about 7 miles northwest of this area, and Hall (1961, p. 55) stated that the Woodside is missing from the northern part of the Upper Gallatin Valley area.

No fossils were found in exposures of the Woodside. Newell and Kummel (1941, p. 205) indicated that diagnostic ceratites and pelecypods date the Woodside (and Dinwoody) as Early Triassic.

Thaynes(?) Formation

Conformably overlying the reddish-brown shaly siltstone beds of the Woodside is about 9 feet of grayish-orange (about 10*YR* 7/4) thin calcareous siltstone with interleaved lenses of grayish-green shale (fig. 9). In Red Canyon these strata form a ledge between the steep slopes of the underlying Woodside and the overlying Sawtooth Formation. To the north, on Skyline Ridge, they are either concealed or absent.

No fossils were found in either the calcareous siltstone beds or the shales, and the age and assignment of these beds is uncertain. On the basis of their stratigraphic position and transitional relations with the Woodside, they may be the thin northeast edge of the Thaynes Formation.

The Thaynes (?) is unconformably overlain by the claystone-limestone sequence of the Sawtooth Formation of Middle Jurassic age.

JURASSIC

ELLIS GROUP

In this area the Ellis Group consists of, in ascending order, the Sawtooth Formation, the Rierdon Formation, and the Swift Formation. The three formations total about 280 feet in thickness; for cartographic reasons they are shown on plate 1 as a single undivided unit, the Ellis Group of Middle and Late Jurassic age.

Sawtooth Formation

The Sawtooth Formation consists of light-gray (*N*7) to moderate-yellowish-brown (10*YR* 5/2) calcareous claystone beds interbedded with light-gray (*N*7) thin dense nodular limestone beds several inches to 2 feet thick. From a distance the unit appears almost white, in contrast with the reddish brown of the underlying Woodside (fig. 9). The Sawtooth commonly forms gentle grass-covered slopes or strike valleys. Near the head of Red Canyon Creek (NW¼ sec. 24, T. 11 S., R. 4 E.) the evenness of the slopes is broken here and there by small steplike ledges formed by the intercalated thin limestone beds. A few of the limestone beds, especially near the top of the Sawtooth contain abundant oolites.

In general, the Sawtooth is somewhat massive, and most of the claystone beds range in thickness from 6 to 15 feet. The claystone is structureless and breaks into angular fragments 1–2 inches on a side which mantle lower slopes.

The Sawtooth unconformably overlies the Thaynes(?) Formation and is conformably overlain by the Rierdon Formation.

The Sawtooth is about 200 feet thick in this area. Regionally, it increases in thickness eastward and decreases northward (Schmitt, 1953, fig. 7; Imlay and others, 1948).

Perhaps the most characteristic feature of the Sawtooth is its abundant fossil content; the claystone and limestone both yield abundant invertebrate remains. The following assemblage (fossil-collection localities 406 (28670), 488 (28800) of fig. 10) was identified by R. W. Imlay (written communs., 1962, 1963), who noted that several of the forms are identical with fossils found in member C of the Twin Creek Limestone of southeastern Idaho and western Wyoming. Member C is considered to be of Middle Jurassic (Bathonian) age (Imlay, 1950).

- Echinoderm:
  - Echinoid, undet.
- Pelecypods:
  - *Astarte meeki* Stanton
  - *Camptonectes platessiformis* White
  - *Gervillia* cf. *G. dolobrata* Crickmay
  - *Gervillia montanaensis* Meek
  - *Grammatodon haguei* (Meek)
  - *Gryphaea planoconvexa* Whitfield
  - *Idonearca? haguei* (Meek)
  - *Lima* (*Plagiostoma*) *occidentalis* Hall and Whitfield
  - *Mactromya?* sp.
  - *Modiolus subimbricatus* Meek
  - *Myophorella montanaensis* (Meek)
  - *Ostrea strigilecula* White
  - *Pholadomya inaequiplicata* Stanton
  - *Pinna kingi* Meek
  - *Pleuromya subcompressa* (Meek)
  - *Pronoella cinnabarensis* (Stanton)
  - *Trigonia elegantissima* Stanton
  - *Vaugonia conradi* (Meek and Hayden)
- Ammonoids:
  - Ammonite, n. gen., n. sp. (much like *Zemistephanus* sp.)
  - Ammonite, n. gen., n. sp.

Rierdon Formation

Conformably overlying the Sawtooth is 40–50 feet of strata which probably correlate with the Rierdon Formation, although this has not been confirmed by fossil evidence. The strata consist mainly of thin limestone beds which alternate irregularly with fissile calcareous shale beds. The basal bed is yellowish-gray (5*Y* 7/2) dense extremely oolitic limestone which thickens and thins erratically but is generally 2–6 feet thick. Although fossiliferous, it is much less so than limestone beds in the underlying Sawtooth. Isolated star-shaped columnal plates of the crinoid *Pentacrinus* are fairly common in this bed. Overlying the basal limestone is light-gray thin-bedded fissile calcareous shale which in places contains an intercalated light-gray thin oolitic limestone.

The contact of the Rierdon with the overlying Swift Formation is sharp and distinct, which suggests that the erosional unconformity of wide extent noted by Cobban (1945, p. 1290, 1291) to the north extends throughout southwestern Montana. The smooth slopes formed on the Rierdon end abruptly at the base of a cliff about 15 feet high formed by the Swift Formation.

The following fossils, collected from the basal part of the Rierdon (fossil-collection locality 488 (28800) of fig. 10) were identified by R. W. Imlay (written commun., 1963), who noted that comparable forms in the Twin Creek Limestone range from member C to member F. These members bracket the boundary between Middle and Late Jurassic; members C and D are assigned to the Bathonian Stage of the Middle Jurassic, and members E and F are assigned to the Callovian Stage of Late Jurassic age (Imly, 1950).

Pelecypods:
- *Lima* (*Plagiostoma*) *occidentalis* Hall and Whitfield
- *Pronoella*? cf. *P. cinnabarensis* (Stanton)
- *Trigonia elegantissima* Stanton
- *Vaugonia conradi* (Meek and Hayden)

Swift Formation

The uppermost unit of the Ellis Group is the Swift Formation, a yellowish-gray (5*Y* 7/2) to brownish-gray (5*YR* 5/1) thin-bedded to platy crossbedded oolitic sandy limestone. It is resistant and normally forms ledges or cap rocks which protect the weaker, underlying Rierdon. Most of the beds are 3–6 inches thick, and the bedding planes die out along strike. Interbedded grayish-green thin claystone beds commonly weather out, accentuating the platy appearance of the unit. Locally the claystone beds thicken and become the dominant lithology. Near the top the formation contains a calcareous sandstone much like the overlying sandstone beds of the Morrison described below.

The sandy limestone beds consist of a heterogeneous mixture of oolites, chitinous shell fragments, angular quartz grains, rounded glauconite grains, and light-brown well-rounded chert grains, all tightly cemented by calcite. In places light-brown chert pebbles ⅜–¾ inch in diameter are scattered throughout the matrix. The oolites range in diameter from 0.30 mm to 0.70 mm; most are about 0.50 mm. Some oolites have angular quartz grains as nuclei.

The contact between the Swift and the overlying Morrison Formation is indeterminate in most places because of poor exposures. For mapping purposes it has been placed between glauconitic sandstone below and nonglauconitic sandstone above.

In this area the Swift has a minimum thickness of about 40 feet. Isopach maps of the Swift indicate that it thickens toward the east and northeast (Schmitt, 1953, fig. 12); in adjacent areas to the north, Hall (1961, p. 59) reported the formation to be about 90 feet thick.

The Late Jurassic (Oxfordian) age of the Swift Formation has been well documented (Imlay, 1950; Schmitt, 1953; Moritz, 1951).

MORRISON FORMATION

The Morrison Formation of Late Jurassic age is a nonresistant unit of structureless claystone and friable sandstone which forms gentle slopes and elongate strike valleys between the ledges developed on both the underlying Swift Formation and the overlying basal conglomeratic sandstones of the Kootenai Formation.

Claystone, the dominant lithology, is normally concealed beneath a thin soil whose color, either grayish red (5*R* 2/2) or light green (about 5*Y* 6/1), reflects the color of the underlying unit. The rock is structureless, breaks into angular fragments ¼–½ inch on a side, and is especially susceptible to mass-wasting where exposed on steep slopes. Near the sandstone beds the claystone grades imperceptibly to a shaly siltstone marked by thin parallel laminae.

The sandstone is light brown (about 10*YR* 7/6) and in most places forms two beds, each 10–25 feet thick, which are interleaved in the claystone sequence. I suspect, however, that these sandstone beds are not continuous and pinch out even as other sandstone beds appear both higher and lower in the section. In places, the more resistant sandstone beds form minor ledges which persist for many hundreds of yards before they become platy, break down, and are inconspicuous. In general, the thicker sandstone beds comprise a number of beds 1–2 inches thick that locally thicken to 1–2 feet.

The sandstone is composed of angular to subround quartz and chert grains which range in size from 0.02 mm to 0.20 mm; most are about 0.10 mm. A few quartz grains have secondary overgrowths, but in general the grains are poorly cemented by silica, and the sandstone is friable. In places calcite acts as weak cement. Although the Morrison sandstone is somewhat similar to that of the underlying Swift, it contains no glauconite grains.

In most places where the contact between the Morrison and the overlying Kootenai is exposed, the uppermost bed of the Morrison is claystone. In a stream floor east of Coal Canyon (center of sec. 25, T. 11 S., R. 4 E.), however, a small isolated exposure of probable Morrison strata is capped by a thin-bedded to platy carbonaceous seam 2–3 feet thick. This seam may be correlative with the well-known and extensive coal bed which forms the top of the Morrison throughout central Montana (Fisher, 1909; Lammers, 1939; Brown, 1946; Vine, 1956). Although Hall (1961) apparently saw no such seams in Morrison strata exposed to the north, Condit (1919, p. 115) noted some 49 feet of "Shale and clay with carbonaceous layers" beneath the conglomeratic sandstone of the Kootenai in a section of Mesozoic and Paleozoic rocks measured in Indian Creek, Mont., some 15 miles to the northwest. The absence of the carbonaceous shale from much of the Tepee Creek quadrangle may be the result of pre-Kootenai erosion.

A widespread and significant unconformity between the Morrison and Kootenai Formations has been recognized throughout southwestern Alberta, Montana, and Wyoming (Lammers, 1939). Although the evidence favoring an unconformity between Morrison and Kootenai strata in the report area is meager and inconclusive, I suggest that one may separate the formations.

Because of poor exposures, the upper and lower contacts of the Morrison are difficult to determine in many places. The thickness is estimated to range from 225 to 400 feet, but the lesser figure may represent foreshortening as a result of structural deformation. A section of the Morrison was measured along the south flank of Skyline Ridge (p. 94 and locality *K*, fig. 10).

No diagnostic fossils were found in the Morrison, although fragments of silicified wood were found in float below Morrison outcrops.

CRETACEOUS

KOOTENAI FORMATION

The Kootenai Formation of Early Cretaceous age is widespread and is known by different names in adjacent areas; it is the Blairmore of southwestern Alberta, and the Cloverly of Wyoming. In this quadrangle the Kootenai is 300–400 feet thick and can be divided into three parts: a basal, light-gray conglomerate and conglomeratic sandstone; a middle, variegated claystone with minor amounts of sandstone; and an upper, limestone and claystone unit.

Probably the Kootenai's most characteristic and persistent feature is its basal, conglomeratic sandstone, the Pryor Conglomerate Member (Roberts, 19?5), which is distinctive in color and lithology. In this area it is about 85 feet thick and is light gray (about *N*6), thick bedded to massive, crossbedded, and medium to coarse grained, and contains many conglomerate lenses. Commonly, the basal part is entirely conglomerate, and this grades upward through conglomeratic sandstone with interbedded conglomerate lenses to a medium- to coarse-grained sandstone.

The conglomerate lenses consist of well-rounded pebbles of dark-gray to black chert and colorless to white quartz and quartzite in a medium- to coarse-grained matrix of similar materials. The pebbles, generally oval, range in size from about 1/4 inch to about 3 inches; most are about 1/2 inch long and about 1/4 inch wide. All nearly parallel the bedding planes.

The sandstone in the basal unit consists of a heterogeneous mixture of angular to subangular chert grains and angular to round quartz grains which range in size from 0.55 mm to 0.60 mm; most are about 0.35 mm thick. The black chert and white quartz impart a "salt-and-pepper" appearance; hence, the term "salt-and-pepper sandstone." Much of the angularity of the quartz seems to be due to secondary overgrowths of quartz. Generally the sandstone is soft and friable, but locally it is well cemented and stands as a low cliff. Silica is the dominant cement.

The middle unit of the Kootenai, about 200 feet thick, consists chiefly of grayish-red (5*R* 4/2) to greenish-gray (about 10*Y* 6/2) claystone and a few beds of light-brown fine-grained sandstone. This sequence commonly forms broad gentle slopes; on steeper slopes, blocks of it have slid downslope to mantle underlying units.

The upper unit of the Kootenai contains two thin ledge-forming limestone beds separated by claystone. The lower limestone, about 13 feet below the top of the Kootenai, is 2 feet thick, light gray (*N*6), dense, and hard, and is composed almost wholly of coiled freshwater gastropod molds. The upper limestone, here considered as the uppermost unit of the Kootenai, is about 6 feet thick, yellowish brown (10*YR* 6/2), dense, and hard, and is composed almost wholly of oolites and pisolites which are firmly cemented by secondary silica. These concretions range in size from 0.20 mm to 2.40 mm and average 0.80 mm; here and there they combine to form compound oolites as much as 3.50 mm in diameter. The intervening claystone, 5 feet thick, is yellowish gray (5*Y* 7/2) and similar lithologically to the massive claystone in the middle part of the formation.

The broad slopes formed on the middle and upper parts of the Kootenai end abruptly at the base of a cliff about 40 feet high composed of brown fine- to medium-grained sandstone, here tentatively correlated with the basal, sandstone member of the Thermopolis(?) Shale of Early Cretaceous age. The two units appear to be unconformable.

No diagnostic fossils were found, but elsewhere in Montana a fossil florule collected from equivalent strata indicates that the Kootenai is Early Cretaceous in age (Brown, 1946, p. 247). The Early Cretaceous age is also substantiated by fresh-water mollusks (Yen, 1951, p. 2) and by nonmarine microfossils (Peck, 1941, p. 287).

THERMOPOLIS(?) SHALE

The youngest consolidated sedimentary rocks are tentatively correlated with the Thermopolis Shale of Early Cretaceous age. Two lithologic units are recognizable: a basal, sandstone, here called the sandstone member which commonly stands as either a steep cliff (fig. 24) or an imposing hogback; and a dark-gray nonresistant fissile shale, here called the upper part of the Thermopolis(?) Shale (table 4). A sandstone bed of variable thickness is intercalated in the lower part of the shale sequence and has been grouped with the upper part of the Thermopolis(?) for cartographic purposes (table 4). These strata, preserved in the center of the Cabin Creek syncline, are exposed chiefly along the flanks, gullies, and draws of Cabin Creek, and there exposures are sparse, as the strata are largely concealed by surficial deposits (pl. 1).

Sandstone member

The sandstone member of the Thermopolis(?) is light-brown (about 10*YR* 5/4) thin-bedded to massive flaggy crossbedded fine-grained quartzose sandstone and seemingly disconformably overlies the oolitic limestone bed selected as the top of the Kootenai Formation. The sandstone is resistant and forms prominent ledges, cap rocks, and hogbacks along the east and north flanks of the valley of Cabin Creek. A few small hillocks on the floor of Cabin Creek are capped by the sandstone. In the northwest corner of the quadrangle, along the west valley wall of Sage Creek, the sandstone underlies volcanic rocks.

The sandstone is characterized by its crossbedding and its light-brown color; crossbedding has been found in all exposures of the unit, and only locally does the color darken to medium brownish gray (5*YR* 5/1). Although the unit appears massive when viewed from a distance, it is composed of thin to thick beds which locally are so firmly cemented that bedding planes are vague. The basal part is hard and quartzitic and normally stands as a sheer cliff. Higher beds are softer and alternate irregularly with shaly siltsone beds. As a result, the massive cliff passes upward into moderate slopes, which in turn become gentle where the very fine grained sandstone beds grade into the basal shaly siltstone beds of the overlying upper part of the Thermopolis(?).

TABLE 4.—*Suggested correlations of the Thermopolis(?) Shale*

Area of this report	Upper Gallatin Valley (Hall, 1961)	Livingston Area (Roberts, 1965)	Northwestern and western Wyoming (Haun and Barlow, 1962)	Mainly northern Wyoming (Eicher, 1960)	Northern Bighorn Basin (Moberly, 1960)	Central Wyoming (Keefer and Troyer, 1964)	Central and northeastern Wyoming (Horn, 1955, 1963)
Thermopolis(?) shale, Upper part: Upper shale unit		Mowry shale: Lower part	Upper Thermopolis shale	Shelf Creek shale	Thermopolis shale:	Thermopolis shale: Upper black shale member	Thermopolis shale:
Thermopolis(?) shale, Upper part: Muddy(?) sandstone member	Muddy(?) sandstone	Thermopolis shale: Upper sandstone member	Muddy sandstone member	Muddy sandstone	Thermopolis shale: Middle sandy member	Thermopolis shale: Muddy sandstone member	Thermopolis shale: Muddy sandstone member
Thermopolis(?) shale, Upper part: Lower shale unit	Thermopolis(?) shale	Thermopolis shale: Middle shale member	Lower Thermopolis shale	Thermopolis shale: Upper shale, Middle silty shale, Lower shale	Thermopolis shale:	Thermopolis shale: Lower part	Thermopolis shale:
Sandstone member	Kootenai formation: "Case-hardened sandstone"	Thermopolis shale: Lower sandstone member	"Rusty beds" Greybull sandstone	Thermopolis shale: Rusty beds	Sykes Mountain formation	Morrison and Cloverly formations undifferentiated: Rusty beds	Cloverly formation: Sandstone
UNCONFORMITY		UNCONFORMITY		UNCONFORMITY	UNCONFORMITY		
Kootenai formation	Kootenai formation	Kootenai formation	Cloverly formation	Cloverly formation	Cloverly formation	Morrison and Cloverly formations undifferentiated	Cloverly formation

The sandstone consists of angular to subround grains of quartz which range in size from 0.05 mm to 0.30 mm; most are about 0.15 mm. These grains are generally well cemented by secondary silica, iron oxide, and calcite; only locally is the sandstone weakly cemented and friable.

The sandstone member is nowhere fully exposed, but it is estimated to range in thickness from 75 to 120 feet. No fossils were found.

In Western Wyoming, sandstone beds which occupy the same stratigraphic position and have much the same color, lithology, and mode of outcrop as this unit were called the rusty beds by Love (1956, p. 76) and assigned to the uppermost part of the Cloverly, the stratigraphic equivalent of the Kootenai (p. 36). Hall (1961, p. 66) referred to equivalent beds exposed north of the Tepee Creek quadrangle as the case-hardened sandstone, and he also considered them as the uppermost unit of the Kootenai. It seems to me, however, that the stratigraphic relations in this area strongly favor assigning this sandstone to the Thermopolis(?) for the sandstone intertongues with the overlying shale and seemingly rests unconformably on the Kootenai.

Much the same stratigraphic relations were noted in the Livingston, Mont., area by Roberts (1965) who designated the sandstone the "lower sandstone member of the Thermopolis Shale" and correlated it with the rusty beds and Greybull Sandstone Member of the Cloverly Formation of central Wyoming, the Fall River Formation of northeastern Wyoming, and the Flood Member of the Blackleaf Formation of northwestern Montana. The unit is also probably correlative with Moberly's (1960, p. 1149) Sykes Mountain Formation exposed at the northern end of the Bighorn Basin (table 4).

Upper part

The upper part of the Thermopolis(?) forms gentle slopes and hills above the ledges and steeply dipping hogbacks of the sandstone member. The shales, which range from dark gray to black, are extremely thin bedded, fissile, soft, and devoid of invertebrate fossils. Scattered irregularly through them are dark-brown ironstone concretions several inches in diameter. The basal beds are shaly siltstone and range in thickness from about ½ inch to 6 inches; these intertongue with the uppermost very fine grained sandstone beds of the underlying sandstone member. Above the basal beds the fissile shale predominates; thin light- to dark-gray siltstone beds are intercalated in places.

These beds total about 125 feet in thickness and probably correlate with the middle shale member of the Thermopolis Shale as used by Roberts (1965) in the Livingston area, the Skull Creek Shale of northeastern Wyoming, the lower part of the Thermopolis Shale of northwestern Wyoming as defined by Haun and Barlow (1962), part of the Thermopolis Shale of Eicher (1960, p. 19), and the Thermopolis(?) Shale as used by Hall (1961, p. 66–68) for the area to the north (table 4).

The next younger unit of the upper part of the Thermopolis(?) is a dark-brown thin-bedded to platy sandstone, generally about 50 feet thick, which forms a prominent hogback along the north side of Cabin Creek valley. This sandstone, much like the underlying sandstone member, is thin bedded to platy and strongly crossbedded and is composed mainly of fine to medium angular to subround quartz grains. It is weakly to moderately well cemented by calcite and commonly friable. Its upper and lower margins are vague, for it intertongues with the confining shale. The thin sandstone beds become finer grained and pass into shaly siltstone beds which grade almost imperceptibly into shale beds. The sandstone changes thickness along strike, probably owing to these gradational lithologic changes. Possibly this sandstone is correlative with the Muddy Sandstone Member of the Thermopolis Shale as used by Horn (1955, 1963) and Haun and Barlow (1962) in central and western Wyoming. It may also be correlative with the "upper sandstone member of the Thermopolis" as used by Roberts (1965) in the Livingston area (table 4).

The top unit of the upper part is a sequence of dark-gray to black soft fissile shales lithologically identical with the shale beds exposed below the Muddy(?) Sandstone Member. These shale beds constitute the youngest sedimentary rocks exposed in the report area; because of erosion their thickness is unknown. Apparently they are not exposed to the north, where Hall (1961, p. 70) reported that nonmarine volcanic detritus overlies the Muddy(?) Sandstone Member.

These beds are here grouped with the Thermopolis(?) and are considered to be equivalent to the unnamed shale unit at the top of the Thermopolis Shale as used by Horn (1955, 1963), and to the upper part of the Thermopolis Shale as used by Haun and Barlow (1962) in Wyoming. Throughout parts of Montana and Wyoming these shales are overlain by the similar-appearing Mowry Shale. Consequently the contract between the two shales is difficult to distinguish, and some geologists have mapped these shales of the Thermopolis with the lower part of the Mowry (table 4) (Roberts, 1965; Keefer and Troyer, 1964).

IGNEOUS ROCKS

The igneous rocks consist of a small group of intrusive bodies—the largest of which is a laccolith—small amounts of andesite breccia and shoshonite flows possibly correlative in age, and extensive pyroclastic flows which cover most of the eastern third of the quadrangle (pl. 1). Field relations suggest that the intrusions were emplaced at some time after the sedimentary beds were folded and broken by thrust faults but before the flows. The sedimentary dome formed by the emplaced laccolith was partly buried by the earliest shoshonite flows.

INTRUSIVE ROCKS

All the intruded igneous rocks are considered to be derived from the same magma chamber and contemporaneous in age. They are the oldest igneous rocks exposed and consist chiefly of sills, plugs, and the Gallatin River laccolith, which was first examined, described, and named by Iddings and Weed during a broad geological study of Yellowstone National Park (1899, p. 84–85). The intrusions, localized near the north edge of the quadrangle, are all composed of dacite porphyry, which consists chiefly of phenocrysts of zoned plagioclase, altered hornblende, and rounded quartz in a fine-grained groundmass of virtually the same minerals. They have an alkali-lime index (SiO_2 value at which $Na_2O+K_2O=CaO$) of about 59.0, and belong, therefore, to the calc-alkalic series as defined by Peacock (1931). Most of these igneous bodies are confined to the Madison Group, and almost all contain altered inclusions of country rock.

GALLATIN RIVER LACCOLITH

Only the south half of the Gallatin River laccolith is in the Tepee Creek quadrangle, and it is exposed chiefly along the Gallatin River valley at the mouth of Snowslide Creek (pl. 1). Although the laccolith has been intensively dissected by both the Gallatin River and its tributary Snowslide Creek, much of it is still concealed beneath sedimentary and volcanic rocks. Scattered exposures indicate that it is oval and elongate northward; its north axis is probably about 4 miles long, and its east axis is probably about 2 miles long. Although the laccolithic floor is not exposed, it may conform to the top of the Precambrian basement complex (cross section $A–A'$, pl. 1); so the laccolith may be planoconvex (mushroomlike) in cross section. Its total thickness is unknown, but I estimate it to have been at least 2,000 feet thick when emplaced; about 1,500 feet is now exposed.

The laccolith is semiconcordant; its roof locally conforms to the confining strata but elsewhere cuts across them. The youngest sedimentary rocks directly overlying the laccolith belong to the Madison Group; the oldest exposed sedimentary rocks lying on the laccolithic borders are part of the undivided Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite.

The dissected sedimentary units that mirror the east flank of the laccolith (along the east wall of the Gallatin River valley) are overlain unconformably mainly by near-horizontal pyroclastic flows, but in one small area the tilted sedimentary strata disappear beneath the toe of a shoshonite flow, which itself is buried beneath the pyroclastic flows (pl. 1).

Extending outward from the laccolithic flanks, much like branches of a tree, are many sills, some as much as 400 feet thick, and minor irregular apophyses. For cartographic reasons only a few of the larger sills are shown on plate 1. Here and there small dikes connect the intrusions.

Xenoliths and autoliths are common along the laccolithic flanks and in the fringing sills but are rare near the center of the laccolith. These inclusions, which range in shape from angular to round and in size from fragments about a quarter of an inch in diameter to blocks about 10 inches on a side, include dark-gray calcareous fissile shale (only slightly altered), quartzitic sandstone, light-gray dolomite, dark-gray amphibolite, medium-gray schist, and well-banded gneiss. These xenoliths are much like the rocks exposed elsewhere in the area. Autoliths of fine-grained dacite porphyry are also common, and they are interpreted as remnants of a chilled zone which was broken and reincorporated in the laccolith.

A chilled zone about 10 feet thick forms part of the laccolithic roof. The large plagioclase phenocrysts (2.0–6.0 mm) characteristic of the bulk of the laccolith end abruptly at the inner edge of the chilled zone and are replaced by small light-gray phenocrysts, most about 0.50 mm in diameter, in a dark-gray fine-grained groundmass. The overlying limestone beds have been conspicuously altered to light-gray, almost white, finely crystalline limestone and dolomite with much epidote. Locally the sedimentary beds along the contact are slightly brecciated and contain small stringers of dacite porphyry several feet long and a few inches wide.

Directly north of the quadrangle, along the west valley wall of the Gallatin River, a small mineralized body, chiefly magnetite and hematite, is at the contact of the laccolith and the overlying limestone beds. This small contact-metamorphic deposit was mined by the Apex Development Co. during the 1920's (p. 86).

Megascopically, the dacite porphyry is bluish gray (about $N5$) dense, and porphyritic; prominent phenocrysts are plagioclase feldspar, quartz, hornblende, and magnetite. The plagioclase phenocrysts stand out sharply as irregularly distributed white angular oblongs; in many, interior reaction zones are clearly visible. What little quartz is present occurs as widely separated discrete minute phenocrysts. The hornblende appears as small dark-gray to black needles and the magnetite as angular specks.

Microscopically, the phenocrysts are seen to be outlined by a groundmass of holocrystalline microgranitic texture. Modal analyses indicate that the groundmass and the phenocrysts occupy about equal volumes; the groundmass commonly occupies about 52 percent and the phenocrysts the remaining 48 percent. The following modal analyses (in percent) are representative. (See fig. 10 for sample locations.)

Field No. / Type intrusion	Wg-340 Small plug	Wg-397b Gallatin River laccolith	Wg-419 Gallatin River laccolith
Quartz	0.1	1.4	2.8
Hornblende	12.2	15.5	12.7
Plagioclase feldspar	28.4	32.3	34.1
Magnetite	2.2	0.9	0.9
Groundmass	57.1	50.0	49.4
Total	100.0	100.1	99.9

Plagioclase feldspar.—Almost all the plagioclase phenocrysts are zoned, the rims commonly being andesine (An_{30}) and the cores as calcic as labradorite (An_{60}). The few unzoned plagioclase grains are andesine, ranging from An_{30} to An_{35}. Most of these plagioclase phenocrysts are euhedral, range in size from about 0.02 mm to 6.0 mm, and are both mottled and outlined by alteration products, which include sericite(?), very fine grained feldspathic material, calcite, and unidentifiable clay minerals. The mottles are irregular shaped, but the reaction rims outlining the phenocrysts are regular and about 0.04 mm thick. Comparable reaction bands outline the composition zones of the zoned feldspars. These rims probably formed in one or more periods of magmatic resorption, during which the already-formed plagioclase phenocrysts reacted with the still-fluid magma to form the alteration products which now outline them. Common inclusions are apatite, magnetite, and amphibole.

Hornblende.—Although X-ray-diffractometer patterns indicate that two amphiboles are present in the rock (Theodore Botinelly, written commun., 1963), only hornblendes were noted in thin section; these are all bright green and subhedral in outline and are almost completely altered to chlorite (penninite), calcite, and magnetite. Some of the grains have been virtually destroyed; only their outlines are discernible in the groundmass. The hornblendes are as much as 3.0 mm long and 0.8 mm wide, although most are about 0.4 mm in diameter. Small crystals of sphene, formed mainly from the titanium released during the alteration of the hornblende, are closely associated with many of the altered hornblende grains. Dark borders of magnetite dust rim some of the hornblende grains, buttressing the evidence of magmatic resorption.

Quartz.—Quartz phenocrysts are rare and widely scattered and commonly appear as clear subround to well-rounded grains. Most are about 0.6 mm in diameter, although some are as much as 1.50 mm in diameter. Common inclusions are fragments of hornblende, blebs of groundmass, and crystals of apatite. Some of the quartz grains are marked by small embayments that break the even continuity of the sides. These embayments, as well as the general rounding of the grains, reaffirm the probability of magmatic resorption.

Magnetite.—Most of the magnetite phenocrysts are opaque and angular, range from 0.05 to 0.1 mm in size, and are dispersed irregularly through the groundmass.

Accessory minerals.—Common accessory minerals include rods and rounded grains of apatite, wedge-shaped grains of sphene, and sparse grains of epidote surrounded by a thin reaction rim of unidentified opaque materials.

Groundmass.—The general texture of the groundmass is holocrystalline micrograntic, but in a few rocks, chiefly in sills which fringe the laccolith, it is pilotaxitic or microfelsitic. Mineralogically, the groundmass seems to consist of an intergrowth of the same minerals that form the phenocrysts, plus myriad grains of quartz and alkalic feldspar. The interstitial quartz apparently dominates the mineral assemblage, as is apparent from the high SiO_2 content of chemically analyzed samples (table 5, Nos. 1 and 2; fig. 10). The plagioclase, probably more sodic than the phenocrysts, is untwinned. The hornblende and magnetite appear chiefly as scattered needles and opaque dust. Apatite and sphene form minute inclusions. Most of the constituents are about 0.005 mm in diameter.

The pyroclastic flows that generally bury the folded and dissected sedimentary beds which reflect the east flank of the laccolith (p. 39) are absent from that part of the Gallatin River valley which bisects the laccolith. This seems highly significant because directly south of there they extend down to, and pass below, the surficial deposits which floor the valleys. This is well shown near the mouth of Bacon Rind Creek (about 3 miles south of the laccolith) and in all the major and tributary valleys south of that point. If that part of the Gallatin valley which bisects the laccolith was in existence when the flows spread across this area, surely the flows would have filled it much as they filled the valleys upstream (south) from this point. The complete absence of the flows from this sector of the valley, and their widespread extent upstream, must mean that the Gallatin River bisected the laccolith after the eruption of the pyroclastic flows.

The comparative widths of the Gallatin River valley where it transects the laccolith and upstream from that point buttress this interpretation. Where the Gallatin River bisects the laccolith, the valley is narrow and is bounded by steep walls. By contrast, upstream from that locality the valley is broad and low walled.

TABLE 5.—*Comparison of chemical and normative compositions of intrusive rocks from the Tepee Creek quadrangle with those of similar rocks collected elsewhere*

[Sample locations are shown in fig. 10]

	Laccolith		Sill		Small intrusion		Average of intrusions	Other rocks		
	1 [1]	2	3	4	5	6 [1]	7	8	9	10
Laboratory No.	160810	D100092	D100091	D100089	D100090	158693				
Field No.	Wg-397	Wg-419	Wg-399	Wg-328	Wg-340	Wg-346				
Analyses (in percent)										
SiO_2	61.0	63.88	59.62	63.78	62.29	67.5	63.01	65.68	63.58	64.27
Al_2O_3	17.4	17.54	17.38	16.84	17.39	16.0	17.09	16.25	16.67	16.87
Fe_2O_3	1.4	2.35	1.95	1.47	1.70	.9	1.63	2.38	2.24	3.13
FeO	2.6	1.46	2.88	1.66	1.89	1.0	1.92	1.90	3.00	2.01
MgO	1.8	.88	1.97	1.11	1.26	.65	1.28	1.41	2.12	1.85
CaO	5.3	3.51	5.06	4.47	4.47	2.5	4.22	3.46	5.53	4.63
Na_2O	4.0	5.22	3.46	4.01	4.36	4.7	4.29	3.97	3.98	3.97
K_2O	1.9	2.61	2.32	2.34	2.69	2.6	2.41	2.67	1.40	1.68
H_2O^+	1.8	.92	2.06	1.42	1.37	2.4	1.82	1.50	.56	1.20
H_2O^-	.95	.12	1.01	.95	.84	----	.73	----	----	----
TiO_2	.39	.38	.44	.31	.39	.24	.36	.57	.64	.38
P_2O_5	.24	.19	.25	.15	.19	.10	.19	1.50	.17	.08
MnO	.15	.07	.13	.11	.11	.15	.12	.06	.11	.06
CO_2	1.4	.57	1.20	1.25	.88	1.1	1.07	----	----	----
Norms (in percent)										
Q	19.46	16.92	19.56	23.59	17.55	27.18	20.54	26.79	19.90	22.05
or	11.22	15.42	13.71	13.82	15.89	15.36	14.24	15.78	8.27	9.93
ab	33.83	44.07	29.19	33.84	36.80	39.75	36.28	33.58	33.66	33.58
an	15.87	12.19	15.60	13.05	15.14	3.80	12.93	7.36	23.49	22.44
C	2.95	1.68	3.48	2.95	1.77	4.06	2.69	4.13	----	.30
di en	----	----	----	----	----	----	----	----	.73	----
di fs	----	----	----	----	----	----	----	----	.39	----
di wo	----	----	----	----	----	----	----	----	1.18	----
hy en	4.48	2.19	4.90	2.76	3.14	1.62	3.19	3.51	4.55	4.61
hy fs	3.25	.24	3.19	1.53	1.63	.98	1.81	.69	2.42	.59
mt	2.03	3.41	2.83	2.13	2.46	1.30	2.36	3.45	3.25	4.54
ap	.57	.45	.59	.36	.45	.24	.46	3.55	.40	.19
il	.74	.72	.84	.59	.74	.46	.68	1.08	1.22	.72
hm	----	----	----	----	----	----	----	----	----	----
C.I.P.W. classification										
Class	II	I	II	I	I	I	I	I	II	I
Subclass	I	I	I	I	I	I	I	I	I	I
Order	4	4	4	4	4	4	4	4	4	4
Rang	3	2	3	2	2	1	2	2	3	3
Subrang	4	4	4	4	4	4	4	4	4	4

[1] Rapid rock analysis.

C.I.P.W. CLASSIFICATION

Class { I. Ratio of salics to femics greater than 7.00.
II. Ratio of salics to femics between 7.00 and 1.667.

Order { 4. Feldspars dominant over quartz.
5. Feldspars extreme over quartz.

Rang { 1. Salic alkalis extreme over salic lime.
2. Salic alkalis dominant over salic lime.
3. Salic alkalis and salic lime equal.

Subrang 4. Na_2O dominant over K_2O.

SAMPLE DESCRIPTION

1. Dacite porphyry from Gallatin River laccolith. Analysts: Paul Elmore, S.D. Botts, Gillison Chloe, Lowell Artis, H. Smith.
2. Dacite porphyry from Gallatin River laccolith. Analyst: D. F. Powers.
3. Dacite porphyry(?) from sill of Gallatin River laccolith. Analyst: D. F. Powers.
4. Dacite porphyry from sill near Bacon Rind Creek. Analyst: D. F. Powers.
5. Small intrusion near Snowslide Creek. Analyst: D. F. Powers.
6. Small intrusion near head of Bacon Rind Creek. Analysts: Paul Elmore, I. H. Barlow, S. D. Botts, Gillison Chloe.
8. Dacite; average of 90 analyses (Daly, 1933, p. 15).
9. Dacite; average of 50 analyses (Nockolds, 1954).
10. Dacite; average of 19 analyses (Johannsen, 1949).

It seems likely that the flows buried a dissected landscape in which the major ancestral streams flowed southward from a highland along, and probably parallel to, the north edge of the quadrangle. This highland, formed in large part by the sedimentary dome underlain by the laccolith, precluded any northward flow of the ancestral Gallatin River. At the time of their eruption (possibly late Pliocene or early Pleistocene, p. 53) the pyroclastic flows filled the ancestral valleys and lapped onto the flanks of the dome underlain by the laccolith, possibly burying it completely. A new drainage pattern formed at that time probably resulted in a northward diversion of the Gallatin River, which in time probably cut through the dome and exposed the laccolith.

