Geology of the Three Forks Quadrangle Montana

GEOLOGICAL SURVEY PROFESSIONAL PAPER 370

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By G. D. ROBINSON

With sections on

Petrography of Igneous Rocks

By H. FRANK BARNETT

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A comprehensive restudy of a classic area of Rocky Mountain geology at the head of the Missouri River

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GEOLOGY OF THE THREE FORKS QUADRANGLE, MONTANA

By G. D. Robinson

ABSTRACT

The Three Forks quadrangle at the head of the Missouri River, has a long stratigraphic succession. The oldest rocks are Precambrian gneiss, derived from arkosic sandstone, and amphibolite, probably altered from mafic sills and lava flows.

The unmetamorphosed sedimentary rocks range in age from late Precambrian to Recent, have an aggregate thickness of about 11,000 feet, and underlie about 85 percent of the quadrangle. The only systems not represented are Ordovician, Silurian, and Triassic.

The oldest unmetamorphosed sedimentary rocks are at least 4,000 feet of marine micaceous arkose and conglomerate, the North Boulder formation of the Belt series of late Precambrian age. The formation is a wedge arkose controlled by a buried east-west fault of pre-Belt or Belt age here named the Willow Creek fault.

The rocks of the North Boulder formation are disconformably succeeded by about 1,500 feet of marine strata of Middle and Late Cambrian age separated into five conformable formations: the Flathead sandstone, 0–100 feet thick (oldest); the Wolsey shale, 200–400 feet thick; the Meagher limestone, 300–500 feet thick; the Park shale, 100–200 feet thick; and the Pilgrim limestone, 300–450 feet thick (youngest).

No rocks of latest Cambrian, Ordovician, Silurian, and earlier Devonian age are recognized. The Upper Cambrian Dry Creek shale and Pebbly limestone of Peale, that overlie the Pilgrim nearby, are not present here. Rather, the Pilgrim is disconformably overlain by the marine Maywood formation, generally less than 50 feet thick, but as much as 200 feet thick, of Late(?) Devonian age.

Succeeding the Maywood conformably is the marine Jefferson dolomite, 350–650 feet thick, of Late Devonian age.

Dolomite is a major component of the Cambrian and Devonian carbonate rocks. The dolomite in mottled carbonate rocks is plainly of replacement origin. The mottling itself, widespread in nonmagnesian limestone, is primarily a structural and not a chemical phenomenon. Those dolomite rocks which are extensive, homogeneous, and have sharp contacts with neighboring limestone seem to have been deposited directly.

Conformably above the Jefferson is the marine Three Forks shale, 100–400 feet thick, of Late Devonian and Mississippian age. Orange calcareous siltstone and sandstone at the top of the Three Forks were separated as the Sappington formation of Mississippian age by Berry. The Early Mississippian age of part of these strata is confirmed. The orange siltstone is gradational with the underlying green shale and is in sharp contact with the overlying Lodgepole limestone. They are retained in the Three Forks, rather than transferred to the Lodgepole, on grounds of priority, lithologic community, and cartographic practicality.

Succeeding the Three Forks are about 2,400 feet of marine Mississippian and Pennsylvanian rocks classed together as Carboniferous because the Mississippian-Pennsylvanian boundary has not been located even approximately. From bottom to top, the formations include the Lodgepole limestone, 350–650 feet thick, and the Mission Canyon limestone, 600–1,400 feet thick, both of the Madison group of early Mississippian age; the Big Snowy formation, 0–400 feet thick, and the Amsden formation, 250–600 feet thick, mapped together and of late Mississippian and early Pennsylvanian age; and the Quadrant formation, 200–400 feet thick of presumed Pennsylvanian age. Field evidence suggests that the Quadrant may be partly of early Permian age. A slight break in sedimentation preceded the Lodgepole, and more significant erosional intervals preceded the Big Snowy and the Amsden. In the pre-Big Snowy interval an extensive zone of solution breccia and locally of karst topography was developed in the upper part of the Mission Canyon limestone.

The Permian is definitely represented only by the marine Phosphoria formation, 75–200 feet thick, that conformably and probably gradationally succeeds the Quadrant.

The Mesozoic record is fragmentary. Triassic time is represented by an unconformity that is covered by the Ellis formation, generally less than 150 feet thick but locally more than 350 feet thick, of Jurassic age. Two other Mesozoic formations have been recognized. The oldest is the landlaid Morrison formation 100 to 250 feet thick, of Late Jurassic age, which conformably(?) succeeds the Ellis and is gradationally overlain by the Kootenai formation, as much as 650 feet thick, of Early Cretaceous age. The Kootenai is succeeded, with probable angular unconformity, by the Elkhorn Mountains volcanics, a thick sequence of andesitic flows and volcanic breccia of Late Cretaceous age. Great thicknesses of marine and continental sedimentary rocks of Late Cretaceous and Paleocene age occupy large areas near the quadrangle but do not crop out within it.

The Tertiary is represented by thick basin deposits of the Bozeman group (herein named, formerly Bozeman lake beds) more than 2,000 feet thick and divided into four formations: Sphinx conglomerate, 0–100 feet thick of Eocene age; the Milligan Creek formation (new name), 0–300 feet thick of Eocene age; Climbing Arrow formation (new name), 750–1,000 feet thick, of late Eocene and early Oligocene age; and the Dunbar Creek formation (new name), 250–800 feet thick, of Oligocene age. The Sphinx conglomerate is an alluvial apron. The Milligan Creek formation is largely lake limestone. The Climbing Arrow formation is mainly streamlaid bentonitic clay and sand. The Milligan Creek and Climbing Arrow contain much volcanic ash, and the Dunbar Creek is made up largely of such debris, blown or washed into the Three Forks basin. Included in the Bozeman group as defined are middle

and upper Tertiary gravels, widespread nearby but present here only in small unmappable patches.

A wide variety of unconsolidated Quaternary deposits completes the stratigraphic succession. Broad benches in the southern and central parts of the quadrangle are capped by rounded gravel deposited in early to late Pleistocene time by the ancestral Jefferson and Madison Rivers. Valleys tributary to the ancient master streams, in the northern part, are filled with subrounded gravel. Fan gravel, of late Pleistocene and Recent age, mantles many benchlands north of the Jefferson River, and floors many short tributary valleys. Late Pleistocene windblown silt thinly coats large upland areas. The present broad flood-plain and channel deposits are largely of very late Pleistocene age.

Masses of calcalkaline intrusive rocks occupy large areas in the quadrangle. All but two are of very Late Cretaceous to very early Tertiary (pre-Bozeman) age. They are mainly monzonitic, dacitic, latitic, and andesitic intrusive rocks in the form of sills, laccoliths, and one semiconcordant pluton emplaced during or after regional folding. Some were probably being intruded while the Elkhorn Mountains volcanics were being extruded. The exceptions are two small bodies of olivine basalt, intrusive into limestone conglomerate at the base of the Bozeman group and probably of early Tertiary age.

The structural history is complex. The earliest recorded structural event was the pre-Belt folding and metamorphism of the lower Precambrian crystalline rocks. Much later, but still in Precambrian time, the southern limit of the embayment that received sediments that formed the Belt series became established apparently through faulting. This buried structural feature, the Willow Creek fault, exerted a powerful influence on much subsequent deposition and deformation. The quadrangle was structurally quiet, though far from dormant, from late Precambrian to Late Cretaceous time.

Most of the folding and faulting were in response to compression along north-south axes in a comparatively short time: very late in the Cretaceous to middle or late Eocene. During this time, referred to as the Laramide interval, large-scale folding occurred first; major thrusting followed, perhaps after a period of quiescence.

A broad belt of thrusting, called the Sixteenmile thrust zone, developed, but only the uppermost thrust is exposed. This thrust consists of the well-known Lombard thrust that trends north-south across the northeast part of the quadrangle, and the equally well known Jefferson Canyon thrust that trends east-west across the west-central part of the quadrangle, linked together by a thrust of intermediate trend, first described in this report and called the Highway thrust, in the center of the quadrangle. The Lombard-Highway-Jefferson Canyon thrust dips northwestward at angles of 40°–60°. Rocks as old as the Belt series are thrust eastward over rocks as young as the Elkhorn Mountains volcanics and even younger latite intrusive rocks; not less than 10,000 feet of stratigraphic throw is involved. Their total displacement is probably greater than the stratigraphic throw because the fault dips in the same direction and at about the same angle as the beds. A minimum displacement of 3 miles seems certain; maximum displacement may be much greater. Principal relative movement on the thrust system seems to have been from west to east so that the Jefferson Canyon thrust segment has probably experienced mostly strike-slip displacement. A cluster of tilted fault blocks above the thrust surface, and several high-angle faults below, are probably contemporaneous with the thrust advance.

Only two steep faults are known or inferred to be distinctly younger than the thrusting but older than the Tertiary basin deposits. Possibly, however, many developed and controlled the general dimensions of the Tertiary basin. If so, they are masked by continental deposits shed from the fault scarps.

All but two of the many intrusive masses were emplaced during or shortly after folding. Time relations between intrusions and thrusting are unknown; a reasonable guess is that the period of intrusion overlapped the period of thrusting.

The Tertiary basin is structurally controlled at least to the extent that it lies on a zone of repeated faulting and its borders cut sharply across deformed older rocks but there is no deciding to what extent it is due directly to faulting or to what extent to erosion of unresistant rocks exposed by diastrophism. There is no evidence to support the theses of Atwood that the region was peneplaned between Laramide deformation and creation of the basin, and that the basin formerly drained southwestward to the Snake River.

Deformation during the Tertiary seems to have been mostly in the form of gentle eastward tilting, recurring or continuing from late Eocene until the close of the Tertiary, and punctuated by shallow local folding. Quaternary deformation has apparently been limited to slight uplift with northward or northwestward tilt.

At the end of Tertiary time the quadrangle seems to have been an area of low relief with scattered hills protruding no more than a few hundred feet above a graveled plain. The development of the present physiography is the history of the dissection of that surface and of the exhumation of a very early Tertiary surface of mature relief that had been buried under the Tertiary basin deposits. The main agent of dissection has been a stream system of large perennial rivers, which derive their discharge from a persistently humid region far beyond the Three Forks basin, and their tributaries, which vary from perennial to nearly dry as the basin climate has varied. The former flood plains of the through streams remain as terrace remnants capped by rounded gravel, and the courses of their tributaries are marked by subrounded gravel fills. Broad benchlands in the northern part of the area that are virtually bare-rock surfaces are identified as pediments. Apparently they have long been isolated from integrated streams and have been slowly lowered by mainly colluvial processes.

The dissection has been episodic, presumably with long periods of lateral cutting and filling by the streams alternating with short ones of vigorous downcutting. The prime driving force in dissection seems to have been uplift of the mountain ranges to the south, with climatic change playing a secondary role. Although uplift must have caused the dissection, discontinuities in uplift were not necessarily responsible for the discontinuities in dissection revealed by the benched topography. The main proximate cause for the development of some or all of the benches may have been differing resistance of strata at the outlet canyon of the Missouri River; major Quaternary climatic changes doubtless contributed, but to what degree is unknown. Climatic change in the past century probably is responsible for deep trenching of the flood plains of tributary streams.

The Three Forks quadrangle has yielded no mineral products other than small quantities of sand and gravel and of glass sand. Its petroleum possibilities have not been tested but they do not seem promising.

INTRODUCTION

The region around the headwaters of the Missouri River rich in agricultural and mineral resources, in handsome scenery and complex geology, has long attracted geologists. More than 70 years ago, the classic Paleozoic stratigraphy of the northern Rocky Mountains was established in this vicinity by A. C. Peale, and many later geologists contributed to the regional literature. Few large-scale studies, however, have been published. The present report portrays the general geology of the Three Forks 15-minute quadrangle, which is at a scale of 1 inch to 1 mile and covers 205 square miles. The quadrangle takes its name from the town of Three Forks, which in turn is named for the three rivers—Jefferson, Madison, and Gallatin—that join nearby to make the Missouri. The quadrangle is bounded by latitudes 45°45′ and 46°0′ N. and by longitudes 111°30′ and 111°45′ W. (See fig. 1.) The quadrangle was mapped during the summers of 1952, 1953, and 1954. In 1952, I was assisted by Thomas P. Lovett, and in 1953 and 1954, by John S. Bader, Jr.

Geology was plotted on 1:24,000 preliminary topographic maps, prepared by multiplex methods, and transferred photographically to publication scale (pl. 1). Most geologic contacts were traced on foot. Color terms used in measured sections follow the "Rock-Color Chart of the National Research Council" (Goddard and others, 1948), but this usage is not consistently followed in the rest of the text.

References to the neighboring Toston quadrangle (see fig. 1*B*) are based on my work there during the summers of 1955–1957. (See Robinson, 1959, for interim report.)

An effort is made to view the Three Forks quadrangle in its regional setting, as part of the Rocky Mountains in western Montana. Little is known about the northern half of this geologic entity, so that the region considered is limited to southwestern Montana; more specifically, Montana south of the latitude of Helena (46°40′) and west of the longitude of the Bridger Range (110°50′). (See fig. 1*A*.)

PREVIOUS WORK

The Three Forks region was one of the first areas studied in the Northwest, perhaps because of Captain Meriwether Lewis' enthusiastic accounts. F. V. Hayden, as a member of the Raynolds Expedition, made the first geologic observations here in 1860 (Hayden, 1869, p. 90). Geologists of the Hayden Survey began working in the region as early as 1868 and a brief account of their findings was published by Hayden (1873, p. 65–85) 5 years later. Paleozoic and Cretaceous fossils collected by Hayden and his colleagues were discussed by Meek (1873). In 1883, members of the northern Transcontinental Survey hastily examined the north half of the quadrangle (Pumpelly and others, 1886). From 1883 to 1889 Peale mapped the 1° quadrangle (45°–46° N. and 111°–112° W.) called the Three Forks sheet, of which the present Three Forks quadrangle is a part (see fig. 1*B*); all unqualified references in this paper to the Three Forks quadrangle are to the 15-minute sheet. Peale published a report on the Paleozoic stratigraphy (1893), and another on the general geology (1896) in folio form at a scale of 1:250,000. While Peale was making his areal map for the Geological Survey, Earl Douglass (1899 and later publications) was independently studying the Tertiary deposits and their vertebrate fossils. Matthew (1903) also collected vertebrate fossils from the Tertiary rocks. The Upper Devonian and Lower Mississippian faunas in the hills north of Three Forks were investigated by P. E. Raymond (1907, 1909, 1912) and somewhat later by W. P. Haynes (1916a). As a byproduct of his paleontologic research, Haynes (1916b) discovered the Lombard thrust.

The Permian phosphatic rocks in the southern part of the quadrangle were traced by D. D. Condit (1918) but not specifically discussed. Frenzel and Mundorf (1942) examined the phosphatic rocks in the northern part of the quadrangle and found diagnostic Permian microfossils in them. The Precambrian crystalline rocks at the south end of the quadrangle were visited by Tansley and Schafer (Tansley, Schafer, and Hart, 1933, p. 8–9, pl. 1). Wood (1933) described a Miocene rhinoceros found a few miles south of Three Forks. Scott (1935) briefly discussed some of the Carboniferous rocks of the quadrangle.

By the late 1930's it was evident that Peale's work, though reconnaissance of a high order, was too generalized for modern use, and several detailed areal mapping and stratigraphic projects were undertaken in the region. Only one of these, however, by G. W. Berry (1943), produced new mapping in the Three Forks quadrangle. Berry's map, covering most of the Jefferson Island and Three Forks 15-minute quadrangles (see fig. 1*C*), is a distinct advance over Peale's. Unfortunately, however, Berry had no topographic base map and had to publish his work at 1:125,000 in black-and-white on a poor planimetric base. His text is brief and barely touches most aspects of the geology. Its main contribution lies in the application of modern names and definitions to many formations.

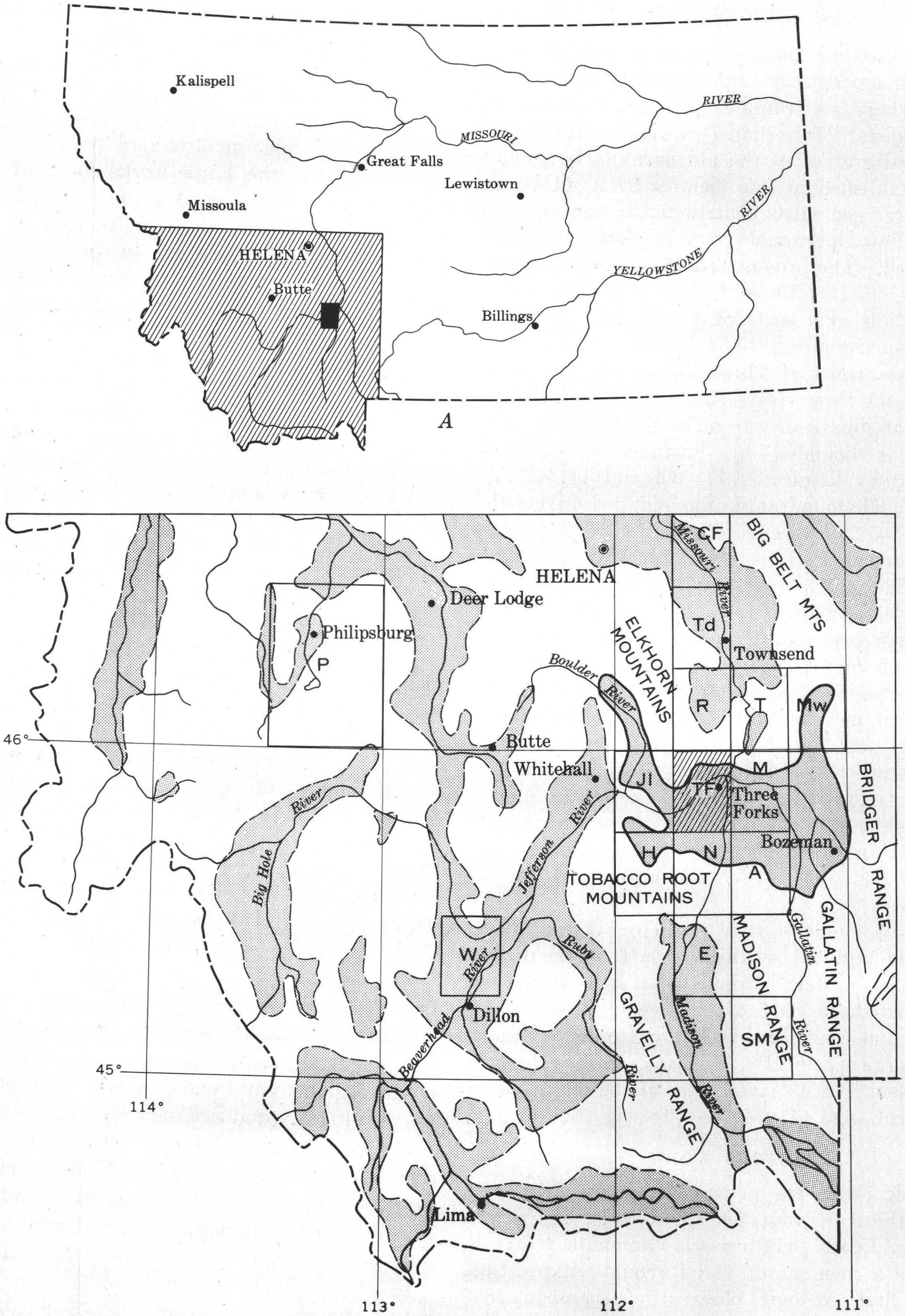

FIGURE 1.—Index maps of the Three Forks quadrangle, Montana. *A*. Outline of Montana, showing southwestern Montana (ruled) and Three Forks quadrangle (black). *B*. Southwestern Montana, showing Cenozoic basins, ranges, rivers, and topographic quadrangles mentioned in this report; basins are shaded and the Three Forks basin is outlined heavily. Key to topographic quadrangles: P, Philipsburg; W, Willis 1; CF, Canyon Ferry; Td, Townsend; R, Radersburg; T, Toston; Mw, Maudlow; JI, Jefferson Island; TF Three Forks; M, Manhattan; H, Harrison; N, Norris; A, Anceney; E, Ennis; SM, Sphinx Mountain.