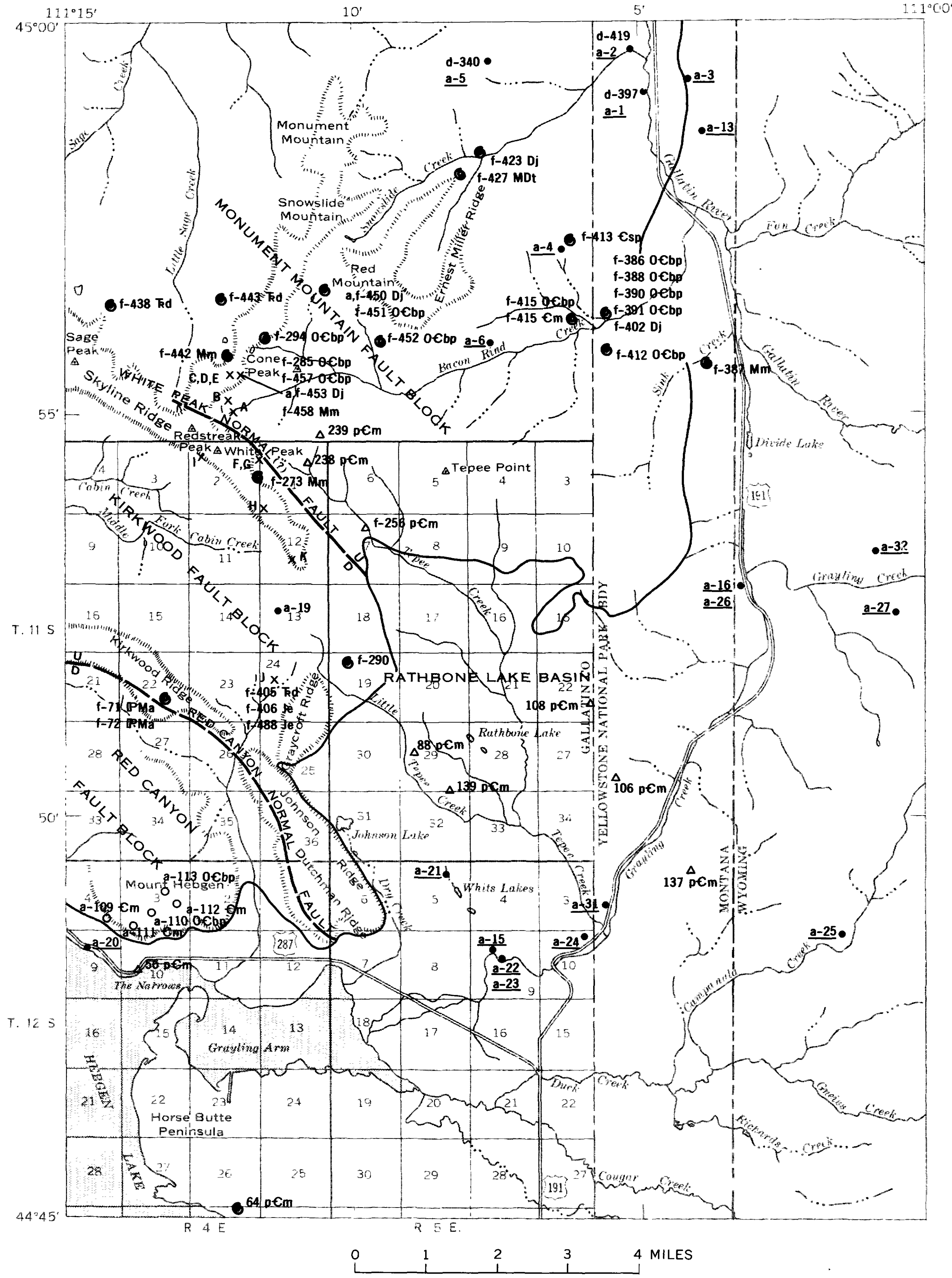


FIGURE 10.—Data map, Tepee Creek quadrangle, Montana-Wyoming.

EXPLANATION

×A

Measured section described in detail

f-457

Fossil-collection locality and field number

○a-109

Sedimentary rock sample, chemically analyzed

● a-27

Igneous rock sample, chemically analyzed

△106 p€m

Rock chip for modal analysis and field number

d-340

Dacite porphyry sample

Possible westward extent of Yellowstone Tuff

U / D

Fault, approximately located

U, *upthrown side;* D, *downthrown side*

Symbols indicating sampled unit

- Je, Ellis Group
- TRd, Dinwoody Formation
- IPMa, Amsden Formation
- Mm, Madison Group
- MDt, Three Forks Formation
- Dj, Jefferson Formation
- O€bp, Bighorn(?) Dolomite, Snowy Range Formation, and Pilgrim Limestone undifferentiated
- €m, Meagher Limestone
- p€m, pre-Belt crystalline metamorphic rocks

OTHER INTRUSIONS

The small sills and dikes along both valley walls of the Gallatin River (near the mouth of Snowslide Creek) are patently offshoots of the Gallatin River laccolith (pl. 1). The broad igneous sheet of unknown thickness, emplaced chiefly in the Park Shale, which caps the divide between Bacon Rind and Snowslide Creeks may be part of a sill extending southwestward from the laccolith. The east edge of this body transects folded Upper Cambrian strata and deforms beds tentatively assigned to the Jefferson Formation.

Another sill, directly north of the quadrangle, is in the Shedhorn Sandstone and may also be related to the Gallatin River laccolith.

Along the northwest wall of Snowslide Creek, the minor igneous intrusions which dome Madison and younger strata are possibly vertical or near-vertical apophyses from the roof of the Gallatin River laccolith.

The intrusions exposed in the valley of Bacon Rind Creek—a sill about 200 feet thick along the south valley wall and a cone-shaped plug along the north valley wall—may be related to another buried semiconcordant body (laccolith?) possibly outlined by the northeastward-plunging Bacon Rind anticline. The cone-shaped plug cuts vertically through Precambrian dolomite and ends against the base of the Flathead Sandstone.

Most of the minor intrusions are rimmed by thin chilled zones which rarely exceed 1–2 feet in thickness and invariably the confining rocks are altered somewhat; for example, the limestone, normally dark grayish blue, alters to a white marble.

In these minor intrusions the groundmass commonly constitutes about 75 percent of the rock and the phenocrysts but 25 percent. The composition of the intrusions is nearly identical with that of the Gallatin River laccolith (table 5); thus there is little doubt that all came from the same magma chamber.

AGE OF THE INTRUSIONS

The age of the laccolith, and by inference the age of the other intrusions, can be fixed within rather narrow limits. The youngest remaining beds bowed up, presumably as a result of the intrusions, belong to the Shedhorn Sandstone of Permian age. The intrusions, therefore, are at least younger than Permian, and may be much younger, for among the xenoliths included in the Gallatin River laccolith are fragments of dark-gray fissile calcareous shale (p. 39) which may have come from the Thermopolis(?) Shale of Early Cretaceous age (p. 37) or any of the several Cambrian units. Moreover, along the east wall of the Gallatin River valley, sedimentary units ranging from the Madison to the Quadrant dip eastward at about 25°, reflecting the roof of the underlying laccolith. In one locality these tilted and dissected sedimentary strata disappear below the toe of the near-horizontal flow (p. 39), which in turn is buried by still younger pyroclastic flows. The shoshonite is probably related to the andesite breccia (p. 45), which is correlative with Hague's "early basic breccia" dated by Dorf (1960, p. 259) as early middle Eocene (p. 45). It would seem, therefore, that the intrusions were emplaced before early middle Eocene.

More specifically, the intrusions were probably emplaced after Laramide folding and thrusting. To the north in the Elkhorn Mountains, the main pluton, the Boulder batholith, apparently was emplaced after major thrusting, which occurred at or near the end of the folding; the age proposed for the emplacement of the Elkhorn Mountains intrusions is very Late Cretaceous or Paleocene (Klepper and others, 1957, p. 60). The intrusions in the Three Forks Basin were emplaced at about the same time (Robinson, 1963, p. 95), and in the Tepee

Creek area the same age may apply, for the Gallatin River laccolith and its satellite intrusions either truncate or penetrate folded strata (p. 39, 43). (Conclusive evidence, however, that some of these folds, such as the Bacon Rind anticline, are Laramide in age is not available; they may reflect some still-concealed intrusion).

In the Gardiner, Mont., area there is clear evidence of two episodes of dacitic intrusion, one correlative with the early acid breccia (and, therefore, older than the early basic breccia), and a second possibly of Oligocene age (G. D. Fraser, oral commun., 1965). If the dacite porphyry intrusions of the Tepee Creek quadrangle are correlative with the earlier dacite intrusions of the Gardiner area, as I think they may be, then their age can be fixed very closely indeed. Dorf (1960, p. 258), on the basis of an elaborate flora, dated the early acid breccia as "spanning the time interval between late early Eocene and early middle Eocene * * *." It seems likely that the intrusions of this area are post-Paleocene but pre-early middle Eocene, and most probably of late early Eocene age.

EXTRUSIVE ROCKS

The extrusive rocks consist of minor amounts of andesite breccia and shoshonite and widespread thick pyroclastic flows of ryholitic tuff, which was named the Yellowstone Tuff by Boyd (1961, p. 390). The andesite breccia, which likely is correlative with Hague's early basic breccia (Iddings, 1899b, p. 275), is probably contemporaneous with the shoshonite. They are the oldest extruded rocks in the quadrangle and are incompletely exposed, being covered in great measure by the tuffs; as a result little is known of their composition, thickness, or extent. By contrast, the younger, Yellowstone Tuff is widely exposed throughout the northwest corner of Yellowstone National Park, where it has been studied in detail by Boyd (1961). This tuff can be traced westward into the Tepee Creek quadrangle, where the general field relations suggest that the flows moved across an eroded terrain of older rocks.

ANDESITE BRECCIA

About 1.3 square miles in the southeast corner of the quadrangle is underlain by gray to brown breccia composed chiefly of angular fragments of andesitic material; the rock has been mapped as an andesite breccia (pl. 1). Dense forest covers the outcrop and exposures are poor. The material is massive, lacks bedding, and normally breaks into fragments a few inches on a side, although in places boulders several feet in diameter have been formed.

The breccia fragments are formed of, and tightly held by, andesitic material that consists chiefly of phenocrysts of subhedral plagioclase (oligoclase-andesine), completely altered hornblende(?) needles, and magnetite in a dense groundmass of plagioclase microlites and magnetite dust. The hornblende(?) seems to have altered to hematite dust which preserves the elongate needlelike nature of the former hornblende crystals.

The exposures, although limited, indicate that the andesite breccia is overlain directly by the Yellowstone Tuff. This interpretation is supported by local exposures of the basal tuff beds that include angular fragments and boulders of the breccia.

SHOSHONITE

Shoshonite, a name applied by Iddings (1895) to a rock much like basalt, is exposed in the northeast corner of the quadrangle and near the mouth of Campanulla Creek (pl. 1). Iddings, (1899a, p. 339) reported that shoshonite crops out "on Grayling Creek and west of The Crags", but none was found in those places during the present study.

The shoshonite in the northeast corner of the quadrangle crops out in a small canyon along the east valley wall of the Gallatin River (locality a–13 of fig. 10) where it overlies eastward-dipping strata of the Madison Group and in turn is overlain on the north, east, and south by the near-horizontal Yellowstone Tuff. The shoshonite outcrop appears as small parallel ridges, each about 100 feet wide, trending generally eastward.

Near Campanulla Creek the rock is scoriaceous and deeply weathered, almost earthy, and is also overlain by the Yellowstone Tuff; the whole unit is almost concealed beneath glacial detritus. The weathered shoshonite is cherry red and ropy and locally shows good flow structure. When viewed microscopically the rock is seen to consist of rounded vesicles enclosed in a holocrystalline groundmass of needlelike plagioclase grains. Scattered through the groundmass are angular phenocrysts of labradorite (An_{60}) as much as 0.20 mm long and 0.02–0.06 mm wide.

Unweathered shoshonite is dark gray to black and dense, and it breaks into angular boulders 2–3 feet on a side. In hand specimen the rock is seen to be porphyritic, with abundant dark-gray angular plagioclase phenocrysts distributed irregularly through a black waxy fine-grained groundmass. Microscopically, the rock is seen to consist of phenocrysts of nearly rectangular plagioclase (chiefly labradorite, An_{60}), both orthopyroxene (hypersthene) and clinopyroxene (augite), altered olivine, and angular specks of magnetite in a tightly interwoven mesh of microlites tentatively identified as andesine. The groundmass probably contains much alkalic feldspar, for the results of chemical analyses suggest that the rock is unusually rich in alkalis and, so, is much like an alkalic basalt (table 6). Scattered through the groundmass are small grains of pyroxene, olivine, and magnetite.

Many of the plagioclase phenocrysts are zoned, and all are fractured to some degree, which suggests somewhat vigorous movement in the magma. Reaction rims are extremely thin, and locally absent. Hypersthene is the most common inclusion in these phenocrysts.

Of the pyroxenes the orthopyroxene hypersthene ($MgSiO_3$=65 percent) is much more plentiful than the clinopyroxene augite (Ca=42 percent, Mg=34 percent, Fe=24 percent). The hypersthene grains are fractured and fairly well rounded and are rimmed by a thin corrosion border. Magnetite is the dominant inclusion.

Olivine grains are rare, occurring as phenocrysts and microphenocrysts surrounded by dark reaction rims of opaque materials.

Two generations of magnetite, occurring as phenocrysts and as dust in the groundmass, are in the rock.

The groundmass consists of felted laths of polysynthetically twinned plagioclase with interspersed grains of alkalic feldspar, granular pyroxene, and olivine and magnetite dust.

Chemically, the shoshonite, when compared with other basic rocks, is high in silica (57 percent) and potash (3 percent), moderately high in soda (3 percent), and somewhat low in lime (6 percent) and magnesia (4 percent) (table 6).

Although the relations between the andesite breccias and the shoshonites are nowhere exposed in this quadrangle, the two rocks are probably correlative and older than similar basaltlike rocks associated with the rhyolites. The equivalency of the shoshonite and andesite breccia was noted by Iddings (1899a, p. 326, 339), who stated in his description of the absarokite-shoshonite-banakite series: "There are certain basaltic-looking rocks associated with the older andesitic breccias * * *." Moreover, in the Gardiner area shoshonite-like rocks both underlie and overlie the andesite breccia and locally may be interbedded in the breccia (G. D. Fraser, oral commun., 1965).

The shoshonite can be distinguished from the younger basalt and basaltlike rocks on the basis of the presence or absence of pyroxene phenocrysts. The shoshonite commonly has both orthopyroxene and clinopyroxene phenocrysts; by contrast, the several basalts and basaltlike rocks interlayered and associated with the rhyolite do not contain pyroxene phenocrysts (Boyd, 1961, p. 401–402).

Although no specific evidence was found to date the andesite breccia–shoshonite complex, the field relations along the east side of the Gallatin River valley indicate that these rocks are younger than the Gallatin River laccolith (late early Eocene?) but older than the Yellowstone Tuff (late Pliocene or early Pleistocene, p. 53). As noted above, the andesite breccia–shoshonite complex is likely correlative with Hague's early basic breccia, which was dated as early middle Eocene by Dorf (1960, p. 259) on the basis of a large composite fossil flora.

YELLOWSTONE TUFF

The youngest extrusive rocks are the westward extension of rhyolitic welded tuffs, named the Yellowstone Tuff by Boyd (1961, p. 390), which mantle much of the northwest corner of Yellowstone National Park. They are probably correlative with the Crown Butte extrusives of Hall (1961, p. 112–120) which crop out north of this quadrangle.

Most of the eastern third of the quadrangle is underlain by these pyroclastic flows, and only here and there is their broad expanse interrupted by knolls and hills composed of older rocks (pl. 1).

Dense forests of lodgepole pine blanket the Yellowstone Tuff, and the boundaries of the unit can almost be determined by plotting the verdant timber stands. By contrast, the relict hills, composed of metamorphic and sedimentary rocks which project through the tuff, support a rather sparse forest cover. This striking difference in vegetation may be attributable to the myriads of small near-surface fractures in the tuff, each fracture apparently serving as a channel for the ground water which feeds the shallow root systems of the lodgepole pines.

The welded tuffs erode to form round hills separated by sinuous valleys which reach far back into the upland. Because of the dense forest cover, adequate exposures are rare and widely dispersed. In some places the Yellowstone Tuff stands as a vertical cliff about 75 feet high, broken by columnar joints; more commonly it forms a steep slope composed of alternating ledges and cliffs, the former covered by debris. This alternation probably reflects the difference in resistance between partly welded and firmly welded tuffs.

The original extent of the flows can only be surmised. Small patches of tuff in the western and northwestern reaches of the quadrangle (pl. 1) suggest that the flows may once have mantled much of the quadrangle. Seemingly identical in all respects with the Yellowstone Tuff, these tuff deposits are near the head of Red Canyon (secs. 13 and 25, T. 11 S., R. 4 E.) (sample 19 of fig. 10 and table 6); near Lakeview (sample 20 of fig. 10 and table 6); north of The Narrows (NE¼ sec. 10, T. 12 S., R. 4 E.); on Horse Butte (SE¼ sec. 22, T. 12 S., R. 4 E.); near the Watkins Creek Ranch (sec. 1, T. 12 S., R. 3 E., and sec. 36, T. 11 S., R. 3 E.) west of this quadrangle; and in the extreme northwest corner of the quadrangle near Sage Creek (pl. 1).

319-733 O—69——4

TABLE 6.—*Comparison of chemical and normative compositions of extrusive rocks from the Tepee Creek quadrangle with those of similar rocks collected elsewhere*

[Sample locations are shown in fig. 10]

	Shoshonite		Glassy Welded tuffs				Devitrified tuffs												Possible Rhyolite flow
	Tepee Creek quadrangle	Sepulchre Mountain Yellowstone Natl. Park	Tepee Creek quadrangel		Average of vitrified tuffs	Mt. Everts Yellowstone Natl. Park	Tepee Creek quadrangle									Average of devitrified tuffs	Norris Geyser Basin core	South end Madison Range	Tepee Creek quadrangle
	13	14	15 [1]	16 [1]	17	18	19 [1]	20 [1]	21 [1]	22 [1]	23 [1]	24 [1]	25 [1]	26 [1]	27 [1]	28	29	30 [1]	31 [1]
Laboratory No.	0100093		158004	158009			153716	158000	158001	158002	158003	158005	158007	158008	158010				158006
Field No.	Wg-404		Wg-98	Wg-147			Wg-29	Wg-58	Wg-92	Wg-94	Wg-95	Wg-100	Wg-135	Wg-146	Wg-149				Wg-102
									Analyses (in percent)										
SiO_2	56.50	51.75	73.4	75.5	74.45	73.15	75.2	74.5	74.9	75.6	75.9	73.4	75.6	74.3	73.5	74.77	76.95	75.6	75.5
Al_2O_3	16.29	17.48	12.4	12.6	12.5	12.13	12.9	13.0	13.1	12.4	12.5	13.1	12.5	12.9	12.6	12.78	12.18	12.6	12.7
Fe_2O_3	1.88	6.42	.6	1.8	1.2	1.12	1.7	1.5	1.6	1.4	1.5	1.8	1.4	1.8	.65	1.48	1.07	1.0	1.5
FeO	5.60	1.46	1.1	.20	.65	.66	.33	.42	.32	.28	.28	.90	.38	.30	1.2	.37	.26	.76	.42
MgO	4.03	4.05	.16	.20	.18	.13	.03	.11	.14	.18	.24	.45	.15	.17	.19	.18	.12	.07	.20
CaO	6.42	8.20	.50	.36	.43	.71	.35	.48	.38	.29	.39	.98	.42	.46	.52	.47	.28	.59	.40
Na_2O	2.90	3.33	3.4	3.5	3.45	4.01	3.4	3.8	3.7	3.5	3.4	3.5	3.5	3.8	3.5	3.57	2.98	3.6	3.4
K_2O	3.31	3.72	5.2	4.8	5.00	5.27	5.2	5.1	5.2	4.9	4.6	4.6	4.8	5.0	4.9	4.92	5.50	5.1	4.8
H_2O+	1.02	2.26	2.3	.53	1.41	1.91	.68	.84	.62	.68	.66	.85	.61	.74	2.2	.88	.24	.36	.66
H_2O-	.29	------	------	------	------	.21	------	------	------	------	------	------	------	------	------	------	.06	------	------
TiO_2	1.00	.86	.13	.15	.15	.14	.13	.16	.16	.12	.13	.24	.13	.17	.15	.15	.13	.12	.15
P_2O_5	.38	.67	.03	.03	.01	------	------	.01	.01	.01	------	.02	.01	.01	.02	.01	.07	.02	.01
MnO	.12	Tr	.03	.03	.45	.04	.08	.03	.04	.02	.05	.06	.06	.04	.03	.05	.01	.04	.04
CO_2	.01	------	.05	.05	.05	------	.07	.05	<.05	<.05	<.05	<.05	<.05	<.05	<.05	<.05	.04	------	.05
									Norms (in percent)										
Q	6.84	------	31.90	35.81	33.88	28.76	34.94	31.74	32.50	35.65	37.34	32.74	35.93	31.87	32.40	34.08	38.11	33.65	36.25
or	19.56	21.98	30.72	28.36	29.54	31.14	30.72	30.13	30.72	28.95	27.18	27.18	33.35	29.54	28.95	29.07	32.49	30.13	28.36
ab	24.30	28.16	28.75	29.60	29.18	33.05	28.75	32.14	31.29	29.60	28.75	29.60	29.60	32.14	29.60	30.19	25.20	30.45	28.75
an	21.78	21.77	2.28	1.59	1.75	------	1.29	2.32	1.82	1.37	1.93	4.73	1.70	2.22	2.45	1.95	.68	2.80	1.92
C	------	------	.34	1.07	.77	------	1.21	.38	.72	.84	1.22	.63	.93	.43	.64	.87	1.08	.14	1.21

di en	1.67	5.24				.32													
di fs	1.22					.35													
di wo	3.01	6.06				.68													
hy en	8.36	3.47	1.76	.50	.45		.08	.27	.35	.45	.60	1.12	.37	.42	.47	.45	.30	.17	.50
hy fs	6.08		.40		.79										1.47			.45	
mt	2.73	2.21	.87	.31	1.74	1.24	.95	.99	.70	.62	.69	2.40	1.04	.60	.94	.92	.49	1.45	1.05
ap	.90	1.59	.07	.07	.02			.02	.02	.02		.05	.02	.02	.05	.02	.17	.05	.02
il	1.90	1.63	.25	.28	.28	.27	.25	.30	.30	.23	.25	.46	.25	.32	.28	.28	.25	.23	.28
hm		4.89		1.59			1.05	.82	1.12	.97	1.02	.14	.68	1.38		.84	.73		.78
C.I.P.W. classification																			
Class	II	II	I	I	I	I	I	I	I	I	I	I	I	I	I	I	I	I	I
Subclass	I	I	I	I	I	I	I	I	I	I	I	I	I	I	I	I	I	I	I
Order	5	5	4	3	4	4	4	4	4	4	3	4	3	4	4	4	3	4	3
Rang	3	3	1	1	1	1	1	1	1	1	1	2	1	1	1	1	1	1	1
Subrang	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3

[1] Rapid rock analysis.

C.I.P.W. CLASSIFICATION

Class { I. Ratio of salics to femics greater than 7.00.
II. Ratio of salics to femics between 7.00 and 1.667.

Order { 3. Quartz and feldspars equal.
4. Feldspars dominant over quartz.
5. Feldspars extreme over quartz.

Rang { 1. Salic alkalis extreme over salic lime.
2. Salic alkalis dominant over salic lime.
3. Salic alkalis and salic lime equal.

Subrang 3. K_2O and Na_2O equal.

SAMPLE DESCRIPTION

13. Shoshonite, east valley wall of Gallatin Valley, Tepee Creek quadrangle, Analyst: D. F. Powers.
14. Shoshonite, lava flow, northeast base of Sepulchre Mountain (Iddings, 1899a, p. 340).
15. Sample of vitric zone at base of flow near mouth of Grayling Creek, Analysts: Paul Elmore, S. D. Botts, Gillison Chloe, and I. H. Barlow.
16. Sample of vitric zone at base of flow near MP 270 along Montana-Wyoming State line. Analysts: Paul Elmore, S. D. Botts, Gillison Chloe, and I. H. Barlow.
18. Glassy welded tuff; base of Yellowstone Tuff, Mt. Everts, Yellowstone National Park (Boyd, 1961, p. 408).
19. Welded tuff at head of Red Canyon Creek. Analysts: I. H. Barlow, M. D. Mack and S. D. Botts.

Samples 20–27 analyzed by Paul Elmore, S. D. Botts, Gillison Chloe, and I. H. Barlow.

20. Welded tuff near Lakeview.
21. Welded tuff near Whits Lakes.
22. Welded tuff near mouth of Grayling Creek.
23. Welded tuff near mouth of Grayling Creek.
24. Welded tuff near mouth of Tepee Creek.
25. Welded tuff along Campanula Creek.
26. Welded tuff along Grayling Creek, near MP 270, Montana-Wyoming State line.
27. Welded tuff along Grayling Creek.
29. Lithoidal welded tuff; Norris Geyser Basin core (Boyd, 1961, p. 408).
30. Rhyolite welded tuff; south end of Madison Range (Hamilton and Leopold, 1962, p. B28).
31. Near mouth of Tepee Creek. Analysts: Paul Elmore, S. D. Botts, Gillison Chloe, and I. H. Barlow.

Most of these remnants range in thickness from 30 to 75 feet and are deeply weathered to a coarse rubble, but, where still intact, the larger remnants are bounded by steep escarpments and exhibit unusually level upper surfaces.

The exposures of tuff along Red Canyon are 8,700–8,900 feet above sea level, and those at Sage Creek are 8,500 feet above sea level. The remnants bordering Hebgen Lake are at altitudes of about 6,800 feet. The main body of rhyolite tuff in the eastern part of the Tepee Creek quadrangle is at an average altitude of about 7,200 feet and a maximum of 8,800 feet. Thus, if all the tuffs here described were once part of a continuous mass of composite flows, they would have had a minimum thickness of nearly 2,000 feet and would have covered all but the highest peaks in the area.

The separate remnants, however, may be nothing more than relicts of eroded local flows which were extruded from widely separated still-concealed fissures. If so, the main flows halted against a sinuous unbroken barrier which trends generally northeastward through the center of the quadrangle, and is formed in part by Precambrian metamorphics and in part by folded and faulted sedimentary rocks (fig. 10). In this interpretation, the flows, unable to crest the barrier, moved south into the former Madison River valley (now occupied by Hebgen Lake) and halted against the Precambrian rocks which form the east flank of the Madison Range.

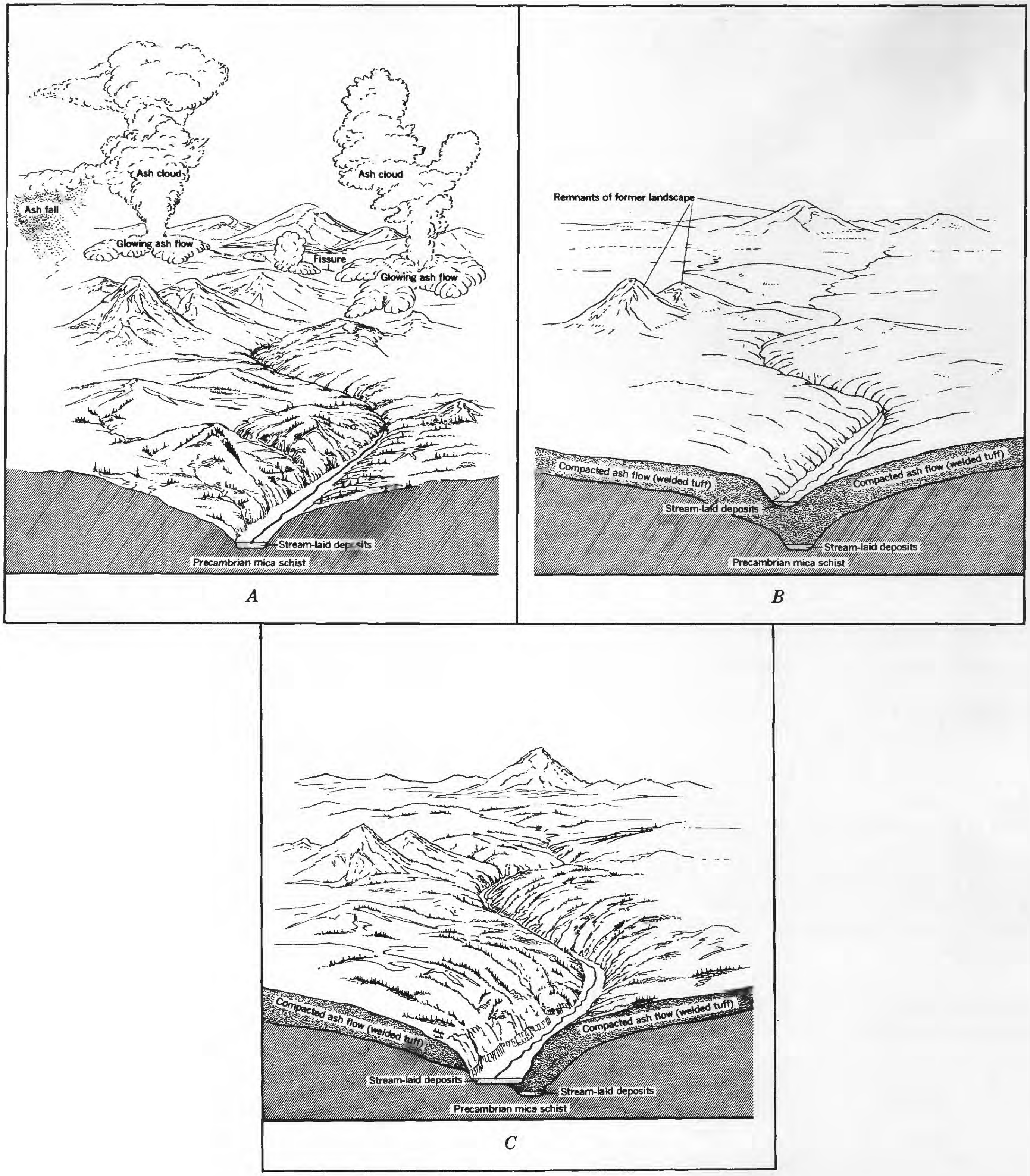
Conclusive evidence is not available to permit a meaningful choice between these two alternatives. The wide distribution of the small tuff deposits and the absence of any definitive fractures or fissures near such exposures makes me favor the view that these localized deposits are merely remnants of a once widespread blanket of welded tuff.

In the eastern third of the area, the present valleys, veneered with welded tuff, probably reflect former stream courses. Likely, compaction was greatest wherever the flows were thickest, as in the valleys. After compaction the valleys probably appeared as broad connected swales of low relief, the whole serving as a poorly integrated drainage system for the streams which were even then beginning to form. Although the coincidence between present and ancestral drainage patterns is close, it is not exact. So locally, as near mile post 32 (labeled MP 32 on pl. 1) along the Yellowstone Park boundary, the stream banks of modern Grayling Creek are formed in part by Precambrian rocks which are interpreted as segments of the exhumed valley walls of ancestral Grayling Creek (figs. 11 and 12). The schist was deeply weathered during a long period of subaerial exposure prior to its burial by a layer of volcanic ash which settled out of the air shortly after the initial explosive volcanic phase. Even as the ash fragments settled they were overridden by a fast-moving incandescent ash flow whose rapidly chilled base is represented by the black glassy vitric zone and whose upper part is represented by the devitrified tuff.

The Yellowstone Tuff is uniform in appearance throughout the area. Characteristically the flows have a streaky foliation (eutaxitic structure) that probably stems from differential compaction of the shards and gas-filled vesicles. This eutaxitic structure is widespread and in places very well developed, appearing almost like continuous flow lines; elsewhere it is crude and difficult to perceive. Where this foliation is well developed the flow can be divided into layers 1–2 inches thick that give it a platy or thin-bedded appearance. Where the lines of foliation are not as closely spaced, the flow is subdivided into layers as much as 6 feet thick, which commonly weather as rounded ledges. This excellent layering closely resembles bedding but apparently is the result of varying degrees of consolidation due to lithostatic load and normally is confined to the basal parts of a flow (Ross and Smith, 1961, p. 20–21).

In general, these foliation lines appear to conform to the buried land surface. Along Grayling Creek, for example, especially near mile post 32 (SE¼ sec. 34, T. 11 S., R. 5 E.) and at the right-angle southward bend of the stream (near MP 270 on the Montana-Wyoming border), the foliation clearly dips away from buried Precambrian metamorphics (fig. 12). This same pattern is repeated on Little Tepee Creek (sec. 29, T. 11 S., R. 5 E.) where the flows abut Precambrian am-

FIGURE 11.—Development of modern Grayling Creek. *A*, During either late Pliocene or early Pleistocene time, eruptions from widely spread fissures in Yellowstone National Park covered a dissected hilly landscape. The first stage in the eruption was a violent explosion which threw great clouds of volcanic ash high into the air. Simultaneously other larger fragments and gas foamed out of the fissures and spread for miles as incandescent ash flows. *B*, In time the former landscape was buried by the flows; only the higher summits showed above a plain in which low sags marked former valleys. One such sag closely paralleled the former course of Grayling Creek. As rains began to run off the new landscape, the sags served as drainageways. *C*, Subsequently the new valley was cut down through the ash to the ancient landscape. The west (left) wall of modern Grayling Creek is formed by the exhumed Precambrian mica schist overlain by welded tuff; the east wall shows only the welded tuff. With continued dissection much more of the ancient valley will be exhumed.


Ash cloud
Ash cloud
Ash fall
Glowing ash flow
Fissure
Glowing ash flow
Stream-laid deposits
Precambrian mica schist
Remnants of former landscape
Compacted ash flow (welded tuff)
Compacted ash flow (welded tuff)
Stream-laid deposits
Stream-laid deposits
Precambrian mica schist
Compacted ash flow (welded tuff)
Compacted ash flow (welded tuff)
Stream-laid deposits
Stream-laid deposits
Precambrian mica schist

A *B* *C*

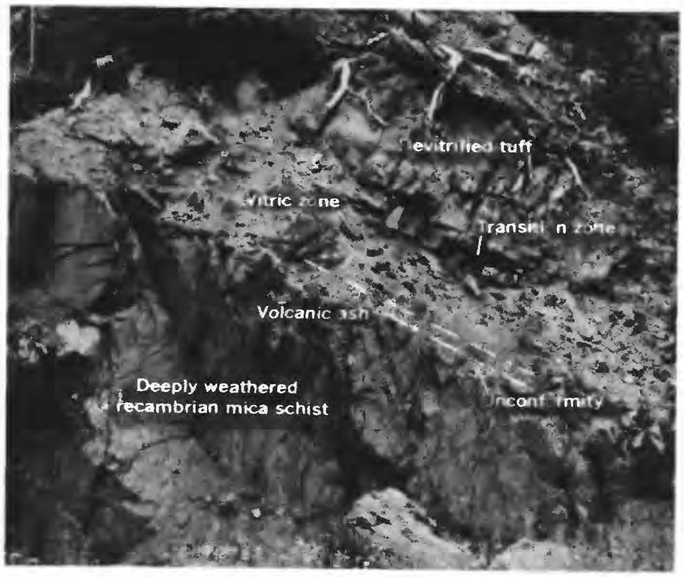


FIGURE 12.—Welded tuff exposure at Grayling Creek. Photograph shows one of several localities where erosion by modern Grayling Creek has exposed Precambrian mica schist overlain by both volcanic ash and welded tuff. The schist, intensely jointed and deeply weathered, is part of a former valley wall buried beneath the ash falls and flows and now partly exhumed.

phibolites (pl. 1). Here and there the flows form broad domelike masses several hundred feet to several hundred yards in diameter, with 20–30 feet of closure. Although these cannot be clearly related to buried topography, they may represent minor irregularities on the surface of an older flow.

In many exposures the Yellowstone Tuff is light brown (about 10*YR* 6/2) or medium brown (about 10*YR* 5/4), although it generally ranges in color from light gray (*N* 7) to medium dark gray (*N* 4); locally it has a pale-red cast. Commonly the tuff is dense and compact and weathers readily to form rounded nodules 1–2 inches in diameter which mantle the surface as a thin scree.

The complete lack of sorting is the most characteristic feature of the tuff. Most exposures are marked by gross variations in texture. Xenoliths, chiefly of andesitic debris, are dark gray and angular and are several inches long on a side. They are irregularly distributed throughout the basal part of the tuff. I estimate that they make up less than 1 percent of the flows. Presumably they are detritus of older andesitic deposits picked up by the overriding ash flows and incorporated with little or no resorption.

Commonly, deposits representative of both ash-fall and ash-flow tuff can be recognized wherever the base of the lowest flow is exposed. The ash fall normally is directly below the base of the flow and appears as a dull olive-gray (about 5*Y* 4/1) earthy bed which consists chiefly of weakly consolidated volcanic ash (*A*, *B*, fig. 13). This is overlain by dark-gray (about *N* 3) to black (*N* 1) unaltered glass or vitrophyre, which forms the base of the flow proper (*C*, *D*, fig. 13). The vitrophyre grades upward into a transitional unit which consists of alternating layers of dark glass that has begun to devitrify and finely crystalline friable tuff marked by streaky foliation. This unit in turn grades imperceptibly into the overlying devitrified tuff which makes up the bulk of the flow (*E*, *F*, fig. 13). Such a sequence is unusually well exposed in the roadcut along U.S. Highway 191, near the sharp, right-angle southward bend of Grayling Creek where the volcanic material overlies the Precambrian mica schist (table 7).

Mineralogically, the Yellowstone Tuff in the Tepee Creek quadrangle, whether as isolated outcrops or as part of the main body of pyroclastic flows, appears identical with the Yellowstone Tuff of Yellowstone National Park as described by Boyd (1961). The tuff of this quadrangle is composed mainly of differentially compacted shards and scattered phenocrysts of clear sanidine ($2V_x=30°–44°$, $n_x=1.520–1.524$, $n_y=1.524–1.528$, $n_z=1.526–1.529$; all index determinations±0.001), rounded grains of quartz, fragments of plagioclase (chiefly oligoclase, An_{23}), grains of magnetite, and fragments of pumice. Also included, but invariably in sparse amounts,

TABLE 7.—*Section of Yellowstone Tuff exposed in roadcut along U.S. Highway 191 near mile post 270 on the State line*

[Figure 13 shows photographs of representative samples of rock types]

Rock type	Thickness	Characteristics
Devitrified tuff	Uncertain; possibly as much as 500 ft.	Gray, dull, lusterless, lithoidal; strong eutaxitic structure. Shards, vesicles, and pumice fragments flattened and elongated; shards are 10–20 times as long as they are thick. Many phenocrysts of sanidine, quartz, oligoclase (An_{10-20}), pyroxene, olivine, and magnetite; locally phenocrysts so plentiful that rock has speckled appearance. In some flows sanidine phenocrysts most common; in others quartz phenocrysts predominate. Pumice fragments somewhat altered to form intergrowth of tridymite(?) and feldspar chiefly in center of fragments.
Transitional zone	Variable; generally 1–4 ft.	Gray; irregular interlayers of vitrified and devitrified tuff. Strong streaky foliation.
Vitric zone	2–3 ft	Dark gray to black; glassy; many light- to dark-gray minute angular phenocrysts. Extremely friable; disintegrates into fine to coarse unconsolidated sand. Shards faintly compacted; crudely alined to form discontinuous nearly parallel laminae broken locally by randomly distributed phenocrysts and pumice fragments. Most shards show typical Y and U shapes; shards invariably wrapped around phenocrysts and pumice fragments. Typical phenocrysts include sanidine, quartz, oligoclase (An_{10-20}), and few grains of pyroxene, and olivine.
Basal ash	1½–2 in	Olive gray, finely laminated, weakly consolidated; abundant phenocrysts, mostly of sanidine, quartz, and oligoclase (An_{10-20}). Shards show little to no compaction; typically Y- or U-shaped. Pumice fragments not compacted. Overlies Precambrian mica schist with high angular discordance.

are grains of clinopyroxene (ferroaugite, $2V_z=56°$, $n_x=1.725$, $n_y=1.732$, $n_z=1.757$; hedenbergite, $2V_z=58°–62°$, $n_x=1.724$, $n_y=1.733$, $n_z=1.753$; all index determinations ±0.001), olivine (fayalite, $2V_x=50°$, $n_x=1.820$, $n_y=1.860$, $n_z=1.880$; ferrohortonolite, $2V_x=56°$, $n_x=1.811$, $n_y=1.847$, $n_z=1.856$), biotite, hornblende, and zircon. A few grains of chevkinite were also noted.

In some tuffs the sanidine phenocrysts are most common, followed in order of abundance by oligoclase, quartz, pyroxene, and fayalite. In others the quartz phenocrysts are the most abundant. Many of the fayalite grains are rimmed by a reaction zone of dark-red material tentatively identified as iddingsite.

The shards of the ash fall are but slightly compacted and are commonly Y- and U-shaped. Normally they are nearly alined except where broken by the scattered phenocrysts and uncompacted pumice fragments.

The shards of the devitrified tuffs are intensely compacted, so much so that in hand specimen the tuff has a strong eutaxitic structure. The lithoidal fine-grained groundmass is broken here and there by flattened vesicles and by a myriad of white, colorless, light-yellow, and black phenocrysts representing, respectively, feldspar, quartz, pyroxene, olivine, and magnetite. In places these phenocrysts are so plentiful that the rock has a speckled appearance and looks almost coarse grained. Elsewhere the phenocrysts are less abundant, and the rock is dull and lusterless—the appearance of the groundmass.

Although some of the pumice fragments are very small, most are about 1 mm long and 0.50 mm wide, and a few are as much as 3 mm long and 0.50 mm wide. All, no matter what their size, have undergone some replacement; the center of the fragments, contain an intergrowth of tridymite(?) and feldspar (fig. 13*F*). In places these intergrowths are comb shaped; elsewhere the minerals form a radiating, spherulitic intergrowth that partly, but not wholly, masks the fibrous structure of the pumice fragments. The phenocrysts are fractured and pulled apart and show little or no orientation. Without exception the shards are wrapped around the phenocrysts and the altered pumice fragments.

Most of the rhyolites in this quadrangle are welded tuffs, but a few may have formed as normal lava flows. These few have an outward appearance, both on outcrop and in hand specimen, somewhat like that of the welded tuffs, but they differ drastically in thin section. The groundmass of these rocks consists chiefly of minute grains (0.02 mm) of feldspar and quartz, which form a tightly intergrown anhedral-granular texture in which phenocrysts of the same minerals are irregularly scattered. Many of the same minerals are in these flows as are in the welded tuffs; this close similarity is even more apparent when their chemical analyses are compared (sample 31, table 6).

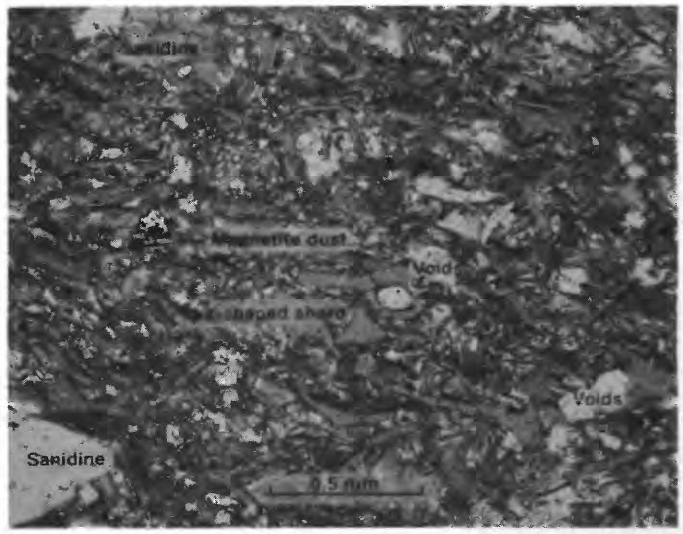
The welded tuffs, whether they occur as isolated outcrops or as parts of a widespread ash flow, are monotonously alike chemically. This similarity is shown in table 6, and in figure 14; analyses and plots of samples from isolated deposits (p. 45) are indistinguishable from analyses and plots of those samples from the main mass of the tuff (fig. 14).

This accordance in mineralogical and chemical composition carries over to different parts of a flow, and even to different flows. A comparison of analyses of samples from the vitric zone (Nos. 15 and 16, table 6) with those of devitrified tuffs shows almost identical relations, and from the chemical analyses one cannot determine whether the rock is a black shiny glass or a dull gray stony devitrified tuff. The rhyolitic welded tuffs that form the south end of the Madison Range have been dated as Oligocene (Hamilton and Leopold, 1962), and are, therefore, presumably older than the flows in this quadrangle (p. 53). Yet an analysis of a sample (sample 30, table 6) collected from these older rocks (Hamilton and Leopold, 1962, p. B28) is almost identical with analyses of the Yellowstone Tuff. This close similarity seems most unusual, for it implies that rhyolitic flows widely separated in time may still be almost identical in composition.

In gross composition the tuffs are high in silica, averaging almost 75 percent; the range is from 73.4 to 75.9 percent (table 6). They are also high in alkalis (fig. 14); the potash averages 4.9 percent and ranges from 4.6 to 5.2 percent; the soda averages 3.7 percent and ranges from 3.4 to 3.8 percent. As might be expected, the tuffs are inordinately low in magnesia, which averages 0.18 percent and ranges from 0.03 to 0.45 percent; lime is also low averaging 0.47 percent and ranging from 0.29 to 0.58 percent. Compared with Daly's (1933, p. 9) average rhyolite, these tuffs are higher in silica, lower in magnesia and lime, and comparable in alkalis. Much the same is also true when these tuffs are compared with Daly's "rhyolite of Yellowstone Park" (1933, p. 26).

In both the intrusive rocks and the rhyolitic welded tuffs, the C.I.P.W. norm classification (tables 5 and 6) patently shows that without exception the salic minerals (such as normative quartz and feldspars) far exceed the femic minerals (such as normative hypersthene, magnetite, ilmenite). If the salic minerals are compared, the normative feldspars invariably dominate the normative quartz in the intrusive rocks; but in the rhyolite tuffs, the normative quartz either equals or dominates the normative feldspars.

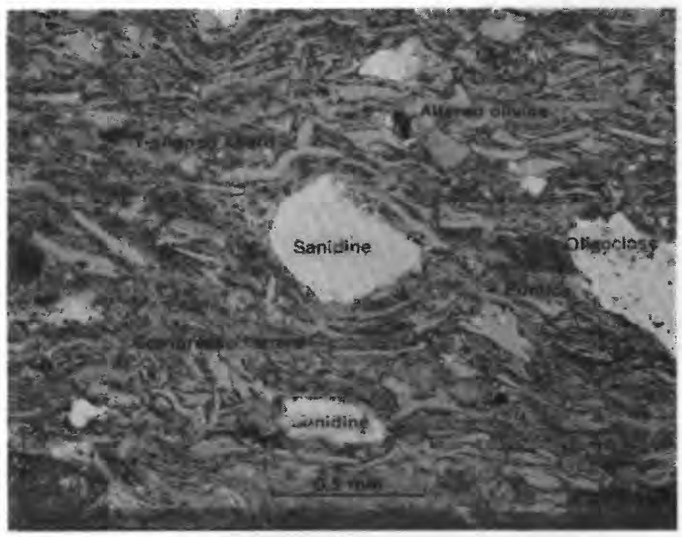
A


Magnetite dust
Voids
Voids
Sanidine
0.5 mm

B

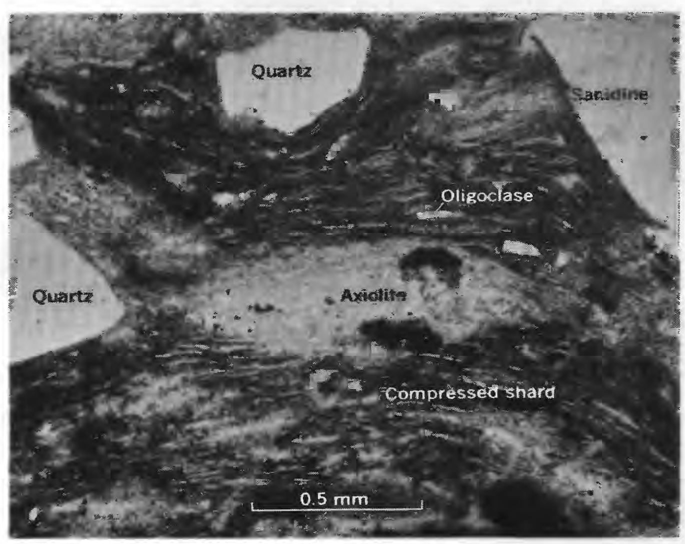
2 cm

C


Sanidine
0.5 mm

D

E


Quartz
Sanidine
Oligoclase
Quartz
Axiolite
Compressed shard
0.5 mm

F

If the oxide molecules, rather than mineral molecules, are contrasted, the salic alkalis are extreme over the salic lime in the tuffs, but are much less so in the intrusive rocks. A comparison of salic potash and salic soda shows them to be equally abundant in the tuffs, whereas the salic soda dominates the salic potash in the intrusives.

In general, the three igneous rock types in this quadrangle the intrusive dacite porphyry, the extrusive shoshonite–andesite breccia complex, and the pyroclastic rhyolite tuffs seem to be but isolated representatives of a much more elaborate sequence of igneous rocks exposed to the east in Yellowstone National Park. This reflects the position of the quadrangle on the western extremity of the volcanic pile which mantles the park. Only a few members of the whole sequence reach this far west, and these, although showing some similarities, show many more dissimilarities. It is much as if three nonconsecutive units of some 10–15 units were available for study; one gains but a glimpse of the petrologic possibilities.

The dacite porphyry intrusions in this quadrangle are probably correlative in age with the earlier dacitic intrusions exposed near Gardiner, Mont. (p. 44), and considered by G. D. Fraser (oral commun., 1965) to be the intrusive equivalent of the early acid breccia. The estimates of the early acid breccia's age range from late early Eocene to early middle Eocene (Dorf, 1960, p. 258). The shoshonite–andesite breccia complex is probably part of the huge pile of early basic breccia, so widespread in the park, and dated as early middle Eocene (Dorf, 1960, p. 259). Hence, the dacite porphyry intrusions and the shoshonite-andesite breccia complex are closely related in time. By contrast, the Yellowstone Tuff of this area, clearly the western extension of identical rocks which cover almost all the northwest corner of the park, is very much younger; all the intervening units are either concealed or never reached this far west.

FIGURE 13.—Rock chips and representative thin sections characteristic of the Yellowstone Tuff. Photographs by R. B. Taylor, U.S. Geological Survey. *A*, Rock chip from a dull-olive-gray earthy weakly consolidated volcanic ash; *B*, Thin section of the volcanic ash in which Y- and U-shaped shards are whole and uncompacted; *C*, Rock chips from the dark-gray to black vitrophyre which is speckled by many light-colored phenocrysts of feldspar and quartz; *D*, Thin section of the vitrophyre showing moderately compacted shards and somewhat alined phenocrysts; *E*, Rock chip of devitrified tuff marked by faint eutaxitic structure; *F*, Thin section of the devitrified tuff in which the shards are extensively compacted and warped around moderately alined phenocrysts. Axiolite is filled with mixture of feldspar and tridymite(?).

In view of the above it seems unreasonable to expect any close similarities in the igneous rocks exposed in this quadrangle.

AGE OF THE TUFFS

The Yellowstone Tuff overlies the shoshonite flows (p. 44) which in turn rest on strata warped and deformed as a result of the emplacement of the Gallatin River laccolith. The tuffs, therefore, are the youngest igneous rocks exposed, but firm data for a specific age assignment are lacking. Boyd (1961, p. 410) considered them to be middle Pliocene, chiefly because they are similar to welded rhyolite tuffs and flows which intertongue with the dated Teewinot Formation south of Yellowstone National Park (Love, 1956, p. 91). The middle Pliocene age of the Teewinot Formation has been confirmed by a potassium-argon date—9.2 million years B.P.—obtained from obsidian from that formation (Evernden and others, 1964, p. 164, 185).

G. D. Fraser, however, suggested (oral commun., 1965) that the Yellowstone Tuff may be as young as late Pliocene or early Pleistocene. His suggestion is based on the assumption that the Yellowstone Tuff and correlative rocks are the youngest tuffs exposed in and near Yellowstone National Park. Thus, the Yellowstone Tuff is the youngest tuff exposed in the stratigraphic sequence of flows and tuffs mapped by Boyd (1961) in the northwest corner of the park. South of the park, however, the youngest tuffs are not in the Teewinot Formation, but are in the overlying Bivouac Formation, of uncertain age but tentatively dated as late Pliocene by Love (1956, p. 93). I concur with Fraser and suggest, therefore, that the Yellowstone Tuff is either of late Pliocene or early Pleistocene age, probably not both, for most likely the tuffs were deposited in a very short time, during a single rapid series of eruptions (Boyd, 1961, p. 417).

South of the Tepee Creek quadrangle, both in the West Yellowstone basin and at the south end of the Madison Range, fossil pollen spores associated with welded rhyolite tuffs have been dated as Oligocene (Hamilton and Leopold, 1962). An obsidian-rhyolite flow along the south flank of the West Yellowstone basin has been dated as late Quaternary (Richmond and Hamilton, 1960). It seems likely, therefore, that rhyolitic tuffs and flows of at least three ages occur in the general area near West Yellowstone: rhyolite tuffs of Oligocene age; widespread ash flow tuffs of either very late Pliocene or early Pleistocene age; and rhyolitic flows of late Quaternary age, which are confined to local valleys.

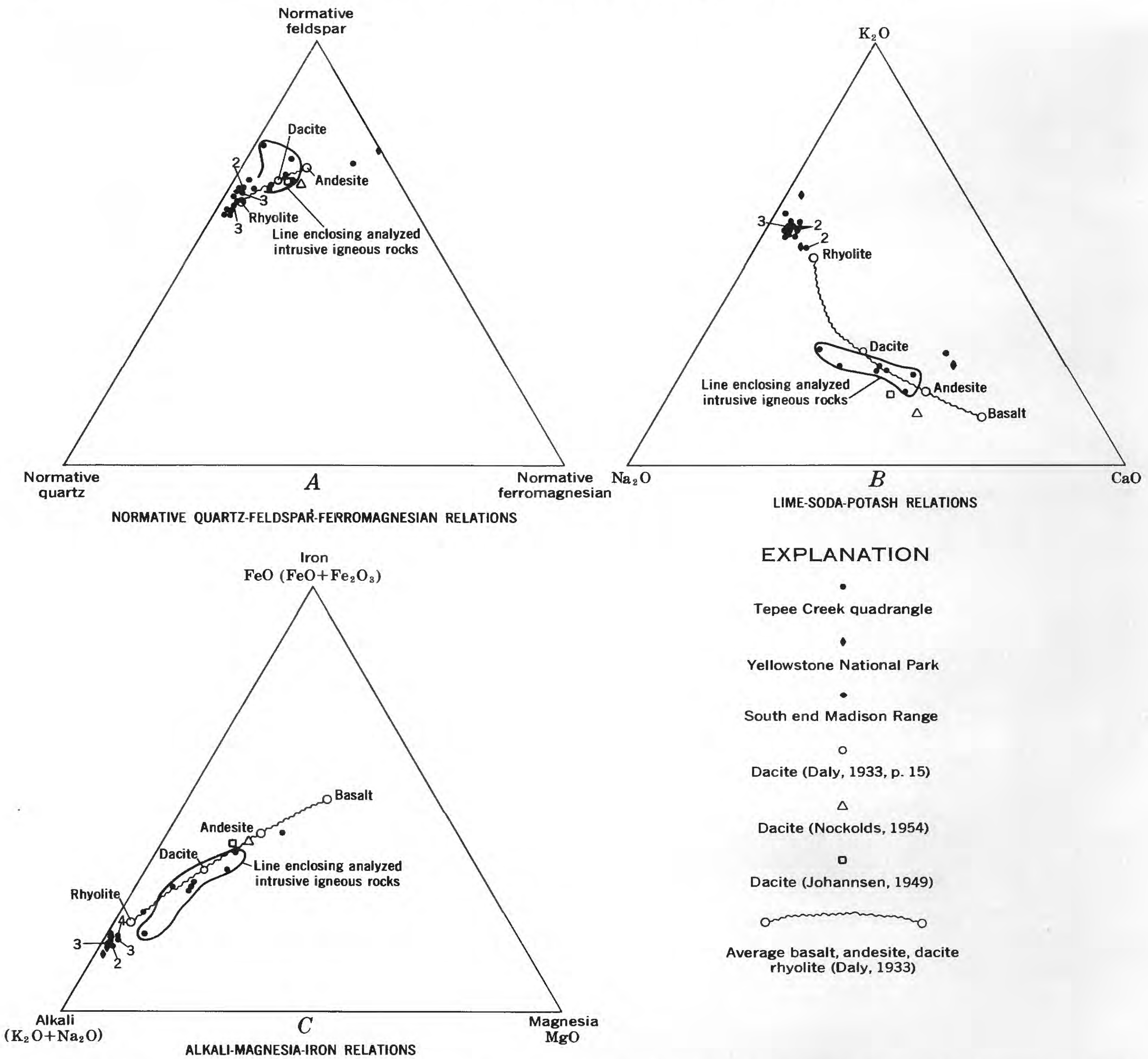


FIGURE 14.—Ternary diagrams showing relations between intrusive and extrusive rocks of the Tepee Creek quadrangle. See tables 5 and 6 for analyzed samples.

SURFICIAL DEPOSITS

PLEISTOCENE DEPOSITS

During at least three separate episodes this quadrangle was covered in whole or in part by glaciers. The morainic debris, both unsorted till and sorted glaciofluvial deposits left behind when these ice masses melted, mantles much of the countryside, but it was mapped only where the deposits are both thick and widespread (pl. 1).

The oldest glacial deposits in the area, possibly representing more than one ice advance, are assigned to pre-Bull Lake age. Most, if not all, of these deposits are likely of pre-Wisconsin, possibly Illinoian age, and most were probably deposited by one ice sheet. This sheet, covering hundreds of square miles and thick enough to override the highest peaks in the area, probably originated in the Absaroka Range and spread westward across the Yellowstone Plateau, perhaps as far as the crest of the Madison Range (Hall, 1959).

The second ice sheet, which was not as widespread, extended westward only into the eastern part of the quadrangle, where its thin distal edge, represented by two broad lobes of ice, filled valleys and locally moved around or ground across low topographic barriers (fig. 15). This ice probably invaded the area either in early Wisconsin time or just in advance of the Wisconsin, and its deposits are assigned to Blackwelder's (1915) Bull Lake stage or to Richmond's (1960) Bull Lake Glaciation.

The third episode of glaciation was the least widespread, for the ice was confined chiefly to valleys (fig. 15). Deposits from this glaciation, assigned to Blackwelder's (1915) Pinedale stage (p. 58) or to Richmond's (1960) Pinedale Glaciation, are plastered along valley walls and clog the mouths of tributary streams. This episode is provisionally dated as late Wisconsin.

PRE-BULL LAKE GLACIATION

Near the Narrows, at the north end of Horse Butte (pl. 1), a small earth slump resulting from the Hebgen Lake earthquake of 1959 exposed a deposit of old till capped by a deeply weathered soil in turn overlain by lacustrine silts. The silts, which are somewhat contorted, possibly as a result of being overridden by younger ice, underlie till of Bull Lake age. The lowermost till, assigned a pre-Bull Lake age by Richmond (1964) chiefly on the basis of its stratigraphic position, is light gray to light brown, sandy, and much like the overlying Bull Lake Till which mantles the flanks of Horse Butte. This is the only till of pre-Bull Lake age observed in the report area. Here and there along the crest of Horse Butte are glacial erratics of gneiss, rhyolite, and basalt; these were probably also deposited by pre-Bull Lake ice, as it was apparently the only ice ever to override Horse Butte completely.

Perhaps the most convincing evidence of the extent and thickness of pre-Bull Lake ice is the many glacial erratics scattered throughout the area. Most of these are isolated angular to subround boulders and fragments of metamorphic or volcanic rock perched on some of the high benches and divides which flank the highest peaks in the area. Thus, glacial erratics are on the flanks of the divide (altitude 9,200–9,500 ft) between Bacon Rind Creek and Upper Tepee basin, on the welded tuffs which floor Rathbone Lake basin (altitude 7,600 ft), and on the isolated hills of metamorphic rock (altitude 8,500 ft) which rise above the basin floor. Neither erratics nor any other indications of overriding glaciers were found on the highest ridges, such as Kirkwood Ridge or Skyline Ridge, to support Hall's (1959) concept that these were once under ice. Their absence, however, does not detract from the hypothesis, for most of the ridge crests are almost knifelike, and any morainic deposits would long since have been removed by erosion; nor would glacial striae last in the relatively soft sedimentary rocks which form the ridges.

BULL LAKE GLACIATION

The Bull Lake Glaciation is represented by two areas veneered with morainic deposits: one along the south edge of the quadrangle and the other along part of the east edge (fig. 15). As the Bull Lake ice sheet flowed westward from the Absaroka Range its distal margin, marked by fingerlike lobes of ice, moved into and down available valleys. One lobe, here called the Horse Butte lobe, spread across the east and southeast edges of the West Yellowstone basin and reached at least as far west as Horse Butte, where it split into two subordinate lobes, one on each side of the butte. The crest of the butte was not covered, but the remainder of the basin was under ice.

A kame terrace was formed between this ice lobe and the north flank of Horse Butte, and this deposit, marked by small sand and gravel hummocks, stands higher (altitude 6,760 ft) than the till deposited by the ice along the terrace flanks (altitude 6,560 ft). Comparable ice-contact deposits, not so well developed however, occur on the south flank of Horse Butte.

The till deposited by the Horse Butte lobe near Grayling Arm (of Hebgen Lake) is moderately well exposed along the north flank of Horse Butte and near the mouths of Grayling and Red Canyon Creeks. Commonly it is light gray to light brown, sandy, and wholly unsorted. The sandy matrix of the till along the north edge of Hebgen Lake suggests modification of a clayey till by outwash waters containing much sand and gravel.

Another lobe, the Grayling lobe, may have extended down the Gallatin River valley, crested the low divide (altitude 7,700 ft) between Fan Creek and the Gallatin River, and reached as far west as the unnamed hill west of Divide Lake (pl. 1). The south margin of this lobe probably flowed into and choked the north-trending valley of the ancestral Grayling Creek, thereupon damming the creek and diverting it southward into a new course (fig. 16).

Grayling Creek probably flowed northward tributary to the Gallatin River before the advance of the Grayling lobe of the Bull Lake ice (fig. 16*A*).

As the Grayling lobe spread into this area, the headwaters of Grayling Creek, the Gallatin River, and Fan Creek were overridden and clogged. The distal edge of the lobe flowed into the north-trending segment of the valley of Grayling Creek but was unable to surmount the west valley wall, formed in part by a large hill. In the subsequent minor readvances and withdrawals of the ice, a somewhat modified terminal moraine was developed in this part of the valley (fig. 16*B*).

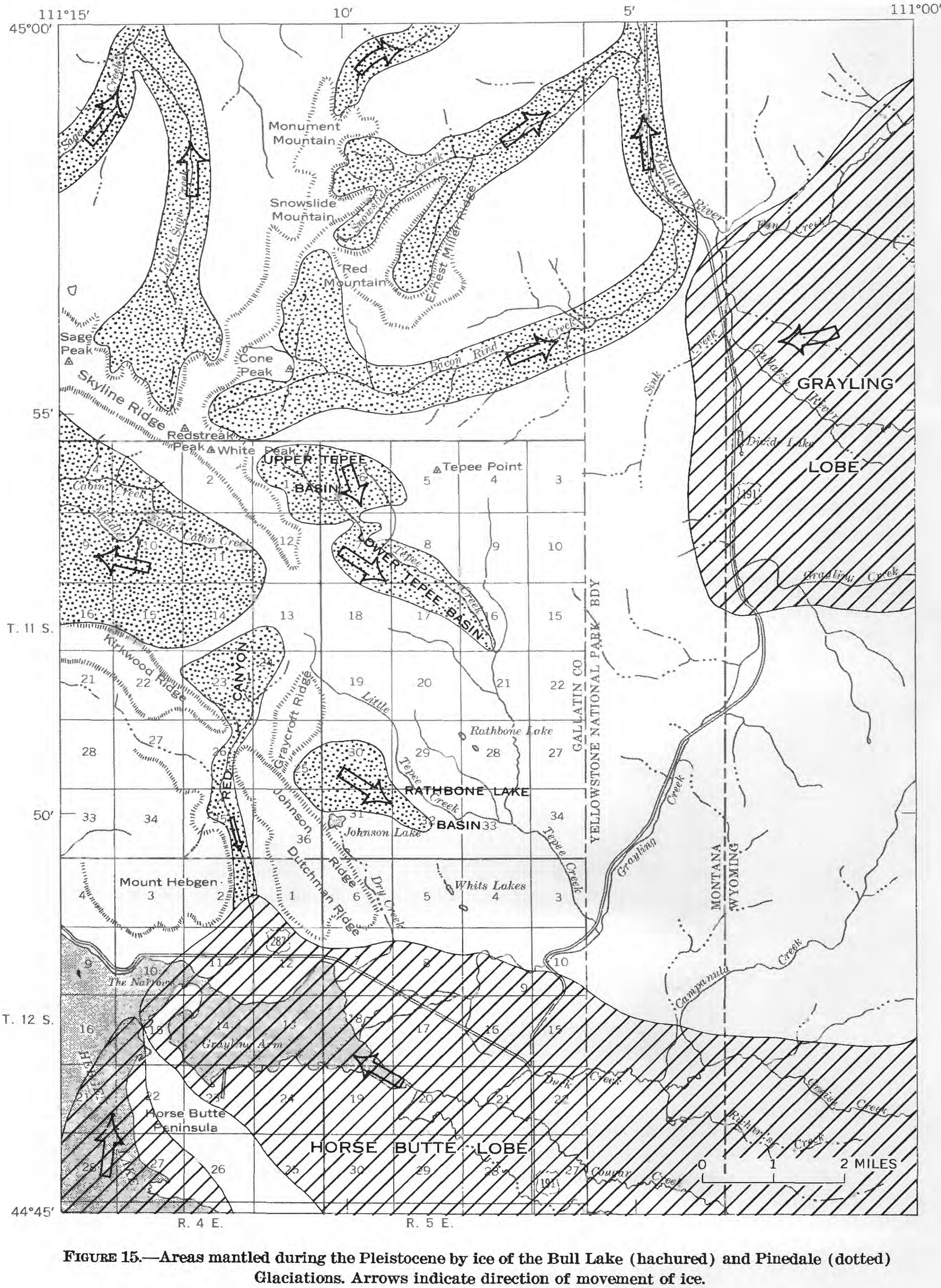


Figure 15.—Areas mantled during the Pleistocene by ice of the Bull Lake (hachured) and Pinedale (dotted) Glaciations. Arrows indicate direction of movement of ice.

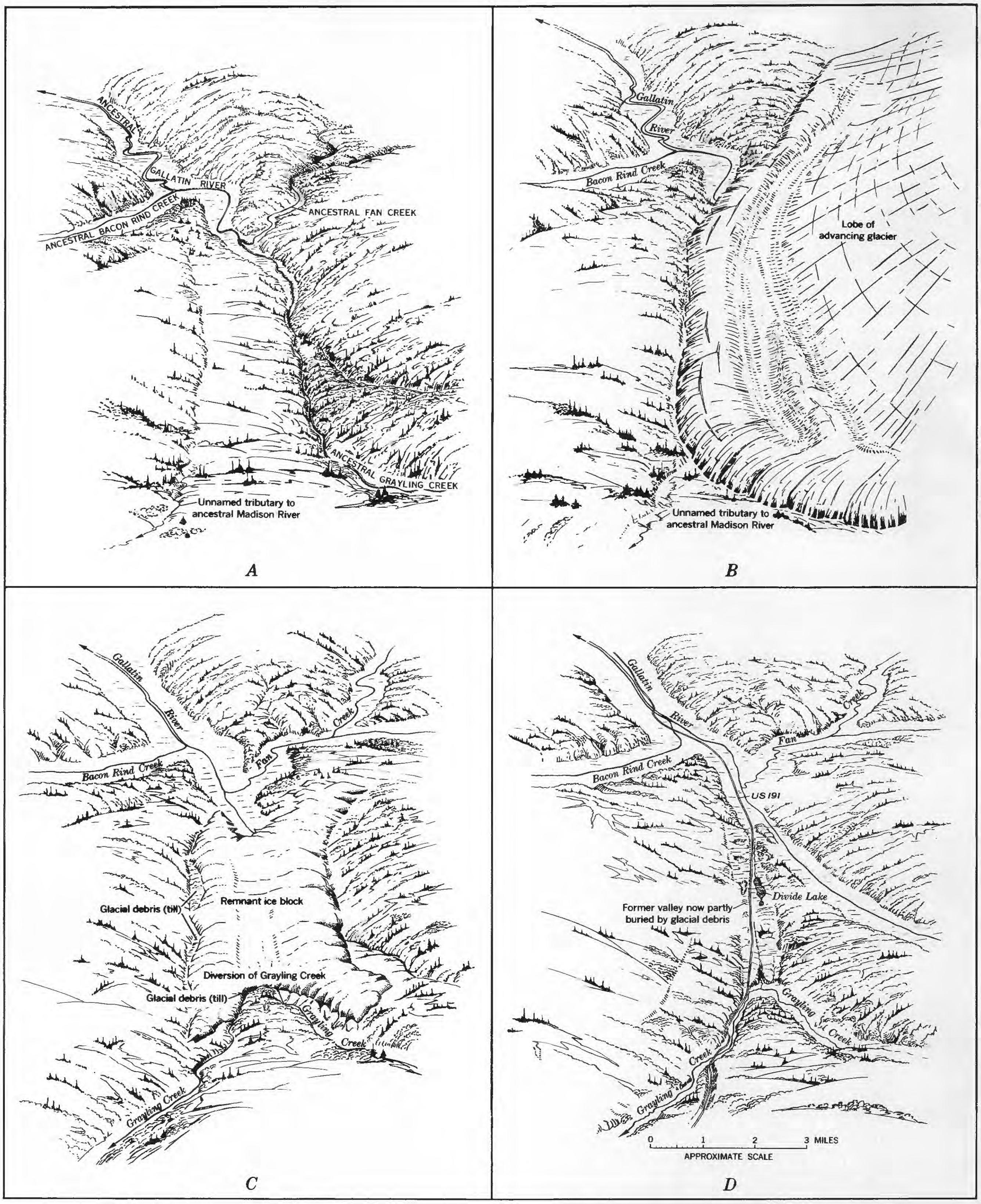


FIGURE 16.—Explanation of the southward diversion of Grayling Creek. *A*, Before the advance of the Grayling lobe of the Bull Lake ice; *B*, During the advance of the Grayling lobe; *C*, During the withdrawal of the Grayling lobe; *D*, At present time.

During the melting and consequent withdrawal of this ice lobe, a large ice block may have been left in the valley. As the ice withdrew, melt waters reoccupied the newly opened former drainage courses, but on Grayling Creek the morainic debris and the remnant ice block effectively dammed the stream. Probably the melt waters ponded until they crested the low divide and then emptied into a preexisting south-trending valley. Within a short time thereafter, the waters had entrenched themselves in their new southward course, and the Gallatin River had lost a major tributary as the Madison River had gained one (fig. 16*C*).

The Gallatin River–Grayling Creek area as it appears now is shown in figure 16*D*. North of the southward bend of Grayling Creek the former valley is floored with almost unmodified glacial debris. Divide Lake, near the midpoint of the valley, is probably a large kettle lake only recently breached along its north edge. This drift-choked valley, about half a mile wide, is comparable in width to the valley of Grayling Creek east of the southward bend, but contrasts sharply with its width, about a quarter of a mile, south of the bend.

Glacial debris of the Grayling lobe veneers the valley of Fan Creek, the upper reaches of both the Gallatin River and Grayling Creek, and the divides between these drainageways. The till extends down to, and passes below, the alluvium flooring these streams, suggesting that the ice for the most part merely followed preexisting valleys.

Probably local glaciers also formed in most of the major valleys surrounding the higher peaks in the western part of the area. As yet, however, no deposits directly attributable to these glaciers have been found in this quadrangle. Whatever drift may have been deposited was likely removed when younger glaciers of Pinedale age scoured these valleys. West of this area moraines of Bull Lake Glaciation are in many of the canyons of the Madison Range (Richmond, 1964, p. 226).

The surface formed on the Bull Lake Till is characteristic of an old till. Most signs of youthful morainic deposits have been somewhat modified; the former knob-and-kettle topography now has a poorly integrated drainage system and is marked by a hummocky surface mantled here and there by a few scattered glacial erratics. The former kettles have been breached, and a few are connected to form uninterrupted drainageways. The till slopes are free of knobs and irregular hummocks. A mature soil 18–24 inches thick has been formed on the till, and in places the soil is veneered with eolian silt (Richmond, 1964).

It seems likely that as the Horse Butte lobe of the Bull Lake ice melted and gradually withdraw eastward, its melt waters flowed westward and gradually filled the West Yellowstone basin with outwash sand and gravel. The outwash laps onto Horse Butte, partly covering the till and kame-terrace deposits formed along the north flank of the butte, and similarly overlaps the till that locally forms the north bank of Hebgen Lake. The westward extent of this outwash is uncertain, for Hebgen Lake now overlies the former outwash plain. Probably the outwash reached into the northwest arm of Hebgen Lake, the entryway to the narrow Madison Canyon. The outwash forms an undulatory plain of low relief, interrupted here and there by small streams along the east part of the West Yellowstone basin. The plain extends unbroken into the mouths of streams such as Gneiss and Cougar Creeks (pl. 1) which cut the west flank of the Yellowstone Plateau, a clear indication that each of these streams carried outwash-charged melt waters.

The outwash ranges in color from dark gray to black. It is unconsolidated and consists chiefly of angular to subround grains of obsidian (vitrophyre) and fragments of devitrified tuff. Also included are rounded to angular grains of sanidine, quartz, pyroxene, and olivine. For much of its extent the outwash is capped by a thin veneer of eolian silt 10–18 inches thick.

Richmond (1964, p. 227) suggested on the basis of drill records that the fill of obsidian sand and gravel is about 90 feet thick and that it rests on lacustrine silts.

The outwash plain in the mapped area shows no evidence, either by being covered by younger till or by having distorted internal bedding, of being overridden by ice, and it is assumed that Pinedale glaciers never reached that part of the outwash plain in the mapped area. Westward, however, at the head of Madison Canyon the obsidian plain is overlain by a lateral moraine which Richmond (1964, p. 232) attributed to ice of Pinedale age.

PINEDALE GLACIATION

Morainic deposits assigned to the Pinedale Glaciation are confined almost wholly to the areas of high altitude in the western part of the quadrangle (fig. 15). These deposits include unmodified till that has been plastered along valley sides to form thin lateral moraines, unsorted drift in the form of low terminal moraines at canyon mouths, and glaciofluvial debris flooring many of the stream valleys both as outwash and as ice-contact deposits.