Since Peale's day there has been difficulty in dating precisely the rocks near the Devonian-Mississippian boundary. Berry's attempted solution to this problem —establishment of the Sappington formation of supposed Mississippian age—led to field examinations by Sloss and Laird (1946, 1947) and by Holland (1952).

Partial stratigraphic sections in the quadrangle have been published by Gardner and others (1946), Imlay and others (1948), and Moritz (1951).

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GEOGRAPHY

The Three Forks quadrangle is in the southeastern part of the Northern Rocky Mountains province of Fenneman. Consisting of deeply dissected mountain uplands separated by broad benchlands, it is a typical, if somewhat subdued, fragment of the province. The brief discussion of geography which follows is designed as background for the geologic description. Further consideration of the landscape is in a later discussion of geomorphology.

TERRAIN

The Three Forks quadrangle is in the northwest part of an irregularly shaped intermontane basin of about 1,000 square miles at the head of the Missouri River. Unlike most of the intermontane basins in Montana, which trend north-south, the Three Forks basin, as it is commonly known, is elongated east-west. (See fig. 1*B*.) It is bounded on the southwest by the Tobacco Root Mountains (also called the Jefferson Range), on the northwest by the Elkhorn Mountains, on the north by the Horseshoe Hills, on the east by the Bridger Range, and on the south by the Madison and Gallatin Ranges.

The landscape of the quadrangle is mainly a series of broad benchlands, from which rise only a few rugged ridges and hills. The total relief is about 1,800 feet, from 4,035 feet at the head of the Missouri River in the northeast corner to 5,800 feet at the west-central edge on the ridge south of Dry Hollow. The lowest benchlands are the flood plains of the Jefferson and Madison Rivers and their tributaries; the broad flood plain of the Jefferson is interrupted by a short canyon reach near its entrance to the quadrangle. A series of benches rising steplike above the flood plains make up the greater part of the area. South of the Jefferson River the benches tend to be broad and smooth topped, and rise 1,000 feet in 5 steps to form the highest surfaces. Those north of the river are small, irregular, and overshadowed by hills.

The few rugged hills are rarely more than 1,000 feet high, with narrow but flattish summits. In most places, the hills are arranged in long parallel ridges separated by narrow valleys or saddles. A few low hilly areas, however, are rolling, without well-defined trends.

With minor exceptions, there is close correspondence between topographic expression and lithology. The long ridges are held up by Paleozoic and Mesozoic carbonate rocks and quartzitic sandstone, the intervening narrow valleys by Paleozoic and Mesozoic clay-rich rocks, the rolling hills by large masses of igneous rocks, and the benchlands by Cenozoic deposits.

FLORA AND FAUNA

The Three Forks area is dry country; the vegetation is meager and the characteristic landscape color is dull grayish yellow. Green patches of brush and woods are scattered about the flood plains of the Jefferson and Madison Rivers as well as along many minor stream courses at lower altitudes. The more common trees are willow, aspen, cottonwood, and poplar; conifers are rare. Some patches of low grassy swamp on the flood plains, especially near the Missouri, mark areas of recurrent flooding and recently abandoned stream channels. Most of the flood plains of the Jefferson and Madison Rivers and Willow Creek are wet farmed, mainly for wheat and alfalfa. Most of the benches are dry farmed, largely for wheat. Uncultivated benches, principally those underlain by gravel, are thinly strewn with sagebrush, cactus, bunch grass, tumbleweed, and an occasional pine. On steeper slopes, such as bench fronts, valley sides, and hill sides, are many bare rock outcrops broken by sagebrush and sparse grasses.

The area is in a game preserve and for this reason wild animals are more abundant than in similar areas nearby. Deer families and small herds of antelope are common as are rabbits, chipmunks, and ground squirrels. A few porcupines, badgers, skunks, and rattlesnakes live near streams. Birds are plentiful and varied.

CLIMATE

The climate is north temperate semiarid, characterized by long windy intensely cold snowy winters, short wet warm springs, hot dry summers, and cool autumns with long spells of "Indian summer." From late June through August midday temperatures are high, often above 90°F and occasionally above 100°F, but nights are cool, and freezing temperatures have been recorded in every month. Maximum daytime temperatures in the winter months are commonly near freezing; minimum winter temperatures are often below –15°F, though rarely below –30°F. Mean annual temperature is near 44.5°F. The highest temperature recorded is 105°F, the lowest –49°F. Average annual precipitation is between 10 and 15 inches, of which a large fraction falls as snow during the winter and early spring. Maximum rainfall is in late May and June. The remainder of the summer and the autumn are likely to be dry, with light rains or snowfall in September and October. Heavy snowfalls are not the rule until November or December. Geologic fieldwork is often feasible from late April through November but is generally comfortable only from early June through mid-October.

These climatic notes are based on the U.S. Weather Bureau's "Climatological Summary" (to 1930) and "Annual Climatological Data" (after 1930). The Weather Bureau maintained a station at Three Forks from 1910 to 1930, and has operated one at Trident on the Missouri, 6 miles northeast of Three Forks, continuously since 1938.

HISTORY AND SETTLEMENTS

The party of Lewis and Clark were the first white men to enter the Three Forks quadrangle. They reached and named the Three Forks of the Missouri in July 1805. Several hunters, traders, and adventurers are known to have crossed the area in later decades, but the first settlement was not made until 1862 (by another account 1864) when Gallatin City was founded just east of the junction of the Madison and Jefferson Rivers. The settlers apparently thought that they were at the head of Missouri navigation and that a mercantile center might be developed there to serve nearby mining camps as well as the agricultural communities which might be expected to develop in the fertile surrounding valleys. When the putative merchants learned about the Great Falls far downstream, Gallatin City was largely abandoned. In 1882 a small settlement was started a mile southwest of Gallatin City, between the Madison and the Jefferson, and called Three Forks. In 1908 the site of Three Forks became a railroad division point and the town began to grow. It soon became clear that the townsite was too low and too close to the river junction for safety and about 1912 the town was moved 1½ miles southwest to its present site. A community was planned for several thousand people, with carefully laid out streets and concrete sidewalks. In the 1930's and early 1940's the town declined, and many of its houses were moved away. On the present outskirts it is surprising to see neat sidewalks framing vacant pasture. The present population is about 900, largely railroad employees, and ranchers who like living in town and commuting to their ranches.

The only other settlement is Willow Creek, 7 miles to the southwest, a farming community with less than 100 inhabitants.

Interesting, occasionally conflicting, discussions of local history are in Vestal (1945), Campbell and others (1915), and Federal Writers Project (1939). These sources were used for history before 1900. History after 1900 is based on these works modified by conversations with elder inhabitants.

TRAVEL

The road system is excellent. Three Forks is connected with Bozeman and the east, with Butte and the west, and with Helena and the north by first-class heavy-duty U.S. Highway 10. From a junction 2½ miles west of Three Forks and northwest of the Jefferson River the highway branches north and west. East of the junction the road is designated simply U.S. Highway 10. The branch leading north through Helena is called 10N; the branch leading west through Butte is called 10S. At the west edge of the quadrangle, paved State Highway 1, which goes south to Ennis and beyond, joins 10S. Good graveled county roads make access easy to every part of the quadrangle.

Both the Northern Pacific and the Chicago, Milwaukee, and St. Paul railroads follow the Jefferson Valley to Three Forks and there turn southeast. Only the Northern Pacific, however, normally stops at Three Forks. Small planes may land on an airstrip southwest of town. The nearest commercial airfield is outside Belgrade on Highway 10 about 20 miles to the southeast. There is no commercial traffic on the Jefferson or Madison Rivers, and even sportsmen rarely try to navigate the swift and dangerous currents.

PRECAMBRIAN METAMORPHIC ROCKS

Metamorphic rocks of Precambrian age include gneiss, derived from sedimentary rocks, amphibolite, probably altered from mafic sills and lava, and hematitic phyllonite. These rocks have no known unmetamorphosed equivalents.

GNEISS AND AMPHIBOLITE

Gneiss and amphibolite underlie about 3 square miles of the southwest corner of the quadrangle. They tend to be deeply weathered, so that outcrops are few, small, and very likely not representative. The best exposures are near contacts with the overlying resistant Flathead sandstone and in Willow Creek canyon (secs. 7 and 18, T. 1 S., R. 1 E.). Typically, layers many feet thick of light-colored gneiss alternate with similarly thick layers of dark amphibolite. Near the Flathead, gneiss is the main rock type; at the south edge of the quadrangle, amphibolite dominates. Banding strikes N. 70°–90° W. and dips 50°–70° N; foliation is parallel to banding. The overlying Flathead sandstone has similar strike but its dip is much lower, 10°–40° N., indicating great angular unconformity.

The dominant rock type is light-colored medium-grained biotite gneiss. Mineral proportions range widely; the specimens studied contain 40–60 percent quartz, 25–45 percent feldspar, 5–20 percent biotite, and 1–2 percent of accessory minerals. Mosaic or sutured texture is typical of quartz aggregates, and almost every quartz grain shows pressure shadows. The main feldspar is oligoclase, which is mostly untwinned and not strikingly oriented. Many of the rocks also contain colorless microcline, some as much as 20 percent. The biotite is brown or olive and is scattered in small bundles and individual flakes. Common accessory minerals are apatite, clinozoisite, sphene, and opaque oxides. Many of the grains of accessory minerals are rounded. A few biotite gneiss layers contain from 5 to 25 percent of faintly pink garnet; except for the garnet these rocks are similar to the other gneisses. Microcline and garnet do not occur together in the small suite examined.

The interbedded amphibolite is dark, almost black, medium grained, and gneissic rather than schistose, despite its large proportion of hornblende. It is mineralogically very simple, averaging about 60 percent hornblende, about 35 percent plagioclase (generally rather calcic labradorite), 1–2 percent each of quartz and of sphene, and a few percent of opaque oxides. Wide departures from these proportions are common. The hornblende is green, moderately pleochroic, and in rather stubby prisms. Feldspar grains are generally clear and little twinned; the edges of some, however, are crowded with sericite flakes and a few are thoroughly saussuritized. Quartz is interstitial and lacks pressure shadows; it developed late, perhaps long after the foliation. The texture of the amphibolite varies a good deal, from definitely nematoblastic to mosaiclike granoblastic. In the granoblastic rocks the feldspars are less twinned, and large porphyroblasts of sphene are common, suggesting a slightly higher metamorphic grade than in the nematoblastic ones.

Biotite-hornblende gneiss, mineralogically transitional between biotite gneiss and amphibolite, crops out sporadically. Its relation to the main types is not known owing to poor exposures.

The stratigraphic thickness represented by the metamorphic rocks is not known. At least several hundred feet and probably several thousand feet of strata are represented.

Small dikes of aplite and granite pegmatite, and veins of milky quartz are widespread both along and across the banding in all the metamorphic rocks. Most of them are but a few inches thick and impersistent, but one large mica pegmatite body (see p. 122) crops out just below the Flathead sandstone in the southeast corner of sec. 9, T. 1 S., R. 1 W. None of these transgressive masses is visibly metamorphosed. Many of them occur near the Flathead contact, but none of like character invade Paleozoic strata.

HEMATITIC PHYLLONITE

Hematitic phyllonite, markedly different from the coarse-grained metamorphic rocks, forms small scattered outcrops between the coarse crystalline rocks and the Flathead sandstone in secs. 8 and 9, T. 1 S., R. 1 W. These rocks, less than 100 feet thick, were not mapped separately.

The contact of the phyllonite with the Flathead is an angular unconformity. The phyllonite strikes roughly parallel to the Flathead, but dips much more steeply: near the contact, dips in the Flathead are 20°–30° N., whereas banding and foliation in the phyllonite dip 60° N. to vertical. Contacts between phyllonite and the main mass of gneissic rocks are not exposed, but in places the phyllonite contains a few thin bands of light-colored gneissic rock much like the underlying biotite gneiss.

The phyllonite is a dark fine-grained rock, composed mainly of hematite, quartz, chlorite, and sericite, but there is much diversity and no two outcrops are identical. The individual bodies differ mainly in degree of foliation.

Most of the outcrops are of medium-dark-gray hard massive faintly foliated spotted rock. The spots are subrounded to angular sand-size clusters of tiny grains of quartz or hematite, or of limonite or calcite replacing hematite. The groundweb consists of microcrystalline hematite, chlorite, sericite, and quartz in laminae, streaks, and patches.

The hematite, variably altered to limonite, is in cross fractures and also forms cement around quartz grains and sericite aggregates, indicating late development. The sericite shows no preferred orientation, even in the more highly foliated types. Rarely present are tiny remnants of feldspar grains, riddled with sericite flakes, and small rounded grains of sphene and epidote-group minerals.

A rare type is strikingly banded, and has half-inch wide light-olive-gray layers, rich in sericite, that alternate with medium-bluish-gray layers, rich in hematite and chlorite; within the color layers, mineral banding is developed down to microscopic scale.

The texture of the banded rock is cataclastic; shearing is less evident in the massive types.

ORIGIN

The gneiss and amphibolite are the product of moderately high grade regional metamorphism, as indicated by their mineralogy and their setting as part of the huge mass of similar rocks that forms the core of the Tobacco Root Mountains and the Madison and Gallatin Ranges.

The gneiss has developed from arkosic sandstone and conglomerate, as indicated by outlines of relict pebbles and cobbles, and by relict graded bedding and crossbedding; sedimentary origin is further suggested by the roundness of such refractory accessories as sphene and by the very high quartz content of many layers.

The amphibolite may represent silica-poor, iron-rich shale or tuff, but more probably was once mafic lava or sills. Hayden's comment (1872, p. 67) on similar rocks exposed in Madison Canyon nearby attests to the antiquity of this interpretation: "Masses of a very compact black hornblende gneiss lie between the strata as if they were old intrusions of trap."

The phyllonite seems to be the product of local cataclastic retrograde metamorphism of gneiss (thus the choice of name; see Williams, Turner, and Gilbert, 1954, p. 199–228). Details of mineralogy and texture leave some questions, but the presence within the phyllonite of gneiss bands, apparently representing unsheared parts, seems conclusive. The phyllonite could scarcely have reached its present state during the same metamorphic cycle that produced the interleaved and bordering gneiss. The fact that the phyllonite belt is in an important tectonic zone active in Precambrian times (pl. 3) encourages the view that the phyllonite is the product of local shearing. The "spots" and the persistently unoriented sericite both suggest an episode of recrystallization after shearing and at higher temperature.

AGE

These metamorphic rocks cannot yet be dated closely. At only three places in southwestern Montana have ancient crystalline rocks been reported in depositional contact with the next younger rocks of the Belt series. Two of these reported occurrences, one south of the Rochester mining district (Sahinen, 1939, p. 14) and the other in the Tendoy Mountains (Scholten, Keenmon, and Kupsch, 1955, p. 353), are reconnaissance observations, and may not hold up under detailed mapping. Only the third occurrence, 6 miles southeast of Whitehall (Lowell, 1956b), seems well established. In many places, however, rocks of the Belt series are faulted against pre-Belt metamorphic rocks, or crop out within a few miles of the older rocks. At or near depositional contacts the Belt is well indurated but unmetamorphosed, or nearly so, whereas the crystalline rocks are high-grade metamorphic products. The crystalline rocks are therefore not metamorphosed Belt rocks but are older by one or more major metamorphic and erosional cycles.

The shearing which made phyllonite from gneiss is younger than the regional metamorphism and older than the Flathead sandstone. Probably it is pre-Belt, but satisfactory evidence for this is lacking.

The age of the many small quartz-rich dikes and veins is uncertain. Because no such bodies occur in the nearby Paleozoic rocks, which contain many intrusive masses of a different character, a Precambrian age may be inferred. Because these rocks are not metamorphosed, they are presumably late Precambrian. Pebbles of vein quartz in the conglomerate of the Belt series show that some quartz veins are pre-Belt. Neither aplite nor pegmatite is represented in the local conglomerate of the Belt, but this negative fact does not prove their post-Belt age.

REGIONAL RELATIONS

In 1873, Hayden (p. 67) said of the crystalline rocks of the Missouri headwaters region: "The work of reducing these metamorphic strata to a system, and connecting them over extended areas, has not yet been attempted, and it seems to me an almost hopeless as well as fruitless task." The task has still not been attempted.

The great bulk of the metamorphic rocks in this region, including those described here, were mapped as a single unnamed unit—gneiss and schist—and assigned an Archean age by Peale (1896, p. 2). In a small area bordering the Madison River, southeast of Virginia City, Peale recognized another, younger group of Precambrian metamorphic rocks—marble, mica schist, and gneiss—which he named Cherry Creek formation and assigned to the Algonkian.

The metamorphic rocks of the same general area were later subdivided by Tansley (Tansley, Schafer, and Hart, 1933) into two series, the Pony (older) and the Cherry Creek, and these terms have been widely used. The Pony is described as "gneiss and schist of igneous and sedimentary origin;" the Cherry Creek is "garnetiferous gneiss, schist, quartzite, and limestone * * * whose sedimentary origin is unquestionable." The Pony is said to contain no limestone or quartzite, and comparatively little garnet-bearing rock. Much of Peale's unnamed Archean unit is included in Tansley's Pony series. Tansley's application of the name Cherry Creek differs so widely from Peale's (compare their maps) that Tansley's use of the name seems unwise.

The metamorphic rocks of the Three Forks quadrangle are included with the Pony series by Tansley (1933, pl. 1, p. 6); the nearest rocks assigned to the Cherry Creek series are 15 miles southwest. If they must be assigned to one of these series, the Pony is preferable. But the Pony and Cherry Creek series of Tansley are based on rapid reconnaissance work, and are likely to be radically modified or abandoned when studied in detail. Reid (1957), for instance, stated that the Pony rocks near their type locality are younger than Cherry Creek rocks. Consequently, the metamorphic rocks of the Three Forks quadrangle are assigned no formal name.

SEDIMENTARY ROCKS

The sedimentary rocks of the Three Forks quadrangle range in age from Precambrian to Recent, have an aggregate thickness of around 11,000 feet, and underlie about 85 percent of the quadrangle. The sequence is summarized in table 1. The only systems not represented are Ordovician, Silurian, and Triassic.

All the Jurassic and older sedimentary rocks are marine. All younger deposits are mostly, if not entirely, continental. Thick sequences of Upper Cretaceous and Paleocene marine and continental sedimentary rocks occupy large areas just outside the quadrangle but do not crop out within it.