The Pinedale Till is dark brown, clayey, and massive. It is wholly unsorted and is mostly filled with fragments and boulders of consolidated claystone, shale, siltstone, and sandstone.

During the maximum Pinedale advance nearly all the northwestern part of the area was covered by ice; only the higher peaks and ridges rose above the ice cap (fig. 15). The broad pattern of glacier movement during this stage appears to have been radial from the high point of Redstreak Peak; the glaciers north of Skyline Ridge moved northward or eastward, those south of the ridge moved westward or southward.

The glaciers spread down the valleys from the cirques and halted at or near the valley mouths. Possibly those glaciers that moved down Bacon Rind and Snowslide Creeks spread into the Gallatin River valley, where they may have coalesced to form a small local glacier which filled part of that valley.

Moraines of Pinedale age are best preserved in Red Canyon and in Upper and Lower Tepee Basins (fig. 15). In Red Canyon the morainic material still marked by knob-and-kettle topography, appears almost the way one imagines it must have looked when the ice melted. The knob-and-kettle topography is mantled by angular to subround glacial erratics of all sizes, and most of the kettles show almost no sign of erosion; only one or two are breached. In the Tepee Creek Basins the morainic irregularities, where not buried by outwash, are emphasized by sinuous eskers and conelike kames; only a few of the larger and more continuous eskers are shown on plate 1.

At Bacon Rind and Snowslide Creeks the narrowness of the canyons has resulted in most of the morainic debris being dissected and removed by the streams, although patches of till, masked somewhat by colluvium, are along the valley walls. At Sage and Little Sage Creeks extensive morainic deposits are along the valley walls. Only patches of till and scattered erratics were observed at Cabin Creek.

South of Johnson Lake (pl. 1) the huge lobe-shaped Grayling earthflow (p. 61) is formed of till deposited by an ice mass which once occupied the west corner of the Rathbone Lake basin. I suggest that the drift, thoroughly saturated by melt water, flowed southward down the valley of Dry Wash to spread out as a lobate fan at the mouth of the creek. The debris, identical in appearance with morainic debris elsewhere in the area (with the possible exception of being somewhat enriched in volcanic detritus), is dark brown, massive, clayey, and wholly unsorted and contains angular to subround fragments and boulders of all sizes and compositions. The topography is hummocky but lacks the distinctive knobs and kettles so common in the fresh moraines.

The wide variety of glacial erratics is impressive as regards both composition and size. One of the largest is an immense boulder of Kootenai sandstone, about 20 by 20 by 10 feet, that was carried by the ice into Red Canyon and then wedged into a narrow gap between Kirkwood and Johnson Ridges. North of this obstacle, evidence of valley glaciation abounds; south of it such evidence is meager. Apparently, once the boulder was wedged into the gap, the southward spread of a narrow ice lobe into Red Canyon was impeded. A polished and striated limestone rib (fig. 17) about 200 yards south of the wedged boulder indicates, however, that the ice overrode the obstacle and extended down Red Canyon to end at some unknown point near the mouth of the canyon. The glacial debris in the NE¼ sec. 2, T. 12 S., R. 4 E., is additional evidence that the ice extended southward beyond the obstacle.

Outwash of Pinedale glaciers is widespread, but much of it is concealed beneath a thin veneer of alluvium (pl. 1). Some of the best exposures of outwash are in the mouth of Bacon Rind Creek, where the small ice-contact deposits which were formed as the ice melted and the huge glacial erratics deposited by the melting ice are partly buried by outwash, which is now being dissected by modern Bacon Rind Creek. The thickness of the outwash is uncertain; I estimate it to be 25–50 feet. Other outwash deposits, not so well exposed, floor parts of Upper and Lower Tepee Basins as well as the valley of Snowslide Creek.

FIGURE 17.—Vertical limestone bed of the Madison Group that was polished and striated when it was overridden by the glacier which moved southward through the narrow gap in Kirkwood Ridge.

MASS-WASTING DEPOSITS

Mass-wasting is one of the most effective erosive processes operating in the region owing to the high relief, the steep valley walls, and the nonresistant nature of many of the rock units. A host of landslides, the general term used for all features of mass-wasting, are in the area, and an attempt has been made to classify them according to the scheme proposed by Sharpe (1938). In general, this classification is based on the kind of movement involved.

The following terms are used:

Creep.—"the slow downslope movement of superficial soil or rock debris, usually imperceptible except to observations of long duration" (Sharpe, 1938, p. 21).

Earthflow.—"A slow flow of earth lubricated with water, occurring as either a low-angle terrace flow or a somewhat steeper but slow hillside flow" (Howell, 1960, p. 92; after Sharpe, 1938, p. 50).

Landslide.—"the perceptible downward sliding or falling of a relatively dry mass of earth, rock, or mixture of the two" (Sharpe, 1938, p. 64).

Rockfall.—"the relatively free falling of a newly detached segment of bedrock of any size from a cliff, steep slope, cave, or arch" (Sharpe, 1938, p. 78).

Rock glacier.—"glacierlike tongue of angular rock waste usually heading in cirques or other steep-walled amphitheaters * * *" Sharpe, 1938, p. 43).

Rockslide.—"the downward and usually rapid movement of newly detached segments of the bedrock sliding on bedding, joint, or fault surfaces or any other plane of separation" (Sharpe, 1938 p. 76).

SLOW FLOWAGE

Colluvium.—Colluvium (pl. 1) is a heterogeneous mixture of loose, weakly consolidated debris. It is plastered along many valley walls and locally forms thick masses at the base of some steep cliffs. This material effectively masks the underlying bedrock; it is a product of creep and represents a slow and gradual accumulation of detritus that has moved downslope under the influence of gravity. In a very real sense it is a mantle of rock and soil creeping downslope at an imperceptible rate, the general direction of flowage being shown by the convex-downslope bend of some of the trunks of trees growing in it.

In general, the colluvial deposits consist of fine to coarse rock debris derived from higher stratigraphic units. This detritus, which locally includes talus, intermingles with, and is adjacent to, morainic and alluvial deposits. As a result, the colluvium merges imperceptibly with these adjacent surficial deposits and contacts between the units are vague and uncertain.

The thickness of the colluvium ranges from a few inches near its margins to possibly as much as 50 feet along and near the base of some slopes; only the largest and thickest deposits are shown on plate 1.

In most places a rich soil formed on the colluvium supports a dense growth of vegetation. Near the mouth of Snowslide Creek, for example, the south valley slope is so densely wooded that the growth forms a pristine forest—not even game trails reach through the dense growth—and the ground surface is completely covered by pine needles and dead and decaying trees.

Rock glaciers.—Several boulder piles lie along the walls of the cirque at the head of Bacon Ridge Creek (pl. 1; fig. 3*C*), and some of these probably are true rock glaciers that have crept downslope. A few are now festooned across the floor of the cirque. Most of the rock glaciers are covered by small to large concentric wrinkles which probably formed by lateral rock movement, possibly as a result of alternate freezing and thawing of moisture in the interstices of the rock debris.

RAPID FLOWAGE

Earthflows.—Typical earthflows are present along the north shore of Hebgen Lake, on both flanks of the Gallatin River valley, and along the valley walls of Sage and Snowslide Creeks; only the largest flows were mapped (pl. 1).

Commonly, the earthflows are tonguelike lobate masses confined almost wholly to tributary stream valleys. At the mouths of some tributaries the earthflows have debouched onto plains of lower gradient and the water-saturated debris has spread out to form a broad lobate hummocky toe (fig. 18) in places as much as 1 mile across. Among the host of materials covered by these toes are terrace deposits, alluvial fans, and alluvium.

The general shape of an earthflow is determined in large measure by the preexisting confining valley. Locally, as along the Gallatin River valley, the flows are pinched at the waist and resemble a dumbbell (pl. 1). This constriction invariably reflects more durable bedrock valley walls. The contact between the earthflow and the overridden units normally is abrupt and relatively sharp, but in many places it has been concealed by subsequent slump and erosion plus the dense foliage so common to this part of Montana.

The surface of the earthflows is markedly hummocky and closely resembles the surface of fresh till. A striking difference, however, between the two surfaces is the complete absence from the surface of earthflows of undrained depressions or lakes. There is no apparent reason why these undrained features could not have formed on earthflows, but none were noted in the mapped area.

FIGURE 18.—Tonguelike earthflow along the east valley wall of the Gallatin River about 1½ miles north of the mouth of Bacon Rind Creek. The earthflow has spread across the flat sand and gravel alluvial fill of the Gallatin River.

A few of the earthflows that may have been more fluid than most have a smooth undulatory surface and closely resemble poorly formed alluvial fans (fig. 18).

The earthflows range in size from about one-twentieth of a square mile to such huge masses as the one near the mouth of Grayling Creek which occupies about three-fourths of a square mile (pl. 1). Their thicknesses differ widely from place to place; I estimate the Grayling earthflow, as an example, to be about 100 feet thick near its toe and about 30 feet thick near its source.

Most of the earthflows consist chiefly of materials which crop out in the parent valley, and some have a distinct concave headwall formed where the earthflow pulled away and began to move downslope. In these, the various strata have been broken and jumbled.

Some earthflows, such as the Grayling earthflow, consist of a heterogeneous mixture of materials of all sizes, shapes, and compositions and represent morainic material that was saturated by water and moved downslope. Such material probably was deposited near the toe of an ice lobe or block and became saturated by the abundant melt waters almost at the time of deposition. Once this liquid mass began to move it probably crept downslope gradually, filling the valley from wall to wall and slowing only where its forward progress was impeded by a constriction of the valley walls. It probably was sufficiently viscous to carry the largest erratics and to pick up and carry along any boulders it overrode. Many of these till-derived earthflows have on their surfaces large erratics which were moved downslope first by ice and are now being moved still farther downslope by gravity.

The earthflows in this area have formed in valleys whose gradients range from 300 to 600 feet per mile, but the gradient, although important, does not seem to have been as significant a factor as the abundance of both surface and ground water. All earthflows mapped have one or more springs near their heads or along their flanks, and many have small streams along one or both flanks. Water is the basic lubricant that starts the flow and keeps it moving.

The earthflows are of Recent age, and some, if not all, are still active. During the Hebgen Lake earthquake of 1959 the Grayling earthflow moved southward, disrupted fence lines, and bowed the once-straight highway (fig. 19).

Apparently earth movement throughout the earthflow is differential; some parts of the flow move more rapidly (and in different directions) than others. In one place on the Grayling earthflow a corner post of a fence line had been pulled from the ground by the unyielding barbed wire attached to it. This suggests that the ground near the head of the earthflow moved at a slower rate than the ground in which the corner post was emplaced. The final result was a corner fence post yanked out of ground—the ground literally moved away from the post.

Rockslides.—Three slides, properly classed as rockslides, are known in the quadrangle. One, athwart Grayling Creek, is named the Grayling landslide; another, very much smaller, masks some of the sedimentary strata exposed along the west flank of Burnt Fork, and hence is called the Burnt Fork landslide. The third is on Snowslide Creek (pl. 1).

FIGURE 19.—Earthquake-induced movement of the Grayling earthflow. During the Hebgen Lake earthquake of 1959 the Grayling earthflow moved southward (toward the right of the photograph) as shown by the bow in the once-straight highway (now U.S. 287). View is eastward from the southeast flank of the Grayling earthflow.

319-733 O—69——5

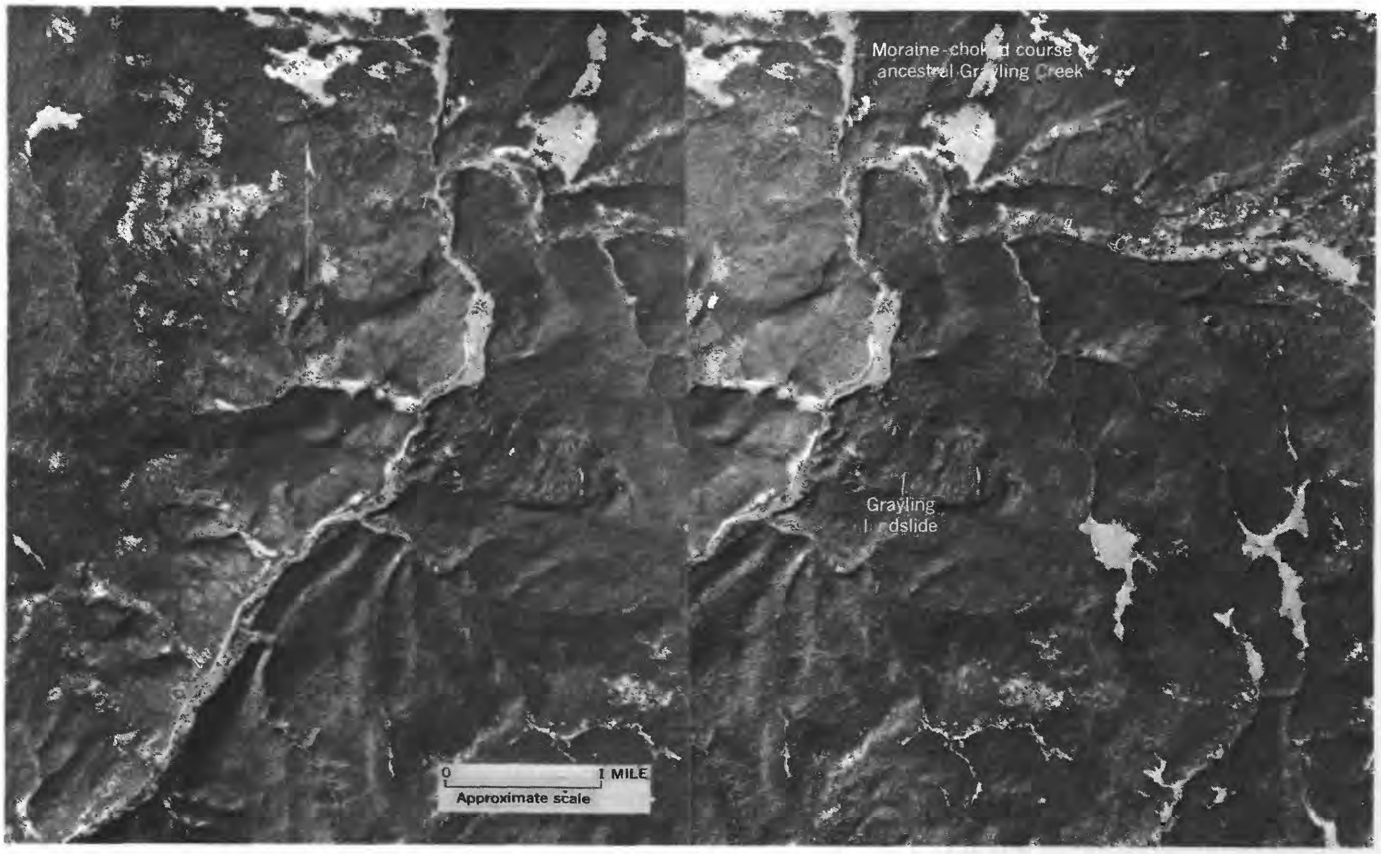


FIGURE 20.—Stereophotographs of the Grayling landslide and adjacent areas in Yellowstone National Park. Since these photographs were taken in 1954 a new route has been constructed across the slide about a quarter of a mile east of Grayling Creek and the former tortuous course of Highway 191 around the toe of the slide (shown as a white line) has been abandoned.

Grayling landslide.—Near mile post 268 on the Montana-Wyoming State line a large rockslide, the Grayling landslide, has constricted the valley of Grayling Creek (pls. 1 and 2). Some time after Grayling Creek had been diverted southward to become tributary to the Madison River (p. 55), a mass of claylike intensely altered rhyolite slid westward into Grayling Creek and temporarily dammed it. In time the ponded waters cut a narrow valley across the toe of the slide, and its width contrasts markedly with the width of the valley both north and south of the slide. Both north and south of the slide the valley is about 2,000 feet wide, but where Grayling Creek has cut across the toe of the slide the valley narrows to about 1,000 feet (fig. 20). U.S. Highway 191, which follows the valley of Grayling Creek northward, once curved around the toe of the slide, but during the late 1950's a new segment of the highway was constructed across the waist of the slide.

The slide trends westward and is about 1.2 miles long, about 0.6 mile wide at its toe, and about 0.3 mile wide near its source. The bulk of the slide is along the east bank of Grayling Creek and occupies about half a square mile. Slide remnants are on the west valley wall about 150 feet above the stream floor, which suggests that the slide was emplaced rapidly enough to cross the valley floor and mount the opposite side. The volume of rock involved is estimated to be about 27 million cubic yards.

The slide is characterized by hummocky "landslide" topography and, in addition, is crossed by transverse concentric ridges (fig. 20). Most of the depressions have been partly filled by slope wash and by vegetation. A dense forest of lodgepole pine covers the slide, masking its general nature; the irregular tilt of some trees suggests that parts of the slide may be moving imperceptibly and spasmodically.

A depression at the head of the slide is delineated in part by steep walls of the decomposed bedrock from which the slide mass separated. The slide bulges at the toe where piles and trains of loosened sliderock have moved over the stable country rock (fig. 20).

The slide is composed chiefly of claylike decomposed rhyolite, which was likely formed as a result of once-active hydrothermal springs and geysers. The whole mass, probably water saturated, moved rapidly downhill, but it is uncertain whether along any particular bedding plane or surface. The mass slid on a gradient of about 7°, and the drop was probably about 600 feet.

The debris undoubtedly blocked Grayling Creek and temporarily impounded the water to form a small lake (fig. 21). The waters eventually topped the dam and cut a new passageway across the toe of the slide.

Differential movement of the slide seemingly persists at the present time, as suggested by the tilted trees and the recurring breaks in Highway 191 where it crosses the slide. There was, however, no appreciable movement caused by the Hebgen Lake earthquake of 1959. Many of the breaks in the highway can be traced to local liquefaction of materials by the ever-present water in the slide, and I suspect that there has been some minor creep of parts or all of the mass, due mainly to the undercutting of the toe of the slide by Grayling Creek.

Burnt Fork landslide.—Along the west valley wall of Burnt Fork near its junction with the valley of Bacon Rind Creek, a small block of Cambrian strata has slid valleyward and now masks the underlying sedimentary units (pl. 1). The slide, known as the Burnt Fork landslide, covers an area 1,300 feet long and 400 feet wide—about 11 acres.

The slide consists of steeply dipping Cambrian and Ordovician strata which range from the Meagher Limestone at its north end to the undivided Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite at its south end. The bedding is largely intact and dips southward 75°–85°. The underlying bedrock, chiefly Cambrian Flathead, Wolsey, and Meagher Formations, dip westward about 45°.

Apparently after the sedimentary block broke loose it moved downslope as a cohesive entity and rotated somewhat; its upper end moved counterclockwise and its lower end pivoted almost in place.

Snowslide Creek slide.—Along Snowslide Creek, debris composed almost wholly of limestone boulders of the Madison Group probably fell as an avalanche from a slope oversteepened by glaciation. The debris, likely shaken loose by an earth tremor, was emplaced rapidly and effectively narrowed, and possibly blocked for a time, the upper reaches of Snowslide Creek. The stream is now confined against the south valley wall,

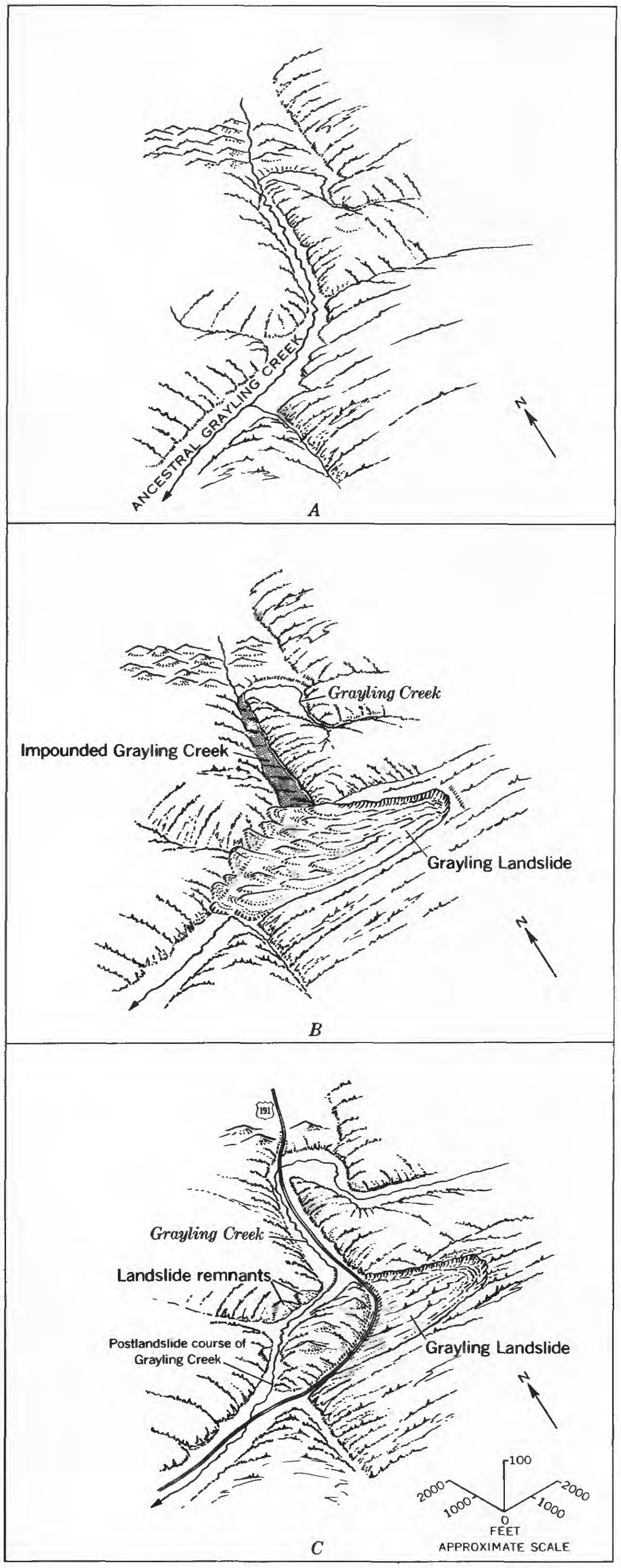


FIGURE 21.—Development of the Grayling landslide. *A*, Southward-flowing Grayling Creek cut a valley through widespread rhyolite flows; *B*, Along one valley wall where part of the flows had been altered by ancient geyser and hot-spring activity, a mass of deeply altered rhyolite slid westward into the valley and dammed the stream; *C*, In time the impounded waters crested the slide and cut a narrow new valley which contrasts sharply with the wider valley both north and south of the slide.

and the slide, heavily forested, makes a remarkably effective barrier across the valley.

Rockfalls.—At the base of many steep cliffs are enormous accumulations of boulders. The largest rockfalls are along the north flanks of Red Mountain, where their great extent and position at the base of the steep cliffs suggest that they represent the gradual accumulation and coalescence of a great many small rockfalls, emplaced during a long interval of time.

In most places it is impossible to determine whether a boulder pile formed as a result of the free fall of a large mass or as the result of gradual accumulation of individual boulders. During the Hebgen Lake earthquake of 1959, for example, an enormous rockfall occurred at Kirkwood Creek (west of the mapped area), where an entire face of a cliff formed by the Quadrant Sandstone was shaken loose. The resultant debris at the foot of the cliff could in no way be distinguished from talus accumulations found elsewhere; only the fresh face of rock on the cliff indicated the rapidity of the event.

ALLUVIUM

Detrital material of Recent age deposited by running water was mapped as alluvium along streams and rivers and as alluvial fans at the mouths of tributary streams.

The general composition of the alluvium reflects the rock units exposed upstream. Along Red Canyon Creek, for example, the alluvium commonly is some shade of red, for it includes debris from the extensive red Woodside Siltstone exposures near the head of the stream. Alluvium along streams such as the Gallatin River and Bacon Rind Creek, which flow across morainal deposits, is more complex, for in addition to materials reflecting the bedrock exposed upstream, it contains exotic pebbles and boulders derived from the drift.

The alluvial fans, although much alike in general shape, differ greatly in slope, reflecting the gradients of the parent streams. The largest fan in the area, at the mouth of Grayling Creek, which has a gradient of less than ½°, slopes about 50 feet per mile. By contrast, the fan at the mouth of Red Canyon Creek, whose gradient ranges from 9° in its upper reaches to about 2° near its mouth, is steeper; it slopes about 110 feet per mile. The small steep fan at the mouth of Monument Creek (in the northeast corner of the quadrangle) has a gradient of approximately 600 feet per mile (a little less than 7°).

REGIONAL STRUCTURE

Southwestern Montana is marked by elongate north- and northwest-trending mountain ranges separated by parallel oval-shaped intermontane basins. This physiographic pattern reflects the underlying regional structural framework—block faulting much like that which formed the Basin and Range Province of the Southwestern United States (Pardee, 1950). In southwestern Montana almost every range represents a relatively upthrown tilted fault block, and the adjacent intermontane basin represents the relatively downthrown block. Between the blocks is a system of high-angle normal faults which invariably dip toward the basin. Locally, Recent deposits which extend from the mountain front onto the plains are offset by small fault scarps, the result of recent movement along these normal faults.

In regional aspect, this part of southwestern Montana is dominated by three intermontane basins and their intervening mountain ranges (fig. 2). First and largest is the Madison Valley, which trends northwestward and is about 55 miles long and 12 miles across at its widest part. The east edge of the valley is sharply defined for most of its length by the Madison Range; this abrupt transition from valley to mountain front is in large measure the result of movement along the Madison Range fault, a system of high-angle normal faults which dip valleyward. In places, Recent fault scarps, which show 10–30 feet of displacement, are along the trace of this fault (Pardee, 1950, p. 370). Parts of the fault south of the mouth of the Madison Canyon may have been active during the Hebgen Lake earthquake of 1959. The west edge of the valley, less well delineated, is bounded by the Gravelly and Tobacco Root Ranges (fig. 2).

The second intermontane basin, almost as large as the Madison Valley, is the Centennial Valley, which trends eastward and is about 47 miles long and 6 miles wide (fig. 2). Although most tectonic elements in this part of southwestern Montana trend northwestward, the Centennial Range and Valley are major exceptions, seemingly cutting directly across the prevailing structural grain of the country. The south margin of the Centennial Valley is well defined by the straight, rugged front of the Centennial Range, a fault block which has been tilted southward. The north edge of this block is bounded by the Centennial fault, another system of high-angle normal faults which probably dip northward below the Centennial Valley (Pardee, 1950, p. 374–375). The north margin of the Centennial Valley is poorly defined by the low foothills which form the south ends of the Snowcrest and Gravelly Ranges (fig. 2).

The third and smallest basin is the West Yellowstone basin, the north edge of which is in the Tepee Creek area (fig. 2). The basin is oval shaped; its long axis trends northwest and is about 18 miles long; its short axis is about 12 miles long. The north edge of the basin is delineated by foothills and lesser mountains which form the east flank of the Madison Range. These mountains are mainly complex fault blocks that have been tilted northeastward along northwest-trending high-angle normal faults, of which the best known are the Red Canyon and Hebgen faults—the active faults probably responsible for the Hebgen Lake earthquake of 1959 (Fraser and others, 1964)[8]. These normal faults dip southward and pass below the West Yellowstone basin, which was tilted northeastward as a result of the 1959 reactivation of these faults. The east and south margins of the West Yellowstone basin are formed by dissected edges of the Yellowstone rhyolite plateau, and the west edge of the basin is outlined by steep heavily wooded slopes formed on the Precambrian crystalline rocks which constitute the backbone of the Madison Range.

Each basin, and its complementary mountain range, probably owes its general shape to two major episodes of deformation. The first involved deep-seated lateral compression which probably began in Late Cretaceous or early Tertiary time during the Laramide deformation. At that time the prism of sedimentary rocks which filled the miogeosyncline was squeezed and the rocks were uplifted, folded, overturned to the northeast, and finally broken along north- and northwest-trending low-angle thrust faults. Much later, possibly during middle Tertiary time, a second episode of deep-seated lateral compression may have arched the area (Pardee, 1950, p. 403). As these compressive forces eased, parts of the uplifted mass settled along high-angle normal faults which were probably guided by the preexisting structures formed during the first or Laramide compressive episode. The result is a remarkable parallelism between the thrust faults of Laramide age and the younger normal faults of middle Tertiary age.

LOCAL STRUCTURE

Three major tectonic elements of Laramide origin can be delineated in and near the Tepee Creek quadrangle (fig. 22). These are, from southwest to northeast: (1) A mass of Precambrian crystalline rocks that was thrust northeastward (upper plate), known as the Madison thrust block; (2) the exposed rocks of the overridden or lower plate, called the Cabin Creek zone; and (3) an area of broad warps and uncertain boundaries distant from the thrust mass in which the orogenic forces were but slightly felt, called the Pika Point zone. This quadrangle overlies the Cabin Creek and Pika Point zones and is 1–7 miles east of the east edge of the Madison thrust block.

Superimposed across these older structures, however, are northeast-tilted fault blocks, the product of mid-Tertiary normal faulting. These include the Red Canyon, Kirkwood, and Monument Mountain fault blocks (fig. 22) within the mapped area, and the Hebgen block to the south and west.

Madison thrust block.—The Madison thrust block, wholly west of this quadrangle, is an enormous plate of Precambrian crystalline rocks, capped here and there by Paleozoic strata, that has been thrust northeastward onto softer, Mesozoic and Paleozoic rocks along the southwest-dipping low- to moderate-angle Beaver Creek thrust fault (fig. 22). Erosion of this thrust block has formed the intensely dissected backbone of the Madison Range. The west edge of the thrust block, bounded by the Madison Range normal fault, rises abruptly from the spacious floor of the Madison Valley (fig. 22); its east edge, determined by the Beaver Creek thrust fault, is not so distinct, for the steep slopes formed on the Precambrian metamorphic rocks of the upper plate pass gradually into a hilly topography formed on the softer, Mesozoic and Paleozoic rocks of the lower plate.

Cabin Creek zone.—The Cabin Creek zone is northeast of the Madison thrust block and includes the intensely folded, overturned, and faulted rocks of the lower plate (fig. 2). Overturned folds and thrusts faults are common in the southwestern part of the zone (nearest the Madison thrust block), but toward the northeast these strongly asymmetric features pass gradually into broad, more moderate asymmetric folds whose axial planes dip steeply southwestward.

The northeast boundary of the Cabin Creek zone is vague and uncertain and is probably best considered as a broad band where the asymmetric folds of the Cabin Creek zone pass into the broad symmetric flexures and swells of the third tectonic element, the Pika Point zone.

Pika Point zone.—The Pika Point zone is marked by broad gentle anticlines, low domes, and wide synclinal troughs. Locally, Mesozoic and Paleozoic rocks exposed along the flanks of these structures are broken by high-angle normal and reverse faults. The Pika Point zone, farthest from the thrust block shows the least effects of lateral compression.

This northeastward diminution of the effects of Laramide deformation can be interpreted in at least two ways. First, the Madison thrust block, shoved northeastward along the Beaver Creek thrust fault by deep-seated lateral compression, deformed and then overrode the strata in front of it. In essence, the thrust block acted as a piston. In the second interpretation the strata in both the Cabin Creek and Pika Point zones are assumed to have been deformed nearly independently of the thrust block, which was emplaced at some time during

[8] The structural significance of these faults is viewed differently by Myers and Hamilton (1964, p. 85–87).

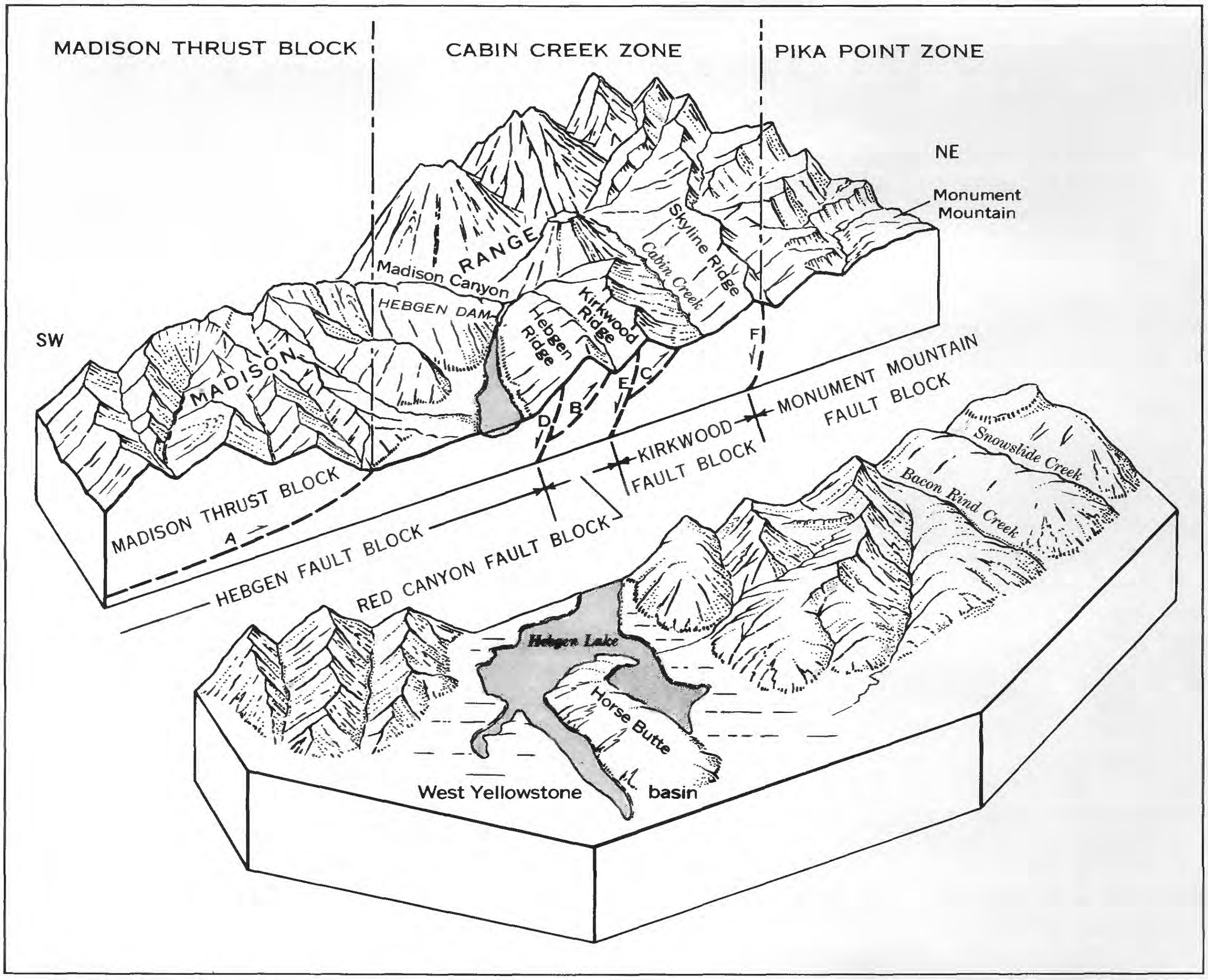


FIGURE 22.—Major tectonic elements in or near the Tepee Creek area: the Madison thrust block, the Cabin Creek zone, and the Pika Point zone. Thrust faults (Laramide): *A*, Beaver Creek thrust fault (beyond mapped area); *B*, Wells thrust fault (beyond mapped area); *C*, Divide thrust fault. Normal faults (middle Tertiary): *D*, Hebgen normal fault (beyond mapped area); *E*, Red Canyon normal fault; *F*, Upper Tepee normal(?) fault.

or after deformation. The thrust block, thus, may have slid into its present position as a result of high fluid pressures much as proposed by Hubbert and Rubey (1959). In this interpretation the proximity of the thrust block to the folded and faulted strata of the Cabin Creek zone is fortuitous.

I prefer the first alternative because there seems to be an integrated genetic relationship between the Madison thrust block, the thrust faults and overturned folds of the adjacent Cabin Creek zone, and the broad warps of the more distant Pika Point zone. It seems most unlikely to me that relations such as these could result from fortuitous juxtaposition.

STRUCTURES OF LARAMIDE AGE

In this quadrangle the basic structural framework consists of two major anticlines (the Kirkwood and Skyline anticlines), overturned in whole or in part, separated by a broad shallow syncline (the Cabin Creek syncline) (fig. 23). All are in the Cabin Creek zone; northeastward from these folds, in the Pika Point zone, the rocks form wide anticlines and synclines.

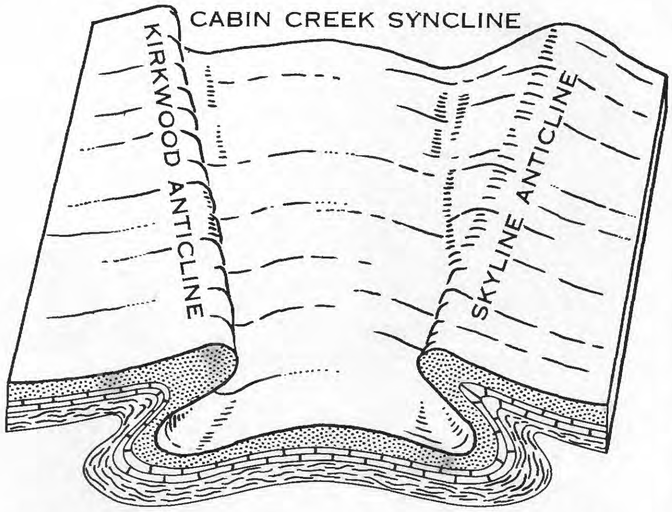


FIGURE 23.—Suggested shape of the major parental folds formed during the Laramide deformation as a result of lateral compression from the southwest. View is northwestward. The Kirkwood anticline is overturned throughout its extent in the quadrangle. The Skyline anticline, by contrast, is upright near its northwest end. The Cabin Creek syncline is actually a low broad minor anticline flanked by two tightly folded synclines.

CABIN CREEK ZONE

KIRKWOOD ANTICLINE

The southwesternmost fold, here called the Kirkwood anticline, is overturned to the north and northeast (fig. 23) and forms a broad curve convex to the northeast (pl. 1); the axial plane probably dips southward about 40°. In map view the fold, about 9 miles long, extends northwestward, westward, and finally southwestward from near the mouth of Grayling Creek (sec. 6, T. 12 S., R. 5 E.) to a point in Kirkwood Canyon (sec. 19, T. 11 S., R. 4 E.) about 1 mile northeast of the canyon mouth (west of the mapped area). The high, prominent range in the southwestern part of the quadrangle, Kirkwood Ridge (pl. 2), represents the eroded remnant of the overturned limb of this fold (fig. 28 and cross sections *C–C′* pl. 1). It is composed of vertical to overturned beds ranging from the basal conglomerate of the Kootenai Formation (Early Cretaceous), which near the west end of the fold (in the mapped area) is almost completely recumbent (dipping 18° southward), to the uppermost beds of the Madison Group (Mississippian).

The north (overturned) limb of the Kirkwood anticline, which forms Kirkwood Ridge, has been broken by the Divide thrust fault and thrust northeastward onto the southwest limb of the Cabin Creek syncline. The south (upright) limb of the Kirkwood anticline, southwest of Kirkwood Ridge, has been downthrown along the Red Canyon normal fault (figs. 24, 28; cross sections *C–C′*, *B–B′*, pl. 1).

CABIN CREEK SYNCLINE

The Kirkwood anticline passes northward into a broad shallow syncline here named the Cabin Creek syncline, which plunges west-northwestward (pl. 1; fig. 23). Topographically the syncline forms a wide, spacious valley occupied by Cabin Creek and its Middle and South Forks (fig. 24).

The syncline is concealed beneath volcanic rocks at the head of Little Tepee Creek (secs. 17, 18, 19, and 20, T. 11 S., R. 5 E., pl. 1); from there it trends west-northwestward for about 9 miles to end near Beaver Creek west of the mapped area.

The youngest strata exposed in the core of the syncline are dark-gray to black shales of Thermopolis(?) age, and the oldest strata exposed at the east end are intensely deformed dolomite beds provisionally assigned to the undifferentiated Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite.

The southwest flank of the syncline is overridden by the Kirkwood anticline along the Divide thrust; by contrast, its northeast flank passes unbroken into the Skyline anticline (figs. 24, 28; cross sections C–C′, B–B′, pl. 1).

SKYLINE ANTICLINE

The Skyline anticline trends mainly northwestward, but, in sharp contrast with most of the other folds in this area, its axial plane dips northward and northeastward at 35°–60° (fig. 23). Near the east end of the fold the beds are steeply overturned and dip northeastward (section C–C′, pl. 1); to the west near White (fig. 8) and Redstreak Peaks the beds are nearly vertical (section B–B′, pl. 1), and at the west edge of the mapped area the beds dip steeply southwestward (section A–A′, pl. 1).

The anticline trends northwestward from near the head of Little Tepee Creek for about 6 miles before it turns abruptly northward (west of the mapped area) for another 2 miles. Near Sage Creek, in the northwest corner of the quadrangle, the anticline passes into the broad gentle warps characteristic of the Pika Point zone. The northeast flank is cut locally by high-angle normal(?) and reverse faults (fig. 26; sections C–C′, B–B′, pl. 1).

The youngest strata exposed are along the southwest flank of Skyline Ridge and consist of the dark-gray to black shale beds of the Thermopolis(?) Shale (fig. 24). The oldest strata exposed form the northeast flank of Skyline Ridge and consist of massive and thin-bedded crystalline limestone and dolomite beds of Cambrian age (fig. 5).

FIGURE 24.—Kirkwood Ridge from the southeast. The approximate trend of the Red Canyon normal fault is shown by the scarp formed during the Hebgen Lake earthquake of 1959. Strata southwest (left) of the fault scarp are part of the downthrown upright limb of the Kirkwood anticline, and strata between the Red Canyon scarp and the Divide thrust fault are part of the overturned limb. The attitudes of the strata which underlie the area shown in this photograph are illustrated in the left (southwest) part of cross section *C–C'*, plate 1 and in figure 28. Montana Highway Department photograph.

RED CANYON ANTICLINE

A small dissected anticline, named the Red Canyon anticline, is exposed near the mouth of Red Canyon in the southwest corner of the quadrangle (pl. 1). The anticline, partly concealed by morainic deposits and colluvium, is about 1½ miles long and 1 mile wide. It plunges northward, and its axial plane dips westward at about 70° (fig. 25*D*). The strata that form the west

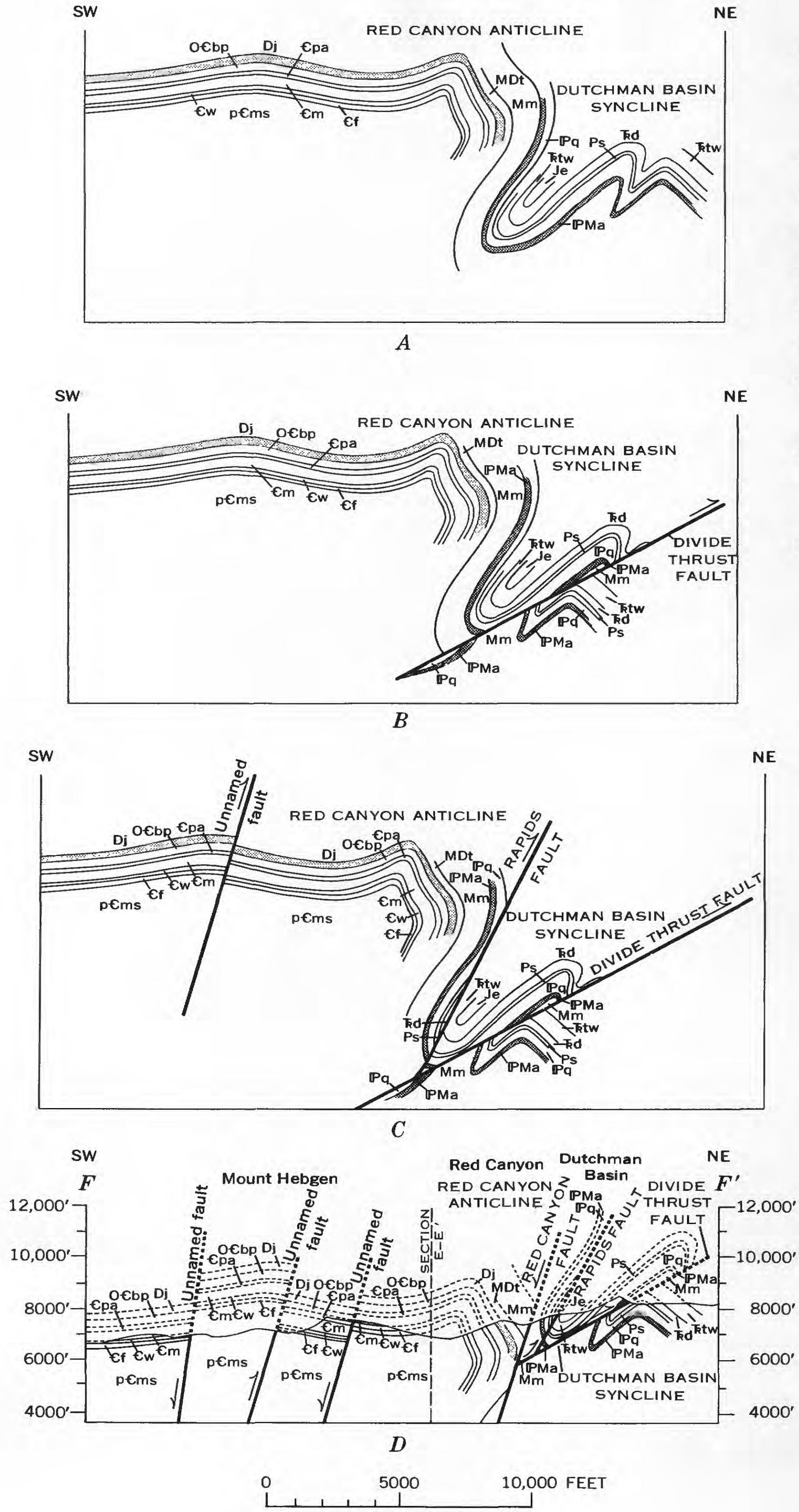
FIGURE 25.—Structural development of the Red Canyon–Dutchman Basin area. For purposes of simplicity, the masking cover of volcanic and surficial deposits is not shown on this cross section. Line of cross section (*F–F'*) is shown on plate 1.

A. The parental folds were overturned to the northeast.

B. Lateral compression caused one of the folds to break along the low-angle Divide thrust fault, and the upper plate was shoved northeastward. The amount of stratigraphic displacement in this specific locality may be about 3,000 feet, although to the northwest near Cabin Creek it is only about 500 feet.

C. Continued lateral compression caused part of the upper plate to break along the minor Rapids fault, a splinter off the larger Divide sole-thrust fault. This slice was shoved upward about 300 feet.

D. Much later, probably during the middle Tertiary, differential subsidence caused the area to break into several fault blocks which are downthrown and tilted northeastward along normal faults, the most important of which is the Red Canyon fault; movement on the Red Canyon fault has probably been continuous since that time; the latest movement resulted during the Hebgen Lake earthquake of 1959.



EXPLANATION

Jurassic:
Je, Ellis Group

Triassic:
Ŧtw, Thaynes(?) Formation and Woodside Siltstone
Ŧd, Dinwoody Formation

Permian:
Ps, Shedhorn Sandstone

Pennsylvanian:
₱q, Quadrant Sandstone

Pennsylvanian and Mississippian:
₱Ma, Amsden Formation

Mississippian:
Mm, Madison Group

Mississippian and Devonian:
MDt, Three Forks Formation

Devonian:
Dj, Jefferson Formation

Ordovician and Cambrian:
OЄbp, Bighorn(?) Dolomite, Snowy Range Formation, and Pilgrim Limestone undivided

Cambrian:
Єpa, Park Shale
Єm, Meagher Limestone
Єw, Wolsey Shale
Єf, Flathead Sandstone

Precambrian:
pЄma, pre-Belt metamorphics; mica schist

flank of the anticline dip westward at low angles which average 5°, but swing imperceptibly to dip northward beneath Mount Hebgen. By contrast, the strata of the east flank dip eastward at 35°–55° and end abruptly against the Red Canyon normal fault (fig. 25*D*; cross section *E–E′*, pl. 1).

Red Canyon Creek follows the axis of the anticline and has exposed Precambrian strata. The youngest rocks involved are basal beds of the Madison Group.

The anticline is but a small part of the Red Canyon fault block, which during the Hebgen Lake earthquake of 1959 was titled northeastward as the result of differential movement on the Red Canyon fault.

DUTCHMAN BASIN SYNCLINE

West of Johnson Lake (W½ sec. 31, T. 11 S., R. 5 E.) downfolded strata on the broad upland between Johnson and Dutchman Ridges form an elongate asymmetric basinlike syncline called the Dutchman Basin syncline (fig. 25*D*; pl. 1).The syncline trends northwestward, and its axial plane dips about 70° southwestward (cross section *E–E′*, pl. 1). The syncline can be traced for about 3½ miles from near Corey Springs (sec. 7, T. 12 S., R. 5 E.) to Red Canyon, where the trace of the axial plane disappears beneath morainic debris (pl. 1). Northwestward beyond this point the axial plane is probably truncated by the Divide thrust fault (cross section *C–C′*, pl. 1).

The syncline is about 1 mile across at its widest part, directly east of the mouth of Red Canyon. Its axis is about a quarter of a mile northeast of the topographic crest of Dutchman Ridge. This ridge, formed mainly by the overturned steeply dipping beds of the Quadrant and Shedhorn Sandstones, is part of the west limb of the syncline (pl. 1).

The youngest strata enfolded in this syncline, exposed in the northwest corner of Dutchman Basin (sec. 6, T. 12 S., R. 5 E.), are beds of the basal conglomerate of the Kootenai Formation. The oldest beds preserved along the synclinal flanks are the uppermost strata of the Madison Group.

The west flank of the syncline is overturned and dips southwestward about 65°; the east limb dips southwestward 40°–60°, although locally those beds near the Divide thrust fault are overturned and dip eastward about 45° (cross section *E–E′*, pl. 1).

The west flank of the syncline is cut by the Red Canyon normal fault and is partly offset by a small high-angle reverse fault called the Rapids fault (figs. 25, 27). The east flank of the syncline is truncated by the Divide thrust and by a small high-angle reverse fault called the Dutchman fault (section *E–E′*, pl. 1).

Quite likely the northwestward extension of the syncline is concealed beneath the overturned limb of the Kirkwood anticline (cross section *C–C′*, pl. 1). The general shape and attitude of the concealed syncline can only be surmised, but its axial plane probably parallels that of the Kirkwood anticline, which strikes northwestward and dips about 40° southwestward.

DIVIDE THRUST FAULT

The Divide thrust fault, along the north flanks of Kirkwood and Johnson Ridges (figs. 25, 28; pl. 1), is almost concealed by glacial debris and colluvium. The fault trace extends for at least 9 miles in a broad curve convex to the northeast, of which about 7 miles is in the report area. Near its east end the fault is concealed beneath the Grayling earthflow (secs. 5, and 8, T. 12 S., R. 5 E., near the mouth of Grayling Creek). From that point it extends northwestward across Red Canyon, and finally westward along Cabin Creek; it ends at some unknown point on Cabin Creek to the west of the mapped area. The fault is best exposed in the SE¼ sec. 15, T. 11 S., R. 4 E., at the head of a tributary valley to Red Canyon Creek, where the overturned beds that form the north flank of Kirkwood Ridge overlie the gently inclined upright beds which form the west flank of the Cabin Creek syncline (fig. 24).

The fault dips westward at angles which I estimate range from 20° to 40°; several smaller faults such as the Rapids and Dutchman reverse faults, interpreted as splinters off the master Divide thrust fault, are much steeper (fig. 25; cross section *E–E′*, pl. 1).

Near the west end of the fault (along Kirkwood Ridge) the basal conglomeratic sandstone beds of the Kootenai Formation, almost completely overturned, rest on the upright claystone beds of the uppermost Kootenai. To the east, they rest on basal strata of the Thermopolis(?); the stratigraphic displacement is about 500 feet. Near Johnson Lake (sec. 31, T. 11 S., R. 5 E.), overturned beds of the Madison and the Amsden rest on upright beds of the Quadrant and Shedhorn Sandstones (cross section *E–E′*, pl. 1). Here the stratigraphic displacement may be on the order of 500 feet—its exact amount is uncertain.

The Divide thrust fault was apparently formed during the Laramide when the Kirkwood anticline was thrust northeastward onto the south flank of the Cabin Creek syncline (fig. 28*B*). Subsequently the thrust fault was most likely cut and offset by the much younger Red Canyon fault (figs. 25*C*, fig. 28*D*; cross sections *C–C′*, *B–B′*, *E–E′*, *D–D′*, pl. 1).

WHITE PEAK FAULT

The northeast edge of the Cabin Creek zone is delineated in part of this area by a high-angle reverse fault called the White Peak fault (secs. 1 and 12, T. 11 S., R. 4 E., and secs. 7 and 18, T. 11 S., R. 5 E., pl. 1). This fault, along with the parallel Upper Tepee fault (p. 79), has determined in large measure the configuration of the northeast flank of Skyline Ridge (fig. 26; cross section *C–C′*, pl. 1).

The White Peak fault can be traced N. 35° W. for about 2 miles. Its southeast end is concealed beneath volcanic rocks in the center of sec. 18, T. 11 S., R. 5 E., and its northwest end is concealed beneath surficial deposits in the SW¼ sec. 1, T. 11 S., R. 4 E. The dip of the fault plane is indeterminate; it may be nearly vertical, for the trace of the fault across well-dissected terrain is almost straight (pl. 1).

I am uncertain how or where the fault ends to the northwest. Its trace is toward the Upper Tepee fault, and possibly it is cut by that fault. An alternative possibility, however, is that the fault planes merge; thus, the White Peak fault plane may be coincident with, and follow the fault plane of, the Upper Tepee fault for a short distance.

Southeastward the fault truncates the overturned northeast flank of the Cabin Creek syncline. Beyond the point where the fault is overlapped by volcanic rocks (center of sec. 18, T. 11 S., R. 5 E.), the beds which form the synclinal nose (secs. 18 and 19, T. 11 S., R. 5 E.) are also steeply overturned and fragmented, which suggests that the fault persists at least that far to the southeast.

The displacement of the fault increases southeastward, and near peak 9,648 (N½ sec. 12, T. 11 S., R. 4 E.) it is estimated to be about 2,000 feet, juxtaposing overturned beds of the Madison Group in the upper plate against overturned beds of the undifferentiated Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite in the lower plate (fig. 26; cross section *C–C′*, pl. 1).

The relation of the White Peak fault to the adjacent Upper Tepee fault is unknown. Two possible interpretations of the field relations are proposed in figure 26. The first hypothesis (figs. 26*A*, *B*) involves two separate and distinct episodes of deformation: an initial episode of lateral compression during which the White Peak fault (high-angle reverse fault) developed, and a second episode of tensional relief during which the Kirkwood fault block settled along the Upper Tepee normal(?) fault. In this interpretation it is assumed that the White Peak fault was formed during the Laramide deformation by the same forces responsible for the Divide thrust fault.

The second interpretation (figs. 26*D*, *E*) suggests that both faults are high-angle reverse faults which dip northeastward; one possibly is a splinter off the other. The faults, thus, are presumed to be contemporaneous in age and the result of lateral compression during Laramide deformation.

I favor the first alternative, that the area was deformed during two separate tectonic episodes, chiefly because this concept conforms to the tectonic history of the region (p. 65), an episode of lateral compression during the Laramide followed by an episode of renewed compression and subsequent tensional relief during the mid-Tertiary.

OTHER FAULTS

Two faults interpreted as splinters off the Divide thrust fault (p. 70) break the rocks in the southwest corner of the quadrangle. The Rapids fault truncates the southwest flank of the Dutchman Basin syncline (fig. 25*D*; pl. 1), and the Dutchman fault cuts the northeast flank of the syncline.

Rapids fault.—The Rapids fault is a small high-angle reverse fault which can be traced for about 2 miles along a trend of about N. 30° W.; it dips southwestward 60°–85° and juxtaposes the Woodside Siltstone (part of the lower plate) and the Dinwoody Formation. Maximum stratigraphic displacement is about 300 feet. The Rapids fault was formed by the same tectonic forces responsible for the Divide thrust fault and is probably a splinter off that sole fault. Although the fault is concealed for much of its length by colluvium and dense foliage, it is well exposed along the southeast valley wall of Red Canyon, where Red Canyon Creek cuts through Kirkwood Ridge (fig. 27).

Dutchman fault.—The Dutchman fault extends for about 1½ miles to the northwest from near the mouth of Dry Wash (NW¼ sec. 8, T. 12 S., R. 5 E.), where it is concealed beneath the Grayling earthflow, to the middle of the west flank of Johnson Ridge (SW¼ sec. 31, T. 11 S., R. 5 E.). It trends about N. 50° W. and dips southwestward at a very high angle; possibly it is vertical for most of its length (section *E–E′*, pl. 1). I estimate the stratigraphic displacement to be about 500 feet. The fault is well exposed along the unnamed ridge which forms part of the southeast wall of Dutchman Basin.

Timber fault.—The Timber fault extends from the NE¼ sec. 34, T. 11 S., R. 4 E., to the Red Canyon normal fault near the center of sec. 36, T. 11 S., R. 4 E. It trends about S. 65° E. and is about 1½ miles long. For much of its length the Timber fault is concealed by a a thick cover of morainic debris.

The exact relations between the two faults are uncertain. Possibly in a few places the Red Canyon fault

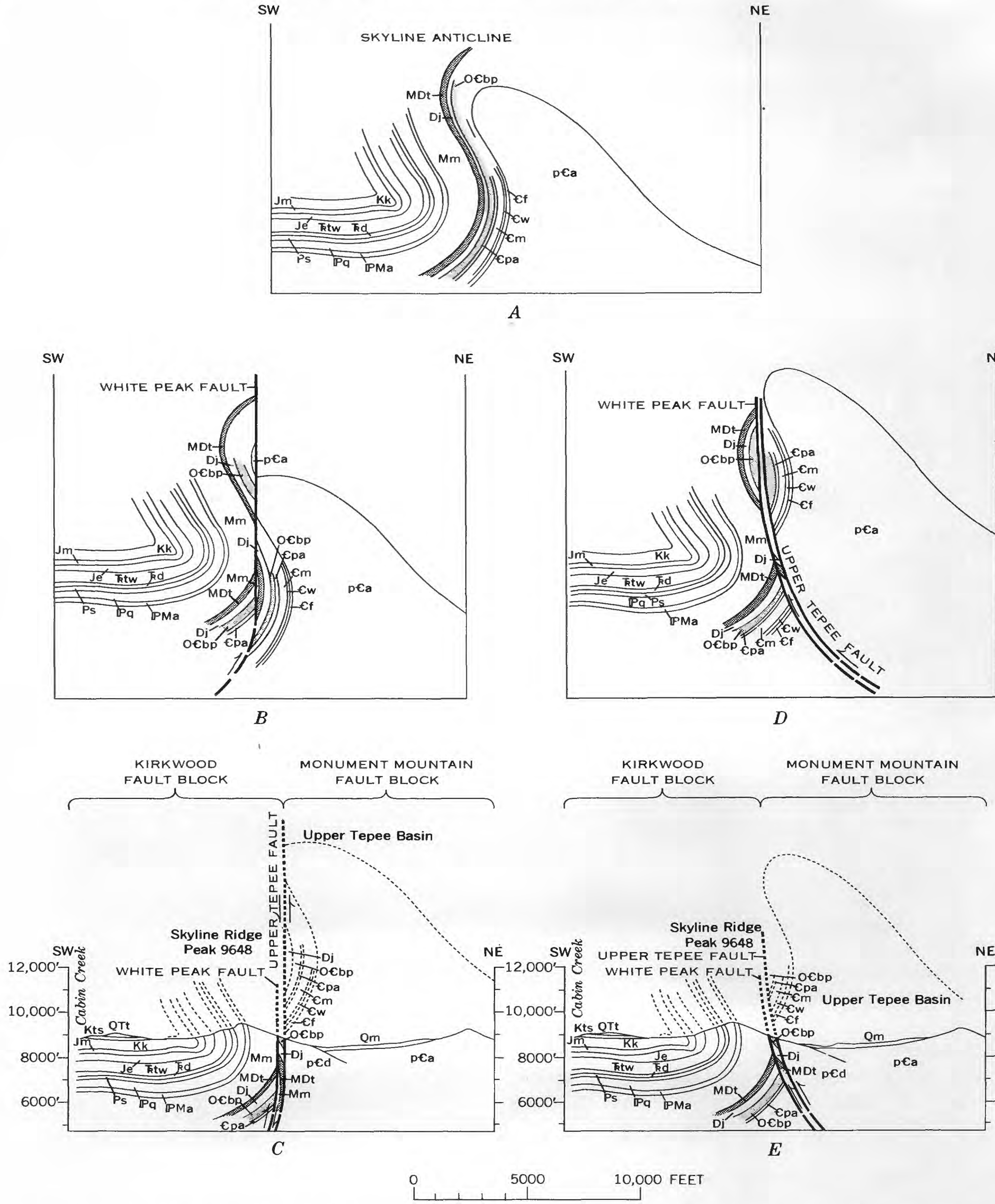


FIGURE 26.—Possible explanation of the structural development of the Cabin Creek–Skyline Ridge–Upper Tepee Basin area. (Explanation on facing page.)

EXPLANATION

Quaternary:
Qm, moraine deposits
Quaternary or Tertiary:
QTt, Yellowstone Tuff
Cretaceous:
Kts, sandstone member of Thermopolis(?) Shale
Kk, Kootenai Formation
Jurassic:
Jm, Morrison Formation
Je, Ellis Group
Triassic:
Ћtw, Thaynes(?) Formation and Woodside Siltstone
Ћd, Dinwoody Formation
Permian:
Ps, Shedhorn Sandstone
Pennsylvanian:
ℙq, Quadrant Sandstone
Pennsylvanian and Mississippian:
ℙMa, Amsden Formation
Mississippian:
Mm, Madison Group
Mississippian and Devonian:
MDt, Three Forks Formation
Devonian:
Dj, Jefferson Formation
Ordovician and Cambrian:
O€bp, Bighorn(?) Dolomite, Snowy Range Formation, and Pilgrim Limestone undivided
Cambrian:
€pa, Park Shale
€m, Meagher Limestone
€w, Wolsey Shale
€f, Flathead Sandstone
Precambrian:
p€a, pre-Belt metamorphic rocks (amphibolite)
p€d, pre-Belt metamorphic rocks (dolomite)

A. Parental fold (Skyline anticline) was overturned to the southwest and rested on beds which form the east limb of the Cabin Creek syncline.

First interpretation—Reverse faults followed by normal faults.

B. The high-angle White Peak reverse fault, almost certainly of the same generation (Laramide) as the Divide thrust fault, raised the upper plate (southwest of the fault).

C. During the middle Tertiary a second episode of orogeny further broke the area into tilted fault blocks along normal faults which paralleled the Laramide faults. The Kirkwood fault block, which includes the White Peak reverse fault, was downthrown along the high-angle Upper Tepee normal fault.

Second interpretation—Reverse faults formed by lateral compression.

D. Lateral compressive forces from the southwest thrust the western block under the overturned fold along two high-angle reverse faults. The overturned beds in the sliver between the faults are presumed to result from drag.

E. The Kirkwood fault block was downthrown in relation to the Monument Mountain fault block.

Inadequate exposures prevent the exact determination of the attitude of either fault plane, but the almost straight traces of the fault planes across dissected terrain suggest that they are vertical or nearly so.

plane is coincident with that of the Timber fault. If so, the Timber fault may have been transected by the Red Canyon fault.

PIKA POINT ZONE

Most of the flexures in the Pika Point zone are minor in comparison with those of the Cabin Creek zone to the south. Generally the Pika Point folds are 2–3 miles long and about 1 mile wide and pass smoothly and evenly into adjacent flexures. Locally, still smaller satellitic flexures deform the flanks of these minor folds.

SNOWSLIDE MOUNTAIN ANTICLINE

The Snowslide Mountain anticline, in the northwestern part of the quadrangle, is about 2 miles long and half a mile wide. Its west flank passes smoothly into the Monument Mountain syncline (pl. 1; p. 74), and its east flank, cut locally by a small high-angle reverse fault named the Skyline fault, flexes sharply downward to form the west flank of the southward-plunging Cold Water syncline (pl. 1; p. 74).

The oldest sedimentary strata exposed in the anticline are beds of the Flathead Sandstone, and the youngest rocks belong to the Madison Group.

RED MOUNTAIN ANTICLINE

One of the more prominent topographic features in this area is horseshoe-shaped Red Mountain (pl. 2), which is structurally a minor asymmetric anticline whose axial plane trends northwestward and dips about 50° northeastward (pl. 1). The structural crest is about three-quarters of a mile southwest of the topographic crest. Dips along the steep southwest and south flanks of the anticline, in the valleys of Burnt Fork and Bacon Rind Creek, range from 40° to 75° southward; by contrast, dips along the north flank are about 10° northward (section *B–B′*, pl. 1).

The oldest sedimentary strata exposed, along the south flank of this well-dissected anticline, consist of beds of the Flathead Sandstone, which here overlie beds of Precambrian dolomite. The youngest strata, preserved on the crest and the north flank of the anticline, are thin limestone beds of the Madison Group.

The beds along the southwest edge of the Red Mountain anticline flex abruptly downward to form the northeast flank of the narrow southward-plunging Cold Water syncline (pl. 1; p. 74). To the east, near Migration Creek, this downwarp disappears, and the northward-dipping strata of the Red Mountain anticline become part of the north limb of the Skyline anticline (section *C–C′*, pl. 1).

It seems likely that the northeastward-dipping axial plane of the Red Mountain anticline gradually changes in attitude eastward to become nearly vertical where it apparently merges with the axial plane of the Bacon Rind anticline (pl. 1).

BACON RIND ANTICLINE

The core of an elongate almost symmetric anticline which plunges northeastward is exposed in the valley of Bacon Rind Creek. This anticline, here named the Bacon Rind anticline, is about 3 miles long and 1–2 miles wide. The fold may be an extension of the Red Mountain anticline, but the two axes cannot be traced continuously through the intervening Precambrian rocks.

Stratigraphic units from the Flathead Sandstone to the Madison Group are exposed in the core. These strata have been pierced here and there by small igneous intrusions and raised by sills, one of which is preserved on the high divide between Bacon Rind and Snowslide Creeks (p. 43).

FIGURE 27.—The Rapids fault exposed along the southeast valley wall of Red Canyon, where Red Canyon Creek cuts through Kirkwood Ridge (NE¼ sec. 26, T. 11 S., R. 4 E.). The fault trends about N. 30° W. and dips southwestward 60°–85°. The Woodside Siltstone, part of the lower plate, dips about 50° northeastward. The Dinwoody Formation, part of the upper plate, dips about 85° southwestward.

OTHER FOLDS

Monument Mountain syncline.—The strata that form the top of Monument Mountain are flexed downward sharply to form an asymmetric syncline, here named the Monument Mountain syncline, which trends for about 5 miles northeastward, northward, and then northwestward in a curve gently concave to the west (pl. 1). The axial plane of the syncline dips about 50° westward.

The apex of the syncline is an oval basin about 1 mile long and half a mile wide, in which the youngest strata preserved are the basal beds of the Woodside Siltstone.

Cold Water syncline.—Squeezed between the Snowslide Mountain anticline on the west and the Red Mountain anticline on the east is a short narrow southward-plunging syncline, here called the Cold Water syncline (pl. 1). The axial plane strikes northward and dips eastward. The valley of Burnt Fork is coincident with the syncline but is so filled with morainic debris that only a few rocks crop out. In general, the youngest strata exposed are dolomite beds of the undifferentiated Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite. These strata dip about 15° eastward and practically abut beds of the underlying Park Shale, which are vertical or dip westward at about 80°. These relations suggest that the syncline may be broken along its east flank by a high-angle reverse fault, although they may reflect an unusually sharp flexure in the form of an asymmetric syncline.

Minor folds.—The rocks in the northwest corner of the quadrangle have been deformed into gentle, almost symmetric anticlines and synclines which trend generally northward. In general, Sage and Little Sage Creeks follow the axes of the anticlines; the unnamed ridges are underlain by synclines.

The folds are 2–3 miles long and about 1 mile wide. The strata range from the Quadrant Sandstone of Pennsylvania age to the Ellis Group of Jurassic age and commonly dip no more than 20° except where they pass eastward into the asymmetric folds associated with the Monument Mountain syncline.

POSSIBLE DETACHMENT OR BEDDING-PLANE FAULTS

One of the most vexing problems in this area is to explain the abrupt change in thickness of the sedimentary units. The Quadrant Sandstone, for example, is 265 feet thick in Red Canyon (NE¼ sec. 26, T. 11 S., R. 4 E.) but is 625 feet thick on Kirkwood Ridge, only about 1½ miles to the northwest. In Red Canyon the Quadrant is part of the northeast limb of the Kirkwood anticline and is either overturned or vertical. On Kirkwood Ridge the Quadrant is near the crest of the fold and is overturned. At both locations the upper and lower contacts of the Quadrant are exposed, and, apparently, no beds are omitted or repeated by faults.

Comparable but not so striking changes in thickness characterize several of the other sedimentary units. Possibly these abrupt variations in thickness mirror surface irregularities in the original basins of deposition of the respective units. Thus where the units are thin they may have been deposited across low knolls; where thick, they may have been deposited in local basins.

These interpretations, based on factors of original sedimentary deposition, seem inadequate to me, and I suggest that a more plausible explanation involves some form of structural deformation: either repetition of strata as a result of one or more detachment faults, or, more likely, minor bedding plane slips which thickened the formation near the fold crests and attenuated it along the flanks.

Possible detachment faults.—The Madison Group shows similar changes in thickness from about 1,000 feet to about 1,450 feet (p. 28).

In only a few localities are the beds which overlie and underlie the Madison Group close enough and well-enough exposed to permit the measurement of meaningful stratigraphic sections. One such locality is along Skyline Ridge near the northwestern corner of Upper Tepee Basin, in the NE$\frac{1}{4}$ sec. 2, T. 11 S., R. 4 E., about half a mile southeast of White Peak (pl. 1; fig. 8, locality *F* of fig. 10). Here the block west of the Upper Tepee fault includes what appears to be a full thickness of the Madison Group. A sliver of the Sappington(?) Sandstone Member of the Three Forks Formation west of the Upper Tepee fault is juxtaposed with Precambrian mica schist which forms the relatively upthrown block east of the fault. The Sappington(?) and overlying Madison strata, part of the southwest limb of the Skyline anticline, are vertical or nearly so near the Upper Tepee fault; about a quarter of a mile southwest of the fault the uppermost beds of the Madison, which underlie the Amsden on Skyline Ridge, are upright and dip about 75° southwestward.

These strata total about 1,000 feet in thickness, which may be the full thickness of the Madison Group in this area; faunal evidence in support of this view is inconclusive. Apparently Zones B and C are recognizable on the basis of their faunal content, and Zones A and D are likely present also (p. 26). So all zones of the Madison may be present; it is uncertain, however, whether each zone is represented by its full thickness. Possibly parts of one or more zones may have been removed by one or more bedding-plane faults.

Although decisive evidence that these beds are broken by such faults is lacking, at least one bedding-plane fault is suspected (measured section *G*), and the White Peak fault seemingly ends against the Upper Tepee fault about half a mile to the southeast (pl. 1).

About a mile northwest of White Peak, in Bacon Rind cirque, the Madison is about 1,450 feet thick. This greater thickness was originally attributed to repetition of Madison beds as a result of two detachment faults, the lowermost of which was thought to have removed both the Jefferson and the Three Forks Formations (Witkind, 1962). It was believed that the upper plate of this lower fault consisted of a slice about 450 feet thick from the middle of the Madison Group, composed of both Lodgepole and Mission Canyon strata, and that this slice rested on red and yellow calcareous siltstone beds of Cambrian and Devonian age. This slice of the Madison Group in turn was interpreted to have been overriden by the upper detachment fault whose upper plate consisted of the full thickness of the Madison Group plus younger strata.

The following corals collected in and near Bacon Rind cirque from the basal 60 feet of the Madison seem to support this concept. They are assigned to "a position in Madison Zone C–1 of Sando and Dutro (1960)." (W. J. Sando, written commun., 1963)[9]. (Fossil-collection localities are shown in fig. 10).

Fossil-collection locality 442 (21405–PC):
- *Homalophyllites* sp.
- *Vesiculophyllum* sp.
- *Zaphrentites* sp.
- *Rylstonia* sp.
- horn coral, undet.

Fossil collection locality 458 (21406–PC):
- *Homalophyllites* sp.
- *Vesiculophyllum* sp.
- *Zaphrentites* sp.
- *Cleistopora placenta* (White)
- horn corals, undet.

Subsequent fieldwork has created some doubt as to the existence of these detachment faults. Fossils collected from beds which underlie the red and yellow calcareous siltstone beds have been tentatively identified as: "pre-Mississippian and probably Devonian and it is probable that all are correctly assigned to the Jefferson * * * " (W. A. Oliver, written commun., 1962). Thus, the red and yellow calcareous siltstone sequence originally assigned to the Cambrian to Devonian may in fact be part of Sandberg's (1965) Logan Gulch Member of the Three Forks Formation (p. 22). (No fossils have been found in the siltstone beds to confirm this inter-

[9] Sando stated in part, "This determination [that the coral assemblage is indicative of Madison coral zone C–1] does not support the field identification of the containing strata as basal Madison. Elsewhere in Montana, northern Utah, and western Wyoming, Zone C–1 occupies approximately the upper two-thirds of the Lodgepole Limestone. I would expect to find this assemblage at least 100 feet above the base of the Madison."

pretation.) Doubt is thus cast on the existence of the lower detachment fault and, also by implication, on that of the upper fault.

How these two proposed detachment faults are related to the lower and middle breccias (p. 27) present locally in lower strata of the Madison is still unknown. The lower breccia, at the base of the Madison group, involves rocks presumably at or near the horizon of the lower detachment fault, and the middle breccia, which is at least 300 feet higher, involves rocks presumably cut by the upper detachment fault(?). Whether this apparent parallelism between breccias and detachment faults is meaningful is unknown. Probably it is fortuitous, for where the evidence for the proposed detachment faults is strongest, in the Bacon Rind cirque, the breccias are absent. Further, two silicified breccias east of the Bacon Rind fault apparently formed at some time after that fault broke the rocks (p. 80). The age of the Bacon Rind fault, based on tenuous evidence, is probably middle Tertiary, and the detachment faults are probably much older, for they are folded and faulted along with the overlying and underlying strata. Hence, the detachment faults are older than the regional folding, which involves beds as young as Paleocene or Eocene (p. 77). Their age, therefore, is provisionally set as post-Early Cretaceous and pre-early Eocene.

As of 1965 the best that can be said concerning the existence of the two proposed detachment faults is that conclusive evidence is not available.