The outcrops of pre-Tertiary sedimentary rocks are largely confined to three widely separated areas in the northeastern, west-central, and southwestern sectors (fig. 2). A major thrust fault system (see pl. 3) cuts across the quadrangle from northeast to southwest, splits the northeast pre-Tertiary mass, and skirts the south edge of the west-central mass. Thus there are two areas of upper-plate rocks, northeast and west-central, and two areas of lower-plate rocks, northeast and southwest. For convenient reference, these masses are designated as follows: Milligan Creek area, west-central upper plate; Mud Spring area, northeast upper plate; Hossfeldt Hills area, northeast lower plate; Willow Creek area, southwest lower plate.

The upper or northwest plate has moved unknown distances to the south and east relative to the lower or southeast plate. The pre-Tertiary formations in the upper plate differ, in some instances strikingly, from their representatives in the lower plate. Ultimately some of these differences may help determine the amount of fault displacement and the character of the original depositional sites. No conclusions as to these matters are reached in this paper, but with structural implications in mind the formations are discussed not in generalities but with emphasis on the similarities and differences within units from place to place.

The pre-Cretaceous marine rocks contain many fossils and the Tertiary beds have a few. With this aid it has been possible to date practically all the pre-Quaternary sedimentary rocks rather closely. Fossil-collecting localities of special interest are shown on plate 2 where each is identified by a brief field number; in the text, a field number may be followed by a formal Geological Survey collection number, in brackets, if one has been assigned.

For simplicity, the thicknesses of formations and their subdivisions are rounded off to the nearest 50 feet, wherever practicable. Measured sections with finer subdivisions and closer measurements are on pages 124–135; their locations are shown on plate 2.

PRECAMBRIAN—BELT SERIES

Unmetamorphosed Precambrian sedimentary rocks, mapped as a single unit, the North Boulder formation, underlie 3 square miles of the west-central part of

TABLE 1.—*Generalized stratigraphic succession, Three Forks quadrangle, Montana*

Era	System	Series	Group	Formation	Thickness, in feet	Description
Cenozoic	Quaternary	Age relations complex; many disconformities.		Young alluvium	0-50(?)	Rounded gravel and finer deposits on present flood plains of master streams; subangular gravel along tributaries.
				Fan gravel	0-50(?)	Fan-shaped sheets of colluvial unsorted angular limestone gravel on benches, and wedges of alluvial subangular gravel in tributary valleys.
				Silt and fine sand	0-15	Unstratified angular silt and fine sand on benches. Eolian.
				Old alluvium Rounded gravel member Subangular gravel member	0-400 0-50 0-150	Deposits of the ancestral Jefferson and Madison Rivers. Rounded gravel on former flood plains of master streams; subangular gravel in tributary valleys.
						Disconformity
	Tertiary	Miocene and Pliocene	Bozeman	Gravel	(?)	Stream gravel (known from one locality only)
		Miocene		Gravel	(?)	Stream gravel (known from one locality only)
				Gravel?	(?)	Deposit not seen in this study. Inferred from report of articulated remains of *Diceratherium armatum* Marsh from one locality.
						Disconformity
		Oligocene		Dunbar Creek formation	250->800	Light-colored tuffaceous siltstone, partly lacustrine, partly eolian, and subordinate stream-channel sandstone and conglomerate.
				Climbing Arrow formation	750-1,000	Olive bentonitic clay and coarse yellow sand. Alluvial.
		Eocene		Milligan Creek formation	0-300	Light-colored tuffaceous lake limestone and interfingering stream-channel sandstone and conglomerate.
				Sphinx conglomerate	0-100	Pebbles and cobbles of Paleozoic limestone in a reddish-orange calcareous matrix. Alluvial.
						Angular unconformity
Mesozoic	Cretaceous	Upper Cretaceous		Elkhorn Mountains volcanics	>1,000(?)	Andesite lava flows and volcanic breccia.
						Angular unconformity(?)
		Lower Cretaceous		Kootenai formation	650(?)	Orange sandstone with subordinate "salt-and-pepper" sandstone, and varicolored mudstone, siltstone, and limestone. Continental.
	Jurassic	Upper Jurassic		Morrison formation	100-250	Red, yellow, and brown mudstone, siltstone, and sandstone. Continental.
		Middle & Upper Jurassic		Ellis formation	75-350	Varied lithology. Basal brown pebble conglomerate common. Other common types are brown sandstone, gray limestone, and yellow siltstone. Marine.
						Disconformity
Paleozoic	Permian	Lower Permian		Phosphoria formation	75-200	Varied lithology. Common types are brown chert, dark phosphatic sandstone, yellow siltstone, light-colored quartzitic sandstone, and dolomitic limestone. Marine.
	Carboniferous	Mississippian and Pennsylvanian		Quadrant formation	200-400	Light-colored quartzitic sandstone and subordinate dolomitic limestone. Marine.
				Amsden formation	250-600	Red sandstone and mudstone grading upward into pink and gray limestone and quartzitic sandstone. Marine.
						Disconformity
				Big Snowy formation	0-400	Varied lithology. Common types are brown and gray siltstone, gray limestone, and dark shale. Marine.
						Disconformity
		Lower Mississippian	Madison	Mission Canyon limestone	600-1,400	Gray thick-bedded limestone. Forms cliffs. Marine.
				Lodgepole limestone	350-650	Dark-gray thin-bedded limestone. Basal beds form ledge. Marine.
	Devonian	Upper Devonian		Three Forks shale	100-400	Upper orange siltstone; medial green shale; lower orange limestone and siltstone. All marine.
				Jefferson dolomite	350-650	Brown fetid dolomite with subordinate black and gray dolomite and gray limestone. Forms cliffs. Marine.
				Maywood formation	0-200	Yellow, orange, and red calcareous siltstone, grading upward into gray limestone. (Some basal siltstone may be uppermost Cambrian.) Marine.
						Disconformity
	Cambrian	Upper Cambrian		Pilgrim limestone	350-450	Gray and orange dolomite, subordinate gray limestone. Forms cliffs. Marine.
		Middle Cambrian		Park shale	100-200	Greenish gray shale. Marine.
				Meagher limestone	300-500	Gray and black limestone with orange mottling. Forms cliffs. Marine.
				Wolsey shale	200-400	Olive and brown micaceous shale. Many interbeds of glauconitic sandstone near base and limestone near top. Marine.
				Flathead sandstone	0-100	Pink quartzitic sandstone. Forms low ledge. Marine.
						Disconformity
Precambrian	Upper	Belt		North Boulder formation	>4,000	Brown coarse micaceous arkose and conglomerate. Marine.
						Angular unconformity
	Lower			Gneiss and amphibolite	Unknown but great	

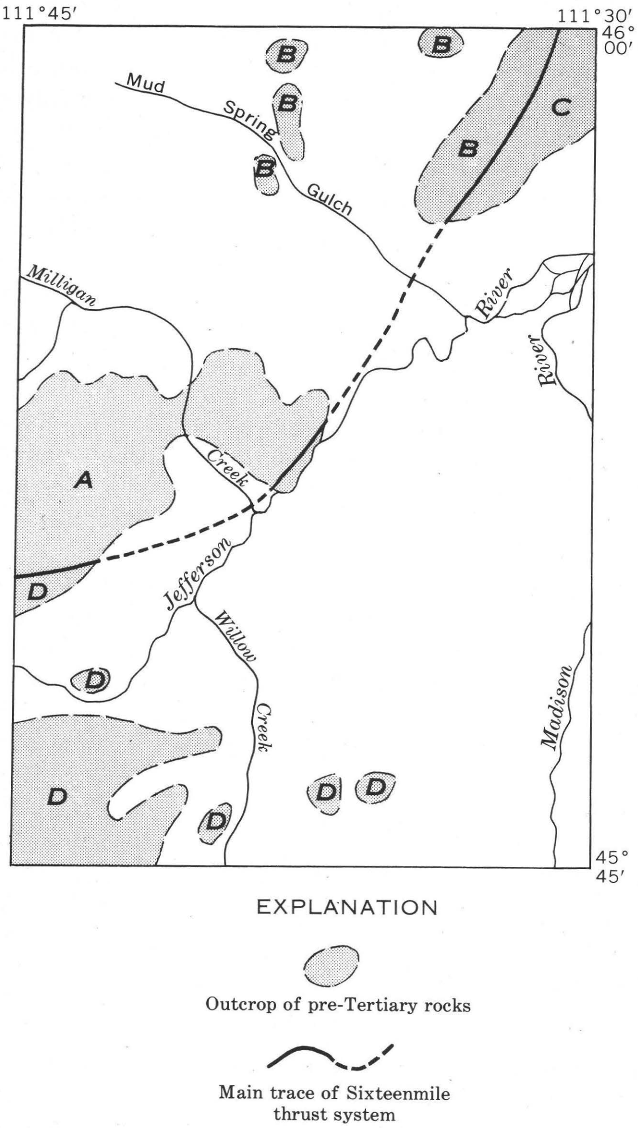

FIGURE 2.—Key map for identifying areas of pre-Tertiary sedimentary rocks. *A*, Milligran Creek area; *B*, Mud Spring area; *C*, Hossfeldt Hills area; *D*, Willow Creek area.

the quadrangle, and are more than 4,000 feet thick. Made up of 75 percent sandstone, 20 percent siltstone, and 5 percent conglomerate, they are nearshore marine deposits of geosynclinal or unstable-shelf type, and are assigned to the Belt series though thoroughly untypical of regional Belt lithology.

NORTH BOULDER FORMATION

Rocks assigned to the North Boulder formation crop out mostly at the west edge of the quadrangle north of Highway 10S, where they form a continuous section dipping northward at angles ranging from 5° to 40°, averaging 30°. They are also exposed in a narrow strip north of the highway in center sec. 5, T. 1 N., R. 1 E.

The North Boulder rocks yield a thin rubbly dark-reddish-brown soil which supports a scrubby growth of conifers and sagebrush through which many low outcrops protrude. The rough and somber landscape thus developed characterizes the formation.

Rocks believed to be Belt were penetrated in at least one of a series of borings put down recently by the Montana Power Co. in the Madison River flood plain and the bordering terraces southeast of Three Forks. A boring in NW¼SW¼ sec. 19, T. 1 N., R. 2 E., started at 4,390 feet, passed through 1,120 feet of Cenozoic continental deposits, entered thin-bedded greenish-gray calcareous arkosic sandstone and chloritic argillite at altitude 3,270 feet, and bottomed in these beds at altitude 3,191 feet. Such rocks are common in the Belt formations of the Horseshoe Hills nearby to the north,[1] but are unknown in other terranes of the region. I am indebted to Montana Power Co. for copies of well logs and for a large core sample of the rocks similar to those of the Belt series.

The North Boulder formation has an exposed thickness of more than 4,000 feet. The local base is the Jefferson Canyon thrust fault. The top is a disconformity, overlain by the Middle Cambrian Flathead sandstone or, where the Flathead is missing, by the Wolsey shale. Consequently, the full thickness of the North Boulder is not present.

The exposed strata are mostly thick-bedded dark-yellowish-brown and reddish-brown coarse micaceous feldspathic sandstone. The average grain size, which approaches very coarse near the local base, decreases upsection and downdip (that is, northward) to medium near the top of the brown sandstone sequence. Many layers show graded bedding, and some others contain numerous flat pebbles parallel to bedding. Poor sorting and poor rounding are characteristic. The sandstone is mineralogically complex, averaging about 40 percent glassy quartz, 30 percent white or colorless plagioclase in the range oligoclase-andesine, 15 percent orange-pink microcline, 5 percent mica, and 10 percent matrix, which is largely greenish-gray chlorite with varying proportions of limonite-hematite, clay minerals, and, locally, carbonate. Most of the quartz grains are strained. The plagioclase is twinned but rarely zoned. The colorless plagioclase grains are unaltered; the white ones are mostly argillized on the surface and along openings but some are more severely altered to saussurite and even to fine-grained mixtures of epidote-group minerals and calcite. Microcline is commonly unaltered and clear, indicating that the pink color is a body color and is not due merely to super-

[1] Verrall, Peter, 1955, Geology of the Horseshoe Hills area, Montana: Princeton Univ., Ph.D. thesis (available on microfilm).

ficial alteration or coating; furthermore, microcline ground to powder retains the color. As with most colored feldspars the source of the color is unknown; if it is due to hematite dust, a reasonable possibility, the dust particles must be submicroscopic for none are visible under high magnification ($\times$ 344). The mica is mainly biotite, variably bleached and chloritized, but includes a little muscovite. Sparse accessory minerals are zircon, apatite, sphene, and opaque oxides. By most definitions this rock is an arkose.

Thin laminae of micaceous olive-gray siltstone separate many sandstone beds, especially in the upper part of the sequence. The siltstone laminae have about the same minerals as the sandstone but in different proportions: micas, chlorite, and clay minerals are dominant, quartz subordinate, and feldspar rare.

Lenses, a few feet thick, of subrounded cobble conglomerate are abundant near the base. The cobbles are mostly vein quartz with lesser proportions of quartzite, granite pegmatite, diorite, gneiss, amphibolite, and diabase.

A distinctive green-flecked reddish-brown to grayish-red coarse sandstone, 100–250 feet thick, is at the top of the exposed section. Good outcrops are in N½ sec. 9, T. 1. N., R. 1 W., and along Highway 10S at the southwest end of the strip of North Boulder formation in sec. 5, which consists entirely of this rock. Its general characteristics and mineralogy are much like those of the underlying brown sandstone but the red rock has less mica, more chlorite, and more clay-sized material, perhaps as much as 25 percent in places, and the matrix is much richer in limonite-hematite. Its plagioclase is thoroughly argillized, though the microcline is usually about as fresh as that in the brown rocks. Along the highway in sec. 5 the rock is especially rich in hematite and is garishly red. The rock is here classed as an arkose though it approaches graywacke.

ORIGIN

The North Boulder formation clearly was derived from the mechanical weathering of rocks like the nearby crystalline rocks, and shows every sign of short transport and rapid deposition. The textures of the coarser fractions suggest transportation by turbidity currents. The landmass from which these sediments were worn must have been mountainous, though not necessarily large. The North Boulder strata are one-cycle immature rocks of the sort common in marine geosynclines, or, as Peale (1896, p. 2) described them, they are "beds of littoral formation."

Southward coarsening of the North Boulder rocks plainly suggests derivation from that direction. The crystalline terrane immediately to the south seems an obvious source. It is not a wholly satisfactory one, however; for one thing, the crystalline rocks lack pink microcline so abundant in the North Boulder formation. The nearest known adequate source of pink microcline seems to have been the granite gneiss near Dillon 60 miles or more to the southwest (see Scholten and others, 1955, p. 351); at least, pink microcline has not been reported by the several workers in intervening areas. There also seems no nearby source for the diabasic fragments in the North Boulder (reported also from the Jefferson Canyon area by Alexander, 1955, p. 29) but this may be due merely to failure of other workers to recognize or mention small masses of such rocks, rather than to their absence. Unmetamorphosed basaltic intrusive bodies are known in Precambrian crystalline rocks near Sheridan about 40 miles to the southwest.[2]

Understanding the origin of the North Boulder formation is complicated by the fact that near Three Forks and almost everywhere else the base of the formation is a thrust fault of evidently large but indefinitely known displacement. Perhaps these rocks were deposited far from their present position. As later discussion of the Laramide thrust faults brings out, movement along the Jefferson Canyon thrust, the local base of the North Boulder rocks, was probably mostly strike slip, with the upper plate having moved relatively east. If so, the mass of North Boulder rocks may be many miles east of its depositional site, but not far south of it.

NAME

The rocks here described are at the east end of a large mass of coarse feldspar-rich clastic rocks that extend westward for 15 miles, nearly to Cardwell. Similar rocks crop out sporadically in a narrow east-west belt for 75 miles from near Melrose on the Big Hole River to the west front of the Bridger Range. These rocks are unique among pre-Flathead strata, but Peale (1896, p. 2) assigned them to the Belt formation. Later, the Belt having been raised to series rank, these rocks were named North Boulder group by Ross (1949), who reported their thickness as more than 6,000 feet at their type locality. Because these rocks have not been subdivided in this area or nearby, the name North Boulder formation seems preferable in this area.

Alexander (1955, p. 17–36) studied these rocks in some detail in the type area, and called them the LaHood formation after LaHood Park, a small settle-

[2] Levandowski, D. W., 1956, Geology and mineral deposits of Sheridan–Alder area, Madison County, Montana: Univ. Michigan, Ph.D. thesis (available on microfilm).

ment near the mouth of the Boulder River in the Jefferson Island quadrangle.

AGE AND REGIONAL RELATIONS

On local evidence the North Boulder formation can be dated only as older than the Flathead and Wolsey formations, which overlie it disconformably, and much younger than the Precambrian metamorphic rocks to the south. As the Flathead grades into the Wolsey and the Wolsey has Middle Cambrian fossils, the North Boulder formation is Middle Cambrian or older. The fact that the Flathead and younger Paleozoic formations overlie thousands of feet of North Boulder rocks north of Highway 10S, but lie directly on pre-North Boulder crystalline rocks a few miles to the south, without extreme differences in facies or thickness, indicates an important erosional, if not diastrophic, interval between the North Boulder and the Flathead. This strongly suggests that the North Boulder is Precambrian.

The regional relations of the North Boulder rocks emphasize their probable Precambrian age. North of Cardwell, the coarse clastic rocks are reported to intertongue northward with the Greyson shale, high in the Belt series (Alexander, 1955, p. 14) ; about 17 miles east of Three Forks, in the Horseshoe Hills, the arkosic material, according to H. D. Klemme[3] intertongues laterally and grades vertically with the Spokane shale which overlies the Greyson; according to Verrall[4] (1955) the intertonguing in the southern part of the Horseshoe Hills is with the Newland limestone and Chamberlain shale, which successively underlie the Greyson, but are still high in the Belt series. These interpretations are not necessarily in conflict; they may all be correct, leading to the conclusion that rocks of North Boulder aspect were being deposited near shore, no doubt intermittently, over a long time interval during which perhaps as much as 10,000 feet of finer grained sediments were accumulating farther from shore. It should, however, be noted that Klemme's reconnaissance interpretation was not confirmed by my detailed mapping in the Toston quadrangle, which indicates that coarse arkosic sandstone similar to that of the North Boulder formation is interbedded with rocks low in the Greyson shale and high in the Newland limestone rather than with the Spokane shale. At present, then, there is no unequivocal evidence that rocks of North Boulder facies were being deposited after Greyson time. The several thick Belt formations with which the North Boulder rocks are intimately related, and other Belt formations, are truncated regionally by the Flathead, in places with marked angular unconformity (Deiss, 1935). It seems safe to conclude that the North Boulder formation is to be correlated with the upper part of the Belt series, and is of late Precambrian age.

The outcrops of Belt rocks on the north side of the Three Forks basin, and the near-certain presence of Belt rocks directly below Tertiary rocks near the south edge of the basin suggest that Belt rocks may be widespread beneath the Cenozoic cover. Precambrian rocks, partly crystalline and partly Belt, bound the Gallatin Valley on the north and south and dominate the western flanks of the Bridger Range, which forms the east boundary of the Three Forks basin. As first noted by Peale (1896, p. 3), the probability is strong that Precambrian, largely Belt, rocks are the principal ones to be expected below the Tertiary deposits throughout the Three Forks basin.