Pending the results of further work I suggest that the Madison Group ranges in thickness from about 1,000 to about 1,450 feet. The lower figure probably represents a local attenuation of the unit, and the higher figure may be the result of either minor bedding-plane slips or some duplication of beds along detachment faults.

Minor bedding-plane slips.—An alternative explanation for the abrupt variations in thickness of the various sedimentary units invokes minor bedding-plane slips or faults, none large enough to be readily detectable in the field, yet each, small though its displacement may be, thickening or thinning the unit a minor amount. The cumulative result is probably reflected in the folds as attenuated flanks and thickened crests.

As the sedimentary strata were flexed and folded each bed slipped somewhat as it adjusted to the newly imposed conditions. This minor movement, which occurred chiefly along bedding planes, left its mark as slickensides, and judging by these, every sedimentary unit in the area was involved.

These bedding-plane slips may have been more common in thin-bedded formations than in thick-bedded or massive formations. The thicker units, being more competent, may have adjusted to the compressive stresses by some other method, such as plastic flow or rotation of grains. Possibly the best analogy would be to compare the thin-bedded Lodgepole Limestone or the thin-bedded Quadrant Sandstone to a deck of playing cards which has been sharply flexed into a convex-upward fold. Each card moved slightly past the underlying one to result in a thickened crest and thinned flanks. The voids in the crest were probably filled largely by "shingling" of the adjacent beds along minor bedding plane slips.

Probably neither thickness represents the true depositional thickness. The lesser thickness probably reflects some attenuation on the limbs of the fold, whereas the greater thickness may well denote a thickening of beds on the crest. This is borne out by the Quadrant Sandstone example first cited: where the unit forms the limb of the Kirkwood anticline (in Red Canyon) it is only 265 feet thick; where it forms the crest of the anticline (on Kirkwood Ridge) it is 625 feet thick.

AGE OF THE LARAMIDE FOLDING

In this area the youngest rocks affected by Laramide deformation are the shales of the Thermopolis(?) Shale of Early Cretaceous age. To the north, however, still younger rocks have been deformed. Thus, northwest of the Three Forks Basin, the Elkhorn Mountains Volcanics of Late Cretaceous age are almost as deformed as the underlying Paleozoic rocks (Robinson, 1961, p. 1007); this suggests that the age of the Laramide is at least post-Late Cretaceous. Further, Lowell and Klepper (1953) indicated that the Beaverhead Formation (of southwestern Montana), part of which is late Paleocene and Eocone as well as Late Cretaceous, was folded and displaced by overthrusting, and that locally it is overlain by thrust sheets of Paleozoic rocks. So, Laramide folding may have persisted through the late Paleocene.

Beck (1960, p. 134) and Hall (1961, p. 145) reported that the Sphinx Conglomerate (in the Madison Range), possibly correlative with the Beaverhead Formation, locally is strongly deformed by sharp folds and low-angle thrust faults. The inference is strong, then, that the Sphinx was involved in Laramide folding. As it rests on beds dated as Paleocene, it has been considered to be late Paleocene or early Eocene in age (Beck, 1960, p. 133) and to be Eocene in age (Robinson, 1963). Thus, beds as young as early Eocene may have been involved in Laramide folding.

In the Livingston area, Roberts (1965) found evidence that the Laramide deformation began possibly as early as Eagle (Late Cretaceous) time and persisted through the Paleocene and possibly through Wasatch (early Eocene) time.

It would seem fair to conclude that in southwestern Montana the Laramide deformation began in very Late Cretaceous time and continued through the Paleocene and probably into the early Eocene.

STRUCTURES OF MIDDLE TERTIARY AGE

At some time during the middle Tertiary (Miocene?) southwestern Montana may have been arched and uplifted by renewed deep-seated lateral compression (Pardee, 1950) and then, as the compressive forces relaxed, broken into huge fault blocks along high-angle normal faults (p. 65). Three of these fault blocks extend into this quadrangle: the Red Canyon, Kirkwood, and Monument Mountain blocks (fig. 22). The north edge of the Red Canyon block is delineated by the Red Canyon fault, and the north edge of the Kirkwood block is delineated in part by the Upper Tepee fault (p. 79).

RED CANYON FAULT

The Red Canyon fault is the best known of the normal faults because its reactivation on August 17, 1959, was most likely a major contributory factor to the Hebgen Lake earthquake (Witkind, 1964b, p. 38–40). The trace of the fault throughout the southwest corner of the quadrangle and to the west beyond the mapped area is now marked by one or more well-defined fault scarps, locally as high as 22 feet and dipping valleyward (fig. 24). The fault is nowhere exposed, for it is concealed by colluvium along its entire length. The scarps, formed as a result of bedrock movement along the concealed fault, are all in surficial deposits. All evidence indicates that the Red Canyon fault block (south of the Red Canyon fault) was downthrown—tilted northeastward—all along that fault.

In plan view the Red Canyon fault is convex to the northeast and outlines, from west to east, the south flank of Kirkwood Ridge, the west flank of Dutchman Ridge, and the nose of the Dutchman Basin syncline (pl. 1). Westward, beyond the mapped area, the fault follows the south flank of Kirkwood Ridge and ends in Kirkwood Canyon in the SW¼ sec. 19, T. 11 S., R. 4 E. Because of inadequate exposures, the southeastward extent of the Red Canyon fault can only be surmised. On the basis of juxtaposition of sedimentary units, the fault reaches as far as Dry Wash, where it disappears below the Grayling earthflow (pl. 1). Southeast of Dry Wash the fault, if present, is concealed beneath surficial deposits and younger volcanic rocks, and its course on plate 1 is based on the fault scarps formed during the earthquake, which end just inside Yellowstone National Park near sec. 22, T. 12 S., R. 5 E.

The northeast flank of the West Yellowstone basin may have been determined by the Red Canyon fault, much as the east flank of the Madison Valley was determined by the Madison Range fault (p. 64). If so, the Red Canyon fault, concealed by glacial outwash and volcanic rocks, extends at least another 5 miles to the east.

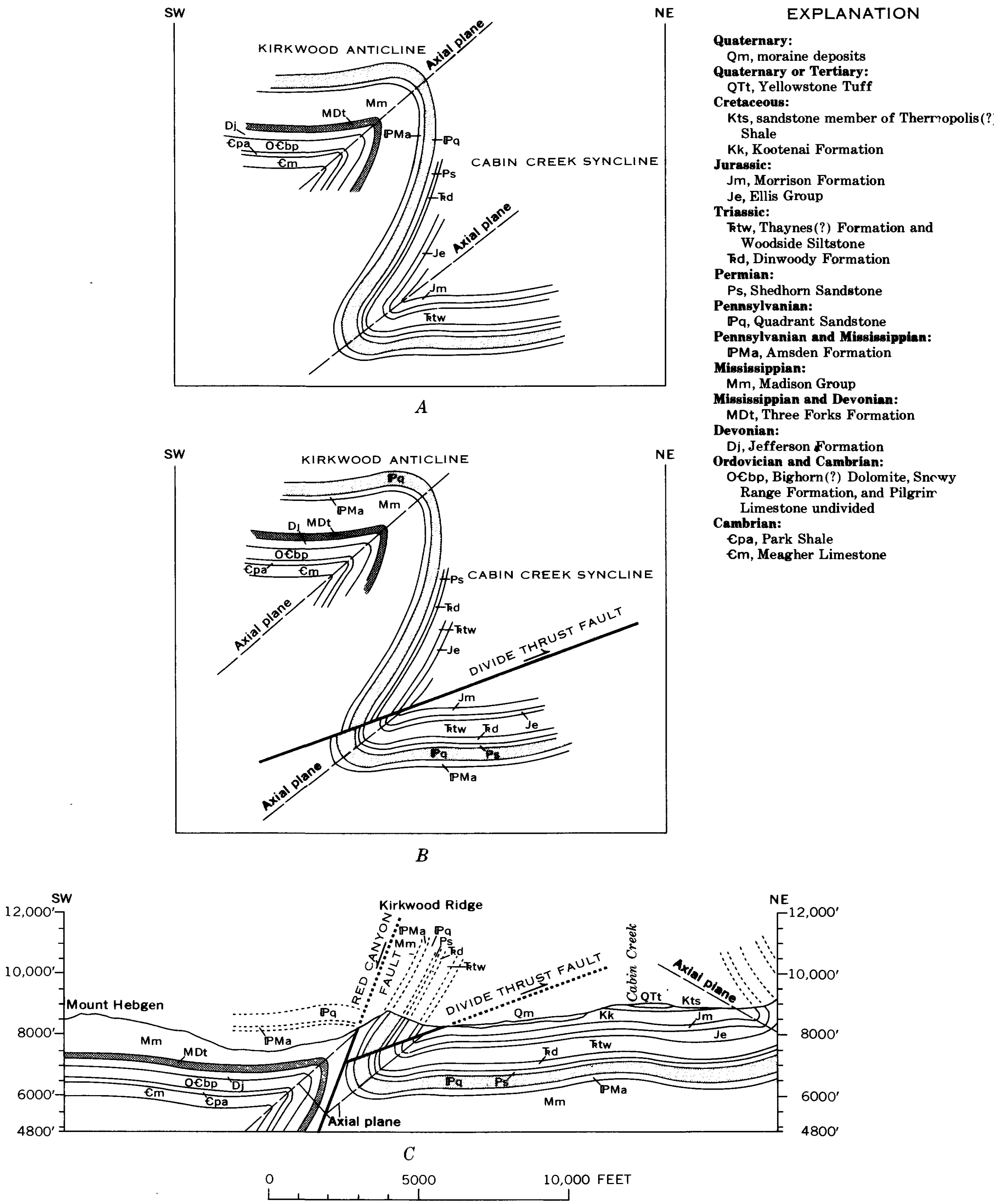
Patently not all parts of a fault system will be active at the same time. A fault system some 60 miles in length, for example, may be reactivated at any instant along only 10–20 miles of its length. The stresses relieved at that time and place must then begin to accumulate elsewhere along the fault, and the next movement will be in a new area of maximum accumulated stress. Over a long period of geologic time the whole length of a fault system will be active, but during specific shorter episodes of geologic time only part of the system will be sufficiently stressed to be considered active. In a sense, movement along the fault is spasmodic, but given enough time, movement will occur along the full length of the fault. In this view, the fault scarps formed during the Hebgen Lake earthquake of 1959 resulted from reactivation of but the northwestern part of the whole Red Canyon fault system.

The Red Canyon fault as traced is about 14 (possibly 19) miles long, but only about 10 miles of this length is in the Tepee Creek quadrangle.

The Red Canyon fault dips valleyward and is normal for its entire length. Its angle of dip is uncertain, although I estimate it to be about 70° southward. The fault scarps generated in the overlying surficial material dip as much as 85° valleyward, but this increased steepness is directly attributable to the unconsolidated surficial deposits that form the scarps.

Throughout much of its length the fault juxtaposes strata of the Madison Group. Commonly the rocks in the downthrown south block are upright and dip 15°–45° into the fault. By contrast, the beds in the relatively upthrown north block are almost invariably vertical, overturned, or steeply dipping. These relations seem explicable only by assuming that the strata south of the Red Canyon fault constitute the upright limb of an overturned anticline which has been downthrown repeatedly along the fault. The steeply dipping and overturned strata north of the fault constitute the overturned limb of the same anticline (fig. 28).

As far as can be determined, virtually all movement on the Red Canyon fault has been dip slip; little or no strike-slip movement has occurred. In essence, the northeast edge of the Red Canyon fault block is slowly subsiding along the Red Canyon fault, so that the block is persistently being tilted to the northeast (fig. 10). This northeastward tilting may characterize the Kirkwood fault block also. If so, the tilting there is a result of dip-slip movement on the Upper Tepee normal(?) fault


SW
NE
KIRKWOOD ANTICLINE
Axial plane
Mm
MDt
Dj
€pa
O€bp
€m
PMa
Pq
CABIN CREEK SYNCLINE
Ps
Ƭd
Je
Jm
Ƭtw
A
SW
NE
KIRKWOOD ANTICLINE
Pq
PMa
Mm
Dj
MDt
O€bp
€pa
€m
Ps
CABIN CREEK SYNCLINE
Ƭd
Ƭtw
Je
Axial plane
DIVIDE THRUST FAULT
Jm
Ƭtw
Ƭd
Je
Pq
Ps
PMa
Axial plane
B
EXPLANATION
Quaternary:
Qm, moraine deposits
Quaternary or Tertiary:
QTt, Yellowstone Tuff
Cretaceous:
Kts, sandstone member of Thermopolis(?) Shale
Kk, Kootenai Formation
Jurassic:
Jm, Morrison Formation
Je, Ellis Group
Triassic:
Ƭtw, Thaynes(?) Formation and Woodside Siltstone
Ƭd, Dinwoody Formation
Permian:
Ps, Shedhorn Sandstone
Pennsylvanian:
Pq, Quadrant Sandstone
Pennsylvanian and Mississippian:
PMa, Amsden Formation
Mississippian:
Mm, Madison Group
Mississippian and Devonian:
MDt, Three Forks Formation
Devonian:
Dj, Jefferson Formation
Ordovician and Cambrian:
O€bp, Bighorn(?) Dolomite, Snowy Range Formation, and Pilgrim Limestone undivided
Cambrian:
€pa, Park Shale
€m, Meagher Limestone
SW
NE
12,000′
10,000′
8000′
6000′
4800′
Kirkwood Ridge
RED CANYON FAULT
PMa
Pq
Mm
Ps
Ƭd
Ƭtw
DIVIDE THRUST FAULT
Cabin Creek
Axial plane
Mount Hebgen
Pq
PMa
Mm
MDt
O€bp
Dj
€m
€pa
Axial plane
Qm
Kk
QTt
Kts
Jm
Je
Ƭd
Ƭtw
Pq
Ps
PMa
Mm
C
0
5000
10,000 FEET

(fig. 28*C*), which bounds part of the northeast flank of the Kirkwood fault block.

The displacement on the Red Canyon fault, as might be expected, increases toward the midpoint of the fault's length and dies out near its ends. Thus, near the southeast end of the fault (around the nose of the Dutchman Basin syncline) the displacement is on the order of several hundred feet. At the west end of Kirkwood Ridge, near the west end of the fault (beyond the mapped area), the displacement is also low, not exceeding several hundred feet. But along the middle segment of the fault, which extends along Kirkwood and Dutchman Ridges, the displacement is on the order of several thousand feet; the exact amount is indeterminable. In cross sections *C–C′*, *B–B′*, *E–E′*, and *D–D′*, plate 1, I have shown estimated displacement of several thousand feet; these figures can be drastically altered by modifying the configuration of the faulted anticline.

This pattern of displacement is also reflected in the fault scarps formed during the 1959 earthquake. The scarps with the greatest displacement—15–22 feet—are confined chiefly to the south and west flanks of Kirkwood and Dutchman Ridges. Away from these places the scarps lessen in height and end finally as small fractures or monoclinal flexures.

UPPER TEPEE FAULT

If the Upper Tepee fault is normal, much as suggested in figure 26*C*, it may delineate part of the northeast edge of the Kirkwood fault block (fig. 10). The fault is exposed along the southwest margins of both Upper Tepee and Tepee Basins (secs. 1 and 12, T. 11 S., R. 4 E.; sec. 7, T. 11 S., R. 5 E.), as well as along the south flank of the Bacon Rind cirque directly north of White and Redstreak Peaks (pl. 1). It can be traced southeastward for about 3 miles from its northwest end near Redstreak Peak to its disappearance below morainic deposits in Tepee Basin. How far it continues southeastward beyond this point is unknown.

FIGURE 28.—Structural development of the Mount Hebgen-Kirkwood Ridge-Cabin Creek area. This figure is part of cross section *C–C′*, pl. 1.

A. The Kirkwood anticline, a major parental fold overturned toward the northeast, rested on the tightly folded southwest limb of the Cabin Creek syncline.

B. As the lateral compressive forces from the southwest continued, the northeast limb of the Kirkwood anticline broke, and the anticline was shoved northeastward along the low-angle Divide thrust fault.

C. During the middle Tertiary the area was arched. Differential settling resulted in normal faults whose general trend and attitude were determined by zones of weakness established during the Laramide tectonic episode. The upright limb of the overturned Kirkwood anticline was let down along the Red Canyon fault.

In general, the Upper Tepee fault trends about N. 45° W. and is 500–1,000 feet northeast of the White Peak fault, which it closely parallels. Its trace along the northeast flank of Skyline Ridge near Tepee and Upper Tepee Basins is almost straight, which suggests that there the fault is nearly vertical; but near Redstreak and White Peaks its trace is faintly curved, convex to the northeast, which suggests a very steep southwestward dip.

The displacement along the Upper Tepee fault increases southeastward, but its amount, as with that of the Red Canyon fault, is uncertain; I estimate it to range from but a few tens of feet near Redstreak Peak to several thousand feet in Tepee Basin. If the fault is normal, the movement apparently was wholly dip slip; no evidence was found for any strike-slip component.

Near its southeastern end the Upper Tepee fault juxtaposes Precambrian tremolite marble and mica schist northeast of the fault with overturned beds of the undifferentiated Pilgrim Limestone, Snowy Range Formation, and Bighorn(?) Dolomite southwest of the fault (cross section *C–C′* pl. 1). At its northwestern end the fault truncates the south limb of the Skyline anticline, formed of strata which range in age from Cambrian to Mississippian. Here the sedimentary beds northeast of the fault are right side up and dip southwestward toward the fault; they are juxtaposed with vertical or overturned beds of Devonian and Mississippian age southwest of the faults (pl. 1). The beds northeast of the fault are part of the northwest-plunging Skyline anticline, whose core has been dissected by streams and ice to form the Bacon Rind cirque (fig. 5).

The exact relation between the Upper Tepee fault and the northwest end of the White Peak fault is concealed beneath talus and morainic deposits. Possibly the Upper Tepee fault cuts the White Peak fault (p. 71).

The Upper Tepee fault was probably formed at the same time and in response to the same tensional forces as the Red Canyon fault. Its age, therefore, is considered to be middle Tertiary; and the fault, much like the Red Canyon fault, has probably been active since that time. Although part of the Red Canyon fault was reactivated in 1959, as far as can be determined no movement occurred then along the Upper Tepee fault.

BACON RIND FAULT

The strata along the north valley wall of Bacon Rind Creek (near its junction with Migration Creek) are broken by the Bacon Rind fault, which trends about N. 25° W. for half a mile and is either vertical or dips southwestward at a very high angle (pl. 1). The strata southwest of the fault are downthrown possibly as much as 200 feet.

The northwest end of the fault passes into folded Madison strata (pl. 1); beyond this point the fault could not be traced with any assurance. May (1950, p. 29), however, suggested that the fault extends to a point north of Snowslide Creek. The southeast end disappears below the colluvium which flanks Bacon Rind Creek. As far as can be determined, the fault does not break the strata along the southeast valley wall.

This fault would seem of but little consequence were it not for the abrupt appearance of two silicified breccias in Lodgepole strata directly northeast of it (p. 27). The strata on the southwest (downthrown) side of the fault are whole and seemingly unaltered. The basal beds of the Lodgepole Limestone rest on a black shale which in turn rests on what seems to be the full thickness of the Three Forks Formation. But directly northeast of the fault, on the relatively upthrown block, two well-developed silicified breccias appear, one at the base of the Lodgepole and the second about 300 feet higher, also in Lodgepole strata. The basal breccia may include boulders of the Sappington(?) Member of the Three Forks Formation (p. 23).

The two breccias can be traced northeastward; the basal breccia passes below the flood plain of the Gallatin River, and the higher breccia continues along the west valley wall of the Gallatin River between the mouths of Bacon Rind and Snowslide Creeks. Neither of the breccias is present along the east valley wall of the Gallatin River, although a remnant of the higher breccia (unsilicified) may be preserved on the small knoll near mile post 275 along the Montana-Wyoming State line.

The abrupt appearance of these breccias east of the Bacon Rind fault and their absence west of the fault suggests that they were formed after the fault had broken the rocks. One possibility is that the Bacon Rind fault served as a channel for silica-rich surface and ground waters and that these waters moved down the fault and then spread laterally downdip (north and northeastward) through permissive beds in the Madison, gradually selectively leaching them and causing them to collapse and form breccias. Silicification of the breccias followed.

The age of the fault is uncertain; as it is normal, I have grouped it with structures of middle Tertiary age. This age assignment (Miocene?) predates the extensive pyroclastic flows tenuously dated as either late Pliocene or early Pleistocene (p. 53).

GEOLOGIC HISTORY

The sedimentary units in southwestern Montana were formed in a tectonic framework dominated by two major structural elements: the Cordilleran geosyncline and the Wyoming shelf (p. 8). Both were established during the Precambrian and persisted throughout the Paleozoic and part of the Mesozoic. In this tectonic setting the Tepee Creek area was but a small part of a platformlike hinge between the deep trough of the geosyncline to the west and the Wyoming shelf to the east. The rocks exposed, dominantly marine carbonates and siltstones, attest that for much of this time this area was beneath shallow marine seas.

From Middle Cambrian time to at least the end of the Triassic the thickness, type, and extent of the sediments deposited on this hinge platform were determined by its temporary oscillations. When the platform subsided the seas spread eastward, and limestones and dolomites rich in marine forms were laid down; when the platform rose, forcing a temporary shoaling of the marine waters, muds, silts, and sands were deposited.

During the Laramide deformation, these structural elements were destroyed, and southwestern Montana, along with much of the western part of the North American continent, was elevated above sea level. In the middle Tertiary(?), epeirogenic uplift arched the area, and blocks of ground settled differentially along normal faults whose general trend was probably determined by zones of weakness established during the compressive episode of Laramide orogeny. Since then, subaerial erosion has carved the folded and faulted Paleozoic and Mesozoic strata into the physiographic features which characterize southwestern Montana today.

The oldest rocks exposed in the area are very ancient Precambrian crystalline metamorphic rocks. These metamorphics, probably correlative with the "Cherry Creek beds" (Peale, 1896) exposed near Ennis, Mont., form the basement complex. They are composed of dark-green amphibolite, gray granitic gneiss, and dark-gray mica schist, interrupted here and there by beds of light-colored dolomite, tremolite marble, and quartzite. West of this area, metamorphosed granites form the core of the Madison Range.

Some 100 miles to the north, near Three Forks, Mont., comparable crystalline metamorphics are overlain by less intensely metamorphosed beds of carbonate, argillite, quartzite, and, locally, coarse clastics, which make up the Belt Series (Robinson, 1963, p. 9–14); but if Belt rocks ever extended as far south as the Tepee Creek area, they have long since been removed. McMannis (1963) suggested that Belt rocks not only never covered this area but that the area may have been part of a major highland which supplied some of the coarse clastics (LaHood Formation) of the Belt Series.

During Early Cambrian time much of Idaho, Montana, Wyoming, and southern Canada was uplifted to form the island 'Montania" (p. 8), which effectively but temporarily divided the Cordilleran geosyncline into north and south parts (Deiss, 1941, p. 1088). Montania may have persisted into the late Paleozoic, but by Middle Cambrian time the once-large island had been sharply curtailed in size, probably as a result of both erosion and tectonic subsidence. In this quadrangle, degradational processes had removed an unknown thickness of the Precambrian metamorphosed crystallines.

During the Middle Cambrian, marine waters spread eastward from the geosyncline across much of southwestern Montana. The Precambrian crystallines were submerged, and the erosional detritus was reworked and then redeposited to form the coarse clastics now known as the Flathead Sandstone (p. 15). As the Wyoming shelf continued to subside slowly but continuously, the strand line migrated eastward, and as the waters gradually deepened over this area finer grained sediments and carbonates were deposited. Thus, the medium- to coarse-grained sandstone beds of the uppermost Flathead grade into the overlying fine-grained sandstone and siltstone beds of the Wolsey (p. 15), which in turn grade into the carbonate sequence of the Meagher (p. 16).

A minor but very temporary uplift caused a brief shoaling of the waters, and the dark-gray to black Park Shale (p. 17) was deposited. Soon the platform once again sank below deeper waters, and the carbonates which now make up the Pilgrim Limestone (p. 18) were precipitated. A transient general emergence resulted in some erosion of Pilgrim strata, but with renewed submergence the limy muds, mud-pebble conglomerates, and glauconite which now form the Snowy Range Formation (p. 19) accumulated.

Shallow marine seas persisted across this area throughout much of the Ordovician, except once during Early Ordovician when a short, rapid emergence resulted in removal of a thin sheet of newly deposited carbonate. By Late Ordovician time a thick wedge of limestone and dolomite (the Bighorn Dolomite) had been deposited across the Wyoming shelf.

At some time after the Ordovician the Wyoming shelf was raised, and it remained above sea level throughout the Silurian and part of the Early Devonian. During this emergence—the first of three major uplifts of this area during the Paleozoic and early Mesozoic—most of the Ordovician and possibly some Upper Cambrian strata were removed. By Middle Devonian time, however, the shelf had subsided, and again the seas migrated eastward to deposit the carbonates now known as the Jefferson Formation (p. 21).

Lagoonal conditions existed briefly across southwestern Montana, probably as the result of a brief rise of the Wyoming shelf, and an evaporite sequence—the Logan Gulch Member (p. 22) of the Three Forks Formation—accumulated. This episode was short lived, and as communication with the open seas was reestablished, fine muds which now form the dark-gray fossiliferous shales (Trident Member, p. 22) of the upper part of the Three Forks Formation were deposited. A minor irregular oscillation of the shelf resulted in the deposition of limy silts and then, as the shelf sank, fine muds. These deposits now form the Sappington Sandstone Member of the Three Forks Formation (p. 22) and a black shale unit (p. 22), respectively.

At this time there probably began an episode of uplift which persisted long enough to permit almost total erosion of the uppermost beds of the Three Forks Formation. Across much of this area almost all the Sappington(?) Sandstone Member plus the underlying characteristic dark-gray fossiliferous shale were shredded and removed, and the area probably appeared as a somewhat monotonous plain speckled here and there with small isolated hills on which were preserved remnants of the uppermost Three Forks strata.

This erosional interval ended as the shelf sank once again, and for much of Mississippian time the area was awash beneath warm shallow seas in which marine life flourished. The thick sequence of fossiliferous carbonates deposited then, now known as the Madison Group (p. 23), dominates the stratigraphic sequence.

During the Late Mississippian, widespread regional uplift occurred, and this area underwent its second major episode of erosion. As a result, a surface of low relief was developed across Madison strata. In time the Wyoming shelf subsided somewhat, and the seas again encroached on the land. In this near-shore environment the silts and fine sands of the Amsden (p. 29) were deposited. Locally, small parts of the shelf were flexed down, and in the resultant deeper basins limy muds settled; these eventually became the mottled limestone so characteristic of the Amsden. The Wyoming shelf remained stable through Amsden time, and this area along with large parts of southwestern Montana and Wyoming was either just above sea level or barely awash.

Clean quartz sand was carried into these shallow waters by streams which drained adjacent highlands. These sands, now consolidated, make up the Quadrant Sandstone (p. 30) of Pennsylvanian age.

In Early Permian time the Wyoming shelf sloped westward into the miogeosyncline, and the clastic sediments which now make up the bulk of the Phosphoria Formation accumulated on this shoaling sea floor (McKelvey and others, 1956, p. 2829). In the miogeosyncline, bedded cherts, phosphatic shales, and mudstones were deposited; these thinned and changed lithology to the north and east. In the report area the dominant lithology is represented by the Shedhorn Sandstone (p. 31).

During Early Triassic time the first stage in what eventually would be the total destruction of the Cordilleran geosyncline occurred far to the south in Nevada, where a north-trending geanticline separated the geosyncline into eastern and western troughs (Nolan, 1943). This tectonic activity, however, restricted almost wholly to Nevada and southern Utah, did not reach as far north as southwestern Montana, where the Early Triassic was marked by two extensive marine inundations separated by an episode of shallow-water deposition. The first marine advance, represented by the Dinwoody Formation (p. 32), was from the Southwest, and the seas reached northward and northeastward across Montana into Canada; siltstone and limestone piled up on the Wyoming shelf. As the seas withdrew southward, deposits from highlands to the east were spread rapidly across the westward-sloping Wyoming shelf, burying the Dinwoody Formation. These continental deposits, chiefly red beds, make up the Woodside Siltstone (p. 32). This interval of shallow-water deposition was halted by the second marine advance, during which the claystone and limestone of the Thaynes Formation (p. 34) were deposited. The Early Triassic closed as the seas withdrew westward into the miogeosyncline.

Middle or Upper Triassic rocks probably were not deposited in this area, but lengthy and widespread pre-Jurassic erosion stripped many of the Lower Triassic rocks from west-central and northwestern Montana, even as the Thaynes or parts of it were locally removed from southwestern Montana. Probably Thaynes strata were once amply represented in this area, but most were worn away during this interval—the third major emergence of the Tepee Creek area.

By Early Jurassic time much of southwestern Montana had been reduced to a low, featureless plain that sloped upward to a highland in western Montana, centered in eastern Beaverhead and western Madison Counties (Moritz, 1951, p. 1805). During Middle Jurassic time, an inland marine sea spread southward from Canada almost to Nevada and inundated much of Montana and adjacent parts of Canada and Wyoming (Imlay, 1956); only central Montana remained above water as an irregular-shaped island of low relief known as "Belt island" (Imlay and others, 1948). Sediments deposited in the shallow waters southwest of this island are represented in this quadrangle by the Ellis Group (p. 34).

Upon withdrawal of the sea in Late Jurassic time, southwestern Montana appeared as a broad plain dominated by "Belt island", which then included all of central and western Montana. Continental fluvial deposits were spread across this plain by large sluggish streams and rivers meandering in tree-lined valleys and heading in the highland; swamps and ephemeral freshwater lakes supporting rich vegetation were widespread. These deposits now compose the Morrison Formation (p. 35).

Near the end of the Late Jurassic the western interior of the United States was gradually elevated as the Cordilleran geosyncline was destroyed. Somewhere west of the Tepee Creek area, possibly along the Pacific Coast (Lammers, 1939, p. 127), a major landmass must have formed, and as erosion attacked the raised areas, much of the recently deposited Morrison Formation and part of the underlying Ellis Group were removed. The coarse clastics shed by this western upland were repeatedly reworked until only the most durable materials remained as well-rounded pebbles and cobbles. These deposits are represented in this quadrangle by the basal conglomerate and conglomeratic sandstone of the Kootenai (p. 36). As this western landmass was lowered, finer sediments were laid down, and these now form the sandstone and claystone parts of the Kootenai (p. 36).

A new marine invasion spread across the western interior, reaching southward from the Arctic and northward from the Gulf of Mexico to join finally as an extensive inland sea. The initial deposits of this transgressing sea appear in this area as massive sandstone, the basal sandstone member of the Thermopolis(?) Shale (p. 37). As the waters deepened, dark-gray to black shales accumulated, and subsequent periodic fluctuations of sea level resulted in differing lithologies. These now compose the upper part of the Thermopolis(?).

The youngest sedimentary rocks exposed in this quadrangle are the black shales of the Thermopolis(?), but north of the mapped area Hall (1961, p. 79–96) reported Tertiary deposits possibly as much as 8,000 feet thick.

During or shortly after the Late Cretaceous the full effects of the Laramide deformation were felt in southwestern Montana. The destruction of the Cordilleran geosyncline, which had begun in Nevada in Early Triassic time, now reached northward. The miogeosyncline of eastern Idaho probably was destroyed during the Late Jurassic, and by Late Cretaceous the geosynclinal prism of sediments in southwestern Montana was slowly being crushed and elevated by lateral compressive forces.

In the general region near the Tepee Creek quadrangle, a mass of Precambrian crystalline rocks was shoved northeastward along the low- to moderate-angle Beaver Creek thrust fault, intensely deforming the sedimentary rocks directly in front of it. As a result, three tectonic elements can be perceived: the thrust block (upper plate), composed mainly of Precambrian rocks and known as the Madison thrust block (p. 65); the intensely deformed Paleozoic and Mesozoic strata (lower plate), formed in front of the thrust block and known as the Cabin Creek zone (p. 65); and the sedimentary strata which were only slightly deformed because they were distant from the thrust block, known as the Pika Point zone (p. 65).

Shortly thereafter, possibly during very Late Cretaceous, Paleocene, or early Eocene time, but likely before early middle Eocene time ((p. 44), the strata in the northeast corner of the quadrangle were deformed by small intrusive bodies of dacite porphyry, including the Gallatin River laccolith (p. 39).

These deformed strata were partly eroded and then buried in early middle Eocene time by flows of basalt like shoshonite (p. 44) which spread across the eastern parts of the quadrangle from Yellowstone National Park. These flows were accompanied by eruptions of andestic material (p. 45).

At some time in the middle Tertiary (Pardee, 1950), possibly during Miocene time, the area was arched as a result of renewed deep-seated lateral compression. As relaxation set in, the area subsided unevenly along high-angle normal faults to form tilted fault blocks. Three such blocks are in this area, and two of these, possibly all three, are tilted northeastward. The southernmost block, named the Red Canyon block, is bounded along its northeast margin by the Red Canyon normal fault (p. 77). Repeated subsidence of this block along that fault has tilted the block decidedly northeastward. The middle block, named the Kirkwood block, in like manner has probably subsided along the Upper Tepee normal (?) fault which bounds part of its northeast margin (p. 79). The northernmost block, called the Monument Mountain block, extends northward beyond the mapped area; I think it has been tilted, or at least warped, northeastward.

The attitude and position of these normal faults, which parallel the earlier Laramide thrusts, seem to have been strongly influenced by zones of weakness established during the Laramide tectonic episode.

In the late Tertiary or early Quaternary (either late Pliocene or early Pleistocene) major fissure eruptions occurred throughout the northwest corner of Yellowstone National Park, and large quantities of volcanic debris, both as ash fall and as ash flow, spread westward into this area. Most of the pyroclastic detritus was spread as incandescent ash flows, which have consolidated into the rhyolitic welded tuffs, named the Yellowstone Tuff (p. 45). The entire quadrangle may have been covered by these flows (p. 48).

Before Pleistocene ice spread across this area, it seems likely that the following major events had occurred: (1) Folding followed by thrust faulting which probably began in this general area during the very Late Cretaceous, persisted through the Paleocene, and may have continued through all or part of early Eocene time; (2) emplacement of intrusions shortly thereafter, most likely during the early Eocene but definitely before the early middle Eocene; (3) partial burial of the eastern part of the quadrangle by andesite breccia and shoshonite flows during the early middle Eocene; (4) epeirogenic uplift accompanied by normal faulting during the middle Tertiary, most likely in Miocene time; and (5) pyroclastic eruptions of rhyolite tuff possibly during either the late Pliocene or early Pleistocene.

During the Pleistocene, all or parts of this quadrangle were overridden by glaciers a minimum of three times: at least once during pre-Bull Lake time, once during Bull Lake time, and once during Pinedale time. The earliest ice (pre-Bull Lake) may have reached from the crest of the Madison Range on the west to the crest of the Absaroka Range on the east (Hall, 1959); if so, most of this area was buried beneath ice at that time.

During the Bull Lake Glaciation, glaciers spread westward from the Absaroka Range and sent lobes of ice down the major valleys. One lobe, the Horse Butte lobe, moved into the West Yellowstone basin and reached as far west as Horse Butte (p. 55); a second lobe, the Grayling lobe, followed the Gallatin River valley as far west as U.S. Highway 191 (p. 55). Grayling Creek, a north-flowing tributary to the Gallatin River, was diverted southward by the Grayling lobe to become tributary to the Madison River (p. 58); fig. 16).

The Pinedale Glaciation is represented by drift deposited by valley glaciers along the floors and walls of the major valleys. These glaciers formed in the high ranges and then flowed radially down adjacent valleys. Locally, a few glaciers reached beyond the mouths of the host valleys and merged; most, however, remained in the valleys (fig. 15).

Since then, several small ice masses have formed in the cirques developed during the Bull Lake and Pinedale Glaciations, but these glaciers never reached much beyond the mouths of the cirques.

Deeply weathered rocks and oversaturated morainic deposits have slid downslope locally to form landslides (p. 62) and earthflows (p. 60). Other mass-wasting deposits are represented by rockslides and rockfalls still forming today. Colluvial debris, chiefly as a thin sheet, and alluvium, both along streams and at their mouths, cover more than 90 percent of the area.

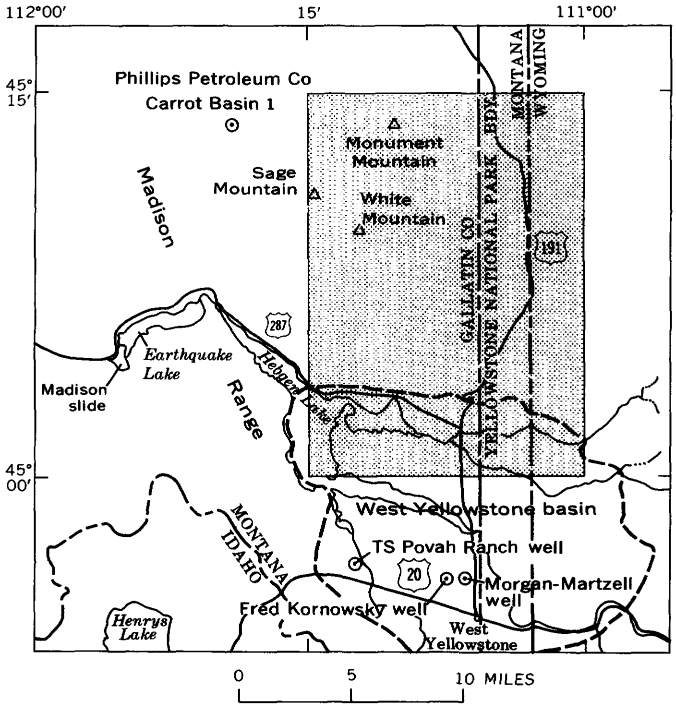


FIGURE. 29.—Location of the four wells drilled to test the oil and gas possibilities of areas near the Tepee Creek quadrangle. The approximate limits of the West Yellowstone basin are shown by heavy dashed line; the Tepee Creek quadrangle is stippled.