The paleogeologic significance of the North Boulder formation and the sharp southern edge of Belt deposits which the formation defines have been widely discussed. Agreement is general that an ancient east-west zone of tectonic activity determined the distribution and affected the lithologic nature of the North Boulder and other Belt rocks of the region; some plainly label it a fault zone, others are less explicit. It is agreed that the source of Belt sediments lay to the south, but opinions diverge widely on such matters as the size, relief, and tectonic habit of this landmass, whether it was ever covered by Belt sediments, and the degree and significance of the unconformity between Belt and Flathead. Recognizing that the Paleozoic formations were deposited over both Belt and pre-Belt crystalline rocks, most authors regard this east-west lineament as structurally unimportant in post-Belt time, but Sloss (1950, p. 430) believed that it "was an important tectonic boundary in Precambrian and Paleozoic time and remains active today," and McMannis (1955, p. 1425) found evidence of important fault movements along it in the Bridger Range "during Laramide deformation." In this report the southern boundary is interpreted as a steep fault zone, and named the Willow Creek fault.

The sole contribution of this study to the confused regional Belt problem is to call attention to a further element of confusion; namely, the possibility that the North Boulder sequence may be largely allochthonous. This is not merely a local possibility. Everywhere west of Three Forks, the base of the North Boulder, where not veiled by Cenozoic deposits, seems to be a northward-dipping thrust fault, with the probable exception of a small area 6 miles southeast of Whitehall where Lowell (1956b, p. 71) found conglomerate

[3] Klemme, H. D., 1949, Geology of Sixteenmile Creek area, Montana: Princeton Univ., Ph.D. thesis (available on microfilm).

[4] See footnote, p. 11.

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of the Belt series lying on rugged topography cut in Precambrian crystalline rocks. The mass east of Divide, shown on the "Geologic Map of Montana" (Ross and others, 1955) as in depositional contact with crystalline rocks, is actually in thrust contact (W. B. Myers, oral communication, 1955). East of Three Forks, the only North Boulder rocks not basally masked by Cenozoic detritus are on the west side of the Bridger Range, where they are in fault (though not necessarily thrust) contact with Precambrian metamorphic rocks (McMannis, 1955, pl. 1, fig. 5, p. 1421). In part, the thrusting seems to represent a reversal of movement on the Willow Creek fault. If these rocks have moved far on a thrust system, they may tell little about conditions in the apparent source area directly to the south. Other aspects of these relations are discussed under the heading "Structure."

PALEOZOIC

The Paleozoic rocks, subdivided into 13 map units, are 4,000–5,000 feet thick. The Paleozoic section is made up of about 60 percent limestone, 10 percent dolomitic limestone and dolomite, 20 percent siltstone and shale, and 10 percent sandstone; the clastic ratio is about 0.4. Typically, very thick sequences of carbonate rocks, mainly chemical but partly clastic, alternate with thinner sequences of carbonate-poor clastic rocks. The earliest Paleozoic rocks are of Middle Cambrian age. A substantial part of the Paleozoic—from very late Cambrian through Ordovician and Silurian to Middle or Late Devonian time—is represented by an erosion surface with little relief. Above this surface the rest of the Paleozoic is represented with only minor breaks. No angular unconformities have been detected. Viewed in the large, the assemblage seems to have been formed under conditions of deposition, and nondeposition, that remained stable for very long periods. In the near view, however, the impression of stability disappears. Within a few miles nearly every formation of Paleozoic age, carbonate as well as noncarbonate, exhibits striking changes in thickness and lithology, only part of which can be laid to later deformation. It would be misleading, therefore, to call them stable-shelf deposits, if this implies strata of monotonous uniformity over large areas.

The clastic components of several of the pre-Carboniferous Paleozoic formations seem to have been derived from land to the west, presumably not far from the present Idaho-Montana border. This landmass was recognized long ago by Willis (1907) and Walcott (1915), and its existence has been confirmed by the more recent work of Ross (1934, 1935), Deiss (1941), and others. The evidence has, however, been overlooked or otherwise interpreted by contemporary paleogeographers, who have visualized geosynclinal conditions in the probable source areas through much if not all pre-Carboniferous Paleozoic time (Rocky Mountain trough of Eardley, 1947; Millard belt of Kay, 1951, p. 7–14; Cordilleran geosyncline of Sloss, 1950).

Notwithstanding some difference of opinion regarding conditions near the Idaho-Montana border, the above-cited paper by Sloss and its companion piece (Sloss and Moritz, 1951) offer broad views of the regional stratigraphy and depositional environments during the Paleozoic era that are invaluable in orienting the small segment that is the Three Forks quadrangle.

The Paleozoic rocks are much deformed. In most places their dips are moderate to steep (20°–60°), but over large areas they are vertical or overturned. Low dips characterize only the Paleozoic rocks at the east side of the Milligan Creek area, and in the axial parts of folds. Strike faults are many, but only the largest can be traced. Dip faults are rare, appearing only in the Milligan Creek sector.

CAMBRIAN

The Cambrian rocks, separated into 5 formations, are 1,200–1,600 feet thick. They are about 35 percent limestone, 25 percent dolomitic limestone and dolomite, 30 percent siltstone and shale, and 10 percent sandstone; the clastic ratio is about 0.7.

Not all of Cambrian time is represented. Lower Cambrian rocks have not been recognized. Four of the mapped Cambrian formations are of known or probable Middle Cambrian age: Flathead sandstone, Wolsey shale, Meagher limestone, and Park shale. Only the fifth, the Pilgrim limestone, is of Late, but not latest, Cambrian age. Some unfossiliferous strata above the Pilgrim may be of very late Cambrian age, but are here included with similar strata of Devonian age.

Comprehensive discussions of Cambrian stratigraphy in southwestern Montana have been presented by Deiss (1936) and Hanson (1952). Cambrian paleoecology in the region has been treated by Lochman (1957).

NAMES AND CORRELATION OF CAMBRIAN FORMATIONS

Peale (1893, p. 20–25) subdivided the Cambrian system of the Three Forks region into two formations, the Flathead and the Gallatin. He further subdivided the Flathead into two members and the Gallatin into five members, which he did not map separately. His Cambrian section is shown on the next page.

			Thickness (feet)
Devonian: Jefferson limestone.			
Cambrian	Gallatin formation	Pebbly limestones	145
		Dry Creek shales	30
		Mottled limestones	260
		Obolella shales	280
		Trilobite limestone	120
	Flathead formation	Flathead shales	290
		Flathead quartzite	125
		Total Cambrian	1, 250

All of Peale's members are readily recognized in the Three Forks quadrangle although some of his measurements of thickness cannot be confirmed; nevertheless, his names are not used. In regional usage, most of his names have been avoided or drastically modified, because they are not carefully defined, or not etymologically acceptable. Accordingly, to conform with current usage, terms are brought in from neighboring areas. Names of the five definitely Cambrian formations are those first applied by Weed (1899a, 1899b) in the Little Belt Mountains, 80 miles northeast of Three Forks, and redefined by Deiss (1936).

These formations seem safely correlated with Peale's five lower members, despite many differences in thickness and lithologic detail. Perhaps the most extreme disparity is between the Pilgrim of the Little Belt Mountains and equivalent rocks (mottled limestones of Peale) at Three Forks. The Pilgrim as originally defined by Weed is largely interbedded flat-pebble limestone conglomerate and gray-green shale, less than 100 feet thick, hardly comparable with the massive dolomite and the banded limestone-and-dolomite sequences, around 400 feet thick, assigned to the Pilgrim in this report. But the rocks fit fairly closely the redefined Pilgrim of Deiss (1936, p. 1280–1281), though they largely lack the intraformational conglomerates emphasized by him (p. 1334) as characteristic of the formation, and are mostly dolomite rather than limestone. If the name Pilgrim is not used for these rocks, the other available choices are Hasmark, from the Philipsburg area (Emmons and Calkins, 1913), 80 miles northwest of Three Forks, or Maurice, from the Yellowstone Park area (Dorf and Lochman, 1940), 100 miles to the southeast. As the Hasmark of Emmons and Calkins probably includes beds elsewhere assigned to the Meagher and Park formations (Hanson, 1952, p. 15 and 17) it is not a desirable term, notwithstanding its acceptance for regional use by Sloss and Moritz (1951, p. 2144–2145). Use of Maurice presents some practical problems, too (see Hanson, 1952, p. 16; Sloss and Moritz, 1951, p. 2144–2145), in addition to the obvious problem of priority. Consequently, the name Pilgrim is used in this paper.

Only Late(?) Devonian rocks of the Maywood formation were definitely identified in the interval in which Peale mapped his two uppermost Cambrian units, the Dry Creek shales and Pebbly limestones. Equivalents of part of the Dry Creek may be present, but cannot be distinguished from the Maywood and therefore are mapped and discussed with it.

FLATHEAD SANDSTONE

The lowest Cambrian rocks are nearly pure quartzitic sandstone assigned to the Flathead sandstone. The Flathead is exposed only in the Milligan Creek area where it overlies the North Boulder formation, and in the Willow Creek area where it rests on Precambrian crystalline rocks. In the Willow Creek area, the Flathead crops out almost continuously in a low ridge for more than 6 miles. Its outcrop width varies greatly: where it is overlain by Wolsey shale, in the west half of the strip, it forms a broad dip slope; where plutonic rocks have taken the place of the Wolsey, in the east half, the dip slope is narrow. Near the Madison-Gallatin County line, in secs. 9 and 10, T. 1 S., R. 1 W., the Flathead outcrop swells to a width of half a mile.

In the Milligan Creek area, the Flathead is largely missing, apparently through nondeposition rather than pre-Wolsey erosion. A small lens of Flathead sandstone is well exposed in a creek bed at the east edge of sec. 10, T. 1 N., R 1. W. The Flathead also appears, along the strike of this lens, at the west edge of the quadrangle, and continues for several miles to the west in the Jefferson Island quadrangle. Where the Flathead is missing, a persistent low ridge, stratigraphically below the Wolsey and thus seemingly representing the Flathead, is actually formed by the upper red arkose of the North Boulder formation.

In most of the Willow Creek area, the Flathead is 30–50 feet thick. Its broad outcrop mostly reflects its tendency to form a dip slope. The very broad outcrop near the county line, however, can only in part be ascribed to parallelism between bedding dip and surface slope. The apparent thickening may be due to deformation, suggested by abrupt strike changes, but it is possible that the formation actually thickens locally to several hundred feet. The lens west of Milligan Creek is less than 50 feet thick.

The formation is of uniform lithology, consisting mainly of thick beds of medium- to coarse-grained subrounded sandstone, cemented by quartz and tinted pinkish or brownish by films of iron oxides. Locally, bedding is thin and in such places crossbedding is common. Lentils of pebble conglomerate, only a few pebbles thick, are abundant near the base and common elsewhere. Rarely, the sandstone is deeply colored in shades of red and brown, due to coating of grains by iron oxides. A few laminae of dark shale appear near the top of the formation, which is placed below the first shale sequence more than 5 feet thick, assigned to the Wolsey shale. This contact is everywhere conformable, and in some places, where sandstone and shale are thinly interbedded for a few feet above thick-bedded Flathead sandstone, is gradational. Where shales do not crop out, the contact is arbitrarily placed at the base of the dip slope on the stratigraphically highest ledge of Flathead sandstone. Two measured sections of the Flathead are given on pages 129 and 135.

Quartz makes up more than 98 percent of the rock, even where iron oxides seem abundant, and approaches 100 percent in many places. The quartz cement is in clear overgrowths, optically continuous with the clastic grains; typically, it only partly fills the pores. Other constituents are wisps of clay and a few rounded grains of zircon and tourmaline. The rock, while firmly cemented, generally breaks around the grains.

The Flathead and Wolsey formations are closely related and their origin, age, and regional relations are discussed jointly in the next section.

WOLSEY SHALE

Rocks assigned to the Wolsey shale lie on the Flathead sandstone in the Willow Creek and Milligan Creek areas, and on the Belt rocks where the Flathead is missing. Outcrops are rare and the trace of the formation is usually marked by a grassy depression. A few good exposures are in steep gulches in SW1/4NW1/4 sec. 11, and SE1/4NE1/4 sec. 10, T. 1 N., R. 1 W., west of Milligan Creek; near the highway in the SW¼ sec. 5, T. 1 N., R. 1 E., east of Milligan Creek; and just west of the quadrangle, along the main northwestward-draining creek in the NW¼ sec. 8, T. 1 S., R. 1 W. (Jefferson Island quadrangle).

In this study, the base of the Wolsey is placed at the first shale beds more than 5 feet thick overlying the massive Flathead sandstone, or overlying the upper red Precambrian arkose where the Flathead is missing. In the Willow Creek area, sandstone bodies thick and continuous enough to be mapped as members appear low in the Wolsey, underlain by perhaps as little as 20 feet of shale. These sandstone beds generally contain a little glauconite; otherwise, they closely resemble the Flathead. In places the thickest of these sandstone bodies, near the base of the formation, is thicker and makes a more impressive outcrop than nearby sandstone mapped as Flathead, and the temptation is strong to include such sandstones and the shale beneath them in the Flathead, as Alexander (1955, p. 39) seems to have done. To do so, however, would lead to unilluminating complexities in the Wolsey-Flathead contact, which would jump up and down in the section to enclose each thick sandstone lens. This problem does not appear in the Milligan Creek area, where the rocks mapped as Wolsey contain no prominent sandstone similar to that of the Flathead.

In the Milligan Creek area the formation ranges in thickness from 200 to 400 feet; it is thinner where the Flathead is absent. In the Willow Creek area, the formation seems to be consistently around 400 feet thick, but measurements are inexact because the formation is riddled with sills. The Wolsey section, typical of both areas, consists of:

	Thickness (feet)
Thinly interbedded olive-gray limestone and subordinate olive shale	100
Olive micaceous shale with thin interbeds of olive-gray limestone	200
Grayish-green and yellowish-brown micaceous shale, with a few beds, some very thick, of reddish glauconitic sandstone	100

Where the formation is thinner, the lower part seems to be absent.

The micaceous material in the shale is largely chlorite, but includes considerable biotite, usually showing signs of marginal alteration to chlorite. Though glauconite is abundant only in the sandstone low in the formation, it is a common minor constituent in sandstone and limestone throughout.

In the Willow Creek area the lower part of the Wolsey is slaty and dark reddish brown, due to high local content of iron oxides. This part of the Wolsey contains many sills. The cleavage and reddening are probably intrusion effects, although the color change may reflect local initially high content of oxidizable iron, as in the red and brown phases of the Flathead.

Bedding planes between many of the micaceous shale beds that make up the bulk of the formation are strewn with flattened cylinders of shale or fine sand, ¼–½ inch wide and 1–3 in. long, that taper slightly at the ends; these forms, widely distributed in early Paleozoic shales, for example, the Cambrian Ophir shale of northern Utah (Calkins and Butler, 1943, p. 12–14 and pl. 9), are generally interpreted as worm trails or casts. While the shale beds themselves rarely crop out, their presence is made plain by soil glittering with flakes of golden-brown mica, and by float fragments bearing worm casts.

The upper quarter of the formation, in which shale becomes subordinate to limestone, may well be thought of as transitional into the overlying Meagher limestone, and on strictly lithologic grounds the beds should no doubt be assigned to the Meagher, as some workers, such as Sloss and Moritz (1951, p. 2142) and McMannis (1955, p. 1393), have done. This partly accounts for the discrepancy between the thickness of the Wolsey as reported by these workers, 125–200 feet, and the much greater thickness reported here. Exposures of the transitional beds are rare; at most places the entire Wolsey is covered and the first outcrop upsection from the Flathead sandstone is thick-bedded limestone from near the base of the Meagher. In mapping, therefore, it is practical to place the contact between the Meagher and the Wolsey slightly below the base of this limestone ledge and arbitrarily to include the transitional beds with the Wolsey, as Klepper and others (1957, p. 8) have done.

For detailed measured sections of the Wolsey shale, see pages 129 and 135.

AGE AND ORIGIN OF FLATHEAD AND WOLSEY FORMATIONS

Many fragments of a species of *Ehmania*, a characteristic Middle Cambrian trilobite (A. R. Palmer, written communication, June 30, 1955) have been recovered from a limestone bed about 200 feet below the top of the Wolsey in the Milligan Creek area (loc. 2, near S½NW¼ sec. 11, T. 1 N., R. 1 W., pl. 2). Nodular shaly limestone beds 30 feet below the top of the Wolsey, in the Jefferson Island quadrangle (loc. 449 [2145–EO], south side of 4698 hill, NE¼NW¼ sec. 8, T. 1 S., R. 1 W., pl. 2) about 1,000 feet west of the Three Forks quadrangle, yielded a definitive middle Middle Cambrian fauna (A. R. Palmer, written communication, October 22, 1956) including a "distinctive and apparently undescribed phosphatic brachiopod resembling *Pegmatreta*" and a trilobite assemblage, listed below, with forms closely related to those reported by Deiss from the Pagoda limestone in northwest Montana:

Tonkinella cf. *T. stephenensis* Kobayashi
Kootenia sp. (with 10 long marginal spines)
Bolaspis sp. indet.
Glyphaspis robusta Deiss

These are the only local age data for both Flathead and Wolsey. Because sedimentation was apparently continuous from the base of the Flathead to the fossiliferous beds, it is a reasonable inference that both Flathead and Wolsey are Middle Cambrian, though an Early Cambrian age is not ruled out for the lower part of the sequence. Fossil evidence from other nearby areas (Hanson, 1952, p. 13–14; Alexander, 1955, p. 40) supports a Middle Cambrian age for the entire Wolsey shale and strengthens the case for assigning the Flathead sandstone to the Middle Cambrian.

The Flathead sandstone represents slow but steady marine invasion across a subaerial erosion surface. This surface represents the removal of very great thicknesses of Belt rocks regionally, as several workers have shown (for example, see Clapp and Deiss, 1931). The source area had considerable relief judging from the general coarseness of grain and the widespread occurrence of conglomerate beds.

In the region around Three Forks the Flathead must have been derived either as first-cycle deposits from erosion of early Precambrian gneiss and schist, or as second-cycle deposits from reworking of Belt sedimentary rocks, or both. In any event, the winnowing process that yielded pure quartz sand near the shore must have concentrated the waste from ferromagnesian and feldspathic components of the source rocks further offshore, as clays or muds. The lower part of the Wolsey shale is, then, probably the offshore time-equivalent of the Flathead sandstone. The local concentration of red hematitic pigment in both Flathead and Wolsey probably represents stripped regolith.

A local topographic high, or island, in the Flathead and early Wolsey sea is suggested by the virtual absence of Flathead and thinning of Wolsey rocks west of Milligan Creek; this island was earlier recognized by Hanson (1952, p. 13) and by Alexander (1955, p. 36). Similar thickness variations in the Flathead are also reported in the Philipsburg area (Emmons and Calkins, 1913, p. 51–52) and in the Gravelly Range (Mann, 1954, p. 6); very likely they are typical of the Flathead throughout western Montana.

Shoreline oscillations are indicated by the reappearance of sandstones similar to the Flathead in the lower Wolsey, but dominant marine transgression and retreat of the shoreline during this episode is clearly shown by the virtual disappearance of sandstone from the upper Wolsey and concomitant increase in the number and thickness of limestone beds. Local evidence of the direction of transgression is lacking, but recent regional studies suggest that it was from west to east (Lochman, 1957, figs. 1–3; Hanson, 1952, p. 13), and not east to west as proposed earlier by Emmons and Calkins (1913, p. 64).

MEAGHER LIMESTONE

The Meagher limestone is the lowest of the great carbonate formations that dominate the Paleozoic column, not only in the Three Forks region but throughout western Montana. The Meagher is exposed in the Milligan Creek and Willow Creek areas, but is absent elsewhere in the quadrangle except for a small thrust wedge in the northeast corner (sec. 19, T. 3 N., R. 2 E.). It forms most of the cliffs skirted by Highway 10S in the center of the quadrangle and the most southerly row of cliffs west of Milligan Creek.

Near Milligan Creek the Meagher limestone is around 500 feet thick and typically consists of·

	Thickness (feet)
Thick-bedded gray limestone; forms ledge	50
Thin-bedded gray limestone, with mottles and bands of orange silty limestone, poorly exposed	250
Thin-bedded orange calcareous siltstone and gray dolomite, generally covered	50
Thick-bedded nearly black limestone with crinkled laminations, and abundant grains of feldspar and quartz sand; forms ledge	100
Thin-bedded gray limestone with yellow shaly partings, generally covered	50

Brecciated, locally cavernous, zones, marked at the surface by dark iron-stained outcrops and slight depressions, appear high in the Meagher at the east side of Milligan Creek sector (secs. 32 and 33, T. 2 N., R. 1 E.). These follow bedding, and are probably due to post-depositional ground-water solution.

In the Willow Creek area the gross thickness and lithologic character are similar, but the typical section is notably different, except for the basal limestone and shale unit:

	Thickness (feet)
Thin-bedded gray limestone with many interbeds of orange silty limestone, poorly exposed	200
Thick-bedded gray limestone; forms ledge	150
Thin-bedded limestone, mottled in gray and orange, poorly exposed	100
Thin-bedded nodular gray and orange mottled limestone and shale	50

Particularly noticeable in the Willow Creek area is the absence of the great black limestone ledge that persists near the base of the section in the Milligan Creek area.

Perhaps the most distinctive characteristic of the Meagher, aside from its stratigraphic position as the lowest Paleozoic carbonate formation, is the mottling of gray microcrystalline limestone with coarser grained, more porous, orange or gray limestone. Rock of this sort was once quarried near Radersburg for decorative stone under the trade name "black and gold marble" (Hanson, 1952, p. 14 and photograph, pl. 3H), and geologists in the region usually refer to this facies informally as mottled blue-and-gold, mottled black-and-gold, or mottled black-and-tan. The orange (equals gold or tan) mottles, which rarely comprise more than 30 percent of any sequence of beds, range from partings in the bedding, to irregularly ramifying networks only crudely related to bedding, to scattered isolated blobs. Individual orange zones are rarely more than half an inch thick. The orange color seems due to films of ferruginous clay. Contacts between the black and the gold phases are sharp; typically, they are not flat but undulating in smooth waves a fraction of an inch in length and amplitude.

Several other features are characteristic of the thin-bedded parts of the Meagher. Thin lenses of yellowish calcite sandstone, locally cross-stratified and laden with fossil fragments, are especially common. Abundant, too, are thin to thick beds of conglomerate made up of flat, angular to subrounded limestone pebbles in a matrix of calcite sand; most of the pebbles lie approximately in the plane of bedding but many are steeply inclined, though without obvious preferred orientation. Rarely, thin zones of edgewise or shingled flat-pebble conglomerate appear near the top of the formation.

Both thin-bedded and thick-bedded parts of the formation have several features in common. Many beds contain abundant rounded grains of glauconite. Oolitic zones are common, though rarely more than a few feet thick or traceable more than 100 feet laterally; the oolites, of slightly darker limestone than their matrix, are of sand size, rarely as much as 1/15 inch long. Much rarer are pisolitic zones, of similar dimensions, but with ovoids ranging from ⅕ to ½ inch long (for photograph see Hanson, 1952, pl. 3B.). Oolites and pisolites do not occur in the same zone. Many of the darker beds have a distinct petroliferous odor when freshly broken. Characteristic, too, of many of the gray and black beds of the Meagher is a distinctly bluish tinge on weathered surfaces.

Where exposures permit, the base of the Meagher is mapped at the contact of thin-bedded limestone with interbedded limestone and shale of the uppermost Wolsey; elsewhere, it is carried about 50 feet stratigraphically below the base of the lowest limestone ledge, in order to include the basal thin-bedded limestone.

The top of the upper thick-bedded limestone makes a dip slope into a narrow grassy valley or saddle underlain by the Park shale. Exposures of the contact are rare, but it nevertheless seems safe to assume that the contact throughout is sharp but conformable, for there are no prominent transitional limestone interbeds in the basal Park, but the Park seems to lie on the same set of Meagher beds everywhere. On the map the contact is arbitrarily placed where the slope made by the upper part of the Meagher limestone begins to flatten.

For detailed measured sections of the Meagher limestone, see pages 128–129, 135.

AGE AND ORIGIN

The Middle Cambrian age of the Meagher limestone near Three Forks is well established by fossil collections from several horizons. An especially well preserved assemblage from the lowermost 10–30 feet of the formation in the Willow Creek area (locs. 435 and 450 [2144 €O], NW¼NW¼ sec. 9, E. 1 S., R. 1 W., pl. 2) contains the following forms (A. R. Palmer, written communications Oct. 7, 1954, and Oct. 22, 1956):

Ptychagnostus atavus (Linnarson)
Tomagnostus sp.
Bathyuriscus cf. *B. formosus* Deiss
Elrathina sp.
Peronopsis cf. *P. scutalis* (Hicks)
cf. Elrathiella sp.
Ptychoparioid gen. and sp. indet.
Prototreta flabellata Bell

The Meagher seems to have been deposited in shallow marine water, but for the most part far from sources of noncarbonate detritus. The widespread pebble beds, clastic zones, and oolitic zones indicate shallow-water deposition despite the thickness of the formation. Probably a large fraction of the formation is of bioclastic origin. It seems likely that the water in which the Meagher was deposited was no deeper than that in which the Wolsey shale accumulated. The change in sedimentation was perhaps due largely to the reduction of relief in the source area that had supplied the detritus in the Wolsey.

Whereas the oolites probably formed as a result of alternate chemical deposition and mechanical rolling in disturbed water, the pisolites, though superficially similar, are probably of very different origin: according to A. R. Palmer (oral communication, 1956) they are the remains of stromatolites of the genus *Girvanella*. Alexander (1955, p. 40) also recognized these forms to be *Girvanella*.

The origin of the mottling in the black-and-gold limestone, which is not confined to the Three Forks quadrangle but is regionally distributed, has not yet been systematically studied. The nearest approach is an interesting but inconclusive investigation by Hanson (1951).

In many places, particularly west of the Three Forks area, the gold mottles in the Meagher are dolomitic limestone or dolomite (Hanson, 1952, p. 14; Klepper and others, 1957, p. 9). Similar mottles and bands in thin-bedded parts of the next Cambrian carbonate unit, the Pilgrim limestone, are consistently of dolomite in the Three Forks quadrangle and elsewhere (Klepper and others, 1957, p. 11; McMannis, 1955, p. 1394–1395). The temptation is strong to regard the mottling as an early product of the dolomitization process. But the absence or paucity of dolomite in the mottled Meagher near Three Forks and in other areas, particularly to the east (see chemical analyses quoted by Hanson, 1952, p. 42) shows that the mottles are not the result of dolomitization; rather, the coarser, more porous gold-colored zones probably served as channelways in some places (but not near Three Forks) for postdepositional dolomitizing solutions. Similar channeling of late dolomitizing solutions has been often interpreted, notably for the Platteville limestone by Griffin (1942) and for certain Alpine carbonate rocks by Sander (1936) who coined the term "belteroporic" to define the process and resultant fabric.

Hanson stated (1952, p. 14) that "the mottling is apparently a primary feature." This interpretation is supported not only by the absence of dolomite but also by the presence of clay in the mottles and by the sharp contacts between the black and the gold phases.

But if primary, meaning depositional, why does the mottling cut across the bedding in so many places and why are many of its contacts wavy? An attractive explanation depends on different lithification rates for the two types of sediment involved; it might therefore be called a diagenetic hypothesis. Suppose that initially thick layers of fine-grained, largely chemical calcium carbonate mud (black) are deposited alternately with thin layers of coarse clayey, largely clastic calcium carbonate mud (gold), and that the original contacts are horizontal and flat. Long after deposition—that is, after removal from the water interface—the sediments will still be wet and plastic. The more porous gold layers could be expected to harden more slowly and, if thick enough, to act as highly viscous liquids after the black layers have begun to act as soft solids. Depending on the viscosity and elasticity relationships that develop, the more competent black layers might merely buckle and sag into the gold ones, thus forming wavy contacts; or they might break up and settle piecemeal, producing a breccia; or a flat-pebble gravel might form if the still-soft black fragments are rounded a little by streaming or by chemical reaction with the gold fluid. This process could operate in a number of layers simultaneously, and requires no external dynamic force other than gravity. It could, therefore, operate far from shore and in horizontal materials without the aid of subaqueous sliding, subaerial desiccation, waves or currents. Angular slabs of carbonate rock in a carbonate mud matrix have been assigned a similar origin and called "inhomogeneity breccia" by Sander (1936).

Intraformational conglomerate, of the sort common in the thin-bedded parts of the Meagher, has generally been interpreted as due either to subaqueous sliding or to penecontemporaneous fragmentation by waves or currents. It seems clear that the nonshingled conglomerate in the Meagher formed far from shore and on a horizontal bottom, as most workers in the same or closely similar rocks have concluded (McKee, 1945, p. 65–69; Lochman, 1957, p. 143–146). Under such circumstances, subaqueous sliding on a large scale seems out of the question. McKee by inference and Lochman explicitly favor fragmentation by waves or currents. Worth considering, however, is the possibility that some of the nonshingled intraformational flat-pebble conglomerate in the Meagher formed as a result of differing rates of lithification in thin-bedded strata, and thus might be called, to extend Sander's usage, inhomogeneity conglomerate. Specifically excluded from such an origin, however, are the rare layers of shingled or edgewise conglomerate, for their high degree of orientation athwart the bedding planes indicates development at a water-sediment interface in the presence of strong currents or waves.

The origin of the few dolomite beds in the Meagher is moot. Some of their field aspects and relations favor direct precipitation. For example, the persistent dolomite beds above the lower black limestone ledge are finely crystalline, even in texture, and without mottling, although the calcite sandstone with which they are interbedded is similar to that in the gold mottles higher in the section. Contacts between clastic calcite rock and seemingly chemical dolomite are sharp. Here the porous sandstone has evidently not acted as a passageway for dolomitizing solutions. The problem is discussed more fully later, in connection with the dominantly dolomitic Pilgrim and Jefferson formations.

Although most of the Meagher was laid down far from the sources of noncarbonate material, there must have been a nearby source for the feldspar and quartz sand common in the lower ledge of the Meagher limestone in the Milligan Creek area. The sand was doubtless derived from the same island responsible for the absence of Flathead sandstone and thinness of Wolsey shale in that area, though its shoreline was presumably farther away than in Wolsey time. The absence of sand in the Meagher of the Willow Creek area coupled with the normal development of Flathead and Wolsey there implies that the island was to the west or north rather than south or east. No such island is indicated by the work of Klepper and others (1957) to the north so it must have been to the west. Middle Cambrian islands may be inferred in or near the Philipsburg quadrangle (Emmons and Calkins, 1913, p. 51–53) 70–100 miles to the west. There, the Flathead and the Silver Hill formation (apparently equivalent to the Wolsey formation although Calkins did not make this correlation) exhibit rapid variations in thickness, especially in the southeast corner of the quadrangle where in places the Flathead is absent, and abnormally thin Silver Hill rocks lie directly on Belt rocks. The principal island mass would thus seem to have been between Three Forks and Philipsburg, where possible evidence for or against it is conveniently concealed by younger igneous rocks.

PARK SHALE

The Park shale, which conformably overlies the Meagher limestone in the Milligan Creek and Willow Creek areas, crops out even more rarely than does the Wolsey shale, but there are good, if incomplete, exposures east of Milligan Creek in the center of the W½NE¼ sec. 5, T. 1 N., R. 1 E., on the west bank of the main valley there; and at two places west of Willow Creek: in the only canyon in the NE¼ NW¼ sec. 12, T. 1 S., R. 1 W., and on the north bank of the creek along the Madison-Gallatin County line between secs. 4 and 9, T. 1 S., R. 1 W.

In the Milligan Creek area the Park shale is 150–200 feet thick; in the Willow Creek area it is somewhat thinner.

The formation is almost wholly greenish-gray clay shale. A few beds are calcareous and many bedding surfaces are strewn with flakes of chlorite and altered biotite. Near the base are scattered layers, less than half an inch thick, of yellow and orange calcareous siltstone and fine-grained clayey limestone. The upper part in the Willow Creek area includes much dark-yellowish-brown mudstone and shale. For measured sections, see pages 128 and 135.

Where exposed, the contact of the Park shale with the overlying Pilgrim limestone is sharp but apparently conformable. In most places, the contact is obscured by float from the basal ledge of the Pilgrim, and on the map it is arbitrarily placed at the base of this ledge.

The Park is a typical marine shale. No fossils were recovered from it in the Three Forks quadrangle, but its sparse fauna from neighboring areas indicates latest Middle Cambrian age (Deiss, 1936, p. 1333; Lochman and Duncan, 1944, p. 5; Hanson, 1952, p. 16). In seeming to lack thick transitional zones with the overlying and underlying carbonate formations, the Park shale in this quadrangle differs from that described in neighboring areas (Hanson, 1952, p. 15). If exposures were better, perhaps the apparent differences would disappear.

The Park shale reflects a brief regressive episode in the dominantly transgressive history of the Cambrian formations of western Montana (Sloss, 1950, p. 432). The nearest likely source area was to the west, where a Cambrian positive element in the general vicinity of the Idaho batholith was recognized half a century ago by Willis (1907, p. 399) noted subsequently by Walcott (1915, p. 198), and discussed in some detail by Ross (1935) and by Deiss (1941, p. 1095–1097). More specifically, central Idaho offers a possible source. Large areas of Belt rocks there could have been exposed to erosion at the required time, as, according to Ross (1934), the basal Paleozoic rocks are of Ordovician age (Kinnikinic quartzite). If the source area was to the south or southwest, it had little relief, as the Park shale apparently does not grade to a sand facies in these directions. On the contrary, it tends to pass southward into a limestone facies (Mann, 1954, p. 8). Further, it does not notably thicken toward a supposed southwestern shore (see Hanson, 1952, pl. 7*B*). Thus, a more distant easterly source may just as reasonably be inferred. Such a source would have been farther east than the Big Snowy Mountains, where Park equivalents are known to be thick (Hanson, 1952, pl. 9) but perhaps still within Montana, as the Park seems to be missing in the Williston basin (Sloss and Moritz, 1951, p. 2140).

PILGRIM LIMESTONE

Carbonate rocks assigned to the Pilgrim limestone crop out in all four areas of pre-Tertiary rocks. Complete sections, however, traceable for long distances, appear only in the Milligan Creek and Willow Creek areas. In the Mud Spring area, the Pilgrim lies on the Lombard thrust, which cuts out the lower one-quarter to one-half of the formation. In the Hossfeldt Hills area (sec. 20, T. 3 N., R. 2 E.), the upper part only of the Pilgrim appears in the core of an overturned anticline, here named the Hossfeldt anticline.

The Pilgrim, 350–450 feet thick, consists of persistent upper and lower sequences of ledge-forming thick-bedded mottled dolomite with an intermediate sequence of poorly resistant thin-bedded limestone and dolomite; but the thickness of the several facies varies widely as do many lithologic details. Characteristic of the Pilgrim throughout are distinctive yellowish pitted weathered surfaces.

The generalized section in the Milligan Creek area is:

	Thickness (feet)
Thick-bedded dolomite mottled in tones of yellowish gray with a few beds of light-gray fossiliferous limestone; ledge	200
Thinly interbedded gray limestone and orange dolomite; generally covered	150
Dolomite with subordinate limestone like upper unit	100

In the Pilgrim near Willow Creek the main rock types are much the same, but their thicknesses are notably different:

	Thickness (feet)
Mottled dolomite, ledge	50
Thin-bedded limestone and dolomite	200
Mottled dolomite, ledge	150

For more detailed measured sections, see p. 128, 134–135.

The Pilgrim near Willow Creek is consistently darker with many yellowish-brown and brownish-gray beds, and includes many thin beds of flat-pebble conglomerate and prominent oolitic and pisolitic zones rare in the Milligan Creek vicinity. Common also in the Willow Creek ledges are scattered twiglike bodies of earthy yellowish-gray calcite, presumably the remains of stromatoporoids or algae. The partial Mud Spring section is much like that near Milligan Creek, though the upper dolomite ledge is thinner. The partial section in the Hossfeldt Hills shares the brownish tones, the twiglike structures, and the conglomerate and oolitic beds of the Willow Creek sector. (For photographs of several of these textural features, see Hanson, 1952, pls. 2 and 3.)

The thin-bedded middle part has distinctive color and compositional banding. Layers from ¼ inch to 4 inches of gray microcrystalline limestone alternate with thinner layers of coarser grained, porous, clayey, orange or brown dolomite, giving a ribboned effect and leading to the informal field name of "banded black-and-gold" or "banded blue-and-gold". Mottled black-and-gold facies like those so common in the Meagher occur also in the Pilgrim, but are rare. Where present, the gold portions are generally dolomite or dolomitic limestone, and the black, limestone.

Exposures of the contact between the Pilgrim limestone and the succeeding Maywood formation are poor and in the 16 miles of contact traced, no suggestion was found either of transitional relation or of angular unconformity. The Pilgrim is overlain by the Devonian Jefferson dolomite in three places for a total of about 2½ miles: above the Lombard thrust in W½ sec. 30, T. 3 N., R. 2 E.; below the thrust on the east flank of the Hossfeldt anticline in the center of sec. 20, T. 3 N., R. 2 E.; and west of Willow Creek from NE¼NE¼ sec. 11 to center NE¼ sec. 12,

T. 1 S., R. 1 W. In none is the Pilgrim deeply eroded. The absence of Maywood rocks in these places suggests a disconformity due to initial nondeposition of the Maywood, on high parts of a post-Pilgrim, pre-Maywood erosion surface. It is possible that the rocks mapped as Maywood include some Late Cambrian beds, conformable with the Pilgrim, and that the disconformity is within the Maywood as mapped rather than between it and the Pilgrim. On the map the Pilgrim-Maywood contact where not exposed is arbitrarily placed where the dip slope on the top of the upper dolomite ledge begins to flatten.

The Pilgrim is like the Meagher in many ways. In structurally complex situations or where exposures are few it may be difficult to tell them apart without the aid of fossils. In the Three Forks quadrangle the acid-bottle is the most useful separatory tool as the Pilgrim is largely dolomite while the Meagher is almost wholly limestone, but this distinction assuredly does not hold regionally. Brownish or yellowish weathered tones suggest Pilgrim, whereas a bluish cast suggests Meagher. Thick sequences of banded black-and-gold beds suggest Pilgrim, whereas much mottled black-and-gold suggests Meagher. Limestone pebble beds are much more common in the Pilgrim.