ECONOMIC GEOLOGY

OIL AND GAS POSSIBILITIES

As of 1963 the oil and gas possibilities of this quadrangle were still untested and thus open to speculation. Four test wells had been completed west and south of the quadrangle, but all were unsuccessful and abandoned (fig. 29). Three of these, in the West Yellowstone basin, are 4–5 miles south of the south edge of the quadrangle and were drilled mainly on the basis of small oil shows in shallow water wells. None of the tests passed through the entire basin fill of volcanic rocks and outwash obsidian sand and gravel. The fourth well, in the SE¼NW¼SE¼ sec. 11, T. 10 S., R. 3 E. (3 miles west of the northwest corner of the quadrangle), tested the gently folded sedimentary strata of the Carrot Basin anticline (fig. 30; p. 85), also known as the Pika anticline (Hall, 1961, p. 146, 173) or Pika Mountain anticline (Hume and Leeder, 1950, p. 26). This test was abandoned as a dry hole after penetrating about 220 feet of the fractured limestone beds of the Madison Group (Tutten, 1960, p. 261).

WEST YELLOWSTONE BASIN

Shortly after the Helena earthquakes of 1935, several shallow water wells in and near West Yellowstone were contaminated by oil, and this renewed interest in the oil possibilities of the West Yellowstone basin. Since then three wells have tested the basin (fig. 29); none penetrated the entire fill, and in not one was there any show of oil or gas.

Morgan-Martzell well.—Of the three, the Morgan-Martzell well (fig. 29), sometimes referred to as the Government No. 1 or the J. H. Morgan well, has supplied most of the available information about the basin fill, for more than 1,000 feet of the total depth of 1,390 feet was cored discontinuously. W. B. Myers, of the U.S. Geological Survey, obtained the cores from William Martzell of West Yellowstone during the intensive geologic investigations of the Hebgen Lake area following the 1959 earthquake.

The Morgan-Martzell well is about 1½ miles north of West Yellowstone in the center of the SW¼SW¼ sec. 22, T. 13 S., R. 5 E. The well was begun in October 1953 and completed in the same year; no shows of oil or gas were reported.

The following log is my resume of a more detailed log prepared from these cores by Warren Hamilton (1964, p. 215–216). The upper 375 feet is taken from the driller's log of the Kornowsky well in the NE¼SE¼ SW¼ sec. 21, T. 13 S., R. 5 E., about three-fourths of a mile west of the Morgan-Martzell well.

Collar of well, altitude about 6,645 ft.	Thickness (feet)	Depth to bottom of unit (feet)
Aluvium	200	200
"Red lava"	100	300
"Lava"	75	375
Basalt or andesite, dark-gray, nonporphyritic	100	475
Rhyolite welded tuff, light-gray; composed of squashed and fused shards with many cavities	550	1,020±20
Perlite, light-gray; contains many phenocrysts of sanidine and sparse ones of oligoclase	70	1,090
Rhyolite welded tuff; crystals of sanidine and fewer of quartz in a matrix of grayish squashed and fused shards of perlitic glass	30	1,120
Water-laid rhyolite tuff; contains angular and unsquashed clasts of glass and fewer of quartz and sanidine in gray to brown matrix of slightly altered vitric ash; abundant diatoms and pollen	35	1,155
Olivine basalt, medium-dark-gray, dense; composed of plagioclase (An_{50}), augite, some olivine, and ilmenite	15	1,170
Rhyolite tuff, medium-gray, stony; not welded	15	1,185
Augite andesite, medium-gray, dense; composed of labradorite phenocrysts in a groundmass of andesine laths; some augite, ilmenite, magnetite, and carbonate	10	1,195

Collar of well, altitude about 6,645 ft.—Continued	Thickness (feet)	Depth to bottom of unit (feet)
Olivine basalt, medium-gray scoriaceous; composed of plagioclase (An_{55}), augite, iddingsite (after olivine), ilmenite, magnetite, and carbonate	20±10	1, 215±10
Augite rhyodacite(?); either one flow of variable texture or four of alternate textures	105	1, 300
Andesite agglomerate; composed of fragments of vesicular andesite in a yellowish-gray matrix of altered rock of the same composition	20	1, 340
Olivine basalt, light-gray; composed of phenocrysts of plagioclase (An_{70}), augite, iddingsite (after olivine), ilmenite, and carbonate	30	1, 365
Altered rhyolite tuff, grayish-yellow; cherty	5	1, 370

Drilling abandoned at this point.

Fred Kornowsky well.—The Fred Kornowsky well, in the NE¼SE¼SW¼ sec. 21, T. 13 S., R. 5 E., about three-fourths of a mile west of the Morgan-Martzell well, chiefly probes unconsolidated outwash sand and gravel and welded ash flows; it does not penetrate the entire basin fill. The well was begun in 1949 and drilled to a depth of about 1,061 feet before it was plugged and abandoned as a dry hole; there were no shows of either oil or gas. In June 1951, interest in the well was revived, and the well was reentered and drilled an additional 77 feet, to a total depth of 1,138 feet, before it was again abandoned as a dry hole in August of the same year.

A driller's log of the upper 375 feet of this well is included as part of the Morgan-Martzell well log.

T. S. Povah Ranch well.—The third well that tested the West Yellowstone basin was drilled on the T. S. Povah Ranch, about 6 miles west-northwest of West Yellowstone, Mont. This well, in the NE¼NW¼ sec. 23, T. 13 S., R. 4 E., penetrated outwash sand and gravel to a depth of 280 feet before drilling was terminated. No shows of oil or gas were reported, and the well now serves as a source of artesian water.

CARROT BASIN ANTICLINE

Phillips Petroleum Co. Carrot Basin 1.—The Phillips Petroleum Co.'s test, known as the Carrot Basin 1, is in the SE¼NW¼SE¼ sec. 11, T. 10 S., R. 3 E. (unsurveyed), at a point about 3 miles west of the northwest corner of this quadrangle (fig. 29). The well was begun in July 1949 and completed as a dry hole about 3 months later. Oil was noted as a show in sandstone beds of the Morrison(?) at a depth of 200–300 feet and as scattered stains in dolomite cores from the Dinwoody Formation at a depth of about 900 feet (Tutten, 1960, p. 263). The well, designed to test the elongate north- and northwest-trending doubly plunging Carrot Basin anticline (fig. 30), was in an area noted for its rugged terrain, remoteness from settled communities, and distance from paved roads. Access was difficult then and still is today (1965). The Phillips Petroleum Co. constructed a graded road about 10 miles long from the Wapiti Ranger Station to the well site. The site is now reached by a graded and locally graveled road which extends westward from Highway 191 via Taylors Fork, Wapiti Creek (Wapiti Ranger Station), and the high bench which forms the north wall of Sage Creek.

The Carrot Basin anticline is about 5 miles long and 3 miles wide and has somewhat more than 600 feet of closure. It occupies the south half of T. 9 S., R. 3 E. (unsurveyed), and the north half of T. 10 S., R. 3 E. (unsurveyed) (fig. 30). The anticline is cut and partly dissected by northeastward-flowing Wapiti Creek and its tributaries.

The collar of the well was estimated to be at an altitude of about 9,200 feet (actual altitude about 9,220

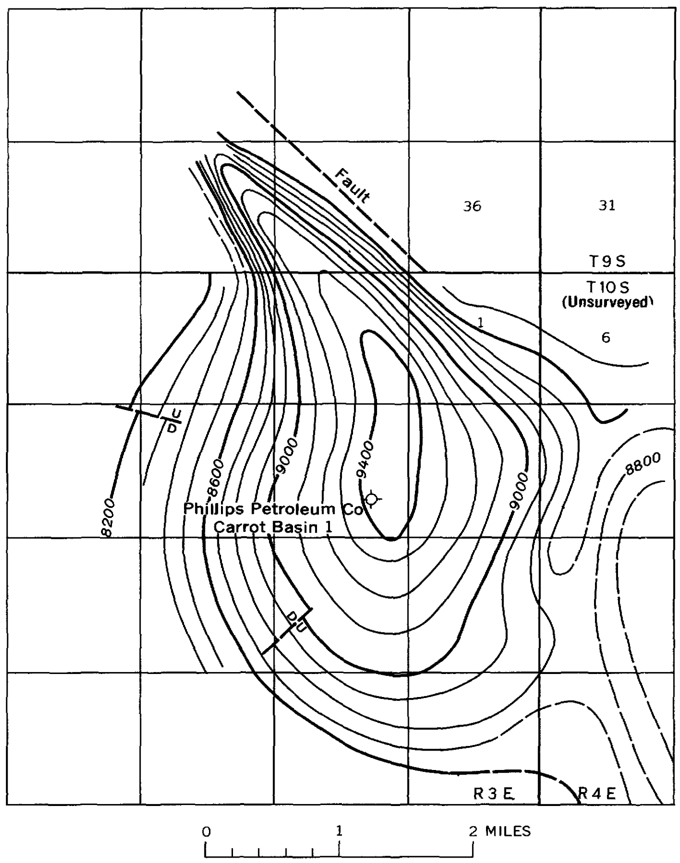


FIGURE 30.—Structure-contour map of the Carrot Basin anticline drawn on top of the basal conglomerate of the Kootenai Formation. Contour interval is 100 feet; datum is mean sea level. Prepared in 1947–48 by the late J. A. Wilsey assisted by A. D. Whitten, Jr. Anticline is about 3 miles west of the northwest corner of the Tepee Creek quadrangle. Map available through the courtesy of the Phillips Petroleum Co.

ft), and the drill reached a total depth of about 2,140 feet (altitude 7,080 ft) before the hole was abandoned, reportedly because of technical difficulties (Austin and Stoever, 1950, p. 99).

The test was begun in the Morrison Formation of Late Jurassic age and penetrated strata of Jurassic, Triassic, Permian, Pennsylvanian, and Mississippian age, to bottom finally in the upper part of the Madison Group. The well log showed the following depths to tops of units:

Unit	*Feet*
Ellis Group	355
Triassic strata	555
Phosphoria	1,163
Tensleep	1,280
Amsden	1,608
Madison	1,920

The oil and gas potentialities of the Tepee Creek quadrangle in particular, and of that part of the Madison Range north of Hebgen Lake in general, are still unknown. Not one of the four test wells seems to have produced meaningful results; those in the West Yellowstone basin did not penetrate the entire basin fill of outwash sand and gravel, and the Carrot Basin well did not test such significant possible oil reservoir rocks as those of the Jefferson Formation.

In the West Yellowstone basin, the source of the oil seeps which contaminated the ground water is still uncertain. One explanation is that fuel oil leaked from one or more storage tanks in or near West Yellowstone and seeped into and contaminated the ground water. Oil could spread for great distances in the loose, unconsolidated fill, as this fill permits almost unimpeded groundwater movement.

Although much of the oil produced in Montana has come from Cretaceous rocks, the Tensleep Sandstone (Pennsylvanian), the stratigraphic equivalent of the Quadrant Sandstone of this area, has produced oil in the Elk Basin field of Montana and Wyoming, and the Madison Group (Mississippian) has yielded much oil in northern Montana (Gardner and others, 1945). Both the Quadrant and the Madison underlie much of this general area. Now (1965) the search for oil has turned toward Devonian strata, chiefly the Jefferson Formation, which is widespread throughout this part of southwestern Montana.

Although suitable oil structures underlain by potential reservoir rocks are known in this general area, the possibilities for oil and gas accumulation seem limited because most structural traps lack closure or are breached to rocks older than the reservoir rocks which commonly contain petroleum elsewhere in southwestern and south-central Montana.

MINERAL DEPOSITS

Mineral deposits were not found, and it seems unlikely that any are present. A small mineralized body composed chiefly of magnetite and hematite occurs in the center of N½ sec. 34, T. 9 S., R. 5 E., about 1 mile north of the north edge of the quadrangle. The deposit is at the contact between the roof of the Gallatin River laccolith and the enclosing limestone beds of the Madison Group and is interpreted as a contact-metamorphic deposit. Access to the deposit, which is on the high ridge northwest of the juncture of Snowslide Creek with the Gallatin River (pl. 1), is via the Yellowstone National Park boundary trail from the Gallatin Ranger Station.

Details of the deposit unfortunately are uncertain, for most of the people who either worked the deposit or knew about it are now dead. From information available to me, it seems likely that the deposit was first discovered in 1913 by Lou C. Bart, a prospector who "built a cabin at the summit of the ridge and made several pits and open cuts nearby" (May, 1950, p. 3). This cabin, now nearly in ruins, is in the center of S½ sec. 34, T. 9 S., R. 5 E., and there is very little evidence near it of any intensive mining activity.

In the early 1920's Bart apparently was joined by David A. Cross and Eugene F. Bunker of Bozeman, Mont., and the three men organized the Apex Development Co. to exploit the property. At about this time another cabin and tool and storage sheds were constructed about 1 mile north of the first cabin, and several adits and open pits were dug. Most of the mining was carried on in this locality, judging by the tailings and collapsed adits. The Apex Development Co. failed shortly thereafter. Since then the cabins and storage and tool sheds have fallen into disrepair, and the adits of the tunnels have caved. Little can be determined about the shape, size, and tenor of the ore body.

The deposit is near the crest of the north flank of the broad upwarp formed by the Gallatin River laccolith; Madison strata here dip northward. On the basis of the distribution pattern of the adits and dug pits it seems likely that the ore body was nearly horizontal and possibly tabular in shape—approximately concordant with the roof of the laccolith. The workings extend over an area 100–150 feet on a side, which suggests that the ore body may have underlain about a quarter of an acre.

Selected samples of ore collected from a specimen box in the cabin indicate that the dominant minerals mined were magnetite and hematite. Analyses of these show iron (as Fe_2O_3) totaling 73.32 percent.[10]

Other minerals noted in small amounts near the mine include hydrogarnet, epidote, pyrite, sphalerite, galena,

[10] Fe_2O_3 content determined volumetrically by W. D. Goss.

and some small fragments of malachite and chrysocolla(?).

The position of the roof of the laccolith is marked by a chilled zone of relatively fine-grained minerals. The limestone beds overlying the laccolith have been baked, hardened, altered from their normal dark-bluish-gray color to white, and recrystallized almost to marble.

PHOSPHATE

Phosphate-bearing strata of Permian age crop out widely in the quadrangle, but as yet no commercial deposits have been discovered. Commonly the beds richest in phosphate are thin and concealed beneath vegetation and talus. Moreover, the area is remote and lacks suitable transportation and power facilities.

Phosphate has been produced in commercial amounts, however, from the Simplot properties in Idaho. These properties are near Odell Creek in the Centennial Mountains along the Montana-Idaho State line (secs. 11, 12, 13, T. 14 N., R. 40 E., and secs. 19, 20, T. 14 N., R. 41 E.), and, therefore, the general region near the quadrangle may be a potential source of phosphate.

SAND AND GRÁVEL DEPOSITS

Small deposits of sand and gravel suitable for road metal are present in the quadrangle and a few pits have been opened to supply material for local road repairs. One small gravel pit is near the abrupt southward bend of Grayling Creek (pl. 1), and several other comparably sized pits, reached by a secondary road, are along the valley wall south of the Gallatin River, about 1 mile northeast of Divide Lake.

Commonly the material in the pits is poorly sorted, ranging in size from sand to boulders. Such diversified compositional types as welded tuff, Precambrian crystallines, and detritus—chiefly limestone derived from Paleozoic and Mesozoic sedimentary units—are common.

Along the north shore of Hebgen Lake, gravel pits have been opened in the broad alluvial fan in the NE¼ sec. 17, T. 12 S., R. 5 E., near the Parade Rest Ranch. Material from these pits, after being washed, crushed, and sized, has been mixed with asphalt and used to resurface Highway 191.

Another small gravel pit has been opened in a deposit of poorly sorted sand and gravel about a quarter of a mile west of the west edge of the quadrangle, in the SW¼ sec. 4, T. 12 S., R. 4 E. This pit is used chiefly by the Montana State Highway Department to supply the small amounts of road metal needed by them for minor road repairs.

QUARRYING

Several of the local ranchers have quarried the welded tuff for building stone, but the rock, despite its pleasing appearance, has not proved popular. The few homes that are constructed of the tuff are of the light-gray or pale-red varieties.

MEASURED SECTIONS

All stratigraphic sections were measured using an Abney hand level, and all data tables were corrected for dip. Location of sections is shown in fig. 10.

A. Wolsey Shale exposed near the mouth of Bacon Rind cirque, about half a mile north of White Peak

	Thickness (feet)
Meagher Limestone.	
Wolsey Shale:	
Covered interval; isolated exposures indicate probable alternating gray thin-bedded dense limestone and light-grayish-green shaly siltstone	104
Sandstone and intercalated thin beds of shaly siltstone, brownish-gray, thin- to medium-bedded, moderately well sorted, chiefly fine-grained; grains subangular to subround; much glauconite; calcite cement; moderately friable; cliff former	38
Covered interval; float suggests light-brown shaly siltstone and light-brown thin-bedded fine-grained sandstone	34
Sandstone, brownish-gray, even-bedded, thin-bedded (most beds 1–2 in. thick), well-sorted, fine-grained; quartz grains subangular; glauconite grains well rounded; quartz about four times as plentiful as glauconite; weak, calcareous cement; friable	4.0
Siltstone, shaly, light-greenish-gray to green, thin-bedded, locally fissile, moderately well sorted; chiefly silt and intermixed fine-grained sand; glauconite and quartz in almost equal amounts; noncalcareous	27
Sandstone, green, fine-grained; almost wholly composed of glauconite; beds irregular, most 2–3 in. thick; glauconite almost four times as plentiful as quartz	2
Shale, greenish-gray, thin-bedded, almost fissile, and intercalated thin, very fine grained sandstone beds; locally becomes a shaly siltstone; glauconite common	10
Total thickness	219
Flathead Sandstone.	

B. Meagher Limestone at the mouth of Bacon Rind cirque, about half a mile north of White Peak

	Thickness (feet)
Park Shale.	
Meagher Limestone:	
Limestone, light-gray, thin-bedded, finely crystalline; pisolitic limestone at the very top of the unit; cliff former; almost concealed beneath soil and talus	98
Limestone, bluish-gray, even-bedded; composed of crenulated beds 2–3 ft thick; parting planes not accentuated; microcrystalline; calcite seams both parallel and cross bedding planes; cliff former	218

B. Meagher Limestone at the mouth of Bacon Rind cirque, about half a mile north of White Peak—Continued

	Thickness (feet)
Meagher Limestone—Continued	
Limestone, light-bluish-gray, many yellow mottles, thin-bedded; each bed ranges in thickness from about ½ to 2 in.; crenulated beds marked by small round nodules about 1½ in. in diameter; finely crystalline; mottles are irregular shaped, about ¼ in. wide and 1 in. long; contains many veins and lenses of coarsely crystalline calcite; forms steep steplike slopes	140
Shale, greenish-brown, thin-bedded, fissile, silty, strongly calcareous; forms moderate slope beneath overlying cliff-forming limestone	2
Limestone, light-brown, thin-bedded; contains mottles which form thin persistant beds; finely crystalline; cliff former	4
Total thickness	462
Wolsey Shale.	

C. Bighorn(?) Dolomite, Snowy Range Formation, and Pilgrim Limestone in Bacon Rind cirque, and about half a mile north of White Peak

	Thickness (feet)
Jefferson Formation.	
Unconformity.	
Bighorn(?) Dolomite:	
Dolomite, light-gray to light-brown, even-bedded, thin-bedded, microcrystalline, dense	35
Sage Member of Grant (1965) of Snowy Range Formation:	
Dolomite, light-tan, even-bedded; beds range in thickness from ¾ to 6 in., most are about 1½ in. thick; beds locally marked by very fine laminae; few lenses of mud-pebble conglomerate	34
Dolomite, light-tan, even-bedded, thin-bedded, almost platy; most beds range in thickness from ½ to 2 in.; many beds consist of a dark-gray limestone interior with a rind of light-brown dolomite, others are completely dolomitized; contains distinctive Snowy Range brachiopod fauna including *Billingsella perfecta, Angulotreta tetonensis,* and trilobites *Taenicephalus* sp., *Parabolinoides* sp., and *Eoorthis* sp	65
Total thickness Sage Member of Grant (1965) of Snowy Range Formation	99
Pilgrim Limestone:	
Limestone, light-grayish-brown, chiefly even-bedded, thin- to medium-bedded; most beds about 3 in. thick, others are as much as 1 ft. thick; limestone generally finely crystalline; mud-pebble conglomerate beds moderately glauconitic	13
Limestone, light-gray, even-bedded, thin-bedded; beds range in thickness from about ¼ to 8 in.; locally contains light-brown calcareous siltstone mottles; many mud-pebble conglomerate beds with much glauconite irregularly interspersed; limestone beds finely crystalline; glauconite-rich lenses coarse grained	29

C. Bighorn(?) Dolomite, Snowy Range Formation, and Pilgrim Limestone in Bacon Rind cirque, and about half a mile north of White Peak—Continued

	Thickness (feet)
Pilgrim Limestone—Continued	
Limestone, grayish-brown, massive to thick-bedded; where bedding planes apparent, unit is almost even bedded; glauconite widespread in upper part and commonly associated with mud-pebble conglomerate beds	12
Limestone, light-gray, even-bedded, thin-bedded; most beds range in thickness from ½ to 6 in.; many edgewise conglomerate beds; limestone beds contain many irregular light-brown dolomite mottles generally ¼–½ in. thick and parallel to bedding planes; limestone beds, locally marked by extremely fine laminae, separated by light-brown fissile calcareous shale beds; many seams of glauconite-rich limestone about 1 in. thick	19
Limestone, grayish-brown, even-bedded, medium-bedded, locally massive; beds range in thickness from about 1 to 4 ft.; medium to coarsely crystalline; composed of oolites of yellow calcite in finely crystalline calcite matrix; contains beds 1–8 in. thick of glauconite-rich oolitic limestone	12
Limestone, light-brown, even-bedded, thin-bedded; most beds dense and about ½ in. thick and separated by calcareous shale laminae; microcrystalline in thin beds, coarsely crystalline in thicker beds. Some beds about 3 in. thick of mud-pebble conglomerate composed of flat angular fragments ½–1½ in. long and ⅛–¼ in. wide of fine-grained limestone in matrix of slightly coarser limestone containing small calcite grains	18
Covered interval; mantled by fragments and boulders; exposures about ½ mile distant indicate light-gray to light-tan mottled limestone and dolomite beds, all about ¾ in. thick; limestone beds are wavy and crenulated	93
Total thickness of Pilgrim Limestone	196
Total thickness, Bighorn(?) Dolomite, Snowy Range Formation, and Pilgrim Limestone	330

D. Jefferson Formation along the north flank of Bacon Rind cirque, about half a mile north of White Peak

	Thickness (feet)
Three Forks Formation.	
Jefferson Formation:	
Dolomite, light-brown, thin-bedded; beds range from 2 to 4 in. in thickness; each bed composed of extremely thin wavy laminae; very rubbly and vuggy surface; contains chert as angular fragments and seams; many calcite seams and small vugs filled with calcite crystals; grades into the overlying Three Forks Formation	93
Covered interval; small widely separated outcrops indicate light-brown thin dolomite beds; forms gentle slope	33

D. Jefferson Formation along the north flank of Bacon Rind cirque, about half a mile north of White Peak—Continued

	Thickness (feet)
Jefferson Formation—Continued	
Dolomite, very light gray, moderately bedded to massive; rubbly surface; finely saccharoidal; contains *Neostringophyllum* sp	35
Dolomite, light-gray to light-brown, medium- to thick-bedded; finely saccharoidal; rubbly surface; broken by fractures healed with calcite; many small angular fragments and seams of light-gray chert; 1-ft-thick ledge about 130 ft above base contains poorly preserved stromatoporoids, and corals *Thamnopora* sp. and *Pachyphyllum* sp	156
Total thickness	317

Bighorn(?) Dolomite, Snowy Range Formation, and Pilgrim Limestone undifferentiated.

E. Three Forks Formation along the north flank of Bacon Rind cirque, about half a mile north of White Peak

Madison Group.
Unconformity.

	Thickness (feet)
Three Forks Formation:	
Sappington(?) Member:	
Siltstone, yellow, thin-bedded; beds 1–3 in. thick; calcareous; forms moderate slopes covered by angular slabs 2–5 ft on a side	18
Trident Member:	
Covered interval; float suggests light-gray to green fissile shale	15
Logan Gulch Member:	
Limestone, gray, finely crystalline, thin-bedded; beds ½–1 ft thick; intercalated between limestone beds are light-brown calcareous siltstone beds and a few thin beds of greenish-gray shale and siltstone	31
Siltstone, yellow-brown to light-brown, shaly, somewhat dolomitic; beds range from ½ to 1 in. in thickness; breaks into angular slabs marked by many calcite seams	63
Siltstone, red, shaly, calcareous; breaks into angular fragments 2–3 in. on a side; forms a distinctive reddish band	35
Total thickness	162

Jefferson Formation.

F. Sappington(?) Member of Three Forks Formation along the southeast slope of White Peak

[Measured by J. T. Dutro, Jr.]

Lodgepole Limestone (unit 1 of Madison section—measured section *G*).

	Thickness (feet)
Three Forks Formation:	
Sappington(?) Member:	
Siltstone to fine-grained sandstone, dolomitic or slightly calcareous, light-brown; weathers dark yellowish orange to dark yellowish brown; argillaceous laminae; beds 0.2–0.5 ft thick; blocky fracturing; worm trails and markings on bedding surfaces; Sappington-like lithology	20

F. Sappington(?) Member of Three Forks Formation along the southeast slope of White Peak—Continued

	Thickness (feet)
Three Forks Formation—Continued	
Sappington(?) Member—Continued	
Limestone, fine-grained, nodular to platy, medium-dark-gray; weathers light brown to pale yellowish brown; argillaceous; beds up to 0.2 ft thick; fossils scattered throughout (USGS 7595–SD, about 4 ft below top of unit)	10
Dolomite, finely crystalline (may be calcareous siltstone), very pale orange to grayish-orange; weathers light brown to pale yellowish brown; irregular beds 0.1–0.3 ft thick; calcite veins abundant	15
Limestone, aphanitic, light-brownish-gray; weathers light bluish gray, two thick beds; laminated; possibly algal	4
Covered interval	15
Sandstone, fine-grained, calcareous or dolomitic, dark-yellowish-orange; weathers pale yellowish brown; limited exposure in creek bed	3+
Total thickness	67

Upper Tepee fault.
Precambrian schist.

G. Madison Group along the southeast slope of White Peak

[Measured by J. T. Dutro, Jr.]

	Thickness (feet)
"Amsden Formation"	100±
Mission Canyon Limestone:	
67. Limestone, very fine grained, aphanitic, light-brownish-gray to light-olive-gray; weathers medium light gray; little or no chert; conchoidal fracture	4.5
66. Dolomite, fine-grained, silty, grayish-orange; weathers dark yellowish orange to pale yellowish brown; beds 0.2–0.4 ft thick	2.5
65. Limestone, as in unit 67; beds 1.5–10 ft thick; thin zone of small light-gray chert nodules about 10 ft above base; lower 10 feet laminated (perhaps algal) and contains some lenses of pelletal debris	31
64. Dolomite, as in unit 66; some beds weather pale red or grayish red	6
63. Limestone, very fine grained, aphanitic, light-brownish-gray; weathers medium light gray to light bluish gray; pinkish mottling in places; cliff-forming unit; beds 5–10 ft thick; 10–15 percent light-brownish-gray to dark-gray chert in beds and lenses 0.2–0.4 ft thick	69
62. Dolomite, fine-grained, silty, pale-yellowish-brown; weathers grayish orange; beds 0.1–0.3 ft thick	5

G. Madison Group along the southeast slope of White Peak—Con.

Mission Canyon Limestone—Continued	*Thickness (feet)*
61. Breccia, polymitic; fragments angular to subrounded, granule size to 0.4 ft across; fragments generally fine-to medium-grained limestones but some dolomite as in units 60 and 62 and rare dark-gray to brownish-gray chert (approximately 5 percent of fragments); no regular bedding; fragments compose ⅔–¾ of unit; matrix predominantly yellowish-gray to grayish-orange or pinkish siltstone; breccia boundaries irregular, cutting upward into units 62 and 63 at least 16 ft; possible 2 ft-wide pipe at top of exposure; lower contact irregular; thickness varies from 15 to 36 ft	25±
60. Dolomite, medium-crystalline, pale-yellowish-brown; weathers pale brown to light brown; beds 3–5 ft thick; thin lenses of brownish-gray chert, about 5 percent stromatolitic structures in lower bed	12
59. Breccia; like that of unit 61 but no reddish silty matrix; faint bedding traces about 3–5 ft apart	21
58. Dolomite, medium to coarsely crystalline, pale-yellowish-brown; weathers pale brown to dark yellowish brown; about 10 percent banded dark-gray and light-brownish-gray chert in lenses 0.1 ft thick and 3–4 ft long and nodules 0.1–0.2 ft in diameter; beds 0.2–2 ft thick	30
57. Limestone, fine-grained, and scattered medium-grained bioclastic debris; light brownish gray; weathers light bluish gray; single massive bed; Syringopora and horn corals (USGS 20327–PC)	4
56. Limestone, dolomitic, and scattered bioclastic debris; dolomitization nearly complete but patchy; medium to coarse grained; brownish gray; weathers pale yellowish brown with bluish gray patches where dolomitized; small dark gray chert nodules (about 5 percent) weather orange; beds 0.1–0.8 ft thick	25
55. Covered interval; lower 25–30 ft covered with dolomite float as in unit 51; about 30 ft above base is 3-ft-thick block of light-brownish-gray aphanitic light-bluish-gray weathering limestone, may be in place; upper 15 ft covered with blocks of dolomitic limestone as in unit 56; calcite vein, 1 ft thick, cuts across strike of beds at angle of 15° near base of interval; possible fault zone	65
54. Limestone, coarse-grained, bioclastic locally pelletal, light-brownish-gray; weathers yellowish gray to light gray; beds 1–3 ft thick; granule-size pellets compose about 20 percent of rock; fossiliferous (USGS 20326–PC about 5 ft below top of unit)	17
53. Dolomite, very fine grained, calcareous, very pale orange; weathers grayish orange; irregular beds 0.1–0.2 ft thick; partly covered	14

G. Madison Group along the southeast slope of White Peak—Con.

Mission Canyon Limestone—Continued	*Thickness (feet)*
52. Breccia; subangular to subrounded limestone fragments, granule to pebble size; mostly light-brownish-gray very fine grained limestone with some medium-grained brownish-gray limestone; bioclastic limestone and dolomite fragments rare, less than 5 percent; matrix of grayish-orange calcareous siltstone; pebbles comprise 70–75 percent of unit; beds 3–5 ft thick	15
51. Covered interval; talus blocks from many overlying units; mainly fine-grained pale-orange dolomite that weathers grayish orange; some slabby blocks of medium-crystalline pale-yellowish-brown dolomite	28
50. Limestone, medium-grained, brownish-gray; weathers light gray; faint laminations of coarser particles; single massive bed	6
49. Covered interval; as in unit 51	25
48. Limestone, bioclastic and oolitic, crossbedded; light-brownish-gray; weathers medium light gray; massive unit, beds 3–5 ft thick; corals, including lithostrotionoids, abundant in upper 15 ft (USGS 20199–PC)	45
47. Dolomite, finely crystalline, light-brownish-gray to pale-brown; weathers pale yellowish brown; thin beds, up to 0.2 ft thick; small quantity (less than 5 percent) orange-weathering small chert nodules	18
46. Limestone, coarse-grained, bioclastic, light-brownish-gray; weathers medium light gray; beds 0.3–0.8 ft thick; abundant crinoidal debris and silicified fossil fragments (USGS 20198–PC)	2
45. Interbedded limestone and dolomitic limestone; limestone is medium grained, bioclastic, brownish gray, weathers medium gray; dolomitic limestone is silt sized, pale brown to dark yellowish brown, weathers pale yellowish brown; limestone beds 0.2–0.4 ft thick grade upward into their platy dolomitic beds; sparsely fossiliferous (USGS 20197–PC from 18 ft below top of unit)	42
44. Limestone, coarse-grained, bioclastic; oolitic in lower 20 ft; crossbedded; brownish gray; weathers medium light gray to yellowish gray; forms massive cliff; beds 3–5 ft thick in lower 20 ft, 0.5–2 ft thick in upper part; beds replaced by irregular patches of finely crystalline dolomitic limestone in upper 15 ft; ooids, abundant in lower part, single and double layered and 0.5–1 mm in diameter; corals, including *Homalophyllites* and *Vesiculophyllum* (USGS 20196–PC, 5 ft below top of unit; USGS 20194–PC, 5 ft above base of unit; USGS 20195–PC)	33
Total thickness Mission Canyon Limestone	545

G. Madison Group along the southeast slope of White Peak—Con.

Lodgepole Limestone:	Thickness (feet)
43. Interbedded limestone and dolomitic limestone; limestone is coarse grained, bioclastic, brownish gray, weathers medium light gray; dolomitic limestone is finely crystalline, brownish gray, weathers pale yellowish brown; gradational contacts between the two rock types; beds 0.2–0.8 ft thick (USGS 20193–PC, lower 6 ft)	13
42. Interbedded limestone and calcareous siltstone; limestone is fine grained, brownish gray to medium dark gray, weathers light bluish gray; calcareous siltstone is pale yellowish brown, weathers moderate yellowish brown; limestone beds 0.1–0.3 ft thick, about 60–70 percent of unit, grade upward into platy 0.1-ft-thick beds of siltstone; limestone beds in upper 5 ft have bioclastic debris in basal parts	18. 5
41. Limestone, medium- to coarse-grained, bioclastic, brownish-gray; weathers medium gray; beds 0.3–0.4 ft thick; fossil fragments; few small silty pebbles in lower few inches	1. 5
40. Interbedded limestone and calcareous siltstone, as in unit 42; bedded limestone about 20 percent of unit, nodular limestone about 20 percent, siltstone about 60 percent; bioclastic material more abundant in upper part; interval partly covered	9
39. Limestone, as in unit 41	1. 5
38. Calcareous siltstone and limestone, as in unit 42; limestone in lenses and discontinuous beds, about 30–40 percent of unit; interval partly covered	16. 5
37. Limestone, as in units 39 and 41; beds 0.2–0.4 ft thick; silty beds, like those of unit 38, in lower 1 ft; both upper and lower contacts gradational	2
36. Calcareous siltstone and limestone, as in units 38 and 42; limestone in lenticular bodies 0.2 ft thick with pale-red-weathering laminae; worm markings, tracks and trails abundant; fossiliferous (USGS 20192–PC, about 4 ft above base of unit)	8. 5
35. Limestone, coarse-grained, bioclastic, brownish-gray; weathers medium light gray; limestone pebbles up to 0.3 ft long; unit grades upward into unit 36	1
34. Calcareous siltstone and limestone, as in unit 36	7. 5
33. Limestone, coarse-grained, bioclastic, cross-laminated, brownish-gray; weathers medium light gray; coarse bioclastic debris at base, contains few limestone pebbles; grades upward through medium-grained bioclastic limestone into unit 34	2. 0
32. Calcareous siltstone and limestone; as in units 34 and 36; limestone about 30 percent of unit; siltstone weathers to thin plates 0.05–0.1 ft thick; worm trails and markings abundant	12. 5

G. Madison Group along the southeast slope of White Peak—Con.

Lodgepole Limestone—Continued	Thickness (feet)
31. Interbedded coarse-grained bioclastic and fine-grained limestones; bioclastic limestone is light gray, weathers medium light gray, contains beds 0.3–0.5 ft thick and composes about 30 percent of unit; fine-grained limestone is medium dark gray to brownish gray; weathers medium gray and contains beds 0.1–0.2 ft thick; fossiliferous (USGS 20191–PC from 4 ft above base)	8 5
30. Interbedded limestone and calcareous siltstone; limestone is fine grained, mottled brownish gray to medium dark gray, weathers light bluish gray, and contains beds 0.2–0.4 ft thick; siltstone is dark yellowish orange, weathers pale yellowish orange, and comprises about 40 percent of unit	5 0
29. Limestone, medium-grained, bioclastic, brownish-gray; contains silicified fossils (USGS 20190–PC)	0. 5
28. Limestone and calcareous siltstone as in unit 30; siltstone about 30 percent of unit; about 5 percent light-yellowish-brown small chert nodules in lower 5 ft	15. 5
27. Limestone, coarse-grained, bioclastic, brownish-gray; weathers medium light gray; beds 0.2–0.5 ft thick; silicified fossil fragments concentrated near top and base of unit (USGS 20189–PC)	4. 5
26. Limestone and calcareous siltstone, as in units 28 and 30; siltstone about 40 percent of total in lower 12 ft, becomes thin partings between 0.1- to 0.3-ft thick limestone beds in upper part; medium-grained bioclastic limestone bed 0.3 ft thick about 10 ft above base (USGS 20188–PC)	20
25. Limestone, coarse-grained, bioclastic; pebbles of fine-grained limestone in lower part; silicified brachiopods (USGS 20187–PC)	1
24. Limestone; graded units of bioclastic (about 50 percent) to fine-grained limestone (about 20 percent) and siltstone (about 30 percent), 0.3–0.8 ft thick; brownish-gray bioclastic limestones with coarse fossil fragments at base grade upward through medium-dark-gray fine-grained limestones to grayish-orange to pale-yellowish-orange calcareous siltstone; *Taonurus* common in siltstones; abundant chonetid brachiopods (USGS 20186–PC, 10–13 ft above base)	18. 5
23. Limestone, coarse-grained, bioclastic, pelletal, brownish-gray; weathers medium light gray; two beds, 0.5 ft below and 1 ft above silicified fossil debris (USGS 20185–PC)	1. 5
22. Limestone and calcareous siltstone, as in units 26, 28, and 30; siltstone about 15 percent of thickness	9. 5
21. Limestone, coarse-grained, bioclastic, brownish-gray; weathers medium light gray; beds 0.2–0.5 ft thick; a few limestone pebbles, as in unit 25; silicified fossil debris (USGS 20184–PC)	1. 5