AGE AND ORIGIN

The Late Cambrian age of the Pilgrim formation in the Three Forks quadrangle is established by the presence of the trilobites *Cedaria* sp., and *Kormagnostus* sp. [D177 ЄO] identified by A. R. Palmer (written communication, April 6, 1955) from the lower and middle parts of the sequence. In the absence of fossil collections from high in the Pilgrim, it is possible that the uppermost part may be of Early Ordovician age, as is the uppermost Pilgrim in southern Montana (Lochman and Duncan, 1944). The possibility is remote, however, as Cambrian fossils have been recovered from beds above the Pilgrim variously assigned to Peale's Dry Creek shale, Calkins' Red Lion formation, or Lochman and Duncan's Snowy Range formation not far from the Three Forks quadrangle (Deiss, 1936, p. 1312 and 1316; McMannis, 1955, p. 1395; Klepper, 1950, p. 32; Klepper and others, 1957, p. 13).

The Pilgrim resembles the Meagher and no doubt formed under similar conditions. Like the thin-bedded phases of the Meagher, the thin-bedded parts of the Pilgrim must have been deposited in disturbed, presumably shallow water, but the thick-bedded, finely homogeneous, clastic-free portions of both formations suggest quiet, presumably deeper, water. It is noteworthy that quiet-water phases of both formations are much better developed in the Milligan Creek area than in the Willow Creek area, and that this reflects the regional pattern (Hanson, 1952, p. 14–17, and pl. 7–8), for quiet-water phases of both formations are dominant west of the Three Forks area to beyond Philipsburg, and disturbed-water lithologies are dominant for great distances to the east. Both formations also thicken rapidly west of Three Forks, and in both dolomite becomes the dominant carbonate to the west; further, the shale and limestone-pebble conglomerate beds that are important constituents of the Pilgrim nearby to the east (Verrall;[5] McMannis, 1955, p. 1394–1395) are thin or absent to the west, and the Park shale is unrecognizable in the Philipsburg area (Hanson, 1952, p. 15). These relations, incidentally, are just the opposite of those often cited for other regions, where, to quote a recent summary (Cloud and Barnes, 1957, p. 185) "rocks that are dolomites near ancient lands are prone to grade to limestones away from the old shore lines"—but the quotation may apply to the western shoreline. The features cited suggest a genetic relation between conditions favoring at least one class of dolomite deposition and quiet, clear, but not necessarily deep water. It has been contended (Hanson, 1952, p. 14, 17) that disturbed-water features and fossils were initially abundant in the Cambrian carbonate rocks west of Three Forks but have been obliterated by regional dolomitization. Far simpler, however, to accept the direct suggestion of quiet clear water, poorly populated perhaps because of its high magnesian content. Initial quiet-water conditions to the west are implied in the westward thinning and disappearance of the shales in and between the Meagher and Pilgrim, which could hardly be the result of dolomitization. Further, in westerly occurrences of these formations disturbed-water features and fossils are just as rare in the few thick dolomite-free limestone beds as in the dolomites.

The dolomite in the banded and mottled beds of the Pilgrim seems to have formed during diagenesis, partly by precipitation of dolomite in openings in the porous gold layers, but mostly by replacement of calcite. Structurally the banded and mottled beds in the Pilgrim are identical with those in the Meagher. The fact that the porous mottles are dolomitic limestone in the Pilgrim but are limestone in the Meagher suggests that the mottling developed similarly in both formations, but that dolomite was introduced into the Pilgrim at a later stage in the lithification process. Whatever the source of the dolomite-bearing solutions which invaded the mottled beds, the introduction of dolomite

[5] See footnote, p. 11.

was so widespread that special sources such as juvenile hydrothermal solutions seem out of the question. Most likely the solutions were connate waters mobilized during compaction and diagenesis. How magnesium was concentrated in this environment, and why dolomite was more stable in it than calcite are unknown.

Although the dolomite in the banded and mottled beds seems to be of postdepositional origin, it does not necessarily follow that the dolomite of the thick-bedded phases represents the same replacement process gone to completion. The field evidence of extensive thick beds of seemingly almost pure dolomite, much of it no coarser in grain than intercalated beds of limestone, and showing only trivial visible signs of intermediate stages of alteration that might be ascribed to dolomitization, suggests that these beds were deposited initially as dolomite.

The primary origin of any stratigraphic dolomite is open to question and with good reason (for recent discussions see Cloud and Barnes, 1957, p. 182–186 and Fairbridge, 1957). In the laboratory, calcite is readily deposited from sea water or similar solutions under conditions expectable in the ocean, but dolomite is not (Van Tuyl, 1916, p. 297–306). In several warm shallow parts of the open ocean (for example, the Bahama banks as discussed by Smith, 1940) limestone has been found now forming as a chemical and organo-chemical precipitate, but dolomite has not. (It has, however, been reported as a "definite authigenic mineral" from the North Atlantic Ocean, both in nearshore deposits in the Gulf of Guinea and in pelagic deposits in the Cape Verde Basin, by Correns (1939, p. 385), and dolomite is forming today on a large scale in coastal saline lakes and a nearby shallow inlet of the sea in the south-east Province of South Australia (Alderman and Skinner, 1957). It is plain, too, that dolomitization has occurred in many mottled carbonate rocks and on a large scale. Further, as Rubey (1952) has reasoned, there have probably been no major shifts in the composition of ocean water since far back into the Precambrian, so that it is hard to visualize physicochemical changes in the open ocean small enough to be reasonable but large enough to cause the deposition of vast amounts of primary dolomite rather than calcite.

It becomes tempting, then, to push the problem of stratigraphic dolomite into the postdepositional history of the rock, though not too far; a penecontemporaneous origin is often postulated, for reasons well expressed by Blackwelder long ago (1913, p. 619–624) based largely on volume considerations. But to decide, as did Blackwelder, and Daly before him (1909, p. 170) and many others after, that dolomite has formed on a vast scale by wholesale replacement of part of the calcium from calcium carbonate "jelly" just after precipitation and before deep burial is merely to veil the problem: the old questions are still there, and some new ones. How did the magnesium, evidently not concentrated enough to precipitate when the calcite jelly was deposited, become sufficiently concentrated to displace part of the calcium merely by being covered with a little very wet sediment? The rise of temperature attending burial could not be expected appreciably to modify the relative solubilities of dolomite and calcite. Did the jelly selectively remove magnesium by sorption from the water that was being squeezed out of the accumulating sediment? Very likely the controlling mechanism was not the ordinary solubility product. Perhaps such agencies as microorganisms or natural earth-currents significantly modified the chemical relationships of calcium and magnesium below the depositional interface. If the magnesium trapped in the jelly was simply the usual magnesium of average sea water, how have any chemical marine carbonates escaped dolomitization? If the substitution of magnesium for calcium happens so readily, how is magnesium sometimes saved from interaction long enough to produce dolomite mottles later in the lithification process, as in much of the Meagher limestone?

And what becomes of the replaced calcium? In this part of the problem there is a little negative field evidence. In the great stratigraphic dolomites calcite is not often trapped in the rock as cement, it does not form large or numerous veins, and it does not, apparently, return to the ocean floor to produce limestone interbeds in a volume proportional to dolomite. Perhaps the freed calcium promptly reacts with pore water to form new calcite that in turn reacts with rest magnesium, on an ever-decreasing scale like the mirrors in Flemish paintings.

Other queries arise but the point is made without them: ignorance about the mechanism of dolomite formation is profound. Under the circumstances, the thesis of direct deposition of stratigraphic dolomite need not yet be abandoned. So little is known about the chemistry of the carbonates, especially in their natural habitat, that current chemical inferences about the origin of dolomite need not be taken as conclusive. Further, the fact that dolomite is not forming in today's oceans may not be especially relevant. An important point in this connection, noted first, perhaps, by Tarr (1920), is that the great dolomite masses of the Paleozoic were not deposited in the open ocean, but rather in bodies of salt water which may have had limited connection with the ocean. The dolomite now forming extensively in coastal South Australia (Alderman and Skinner, 1957) is an impressive if rare modern example. Epeiric seas like those in which the great stratigraphic dolomites were deposited do not exist today. The closest parallel might be along the northwestern coast of Australia, on the Sahul shelf and related shelves, which is little known with respect to sedimentation (see Fairbridge, 1953, for summary of present knowledge).

The direct physical evidence on record, both microscopic and macroscopic, is that dolomite deposits share every feature of limestone deposits though perhaps not in the same proportions. In the only exhaustive published investigation of the petrography of interbedded limestone and dolomite rocks known to me, Sander (1936) found that in the rhythmically bedded Triassic carbonate rocks of the Alps every fabric developed is present in both limestone and dolomite, and that, as far as the petrographic evidence goes, dolomite may form in all the ways that calcite may.

DEVONIAN AND LOWER MISSISSIPPIAN

The Devonian rocks, 600–1,000 feet thick, are separated into three formations: The Maywood formation, at the base; the Jefferson dolomite; and the Three Forks shale. Some of the unfossiliferous beds at the base of the Maywood as mapped are conceivably pre-Devonian, and near the top of the Three Forks shale beds containing a Late Devonian fauna are transitional with and cannot be mapped separately from beds containing an Early Mississippian fauna. Thus, neither the upper nor lower limit of the Devonian section is accurately located, but the uncertainty involves only a few tens of feet of strata at each end.

The Devonian rocks are about 50 percent dolomite, 20 percent dolomitic limestone and limestone, 20 percent shale, and 10 percent siltstone and sandstone; the clastic ratio is about 0.4. Included high in the section is a small thickness of evaporite beds, the oldest such deposits in the quadrangle.

Although the Devonian has not received nearly so much attention as the Cambrian, regional treatment of Devonian stratigraphy has been given by Sloss and Laird (1947) with emphasis on correlation, and by Andrichuk (1951) with emphasis on the sedimentational environment.

MAYWOOD FORMATION

Assigned to the Maywood formation (Emmons and Calkins, 1913, p. 64–65; age emended from Silurian(?) to Devonian after Lochman, 1950, p. 2213) are poorly exposed yellow, orange, and red calcareous clastic rocks locally grading upward into, and interbedded with, light-colored limestone that occupies a narrow grassy valley or saddle between the upper dolomite ledge of the Pilgrim limestone and the ledge-forming brown dolomite at the base of the Jefferson dolomite. In the 13 linear miles along which this stratigraphic interval has been traced, outcrops and float fragments suggest that it is occupied entirely by brightly colored siltstone and mudstone for about 9 miles, and that limestone appears in the upper part of the interval along about 4 miles.

The interval is markedly thicker and contains much more limestone in the Milligan Creek and Mud Spring areas. In these areas its thickness commonly ranges from 30 to 100 feet, though it is missing for about half a mile in sec. 30, T. 3 N., R. 2 E., and swells to about 200 feet in the southwest corner of sec. 31, T. 3 N., R. 2 E. In the Hossfeldt Hills area it is confined to the crestal zone of the anticline (sec. 20, T. 3 N., R. 2 E.). On the upright limb to the west of the crest the sequence is persistent but less than 50 feet thick, and lacks prominent limestone beds; on the overturned limb it is absent, perhaps due to squeezing rather than to nondeposition or erosion. In the Willow Creek area the Maywood is missing for a mile west of the Principal meridian and is generally less than 30 feet thick elsewhere; limestone is virtually absent.

In only two places is the entire sequence well exposed. The best exposure is one-third mile east of Milligan Creek on the west wall of the only large canyon in the NE1/4NE1/4 sec. 1, T, 1 N., R. 1 W., where 60 feet of these beds appear in a small cliff (measured section E, pl. 2). The sequence may conveniently be summarized as:

	Thickness (feet)
Limestone and siltstone. Thin-bedded fossiliferous yellowish and bluish-gray limestone and orange calcareous siltstone	20
Mudstone and limestone. Alternating beds 1–3 feet thick of calcareous mudstone, mottled in orange and red, and medium-gray limestone	15
Mudstone and fine sandstone. Beds 1–4 feet thick, calcareous, vuggy; orange with scattered reddish mottling	25

The clastic beds throughout are much alike and there are no overt breaks in the sequence.

The second good exposure is 1⅓ miles northeast of the first, in SW¼NW¼ sec. 32, T. 2 N., R. 1 E. Here, on the north wall of the eastward-draining gulch at the south base of hill 4797, the interval is 30 feet thick. The upper 12 feet consists of thinly interbedded gray limestone and yellowish-gray shale. The lower 18 feet is yellowish-gray siltstone and shale, mottled in orange, with 1 foot of dark-purplish-gray shale at the top. These rocks, though broadly similar to those in the first exposure, are markedly harder, more fissile, and more somber, perhaps due to differences in weathering rather than in initial sedimentation.

In the area of maximum thickness (sec. 31, T. 3 N., R. 2 E.), the upper half of the sequence is almost entirely thick-bedded light-yellowish-gray limestone, and the lower half, poorly exposed, seems to be interbedded brownish-gray limestone, and yellowish-brown and yellowish-gray siltstone and shale. Scattered in the lower half are thin lenses of conglomerate, the pebbles of which are red siltstone.

The contact relations of the Maywood formation are not wholly clear. The unit as mapped overlies the Pilgrim limestone with erosional unconformity. The upper contact with the Jefferson dolomite is rarely exposed and then only where there is limestone in the Maywood. At such places the contact is conformable. In the few places where light-colored limestone alternates with dark dolomite near the base of the Jefferson, the contact might even be called gradational. The local absence of Maywood between the Pilgrim and the Jefferson may indicate overlap of the sea during Jefferson time on pre-Devonian topographic high places not buried by the Maywood, but could be due to local or widespread post-Maywood pre-Jefferson erosion due to uplift. Conformity with overlap seems the better explanation.

On the map the upper contact where not exposed is arbitrarily drawn at the break in slope between the lowest outcrop of fetid dolomite, regarded as Jefferson, and the highest appearance of red, orange, or yellow siltstone float.

For measured sections of the Maywood, see pages 128, 131, and 134.

AGE, NAME, AND CORRELATION

The only identifiable fossil from the Maywood are brachiopods "similar to the genus *Allanaria* * * * most likely of Late Devonian age" (A. J. Boucot, written communication, Apr. 14, 1955) recovered from the limestone beds 12–16 feet below the top in the exposure one-third of a mile east of Milligan Creek (loc. 74 [D38 (SD)], p. 2). As all the beds assigned to the Maywood seem to reflect unbroken deposition, they are presumably of the same age. Pending better faunal evidence, the age is tentatively considered Late(?) Devonian.

The clastic rocks of this sequence are those designated Dry Creek shale (member of the Cambrian Gallatin formation) by Peale (1893, p. 24; 1896, p. 2), but this name is not well applied to rocks with Devonian fossils. This is not to deny the existence of Peale's Dry Creek shale, for Cambrian fossils have been found in beds of Dry Creek lithology above the Pilgrim limestone nearby (Deiss, 1936, p. 1312 and 1316; McMannis, 1955, p. 1395; Klepper, 1950, p. 32; Klepper and others, 1957, p. 13). Whether the name should be used even for Cambrian rocks above the Pilgrim is debatable. (See Lochman, 1950, p. 2204–2205.)

Above the Dry Creek shale and below the black carbonate rocks of the Jefferson formation, Peale (1893, p. 24–25) recognized an even later Cambrian unit, the "Pebbly limestones" (member of the Gallatin formation). This is the only stratigraphic unit of Peale that might be applied to the limestone parts of the Maywood. There is little lithologic similarity between Peale's description and the limestone of the Maywood, and it is more likely that Peale simply did not see any of the rare outcrops of Maywood limestone in his widely spaced traverses. This is not to deny the existence of any Cambrian rocks equivalent to the Pebbly limestones, as Deiss (1936) did. A more reasonable interpretation (Sloss and Laird, 1947, p. 1409) is that the Pebbly limestones of Peale exist, at least on Dry Creek where Peale measured them, but are lacking in many other places including the Three Forks quadrangle owing to pre-Maywood erosion.

While assigning the entire sequence to the Maywood is reasonable, it may oversimplify the situation. The limestone-bearing upper parts of the interval are all no doubt part of the same depositional sequence of Late(?) Devonian age. It is possible, however, that in places the clastic rocks of the sequence are partly, or even wholly, of Cambrian age and that an undetected irregular erosion surface separates similar rocks of two systems of widely different age. An extreme interpretation is that only the prominent limestones are Devonian and that the bulk of these rocks is Cambrian. If this is so, pre-Jefferson Devonian rocks may be confined to the upper plate of the thrust system in the Three Forks quadrangle.

ORIGIN

The Maywood formation is apparently marine, but both the texture and color of its clastic rocks reflect a geologic history different from that of the Cambrian marine clastic Wolsey and Park formations. The differences may date from the time of deposition or may be the result of weathering long after deposition. Whether or not any Dry Creek equivalents occur in the quadrangle itself, fine-grained red, orange, and yellow clastic rocks of Late Cambrian age are widespread nearby. Probably, as Lochman (1950, 1957) suggested, the Dry Creek was deposited as gray and green shale like the Park and Wolsey formations. Impersistent shales in the Dry Creek, much like those in the Park and Wolsey, crop out in the Toston quadrangle and were also reported by McMannis (1955, p. 1395) and by Sloss and Laird (1947, p. 1411–1413).

Later the Dry Creek was partly eroded and deeply weathered to its present colors and textures during continuous or recurrent emergences in the well-known stillstand in the region during Ordovician, Silurian, and earlier Devonian time. Reworking of the weathered Dry Creek strata by the transgressing later Devonian sea may have produced the clastic beds of the Maywood. As the shoreline retreated, in a direction as yet unknown, limestone deposition gradually prevailed over clastic deposition. Such a history could explain the similarity between the clastic facies of the Dry Creek and Maywood despite their great difference in age.

It is possible, of course, that the similarity is fortuitous, and that the Maywood clastic deposits were derived from other, more remote, sources. A reconstruction of pre-Middle Devonian paleogeology by Sloss (1950, fig. 6) indicates that non-Cambrian rocks, mostly Ordovician, were exposed in pre-Maywood time within a hundred miles of Three Forks to the west, south, and east, but not to the north. These rocks to the south and east are almost wholly carbonate (Big Horn dolomite) unsuitable as clay sources. The Ordovician of central Idaho, however, contains a high proportion of suitable source material (Saturday Mountain and Kinnikinic formations of Ross, 1934).

JEFFERSON DOLOMITE

A thick sequence of dolomite and limestone assigned to the Jefferson dolomite (Peale, 1893, p. 27–29) crops out in all the areas of pre-Tertiary rocks. The lower part of the Jefferson, except where dips are nearly vertical, makes the highest of the series of cliffs that begin in the Cambrian limestones; the upper part of the formation descends in a long dip slope, broken by low ledges, to a broad valley underlain by Three Forks shale.

The Jefferson is a distinctive formation although it varies considerably in lithologic detail and in thickness. The dominant rock type is rather coarsely crystalline sugary fetid dark-brown dolomite in beds a few feet thick. The color and odor are due to hydrocarbons in films on individual grains and in tiny interstitial pellets. At the top of most sections is distinctive light-colored coarse-grained dolomite 30–60 feet thick. Typical sections also include many thick to thin beds of black dolomite, of yellowish-gray dolomite, and of gray limestone and dolomitic limestone that are finer in grain than the dominant dolomite type and that tend to form ledges.