G. Madison Group along the southeast slope of White Peak—Con.

Lodgepole Limestone—Continued	Thickness (feet)
20. Limestone and calcareous siltstone, as in units 22, 26, 28 and 30; lower 2 ft contains some fine- to medium-grained limestone with finely comminuted cross-laminated fossil debris and weathers pale red; 0.4-ft-thick bed of intraformational conglomerate, like unit 19, about 10 ft above base; (USGS 20183–PC from lower half of unit)	22
19. Intraformational conglomerate: matrix of coarse grained bioclastic limestone; flat pebbles of fine-grained limestone and calcareous siltstone, as in unit 18; limestone is brownish gray to grayish red and ferruginous and weathers reddish brown; siltstone weathers yellowish and is laminated; pebbles up to 0.3 ft long and 0.1 ft thick and parallel bedding; pelletal grains make up about half of matrix; (USGS 20182–PC from lower 1 ft)	3
18. Interbedded limestone and calcareous siltstone; limestone is fine grained and mottled brownish gray to medium dark gray and weathers light bluish gray; siltstone is dark yellowish orange and weathers pale yellowish orange; limestone beds 0.2–0.4 ft thick; siltstone about 20–25 percent of unit; lower 3 ft slightly coarser limestone than rest of unit; (USGS 20181–PC from 2 ft above base)	22
17. Limestone, coarse-grained, bioclastic, light-brownish-gray; weathers medium light gray; in two graded beds, lower 2 ft thick, upper 3 ft thick; silicified brachiopods; (USGS 20180–PC from lower bed)	5
16. Limestone, fine-grained, brownish-gray; weathers medium light gray; beds 0.05–0.3 ft thick; silty partings about 10 percent of thickness; thin nodules, 0.1 ft thick and 0.3 ft long, of light-brown chert weather "tannish" and compose about 20 percent of thickness; silicified fossil debris near base and middle of unit (USGS 20179–PC, from lower 1 ft)	6
15. Limestone, coarse-grained, echinodermal, bioclastic, and pelletal, crossbedded, light-brownish-gray; pockets and lenses of silicified fossils, mainly brachiopods and horn corals; beds 0.8–3 ft thick; (USGS 20178–PC in lower 2 feet of unit)	6.5
14. Limestone, as in unit 16; chert about 15 percent of thickness; several layers of silicified fossils (USGS 20177–PC from upper 1 ft; 20176–PC from 4 ft below top of unit; 20175–PC from basal bed)	8.5
13. Limestone, coarse-grained, bioclastic, oolitic, crossbedded, light-brownish-gray; weathers medium light gray; single massive bed; corals especially abundant in upper half, silicified brachiopods in upper 0.1 ft (USGS 20174–PC, brachiopods from upper 0.1 ft; USGS 20173–PC, corals from upper half; USGS 20172–PC from lower 1 ft)	3

G. Madison Group along the southeast slope of White Peak—Con.

Lodgepole Limestone—Continued	Thickness (feet)
12. Dolomite, fine-grained, brownish-gray, weathers light brown to moderate yellowish brown; beds 0.1–0.2 ft thick	1.5
11. Limestone, coarse-grained, bioclastic, light-brownish-gray; weathers medium light gray; irregular beds 0.8–2.0 ft thick; pelletal in uppermost bed; undulatory basal surface with 0.5–0.7 ft relief; abundant silicified brachiopods and corals, especially *Homalophyllites* (USGS 20171–PC from lower 2 ft)	4
Possible bedding plane fault, no displacement apparent.	
10. Limestone, fine-grained, light-brownish-gray; weathers medium light gray to olive gray; beds 0.1–0.2 ft thick; about 10 percent light-brownish-gray chert nodules, 0.1 ft thick and 0.2–0.5 ft long	3
9. Limestone, fine-grained, light-brownish-gray to brownish-gray; weathers medium light gray to bluish gray; as much as 20 percent scattered bioclastic debris; 15–20 percent chert as in unit 10; beds 0.3–0.8 ft thick; silicified fossils 4–12 ft below top of unit (USGS 20170–PC)	20
8. Limestone, fine-grained to subaphanitic, brownish-gray; weathers light gray to light olive gray; beds 0.2–0.8 ft thick, but unit forms massive ledge because of intense silicification; light-brown chert in ragged lenses and stringers, about 60 percent of thickness; (USGS 20169–PC, from 8 ft above base)	21
7. Limestone, fine-grained, brownish-gray; weathers medium bluish gray; beds 0.1–0.4 ft thick; basal bed 1.5 ft thick; silty partings along bedding planes; about 15 percent dark-gray to light-yellowish-gray chert (USGS 20168–PC from upper 1 ft)	6
6. Limestone and chert, as in unit 7; chert about 50 percent in ragged beds and lenses; forms small cliff (USGS 20167–PC, 9 ft above base; 20166–PC, 6 ft above base; 20165–PC, 2 ft above base)	10
5. Limestone and chert, as in unit 7; chert about 20 percent in small nodules; beds 0.2–0.5 ft thick; conchoidal fracturing	9
4. Limestone and chert, as in unit 6; chert about 40 percent, in large irregular patches; beds 0.3–1 ft thick	6
3. Limestone, fine-grained to subaphanitic, light-brownish-gray to brownish-gray; weathers light gray to light olive gray; crinoidal debris in lenses; rather uniform beds 0.2–0.4 ft thick; 40 percent light-gray to brownish-gray chert in scattered ragged lenses, decreasing upwards; silty partings along bedding planes; (USGS 20164–PC, float from entire unit; 20163–PC, 25 ft above base)	36
2. Limestone, as in unit 3; silicified crinoidal debris about 20 percent of thickness; 50 percent chert nodules and lenses, 0.1 by 0.2 ft, alined parallel to bedding; small horn corals (USGS 20162–PC)	16

G. Madison Group along the southeast slope of White Peak—Con.

	Thickness (feet)
Lodgepole Limestone—Continued	
1. Limestone, fine-grained, nodular, pale-yellowish-brown to light-brownish-gray; weathers light bluish gray to light olive gray; silty laminae, bioclastic debris about 50 percent; beds 0.2–1 ft thick; single layer of small, black chert nodules about 3.5 ft above base; basal contact disconformity	5
Total thickness, Lodgepole Limestone	393
Total thickness, Madison Group	938
Three Forks Formation	18±

H. Amsden Formation along the southeast flank of peak 9830 (NW¼; sec. 12, T. 11 S., R. 4 E.).

	Thickness (feet)
Quadrant Sandstone.	
Amsden Formation:	
Siltstone, red; partly concealed by debris; uppermost beds composed of light-tan calcareous siltstone; few thin limestone nodules	18
Limestone, light-gray; surface covered by pale-red mottles; thin to medium bedded, most beds 1–2 ft thick; dense; finely crystalline; much light- to dark-gray chert as both nodules and veins chiefly along bedding planes; few fossils, chiefly solitary coral fragments	29
Siltstone, red, thin-bedded, almost fissile; forms strike valley; supports thin grass cover	15
Limestone, light-gray; beds about 4 ft thick; dense; finely crystalline; cut by many minute calcite veins	10
Siltstone, red, medium-bedded; beds about 4 ft thick; forms elongate strike valley, basal 5 ft consists of red breccia and thin-bedded siltstone firmly cemented by calcite	87
Total thickness	159
Unconformity.	
Madison Group.	

I. Shedhorn(?) Sandstone and equivalent strata on the southwest flank of White Peak (NW¼ sec. 2, T. 11 S., R. 4 E.)

	Thickness (feet)
Dinwoody Formation.	
Upper member of the Shedhorn(?) Sandstone:	
Sandstone, dark-brown to grayish-brown, thin- to medium-bedded; fine- to medium-grained; contains much light-gray chert as elongate nodules and as concentric ribbed cylinders; forms steep slope	62

I. Shedhorn(?) Sandstone and equivalent strata on the southwest flank of White Peak (NW¼ sec. 2, T. 11 S., R. 4 E.)—Continued

	Thickness (feet)
Tosi Chert Member of the Phosphoria(?) Formation:	
Chert, light-gray, thin-bedded; each bed about 2 in. thick and separated from adjacent beds by thin layers of gray fissile shale; even bedded; stands as steep slope	38
Lower member of the Shedhorn(?) Sandstone:	
Sandstone, light-brown to grayish-brown, thin- to medium-bedded; most beds about 8 in. thick, even bedded, crossbedded; ranges from very fine to coarse grained, and locally is conglomeratic; contains much chert as rounded nodules and as small ribbed cylinders about 1½ in. in diameter and 6 in. long	22
Grandeur Tongue of the Park City(?) Formation:	
Dolomite, light-brown, thin-bedded; each bed ranges in thickness from 1½ to 3 in.; contains many angular to well-rounded light-gray chert nodules which nearly parallel the bedding planes; stands as steep slope	39
Total thickness	161
Quadrant Sandstone.	

J. Woodside Siltstone near the head of Red Canyon Creek (NW¼ sec. 24, T. 11 S., R. 4 E.)

	Thickness (feet)
Thaynes(?) Formation.	
Woodside Siltstone:	
Siltstone, moderate-reddish-brown, thin-bedded to shaly; many intercalated layers of grayish-red thin-bedded very fine grained sandstone; friable; forms steep slope broken here and there by small steplike ledges	253
Siltstone, moderate-reddish-brown, thin-bedded to shaly; locally becomes a very fine grained sandstone; friable; locally crossbedded; ripple marked; beds somewhat thicker near top; forms steep slope	182
Sandstone, moderate-reddish-brown, thin-bedded, very fine grained; locally becomes shaly siltstone; concealed beneath scree in places	198
Sandstone, light-yellowish-gray, very fine grained; thin-bedded; locally becomes shaly siltstone; forms ledge	2
Silstone, moderate-reddish-orange, thin-bedded to shaly; forms moderate slope	55
Covered interval; isolated exposures indicate moderate-reddish-brown thin-bedded shaly siltstone	34
Total thickness	724
Dinwoody Formation.	

319-733 O—69——7

K. Morrison Formation near the head of Cabin Creek along the southwest flank of Skyline Ridge (S½ sec. 12, T. 11 S., R. 4 E.)

Strata overturned. Because the units change thickness along strike and because the sandstone beds may pinch out, only approximate thicknesses are cited]

	Thickness (feet)
Kootenai Formation.	
Unconformity(?).	
Morrison Formation:	
Claystone and shaly siltstone, light-green, locally thin-bedded; forms gentle slope partly covered by scree and soil	100
Sandstone, light-brown, thin-bedded to platy, very fine grained; locally becomes siltstone; calcite cement; friable; ledge former	10–25
Claystone, light-green, amorphous; breaks into small angular fragments; forms gentle soil-covered slope	100
Claystone, grayish-red, amorphous; differs from overlying unit only in color	75
Sandstone, light-brown to light-green, thin-bedded to platy, very fine grained; locally becomes siltstone; calcite cement, friable; ledge former	10–25
Claystone, light-green, amorphous; forms gentle slope	50
Sandstone, light-brown, very fine grained, friable; poorly cemented by calcite; forms gentle slope	25
Approximate total thickness	400
Swift Formation of Ellis Group.	

BIBLIOGRAPHY

Andrichuk, J. M., 1955, Mississippian Madison Group stratigraphy and sedimentation in Wyoming and southern Montana: Am. Assoc. Petroleum Geologists Bull., v. 39, no. 11, p. 2170–2210.

Austin, W. H., Jr., and Stoever, E. C., Jr., 1950, Reconnaissance geology of the south flank of Cinnamon Mountain, Gallatin County, Montana: Michigan Univ., unpub. M.S. thesis.

Beck, F. M., 1960, Geology of the Sphinx Mountain area, Madison and Gallatin Counties, Montana, *in* Billings Geol. Soc. Guidebook 11th Ann. Field Conf., West Yellowstone Earthquake area, 1960: p. 129–134.

Blackwelder, Eliot, 1915, Post-Cretaceous history of the mountains of central western Wyoming: Jour. Geology, v. 23, p. 97–117, 193–217, 307–340.

——— 1918, New geological formations in western Wyoming: Washington Acad. Sci. Jour., v. 8, p. 417–426.

Boyd, F. R, 1961, Welded tuffs and flows in the rhyolite plateau of Yellowstone Park, Wyoming: Geol. Soc. America Bull., v. 72, no. 3, p. 387–426.

Brown, R. W., 1946, Fossil plants and the Jurassic-Cretaceous boundary in Montana and Alberta: Am. Assoc. Petroleum Geologists Bull., v. 30, no. 2, p. 238–248.

Cobban, W. A., 1945, Marine Jurassic formations of Sweetgrass Arch Montana: Am. Assoc. Petroleum Geologists Bull., v. 29, no. 9, p. 1262–1303.

Condit, D. D., 1919, Relations of late Paleozoic and early Mesozoic formations of southwestern Montana and adjacent parts of Wyoming: U.S. Geol. Survey Prof. Paper 120–F, p. 111–121.

Cressman, E. R., and Swanson, R. W., 1956, Shedhorn sandstone *in* McKelvey, V. E. and others: Am. Assoc. Petroleum Geologists Bull., v. 40, no. 12, p. 2852–2854.

——— 1959, Shedhorn sandstone, *in* McKelvey, V. E., and others: U.S. Geol. Survey Prof. Paper 313–A, p. 31–38.

——— 1960, Permian rocks in the Madison, Gravelly, and Centennial Ranges, Montana, *in* Billings Geol. Soc. Guidebook 11th Ann. Field Conf., West Yellowstone Earthquake area, 1960: p. 226–232.

Daly, R. A., 1933, Igneous rocks and the depths of the earth: New York, McGraw-Hill Book Co.

Deiss, C. F., 1936, Revision of the type Cambrian formations and sections of Montana and Yellowstone National Park: Geol. Soc. America Bull., v. 47, no. 8, p. 1257–1342.

——— 1941, Cambrian geography and sedimentation in the central Cordilleran region: Geol. Soc. America Bull., v. 52, no. 7, p. 1085–1115.

Dorf, Erling, 1960, Tertiary fossil forests of Yellowstone National Park, Wyoming, *in* Billings Geol. Assoc. Guidebook 11th Ann. Field Conf., West Yellowstone Earthquake area, 1960: p. 253–260.

Dorf, Erling, and Lochman, Christina, 1940, Upper Cambrian formations in southern Montana: Geol. Soc. America Bull., v. 51, no. 4, p. 541–556.

Eicher, D. L., 1960, Stratigraphy and micropaleontology of the Thermopolis Shale: Yale Univ. Peabody Museum of Nat. Hist. Bull. 15, 126 p.

Evernden, J. F., Savage, D. E., Curtis, G. H., and James, G. T., 1964, Potassium-argon dates and the Cenozoic mammalian chronology of North America: Am. Jour. Sci., v. 262, no. 2, p. 145–198.

Fisher, C. A., 1909, Geology of the Great Falls coal field, Montana: U.S. Geol. Survey Bull. 356, 85 p.

Fraser, G. D., Witkind, I. J., and Nelson, W. H., 1964, A geological interpretation of the epicentral area—the dual-basin concept, *in* The Hebgen Lake, Montana, earthquake of August 17, 1959: U.S. Geol. Survey Prof. Paper 435, p. 99–106.

Freeman, L. B., Sweet, J. M., and Tillman, C. G., 1949, Geology of the Henrys Lake Mountains, Fremont County, Idaho, and Madison and Gallatin Counties, Montana: Michigan Univ., unpub. M.S. thesis.

Gardner, L. S., Hendricks, T. A., Hadley, H. D., and Rogers, C. P., Jr., 1945, Columnar sections of Mesozoic and Paleozoic rocks in the mountains of south-central Montana: U.S. Geol. Survey Oil and Gas Inv. Prelim. Chart 18.

Goddard, E. N., chm., and others, 1948, Rock-color chart: Washington, Natl. Research Council (repub. by Geol. Soc. America, 1951), 6 p.

Grant, R. E., 1965, Faunas and stratigraphy of the Snowy Range Formation (Upper Cambrian) in southwestern Montana and northwestern Wyoming: Geol. Soc. America Mem. 96.

Hadley, J. B., 1960, Geology of the northern part of the Gravelly Range, Madison County, Montana, *in* Billings Geol. Soc. Guidebook 11th Ann. Field Conf., West Yellowstone Earthquake area, 1960: p. 149–153.

——— 1964, Landslides and related phenomena accompanying the Hebgen Lake earthquake of August 17, 1959, *in* the Hebgen Lake, Montana, earthquake of August 17, 1959: U.S. Geol. Survey Prof. Paper 435, p. 107–138.

Hague, Arnold, Iddings, J. P., Weed, W. H., Walcott, C. D., Girty, G. H., Stanton, T. W., and Knowlton, F. H., 1899, Geology of the Yellowstone National Park: U.S. Geol. Survey Mon. 32, pt. 2.

Hague, Arnold, Weed, W. H., and Iddings, J. P., 1896, Yellowstone National Park sheets: U.S. Geol. Survey Geol. Atlas, Folio 30.

Hall, W. B., 1959, Multiple glaciation and high-level till deposits in southwestern Montana [abs.]: Geol. Soc. America Bull., v. 70, no. 12, pt. 2, p. 1778.

——— 1961, Geology of part of the upper Gallatin Valley of southwestern Montana: Wyoming Univ. unpub. Ph. D. thesis.

Hamilton, Warren, 1964, Volcanic rocks of the West Yellowstone and Madison Junction quadrangles, Montana, Wyoming and Idaho, *in* the Hebgen Lake, Montana, earthquake of August 17, 1959: U.S. Geol. Survey Prof. Paper 435, p. 209–221.

Hamilton, Warren, and Leopold, E. B., 1962, Volcanic rocks of Oligocene age in the southern part of the Madison Range, Montana and Idaho, *in* Short papers in geology, hydrology, and topography: U.S. Geol. Survey Prof. Paper 450–B, p. B26–B29.

Hanson, A. M., 1952, Cambrian stratigraphy in southwestern Montana: Montana Bur. Mines and Geology Mem. 33.

——— 1953, Upper Cambrian formations in northern Utah and southeastern Idaho, *in* Intermountain Assoc. Petroleum Geologists 4th Ann. Field Conf., Guide to the geology of northern Utah and southeastern Idaho, 1953: p. 19–21.

Haun, J. D., and Barlow, J. A., Jr., 1962, Lower Cretaceous stratigraphy of Wyoming, *in* Wyoming Geol. Assoc. Guidebook 17th Ann. Field Conf., Symposium on Early Cretaceous rocks of Wyoming and adjacent areas, 1962: p. 15–22.

Hayden, F. V., 1873, Sixth annual report of the U.S. Geological and Geographical Survey of the Territories for the year 1872: Washington, 844 p.

Heinrich, E. W., 1953, Pre-Beltian geologic history of Montana [abs.]: Geol. Soc. America Bull., v. 64, no. 12, pt. 2, p. 1432.

——— 1960, Geology of the Ruby Mountains and nearby areas in southwestern Montana: Montana Bur. Mines and Geology Mem. 38.

Heinrich, E. W., and Rabbitt, J. C., 1960, Pre-Beltian geology of the Cherry Creek and Ruby Mountains areas, southwestern Montana: Montana Bur. Mines and Geology Mem. 38.

Horn, G. H., 1955, Geologic and structure map of the Sussex and Meadow Creek oil fields and vicinity, Johnson and Natrona Counties, Wyoming: U.S. Geol. Survey Oil and Gas Inv. Map OM–164.

——— 1963, Geology of the east Thermopolis area, Hot Springs and Washakie Counties, Wyoming: U.S. Geol. Survey Oil and Gas Inv. Map OM–213.

Howell, J. V., chm., 1960, Glossary of geology and related sciences, with supplement [2d ed.]: Washington, Am. Geol. Inst., 397 p.

Hubbert, M. K., and Rubey, W. W., 1959, Mechanics of fluid-filled porous solids and its application to overthrust faulting, [Pt.] 1 *of* Role of fluid pressure of mechanics of overthrust faulting: Geol. Soc. America Bull., v. 70, no. 2, p. 115–166.

Hume, J. D., and Leeder, R. W., 1950, Geology of the Pika Mountain area, Gallatin Co., Montana: Michigan Univ., unpub. M.S. thesis.

Hunt, C. B., 1953, Geology and geography of the Henry Mountains region, Utah: U.S. Geol. Survey Prof. Paper 228, 234 p. [1954].

Iddings, J. P., 1895, Absarokite-shoshonite-banakite series: Jour. Geol. v. 3, p. 935–959.

——— 1899a, Absarokite-shoshonite-banakite series, *in* Hague and others: U.S. Geol. Survey Mono. 32, pt. 2, p. 326–355.

——— 1899b, Early basic breccia and associated basaltic flows, *in* Hague and others: U.S. Geol. Survey Mono. 32, pt. 2, p. 275–281.

Iddings, J. P., and Weed, W. H., 1899, *in* Hague and others: U.S. Geol. Survey Mono. 32, p. 1–59, 84–85, 149–157.

Imlay, R. W., 1950, Jurassic rocks in the mountains along the west side of the Green River basin, *in* Wyoming Geol. Assoc. Guidebook 5th Ann. Field Conf., southwest Wyoming, 1950: p. 37–45.

——— 1956, Paleogeographic maps of the United States during the Jurassic period, *in* McKee, E. D., and others, Paleotectonic Maps, Juraissic System: U.S. Geol. Survey Misc. Geol. Inv. Map I–175.

Imlay, R. W., Gardner, L. S., Rogers, C. P., Jr., and Hadley, H. D., 1948, Marine Jurassic formations of Montana: U.S. Geol. Survey Oil and Gas Inv. Prelim. Chart 32.

Jacques R. D., 1949, Geology of the northwest corner of Yellowstone National Park, Montana: Michigan Univ., unpub. M.S. thesis.

Johannsen, Albert, 1949, A descriptive petrography of the igneous rocks: Univ. Chicago Press, v. II, p. 398.

Kay, Marshall, 1947, Geosynclinal nomenclature and the craton: Am. Assoc. Petroleum Geologists Bull., v. 31, no. 7, p. 1289–1293.

Keefer, W. R., and Troyer, M. L., 1964, Geology of the Shotgun Butte area, Fremont County, Wyoming: U.S. Geol. Survey Bull. 1157, 123 p.

Klepper, M. R., Weeks, R. A., and Ruppel, E. T., 1957, Geology of the southern Elkhorn Mountains, Jefferson and Broadwater Counties, Montana: U.S. Geol. Survey Prof. Paper 292, 82 p. [1958].

Knechtel, M. M., Smedley, J. E., and Ross, R. J., Jr., 1954, Little Chief Canyon member of Lodgepole limestone of early Mississippian age in Montana: Am. Assoc. Petroleum Geologists Bull., v. 38, no. 11, p. 2395–2400.

Kummel, Bernhard, 1954, Triassic stratigraphy of southeastern Idaho and adjacent areas: U.S. Geol. Survey Prof. Paper 254–H, p. 165–194.

Lammers, E. C. H., 1939, The origin and correlation of the Cloverly conglomerate: Jour. Geol., v. 47, no. 2, p. 113–132.

LeVan, D. C., and McLean, W. F., 1951, Structure and stratigraphy of the Red Canyon area, Gallatin County, Montana: Michigan Univ., unpub. M.S. thesis.

Liddicoat, W. K., and May, P. R., 1950, Geology of the Monument Mountain–Gallatin River area, *in* May, P. R., 1950, Geology of the Monument Mountain–Gallatin River area, Yellowstone National Park, and Gallatin County, Montana: Michigan Univ., unpub. M.S. thesis.

Lochman, Christina, 1950, Status of Dry Creek shale of central Montana: Am. Assoc. Petroleum Geologists Bull., v. 34, no. 11, p. 2200–2222.

Love, J. D., 1956, Cretaceous and Tertiary stratigraphy of the Jackson Hole area, northwestern Wyoming, *in* Wyoming Geol. Assoc. Guidebook 11th Ann. Field Conf., Jackson Hole, 1956: p. 76–94.

Lowell, W. R., and Klepper, M. R., 1953, Beaverhead Formation, a Laramide deposit in Beaverhead County, Montana: Geol. Soc. America Bull., v. 64, no. 2, p. 235–244.

May, P. R., 1950, Geology of the Monument Mountain-Gallatin River area, Yellowstone National Park and Gallatin County, Montana: Michigan Univ., unpub. M.S. thesis.

McKelvey, V. E., and others, 1959, The Phosphoria, Park City, and Shedhorn Formations in the western phosphate field: U.S. Geol. Survey Prof. Paper 313–A, 47 p.

McKelvey, V. E., Williams, J. S., Sheldon, R. P., Cressman, E. R., Cheney, T. M., and Swanson, R. W., 1956, Summary description of Phosphoria, Park City, and Shedhorn formations in western phosphate field: Am. Assoc. Petroleum Geologists Bull., v. 40, no. 12, p. 2826–2863.

McMannis, W. J., 1955, Geology of the Bridger Range, Montana: Geol. Soc. America Bull., v. 66, no. 11, p. 1385–1430.

——— 1962, Devonian stratigraphy between Three Forks, Montana and Yellowstone Park, *in* Billings Geol. Soc. Guidebook 13th Ann. Field Conf., Three Forks–Belt Mountains area, and symposium, the Devonian system of Montana and adjacent areas, 1962: p. 4–12.

——— 1963, LaHood Formation—a coarse facies of the Belt series in southwestern Montana: Geol. Soc. America Bull., v. 74, no. 4, p. 407–436.

Moberly, Ralph, Jr., 1960, Morrison, Cloverly, and Sykes Mountain formations, northern Big Horn Basin, Wyoming and Montana: Geol. Soc. America Bull., v. 71, no. 8, p. 1137–1176.

Moritz, C. A., 1951, Triassic and Jurassic stratigraphy of southwestern Montana: Am. Assoc. Petroleum Geologists Bull., v. 35, no. 8, p. 1781–1814.

Myers, W. B., and Hamilton, Warren, 1961, Deformation accompanying the Hebgen Lake, Montana, earthquake of August 17, 1959—single-basin concept, *in* short papers in the geologic and hydrologic sciences: U.S. Geol. Survey Prof. Paper 424–D, p. D168–D170.

——— 1964, Deformation accompanying the Hebgen Lake, Montana earthquake of August 17, 1959, *in* The Hebgen Lake, Montana, earthquake of August 17, 1959: U.S. Geol. Survey Prof. Paper 435, p. 55–98.

Newell, N. D., and Kummel, Bernhard, 1941, Permo-Triassic boundary in Idaho, Montana, and Wyoming: Am. Jour. Sci., v. 239, p. 204–208.

——— 1942, Lower Eo-Triassic stratigraphy, western Wyoming and southeast Idaho: Geol. Soc. America Bull., v. 53, no. 6, p. 937–995.

Nockolds, S. R., 1954, Average chemical composition of some igneous rocks: Geol. Soc. America Bull., v. 65, no. 10, p. 1015.

Nolan, T. B., 1943, The Basin and Range province in Utah, Nevada, and California: U.S. Geol. Survey Prof. Paper 197–D, p. 141–193.

Nordquist, J. W., 1953, Mississippian stratigraphy of northern Montana, *in* Billings Geol. Soc. Guidebook 4th Ann. Field Conf., Little Rocky Mountains, Montana, southwestern Saskatchewan, 1953: p. 68–82.

Norton, G. H., 1956, Evidences of unconformity in rocks of Carboniferous age in central Montana, *in* Billings Geol. Soc. Guidebook 7th Ann. Field Con., central Montana, 1956: p. 52–66.

Pardee, J. T., 1950, Late Cenozoic block faulting in western Montana: Geol. Soc. America Bull., v. 61, no. 4, p. 359–406.

Peacock, M. A., 1931, Classification of igneous rock series: Jour. Geology, v. 39, no. 1, p. 54–67.

Peale, A. C., 1873, Madison Valley geyser-basins to Gallatin City, Cherry Creek mines, *in* Hayden, F. V., 1873, Sixth Annual report of the U.S. Geological and Geographical Survey of the Territories for the year 1872: Washington, p. 159–167.

——— 1896, Description of the Three Forks sheet, Montana: U.S. Geol. Survey, Geol. Atlas, Folio 24.

Peck, R. E., 1941, Lower Cretaceous Rocky Mountain nonmarine microfossils: Jour. Paleontology, v. 15, no. 3, p. 285–304.

Rau, J. L., 1962, The stratigraphy of the Three Forks Formation *in* Billings Geol. Soc. Guidebook 13th Ann. Field Conf., Three Forks-Belt Mountains area, and symposium, the Devonian System of Montana and adjacent areas, 1962: p. 51–66.

Reid, R. R., 1957, Bedrock geology of the north end of the Tobacco Root Mountains, Madison County, Montana: Montana Bur. Mines and Geology Mem. 36.

Reiter, F. H., 1950, Geology of a portion of the Gallatin Range in southwestern Montana: Michigan Univ., unpub. M.S. thesis.

Richards, P. W., 1957, Geology of the area east and southeast of Livingston, Park County, Montana: U.S. Geol. Survey Bull. 1021–L, p. 385–438.

Richards, P. W., and Nieschmidt, C. L., 1961, The Bighorn Dolomite and correlative formations in southern Montana and northern Wyoming: U.S. Geol. Survey Oil and Gas Inv. Map OM–202.

Richmond, G. M., 1960, Correlation of alpine and continental glacial deposits of Glacier National Park and adjacent High Plains, Montana: U.S. Geol. Survey Prof. Paper 400–B, p. 223–224.

——— 1964, Glacial geology of the West Yellowstone basin and adjacent parts of Yellowstone National Park, *in* the Hebgen Lake, Montana, earthquake of August 17, 1959: U.S. Geol. Survey Prof. Paper 435, p. 223–236.

Richmond, G. M., and Hamilton, Warren, 1960, The late Quaternary age of obsidian-rhyolite flow in the western part of Yellowstone National Park, Wyo., *in* Short papers in the geological sciences: U.S. Geol. Survey Prof. Paper 400–B, p. B224–B225.

Roberts, A. E., 1961, Insoluble residues and Ca:Mg ratios in the Madison Group, Livingston, Montana, *in* Short papers in the geologic and hydrologic sciences: U.S. Geol. Survey Prof. Paper 424–B, p. B294–B296.

——— 1964, Geology of the Brisbin quadrangle, Montana: U.S. Geol. Survey Geol. Quad. Map GQ–256.

——— 1965, Correlation of Cretaceous and lower Tertiary rocks near Livingston, Montana, with those in other areas in Montana and Wyoming, *in* Geological Survey research, 1965: U.S. Geol. Survey Prof. Paper 525–B, p. B54–B63.

——— 1966, Stratigraphy of Madison Group near Livingston, Montana, and discussion of karst and solution breccia features: U.S. Geol. Survey Prof. Paper 526–B, 23 p.

Robinson, G. D., 1961, Origin and development of the Three Forks Basin, Mont.: Geol. Soc. America Bull., v. 72, no. 7, p. 1003–1014.

——— 1963, Geology of the Three Forks quadrangle, Montana: U.S. Geol. Survey Prof. Paper 370, 143 p.

Rogers, C. P., Jr., Gardner, L. S., and Hadley, H. D., 1945, Maps showing thickness and general distribution of Mesozoic and Paleozoic rocks in south-central Montana: U.S. Geol. Survey Oil and Gas Inv. Prelim. Map 43.

Ross, C. S., and Smith R. L., 1961, Ash-flow tuffs—their origin, geologic relations, and identifications: U.S. Geol. Survey Prof. Paper 366, 81 p.

Runner, J. J., and Thomas, L. C., 1928, Stratigraphic relations of the Cherry Creek group in the Madison Valley, Montana [abs.]: Geol. Soc. America Bull., v. 39, no. 1, p. 202–203.

Sandberg, C. A., 1963, Dark shale unit of Devonian and Mississippian age in northern Wyoming and southern Montana, *in* Short papers in geology and hydrology: U.S. Geol. Survey Prof. Paper 475–C, p. C17–C20.

——— 1965, Nomenclature and correlation of lithologic subdivisions of the Jefferson and Three Forks Formations of southern Montana and northern Wyoming: U.S. Geol. Survey Bull. 1194–N, 18 p.

Sandberg, C. A., and McMannis, W. J., 1964, Occurrence and paleogeographic significance of the Maywood Formation of Late Devonian age in the Gallatin Range, southwestern Montana, *in* Geological Survey research, 1964: U.S. Geol. Survey Prof. Paper 501–C, p. C50–C54.

Sando, W. J., and Dutro, J. T., Jr., 1960, Stratigraphy and coral zonation of the Madison Group and Brazer Dolomite in northeastern Utah, western Wyoming and southwestern Montana, *in* Wyoming Geol. Assoc. Guidebook 15th Ann. Field Conf., Overthrust belt of southwestern Wyoming and adjacent areas, 1960: p. 117–126.

Schmitt, G. T., 1953, Regional stratigraphic analysis of Middle and Upper marine Jurassic in northern Rocky Mountains–Great Plains: Am. Assoc. Petroleum Geologists Bull., v. 37, no. 2, p. 355–393.

Scholten, Robert, 1955, Pony–Cherry Creek age relations in southwestern Montana—a reply: Geol. Soc. America Bull., v. 66, no. 10, p. 1317.

Scott, H. W., 1935, Some carboniferous stratigraphy in Montana and northwestern Wyoming: Jour. Geology, v. 43, no. 8, pt. 2, p. 1011–1032.

Severson, J. L., 1952, A comparison of the Madison Group (Mississippian) with its subsurface equivalents in central Montana: Wisconsin Univ., unpub. Ph. D. thesis.

Sharpe, C. F. S., 1938, Landslides and related phenomena: New York, Columbia Univ. Press, 136 p.

Sloss, L. L., 1950, Paleozoic sedimentation in Montana area: Am. Assoc. Petroleum Geologists Bull., v. 34, no. 3, p. 423–451.

Sloss, L. L., and Laird, W. M., 1947, Devonian system in central and northwestern Montana: Am. Assoc. Petroleum Geologists Bull., v. 31, no. 8, p. 1404–1430.

Sloss, L. L., and Moritz, C. A., 1951, Paleozoic stratigraphy of southwestern Montana: Am. Assoc. Petroleum Geologists Bull., v. 35, no. 10, p. 2135–2169.

Spath, L. F., 1930, The Eo-Triassic invertebrate fauna of East Greenland, Medd. om Grønland, v. 83, p. 1–87.

——— 1935, Additions to the Eo-Triassic invertebrate fauna of East Greenland, Medd. om Grønland, v. 98, no. 2, 115 p.

Strickland, J. W., 1956, Mississippian stratigraphy, western Wyoming, *in* Wyoming Geol. Assoc. Guidebook 11th Ann. Field Conf., Jackson Hole, 1956: p. 51–57.

Tansley, Wilfred, Schafer, P. A., and Hart, L. H., 1933, A geological reconnaissance of the Tobacco Root Mountains, Madison County, Montana; Montana Bur. Mines and Geology Mem. 9.

Thom, W. J., Jr., Hall, G. M., Wegemann, C. H., and Moulton, G. F., 1935, Geology of Big Horn County and the Crow Indian Reservation, Montana: U.S. Geol. Survey Bull. 856, 200 p.

Tutten, W. D., 1960, Carrot Basin anticline, Gallatin County, Montana, *in* Billings Geol. Soc. Guidebook 11th Ann. Field Conf., West Yellowstone Earthquake area, 1960: p. 261–264.

U.S. Geological Survey, 1964, The Hebgen Lake, Montana, earthquake of August 17, 1959: U.S. Geol. Survey Prof. Paper 435, 242 p.

U.S. Weather Bureau, Climatological data for Montana [Annual Summary].

Vine, J. D., 1956, Geology of the Stanford-Hobson area, central Montana: U.S. Geol. Survey Bull. 1027–J, p. 405–470.

Witkind, I. J., 1961, Deformation of the epicentral area, Hebgen Lake, Montana, earthquake of August 17, 1959—dual-basin concept, *in* Short papers in the geologic and hydrologic sciences: U.S. Geol. Survey Prof. Paper 424–D, p. D165–D168.

——— 1962, Possible detachment faults in the Tepee Creek quadrangle, Gallatin County, Montana, *in* Short papers in geology, hydrology, and topography: U.S. Geol. Survey Prof. Paper 450–B, p. B6–B8.

——— 1964a, Preliminary geologic map of the Tepee Creek quadrangle, Montana-Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I–417.

——— 1964b, Reactivated faults north of Hebgen Lake, *in* the Hebgen Lake, Montana, earthquake of August 17, 1959: U.S. Geol. Survey Prof. Paper 435, p. 37–50.

——— 1964c, Geology of the Abajo Mountains area, San Juan County, Utah: U.S. Geol. Survey Prof. Paper 453, 110 p.

Witkind, I. J., Myers, W. B., Hadley, J. B., Hamilton, Warren, and Fraser, G. D., 1962, Geologic features of the earthquake at Hebgen Lake, Montana, August 17, 1959: Seismol. Soc. America Bull., v. 52, no. 2, p. 163–180.

Yen, Teng-Chien, 1951, Fresh-water mollusks of Cretaceous age from Montana and Wyoming: U.S. Geol. Survey Prof. Paper 233–A, 20 p. [1952].

INDEX

[Italic page numbers indicate major references]

U.S. GOVERNMENT PRINTING OFFICE: 1969 O—319-733