Widespread, too, though of small volume are layers of dolomite pebble conglomerate, pebble breccia, and "spaghetti beds." The pebble conglomerate typically consists of discoidal subrounded pebbles of dark sugary dolomite, generally black or brown, but rarely red, in a lighter colored matrix of finer grained dolomite or dolomitic limestone. The pebbles are oriented parallel to bedding and generally do not touch but float in the matrix. Much less common though widely present are thin beds with pebble-size fragments so angular that the rock must be termed breccia. The pebbles typically are of dark dolomite and are crowded together with only rude orientation in a light-colored limestone matrix. Possibly the conglomerate and breccia grade into each other but intermediate types were not seen. The "spaghetti beds" consist of closely packed slender tubes, about ½ inch long and 1/20 inch across, of grayish-yellow finely crystalline calcite imbedded in a brown dolomite or dolomitic limestone matrix of coarser grain. The tubes are patently organic, but in the few thin sections seen, any cell structure that they may have had has been obscured by recrystallization. The "spaghetti" structure has been variously interpreted as representing algal colonies, worm burrows, and poorly preserved corals of the genus *Cladopora*. According to Helen Duncan (written communication, June 28, 1956) however, the "spaghetti beds" of the Jefferson were probably formed by dendroid stromatoporoids of the genus *Amphipora* (Lecompte, 1952, p. 321–331, pls. 67–70).

The proportions of limestone, conglomerate, and "spaghetti" beds vary widely. Limestone is markedly more abundant in the lower part of the formation in the Milligan Creek and Willow Creek areas, but in the Mud Spring area it is more abundant in the upper part. Conglomerate and "spaghetti" beds are rare in the Milligan Creek and Mud Spring sectors, but prominent in the Willow Creek area.

The Jefferson dolomite is thickest in the Milligan Creek area, where it averages about 600 feet. Apparent thicknesses are locally as great as 750 feet, but in such places some duplication by faulting is suspected. The typical section in this area includes:

	Thickness (feet)
Grayish-yellow thick-bedded dolomite	50
Interbedded grayish-yellow dolomite and yellowish-brown dolomite, base forms ledge	300
Gray limestone thinly interbedded with brownish-gray dolomite, poorly exposed	200
Dark-yellowish-brown dolomite, forms ledge	50

Limestone interbeds are considerably more abundant east of Milligan Creek than west. Zones of coarse solution breccia, widespread in the Jefferson elsewhere, are prominent in the Three Forks quadrangle only in the northeast side of the Milligan Creek sector, especially sec. 29, T. 2 N., R. 1 E.

In the Mud Spring area the formation seems much thinner, averaging around 400 feet, and less magnesian, but here the Jefferson is much affected by folding and faulting, and the differences may be more apparent than real. Furthermore, the formation has been widely altered by igneous intrusions in the northeast part of the Mud Spring area. Near the 1ON pluton in sec. 25, T. 3 N., R. 1 E., the Jefferson is bleached to pale tones of yellowish gray and is not fetid. On the other hand the Jefferson in the high ridge east of this intrusive is unusually dark and petroliferous, probably because of condensation of hydrocarbons distilled from the formation close to the intrusion. This sensitivity to intrusive igneous rocks is apparently a regional characteristic of the Jefferson (see Emmons and Calkins, 1913, p. 66; Klepper and others, 1957, p. 15; Calkins and Butler, 1943, p. 20).

Seemingly straightforward sections in the Mud Spring area (SE cor., sec. 36, T. 3 N., R. 1 E., and NE cor., sec. 1, T. 2 N., R. 1 E.) show:

	Thickness (feet)
Thick-bedded yellowish-gray and yellowish-brown dolomite (2/3), and yellowish-gray limestone, spottily dolomitic (1/3), poorly exposed on dip slope	300
Thick-bedded dark-yellowish-brown dolomite, upper part forms ledge	100

In the Hossfeldt Hills the Jefferson appears only near the crest of the overturned anticline. Exposures are poor, but the formation seems to be about 500 feet thick and not notably different from the Mud Spring area.

The Jefferson dolomite in the Willow Creek vicinity is comparatively thin, rarely exceeding 350 feet, and typically consists of:

	Thickness (feet)
Grayish-orange and yellowish-gray thin- to thick-bedded dolomite	100
Yellowish-brown and brownish-gray thick-bedded dolomite with "spaghetti" and pebble conglomerate beds; forms ledge	100
Orange dolomitic limestone with pebble conglomerate beds, poorly exposed	50
Olive to brownish-gray thick-bedded dolomite, forms ledge	50
Yellowish-brown dolomitic limestone with many chert masses, poorly exposed	50

Although the Jefferson dolomite as a unit is distinctive and easy to recognize, similar rock types occur sparingly in other formations. For example, in the Willow Creek sector, and also east of the Lombard thrust in the neighboring Toston quadrangle, the upper part of the Pilgrim is dark, coarse grained, and fetid with many twiglike bodies. Dark, fetid, dolomitic beds also occur locally in the Meagher limestone and in the Lodgepole limestone of the Madison group. Consequently, strata should not be assigned to the Jefferson on the basis of single small exposures, however "typical" they might appear.

The contact with the overlying Three Forks shale is sharp but conformable. Where exposures permit, it is placed between the light-colored thick-bedded dolomite at the top of the Jefferson and the thin-bedded orange siltstone and limestone at the base of the Three Forks. This is generally close to the valley bottom, for stream courses tend to follow the unresistant and intricately fractured lowest beds of the Three Forks. Where the contact is not exposed it is arbitrarily placed at the foot of the dip slope on the uppermost Jefferson ledge.

Detailed measured sections of the Jefferson are given on pages 127–128, 130–131, and 134.

NAME, AGE, AND CORRELATION

In the more casual usage of his day, Peale (1893) referred to these rocks variously as Jefferson formation and Jefferson limestone although the chemical analyses he quoted show the rock to be dolomite. The term "Jefferson formation" is widely used, but because dolomite is the dominant rock type in the Three Forks quadrangle and regionally, I prefer the name Jefferson dolomite.

Except for the *Amphipora* beds, and a few recrystallized corals, fossils were not recognized in the Jefferson dolomite. The formation is lean in fossils everywhere, a feature no doubt closely related to its high magnesian content. Despite this, its age is regarded as definitely Late Devonian, for the underlying Maywood formation has a probable Late Devonian fauna and the overlying Three Forks shale has a rich Late Devonian fauna. The presence of *Amphipora*, according to Helen Duncan (written communication, June 28, 1956), permits but does not demand a Late Devonian age:

> * * * [*Amphipora*] are a characteristic feature of the upper Middle and lower Upper Devonian dolomites in western United States. In the Great Basin, banks of *Amphipora* are such distinctive lithologic features that they are widely referred to as "spaghetti beds." Similar beds have not been described in the older literature on the Jefferson of southwestern Montana, but Jean M. Berdan and I made a point of looking for *Amphipora* when we examined the Logan section in 1951. We found several impersistent beds of this fossil in the Jefferson at that locality and have also identified *Amphipora* in samples collected from the Jefferson near Melrose in Silver Bow County, and in the Madison Range, Ennis quadrangle. *Amphipora* beds probably are far more widely developed in the Jefferson of the region than collections in hand indicate.
>
> Ordinarily these fossils were not collected because, until recently, most American geologists and paleontologists thought they were corals (especially *Cladopora*), sponges, bryozoans,

or some other organism too poorly preserved for identification and therefore of no value for age determination. In spite of their unpromising appearance, specimens rarely are too thoroughly recrystallized for generic identification. The general prevalence of *Amphipora* in calcareous facies of the Devonian all over the world makes this one of the more useful fossils in stratigraphic work.

The Late Devonian age of the Jefferson is accepted by most workers (see Devonian correlation chart of Cooper and others, 1942; Sloss and Laird, 1947) although both Berry (1943, p. 10–14) and Verrall [6] favor late Middle Devonian age for the lower part of the Jefferson dolomite in this vicinity.

According to Sloss and Laird (1946; and 1947, p. 1407–1410) "* * * at its type locality near Logan, Montana and elsewhere, the Jefferson is divisible into an upper dolomite member and a lower limestone member." The two members, however, have not been recognized in the Three Forks quadrangle, in the Bridger Range (McMannis, 1955, p. 1397), in the Elkhorn Mountains (Klepper and others, 1957, p. 14), or even in the southern Horseshoe Hills that include Logan (Verrall).[7] On the other hand, the lower limestone unit is reported in sections measured in several areas by Sloss and Laird, and by Andrichuk (1951, p. 2371–2374); Klemme [8] noted that limestone dominates the lower part of the Jefferson in some parts of the Sixteen Mile Creek area (Toston and Maudlow quadrangles) but is subordinate in other nearby parts; the entire formation is dolomitic limestone in the Canyon Ferry quadrangle (Mertie, Fischer, and Hobbs, 1951, p. 25) and far south in the Gravelly Range (Mann, 1954, p. 10) it is described as limestone throughout; and in the Three Forks quadrangle limestone is an important component of the lower Jefferson in the hills east of Milligan Creek, though a minor component west of the creek. These observations suggest that the Jefferson is characterized by rapid lateral changes in the relative proportions of limestone and dolomite, and that dividing the formation in the Three Forks region into limestone and dolomite members is at best of strictly local value.

The boundaries of the Jefferson dolomite chosen in this report are believed to be those of Peale (1896, p. 2), who described the Jefferson as "black or mud-colored limestone, which is generally crystalline and magnesian from top to bottom." There is room, however, for other choices in placing the upper boundary of the Jefferson, as Peale's description of the rocks near the contact with the Three Forks shale is somewhat cloudy and his reported thicknesses are hard to confirm. Thus, Sloss and Laird (1947, p. 1409–1410) felt impelled to assign to the Jefferson the unresistant red, yellow, and orange "breccias and associated shales," about 100 feet thick, above the massive dark dolomite, despite their closer physical affinity to the clastic beds of the Three Forks, on the grounds that to do otherwise would violate Peale's original classification. This approach led Sloss and Laird to report significantly greater thicknesses of Jefferson formation than those stated here, and correspondingly smaller thicknesses of Three Forks shale.

[6] See footnote, p. 11.
[7] See footnote, p. 11.
[8] See footnote, p. 13.

ORIGIN

The Jefferson dolomite is a marine deposit probably laid down in shallow water subject to considerable current action, but far from sources of land-derived sediment. In thickness and in many lithologic details the Jefferson is similar to the Meagher and Pilgrim limestones, but in several respects it is strikingly different and must have had a correspondingly divergent origin or history. The most vivid differences are the sugary texture, the rich content of hydrocarbons reflected in color and odor, and the color contrasts between limestone and dolomite.

The sugary texture may result from recrystallization, but more likely it is a clastic fabric due to reworking of preexisting dolomite. A clastic origin is favored by the occasional presence of cross-stratification and channeling in the sugary dolomite beds.

The sugary texture and the rich content of hydrocarbons may be related, for in general the most petroliferous parts of the formation are the most sugary. There are exceptions, however, particularly in the upper part of the formation where many beds of sugary dolomite are very pale and poor in hydrocarbons. If texture and hydrocarbon content are related, the relation probably is epigenetic and structural rather than syngenetic. Very likely the volatile hydrocarbons, whatever their origin and initial distribution in the sediment, have tended to migrate during lithification (and possibly tectonism) into the porous sugary parts of the formation.

Surprisingly, the Jefferson has little of the mottling so widespread in the Cambrian carbonate rocks; individual beds retain their color, texture, and composition for long distances. Whether the individual bed is limestone, dolomitic limestone, or dolomite, it seems to have reached a stable composition during or soon after deposition, and was only locally subjected to postlithification attack by migrating dolomitizing solutions. Although homogeneous fabric in dolomites may mean that the metasomatic process has gone to completion, this is questionable where intermediate products of the process are rare or lacking.

The "spaghetti beds" offer some evidence on the problem of dolomite origin. The stromatoporoid tubes are porous and are finer grained than their dolomite matrix. Had dolomitizing solutions entered the rock after deposition, they would presumably have entered the tubes, but the "spaghetti" is wholly calcite. This suggests that the dolomite of the matrix was not produced by postdepositional reactions but was already dolomite when it enclosed the *Amphipora* tubes. The point is weakened, however, by the possibility that the sugary dolomite is a clastic rather than a direct chemical deposit.

Beales (1953) studied the widespread dolomitic mottling in the Palliser limestone in Alberta, which resembles the Jefferson and is also of Late Devonian age. In beds exactly like those here called "spaghetti beds" (see his fig. 4), tubes which Beales ascribed to worm burrows and algal colonies are dolomite, and served as centers from which dolomitizing solutions worked outward into the surrounding limestone, just the reverse of conditions at Three Forks. The ready conclusion at this time is that stratigraphic dolomites can form in more than one way. There seem to be stratigraphic dolomites and stratigraphic dolomites.

The origin of a rock mass which is here limestone and there dolomite, with a zone of mottling or intercalation or imperceptible chemical gradation between can be approached in the same way as the origin of granite: as a problem in contact relations between initially unlike materials, or as a problem in the transformation of a single original material. The dramatic appeal of transformation or replacement hypotheses and their current popularity makes it easy to neglect other possibilities, but the state of knowledge does not. At this writing, it is at least as reasonable to visualize simultaneous deposition of some limestone and dolomite in separate semiconnected parts of a sea, with both lateral and vertical gradations between due to shifts of chemical and physical variables, as to picture all stratigraphic dolomite as due to postdepositional replacement of limestone.

The thick, extensive, and well-exposed Devonian and Cambrian carbonate formations of the region around Three Forks are well suited for detailed study of the dolomite problem. I suspect, though, that field studies, however detailed and supported by petrographic and petrochemical data, will bring little progress unless further supported by experimental data. In whatever ways dolomites form, it must be under conditions well within the means of modern laboratories to duplicate. Is it not time to try again to make laboratory analogs of stratigraphic dolomite?

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The differences in detail between the Jefferson and the Cambrian carbonate formations may be related to gross differences in the environment of deposition. Regional studies indicate that the Cambrian limestones pass shoreward into normal marine clastic rocks. Much of the Jefferson, on the other hand, grades, at least northward from the Three Forks area, into thick evaporite sequences (Potlatch anhydrite of Perry, 1928, and equivalent units) and the Jefferson in the Three Forks quadrangle is conformably succeeded by evaporitic rocks. If shoreline environments were in marked contrast, then offshore depositional conditions probably differed too, but how is not yet known.

THREE FORKS SHALE

The Three Forks shale (Peale, 1893, p. 29–32), a highly varied assemblage of clastic and carbonate rocks, appears in all the areas of pre-Tertiary rocks. Composed mainly of clay shale, the formation is unresistant and forms deep valleys between ridges of Jefferson dolomite and Lodgepole limestone. Outcrops are rare, but the formation is nevertheless one of the most readily recognized in the region, even when faulted out of its normal stratigraphic position, because of its unique succession of rock types and its unique richness in invertebrate fossils, both apparent from float where outcrops are lacking.

Throughout the quadrangle, the Three Forks consists of a lower unit of thin-bedded, crumpled, and brecciated orange limestone and siltstone, a medial unit of shale with many thin limestone beds, and an upper unit of flaggy orange siltstone, but the thickness and lithologic detail of each of these units range widely. Generally, a valley bottom follows the lower brecciated unit, a gently rolling grassy slope marks the medial shale, and a steep slope is developed on the upper unit, partly because it is more resistant and partly because it is protected by the basal ledge of Lodgepole limestone.

In the Milligan Creek area the formation is exceptionally well exposed, particularly along Dry Hollow. A generalized section in the Milligan Creek vicinity where the Three Forks is 300–400 feet thick comprises:

	Thickness (feet)
Thin- to thick-bedded orange siltstone and sandstone	50
Thinly interbedded dark-green shale and dark varicolored limestone with some layers of orange siltstone in upper part	100
Olive shale with subordinate thin beds of orange siltstone	150
Thin-bedded orange and yellow limestone and mudstone, brecciated and contorted	50

The basal breccia is riddled with cavities; its fragments are generally no more than a few inches long, angular to subangular, and spottily cemented by a carbonatic clay. It grades laterally into and is interbedded with crumpled beds of similar color and composition, with contortions of wavelength and amplitude ranging from a few inches to several feet. Red layers were not seen in outcrop, but bits of red mudstone appear sporadically in the float from this interval and red soil is developed on it.

The green and olive shale that makes up the bulk of the formation is almost all calcareous clay shale, with little of the mica and glauconite that characterize the Cambrian green shales. The dominant green and olive types grade locally into black, purple, brown, and orange varieties, presumably related to local concentrations of organic and ferriferous material. Black shale is especially common in the green-shale interval in sec. 31, T. 2 N., R. 1 E., east of Milligan Creek, where many prospect pits have been dug, apparently in the hope of finding coal or possibly oil shale. Black shale is also rather common just west of the Lombard thrust in the Mud Spring sector.

All the shale varieties are locally rich in fossils, especially small thin-shelled brachiopods; a few dark shale beds, lean in brachiopods, contain nautiloid and goniatite cephalopods. The brachiopod shells are almost entirely carbonate, with fillings of finely crystalline carbonate or of shale, whereas nearly all the cephalopod tests are replaced by dark-reddish-brown to black hematite with fillings of hematite, or, rarely, of pyrite. Many of the replacements are beautiful pseudomorphs with smooth shiny surfaces. More commonly, though, replacement has gone beyond the pseudomorph stage to yield, first, shells partly crusted with crystals of hematite, and finally reniform masses of hematite as much as 3 inches long, with rough surface coatings of tiny cubes and octahedra of hematite. The cores of these masses are generally hematite and hydrous iron oxides, or, uncommonly, finely crystalline pyrite. Associated with the fossil pseudomorphs and nodules are cubes and octahedra of hematite as much as half an inch in diameter, and small clusters of such crystals. Evidently the crystals, and the hematitized fossils too, were formerly pyrite, now replaced by hematite and limonite.

These occurrences are not confined to the Three Forks shale of the Three Forks quadrangle, as "pyritized fossils, chiefly cephalopods" have been collected from the adjoining Toston quadrangle (Haynes, 1916a, p. 16) and "pseudomorphic crystals of limonite after pyrite are common" in the Spokane Hills 50 miles to the north (Mertie, Fischer, and Hobbs, 1951, p. 26). The interbedded siltstone is usually a few inches thick and contains about equal amounts of angular grains of quartz and rhombs of calcite. The siltstone is porous but is firmly cemented with iron-stained carbonatic clay.

The limestone beds intercalated with the clay shale are generally less than a foot thick, are colored in dark tones of red, brown, and gray, and are largely made up of fossil fragments and calcite sand in a matrix of clayey finely crystalline limestone. Some are so fossiliferous as to be classed as coquina. As in the shale, the dominant type of fossil material is small thin-shelled brachiopods but the limestone is rich also in larger, thicker shelled, more tumid brachiopods, as well as bryozoan and crinoid fragments; cephalopod remains are sparse. Widely scattered in both shale sequences are lenses a few inches to a few feet thick of yellow and orange limestone breccia, like that in the basal unit.

In the upper 30 to 50 feet of the green shale sequence, siltstone beds like those in the olive shale interval become increasingly common, thicker, and coarser in grain, eventually grading into the upper siltstone and sandstone unit. This upper unit ranges widely in thickness; just east of Milligan Creek (center sec. 36, T. 2 N., R. 1 W.) it is nearly 100 feet thick, and at the east edge of the Milligan Creek sector it is 70 feet thick, but it is much thinner between these areas, and also thins west of Milligan Creek, so that it is less than 30 feet thick at the west edge of the quadrangle. These changes may be due not so much to decrease in the total amount of siltstone in the upper part of the formation as to interruption of the light siltstone succession by interbeds of dark shale.

The upper siltstone and sandstone is mineralogically much like the siltstone lower in the formation, composed of about half quartz and half calcite, but individual beds tend to be thicker and the average grain size is greater, with many beds in the fine sand–very fine sand range. As is generally true of rocks containing both silt and sand fractions, the sand is better rounded than the silt. Both cross-stratification and graded bedding are rare. Most of these beds lack fossils, but a few contain assemblages dominated by large tumid thick-shelled brachiopods.

The Three Forks is intruded by andesite and diorite sills east of Milligan Creek. The sills plainly do not replace but displace the shale, for the stratigraphic interval between the Jefferson dolomite and the Lodgepole limestone increases to match the thickness of the sills. The shale is unaffected by the sills, except for slight hardening and reddening within a few feet of an intrusive contact; the upper siltstone is locally bleached and epidotized.

In the Mud Spring sector, the Three Forks is poorly exposed except for a short stretch in the center of sec. 1, T. 2 N., R. 1 E. Here, with a thickness close to 300 feet, the Three Forks includes:

	Thickness (feet)
Thick-bedded yellowish-brown siltstone	50
Shale, mostly olive gray with thick lenses of black shale, and lentils of red limestone coquina; a few orange siltstone layers near the top	150
Very thin bedded yellowish-gray and orange siltstone and limestone, with minor gray shale; crumpled and brecciated	100

This section, much like that in the Milligan Creek area, differs in having markedly less dark-green shale, more black shale, and fewer transitional siltstone beds between the medial shale and the upper siltstone.

It is unfortunate that the Three Forks interval in the N½ sec. 36, and NE¼ sec. 25, T. 3 N., R. 1 E., is almost unexposed, for the rare outcrops are of rock types unique in the formation, apparently recrystallized during emplacement of the neighboring large monzonite mass and its apophyses. In the center of NE¼ sec. 36, exposed in the bed of the largest northward-flowing stream, are a few feet of olive-gray to dark-greenish-gray impure chert or microcrystalline quartzite with splotches of grayish red and yellowish green, crowded with ovoids that are lighter in color and, being slightly more resistant, weather into seed-like lumps. Here and there slaty partings separate ovoid-bearing layers. The ovoids range in length from about 1/30 to 1/3 inch, but are mostly in the range 1/8 to 1/4 inch. They have nearly equal intermediate axes about half as long. In most of the ovoids the long axis is subparallel to bedding, with random orientation in the bedding plane. In most places the ovoids cut across fine details of bedding; in a few, bedding elements bend around the ovoids. Oolitic or ovulitic beds are unknown in unmetamorphosed Three Forks, and the ovoids are probably incipient crystals of some prismatic metamorphic mineral. The rock seems to be recrystallized and perhaps silicified quartzose shale.

Exposures in the Three Forks interval in sec. 25 are confined to a few outcrops of greenish-gray finely banded hornfels and a single thick, massive outcrop of pinkish-gray quartzite. The hornfels seems composed largely of microcrystalline clay minerals and calcite, the paper-thin banding reflecting varying proportions and grain sizes of the two main constituents; it was probably originally calcareous shale. The quartzite consists of subrounded to subangular quartz sand, with calcite in scattered grains and as intergranular cement. This rock is recrystallized orange sandstone or siltstone. Such rocks in thick beds are most common in the upper unit of the Three Forks, but this outcrop is near the base of the formation.

The V-shaped valley cut in folded Three Forks shale in the Hossfeldt Hills area is perhaps the most striking topographic feature in the quadrangle. Outcrops along the valley are sufficient to identify the formation, but not continuous enough for detailed measurements of the section, which is around 300 feet thick. The section seems much like that in the Mud Spring sector with two differences: the orange siltstone at the top is more clayey and the sequence is distinctly thinner; and a gray limestone sequence, thick and resistant enough to produce a sharp ridge, appears between the lower orange siltstone and the olive shale in the extreme northeast corner of the quadrangle (sec. 20, T. 3 N., R. 2 E.). This ledge-forming limestone characterizes the lower part of the Three Forks in the Logan area (Verrall)[9] and farther east (McMannis, 1955, p. 1397).

The Three Forks shale in the Willow Creek vicinity differs markedly from that in other sectors and is so variable that a "typical" section cannot be given. Here it is much thinner, being rarely more than 200 feet thick and in places, such as SE¼ sec. 3, T. 1 S., R. 1 W., less than 100 feet thick. The lower contorted orange siltstone and limestone unit is poorly developed throughout and is missing for long stretches; masses of thick-bedded orange and gray limestone with many solution breccia zones sporadically overlie the basal unit near the west edge of the quadrangle but do not appear in most of the sector. The olive and green shale sequences that dominate the formation elsewhere are reduced to thicknesses of as little as 20 feet in places and largely lack the dark fossiliferous limestone interbeds so common elsewhere. The upper siltstone unit is relatively thick, as much as 100 feet, near the west edge of the quadrangle but is thin or missing to the east near the Principal meridian, and it is distinctly more clayey and less quartzose than in other sectors.

For detailed measured sections of the Three Forks shale see pages 127, 130, and 134.

The lower boundary of the Three Forks, discussed previously, is obscure, but the upper boundary, the contact with the overlying Lodgepole limestone, is widely exposed especially in the Milligan Creek sector. In all exposures, dark thin-bedded Lodgepole limestone seems to lie conformably on the orange siltstone or sandstone of the upper part of the Three Forks. The only direct sign of hiatus is a little local channeling and grooving of the top of the Three Forks, to depths of perhaps an inch. The variable

[9] See footnote, p. 11.

thickness of the upper unit is probably due more to primary depositional factors than to post-Three Forks erosion.

NAME

The name Three Forks shale in this paper follows Peale's original usage, although limestone and siltstone are important in the sequence and justify the more general term "Three Forks formation," as first used by Haynes (1916b, p. 281) and subsequently by several others (such as, Sloss and Laird, 1946 and 1947; Andrichuk, 1951; Berry, 1943; McMannis, 1955). The general sequence within the formation as described here is identical with Peale's description as to lithology, but not as to thickness: present total measurements are consistently greater than Peale's, which total less than 150 feet. As noted before, difficulties in reconciling thickness measurements are largely responsible for wide divergences in published descriptions of relations in the upper part of the Jefferson and lower part of the Three Forks. Haynes' measured sections (1916b, p. 281), made a few miles east of the Three Forks quadrangle, give thicknesses of 222 and 287 feet, which are comparable to those of this report. The "seven fairly distinct lithologic divisions" he recognized in the region were not, however, identified in the Three Forks quadrangle. Despite the considerable difference in reported thickness and in details of description, Peale and Haynes chose the same, thoroughly practicable, boundaries for the Three Forks and these are the boundaries used here.

FOSSILS

Parts of the Three Forks shale are richly fossiliferous, and paleontological literature about the Three Forks is extensive. The stratigraphy and age of the upper part of the formation have, nevertheless, been in debate ever since Berry (1940) concluded that the upper siltstone is of Early Mississippian age and separated it, as the Sappington sandstone, from the wholly Devonian Three Forks shale (restricted). Extensive faunal collections were made for this report partly in an effort to illuminate if not to eliminate this controversy, which is discussed later. The reasonably well identified components of collections whose stratigraphic position is certain to within a few feet are listed in table 2. Collections Mil–1 through Mil–6 were made by R. J. Ross, Jr. in June 1953; collection 57MDu35 was made by J. T. Dutro, Jr. in July 1957; the remainder were made by me in the period 1952–54.

Brachiopods and cephalopods are the most abundant faunal elements. The brachiopods were studied by J. T. Dutro, Jr. (written communications June 16, 1958, and July 6, 1959). The cephalopods were studied by Mackenzie Gordon, Jr. (written communications August 22, 1955, and August 24, 1955). In addition, pelecypods, ostracodes, and fragments of crinoids and bryozoans are present but were not studied systematically for this report. Present, but rare, are conodonts, not further identified, and bits of primitive vertebrate jaws (in collection 23; see table 2, and pl. 2), identified as such by David Dunkle, U.S. National Museum (written communication, Jan. 26, 1955).

Table 2, used with plate 2, is largely self-explanatory, but a few comments may be helpful.

The interval represented by each collection is given in feet below the base of the Lodgepole limestone, and also in feet above (+) or below (−) the base of the upper siltstone unit if exposures permit; single numbers indicate an interval of 1 foot or less. Great significance should not be attached to either of these reference levels, as the upper one is at a slight disconformity and the lower is gradational, and thus rather subjective.

The best collections of Three Forks faunas, from the standpoint of stratigraphic control, are from the east side of the Mud Spring sector, in sec. 1, T. 2 N., R. 1 E., (loc. 252, pl. 2; the stratigraphic position of each element of collection 252, *a* through *i*, is shown in measured section B, p. 130.

In the strip of Three Forks shale on the west limb of the Hossfeldt anticline, a few of the rare outcrops provided excellent collections, mostly from the upper siltstone unit (collections 425, 426, 233, 244). The shale below the siltstone in this strip has yielded vast numbers of fossils in float, but few in place. An exception, the same locality that produced collections 425 and 426 from the siltstone, provided three collections—232, 351, and 424.

A few well-located collections have been made in the upper Three Forks shale of the Milligan Creek area. Only one locality west of Milligan Creek has yielded good material from both the upper siltstone and the medial shale units. This is locality 25, on the north bank of Dry Hollow (SE¼SW¼ NE¼ sec. 3, T. 1 N., R. 1 W.). Other localities upstream have yielded fossils only from the shale. Collection A–29, made from the south bank of Dry Hollow, downstream from locality 25 (W½ sec. 2, T. 1 N., R. 1 W.), is from the top 30 feet of unit A–29 in measured section A (see p. 127) 200–230 feet below the Madison and squarely in the middle of the Three Forks, which may be partly duplicated by faulting.

Just west of Milligan Creek, in sec. 36, T. 2 N., R. 1 W., are fine exposures of the upper siltstone unit, here 100 feet thick (see fig. 3, column 2). This is the type area of Berry's Sappington sandstone. The upper 60 feet yielded only a few specimens, collection 348, of one species, from 40–50 feet below the top. The lower 30 feet yielded the rather sparse collections 56, 55, 348, Mil–1, Mil–2, Mil–3, and 57MDu35. Collection Mil–2 is omitted from table 2 because it is contaminated by float; collection 56 includes the same interval but shares only a few forms. The exposures had been previously sampled, which may account for the relative paucity of fossils.

Collection Mil–4, from 12 feet of greenish-gray shale just below the yellow siltstone unit, is omitted from table 2 because of contamination by float; these strata, unfortunately, are not otherwise represented at this locality.

Collection 50c, NW¼SE¼ sec. 36, is included although comprised solely of float specimens; they must have come from beds no lower stratigraphically. The specimens of collection 50a were collected in place, but indiscriminately from both shale and siltstone. The same is true of collection 54, from 300 feet farther east.

The Three Forks shale in the vicinity of collections 81–83 from S½SE¼ sec. 30, T. 2 N., R. 1 E., is slightly silicified and the strata look a little different from their unaltered equivalents.

In the Willow Creek sector, the only good exposures of fossiliferous beds in the Three Forks are in S½N½ sec. 4, T. 1 S., R. 1 W., in a stretch about 500 feet long in the valley of the creek that flows northeastward across the section. The calcareous siltstone and nodular clayey limestone that yielded collections 284–287 are continuously exposed for a thickness of about 60 feet from the base of the Madison to the creek bed (this is the upper ⅔ of unit D–38, measured section *D*, p. 134). The strata are mainly orange, but a few near the base are black, gray, or reddish brown. More clayey and less quartzose than the upper siltstone unit in other sectors, they resemble the common upper siltstone in color, general texture, and apparent stratigraphic position. The base of the upper siltstone unit seems to be about 30 feet lower, to judge by the position of nearby outcrops of shale (D–37 in measured section *D*, p. 134.

Collection 286 is from a prominent zone, 3 to 6 feet thick, rich in potato-shaped limestone nodules that are 1–4 inches long and are reddish brown on the outside and gray inside. According to R. C. Gutschick (oral communication, 1958) these nodules are fossils, too; some are spongiostromatid algae, and others are isolated sponges of the class Calcispongea(?).

AGE

These collections contain both Late Devonian and Early Mississippian elements. Fixing the Devonian-Mississippian boundary with respect to either physical stratigraphy or morphologic paleontology is not, however, easy.

Collections regarded as assignable to the Early Mississippian include: 252i, 426, 425, 233, 244, 25c, 81, and 285. Without exception, they are from light-colored siltstone or fine sandstone. Most are from near the middle of the upper siltstone unit; none is less than 20 feet below the Madison and most are more than 20 feet above the base of uninterrupted siltstone.

Collections assignable to the Late Devonian include: 252g, 252f, 252d, 252e, 260a, 252c, 252b, 252a, 232, 351, 424, 25a, 23, 27, A–29, 56, 348, 55, 39, 50, 54, 83, 284, 286,[10] 287, 288, 435T6, 57MDu35, Mil–1, Mil–3, Mil–4.5, Mil–5, and Mil–6. Practically all the cephalopods are from dark shale, but the brachiopods are not closely confined to a particular rock type. Most of the Devonian brachiopod collections are from dark shale and associated limestone 60 feet to as much as 230 feet below the Lodgepole and below the sequence of uninterrupted light-colored siltstone, but several are from the upper siltstone (256, 284, 286, 287, 57MDu35, Mil–1, Mil–3), and others (252a, 50a, 54, Mil–6) are from siltstone or thinly interbedded siltstone and dark shale from well below the upper siltstone unit. An especially striking instance in which the same brachiopod assemblage occurs in two distinct lithofacies is in the type area of the Sappington sandstone of Berry, where collection 57MDu35 from the lower part of the orange siltstone unit contains many of the same species as collection Mil–5 from dark shale 50 feet lower.

According to R. J. Ross, Jr. (written communication, Nov. 30, 1954), several of the brachiopod genera and probably even some of the species are present in the Percha shale of New Mexico and the Ouray limestone of southwestern Colorado. Both of these formations are of latest Devonian age.

According to Mackenzie Gordon, Jr. (written communication, Aug. 22, 1955), the goniatite cephalopods listed are all members of a well-known Late Devonian Rocky Mountain fauna; all the known species are present. *Raymondiceras* and *Tornoceras* are not regarded as of value for precise dating, but the representatives of *Platyclymenia* are related to *Platycly-*

[10] R. C. Gutschick (oral communication, 1958) regards the fossil algae and sponges from the same stratum as of Early Mississippian (Kinderhook) age, but the age significance of such fossils is debatable.

menia (*Pleuroclymenia*) *crassa* Schindewolf, which occurs in the so-called *Prolobites-Platyclymenia* Stufe of the German Upper Devonian; two goniatite zones, (not represented in the Three Forks shale) the *Orthoclymenia* and *Wocklumeria*, lie between the *Prolobites-Platyclymenia* zone and the *Gattendorfia* zone which has been accepted by the Heerlen Congress as the base of the Lower Carboniferous in Europe.

In discussing the ages of the U.S. Geological Survey collections from the type area of the Sappington sandstone of Berry, NE cor. SW¼ sec. 36, T. 2 N., R. 1 E., Dutro (written communication, June 16, 1958) stated:

> There is a gradual change in the composition of the assemblages * * * up the section. Mil-7 represents the typical molluscan assemblage described by Raymond (1909). It is this fauna that has been correlated with the *Prolobites-Platyclymenia* zone of the German usage (approximately middle Famennian).
>
> Collections Mil-6 upward through Mil-4.5 contain about the same fossils and are representative of the "Three Forks fauna" described by Haynes (1916).
>
> In Mil-3, and Mil-1, the cyrtospiriferids appear to be progressively less like *C. whitneyi* and *C. monticola.* Two types of syringothyrid occur in Mil-1. *Nudirostra* is found in Mil-3. A different species of *Schuchertella*, from that found below in Mil-4.5 and Mil-6, appears in Mil-1.
>
> These changes are what I would expect to find in progressively younger Famennian faunas.
>
> Early Mississippian conodont faunas have been reported from shaly zones below the Lodgepole at several places in the northern Rockies. A similar situation may exist in or near Milligan Canyon. Fossils of undoubted Early Mississippian age do occur in the Sappington lithology nearby (colls. 25c and 81). Therefore, the Devonian-Mississippian boundary lies somewhere in the silty Sappington interval. Exactly where is of no particular importance.

The sparse faunas in collections 252h and 82 are insufficient to permit firm age assignment.

In all but one of the few places where both Mississippian and Devonian faunas have been recovered from the same locality, all of the Mississippian faunas overlie all the Devonian. Generally in such localities, several tens of feet of unfossiliferous rocks intervene between the lowest occurrence of definitely Mississippian faunas and highest occurrence of Devonian faunas.

An exception to the customary sequence appears in the Willow Creek area, where Mississippian collection 285 was recovered from beds 20–25 feet below Devonian collection 284. This reversal of the customary faunal sequence was noted in the field, confirmed in preliminary paleontologic reports, and checked in the field by several of the paleontologists who are cited. Conceivably, the inversion is due to faulting, but there are no overt signs of deformation, nor, more significantly, of the repetition of specific identifiable beds. The strata seem indeed to be in the sequence of deposition. The many gaps in the Three Forks section may conceal similar alternations elsewhere in the quadrangle.

The anomalous position of collection 284 is possibly the result of reworking of Devonian fossiliferous strata by Mississippian ocean currents. The fossils themselves, however, show no signs of special wear, either in surface abrasion or in degree of disarticulation.

A better possibility is that the alternation of faunas reflects an alternation in primary depositional conditions during a time when earliest Mississippian and latest Devonian faunas coexisted, each adapted to a slightly differing environment. There are several known instances of alternating faunal zones where biota sensitive to lithofacies—"facies fossils"—are involved; there is no reason why they should not occasionally develop at conformable systemic boundaries. Under such circumstances, the systemic boundary might be selected at the first (lowest) appearance of the "younger" fauna, the last (highest) appearance of the "older" fauna, or regarded as a zone between these two limits.

If it should be argued that collection 284 is not really Mississippian but Devonian, or that collection 285 is not Devonian but Mississippian, the fact of faunal alternation is not changed but the matter becomes another possible case of facies control within a system.

Whatever is made of the relations in the Willow Creek sector, the paleontologic evidence presented for the quadrangle as a whole shows that the Devonian-Mississippian transition occurs high within the Three Forks shale. In the few places where exposures and fossil collecting are both fairly good, the transition seems to occur well above the base of the upper siltstone. If the siltstone grades laterally into dark shale, however, as seems likely, there must be places where the transition is accomplished wholly within shale and others where the lithologic boundary coincides with the time boundary.

THE SAPPINGTON PROBLEM

Difficulties with the exact placing of the Devonian-Mississippian boundary in southwestern Montana have been recognized and discussed ever since the pioneer work of Peale (1893). Usually, questions about the age of a formation arise from too little paleontologic evidence, but in the Three Forks shale the problem stems partly from an embarrassment of fossil riches. Peale (p. 31–32) noted that the prolific fauna in the upper part of the Three Forks shale contains many Upper Devonian organisms and a few Carboniferous ones. This led him (1896, p. 2) to conclude that "the shales are near the border line of the Devonian and Carboniferous * * * but the predominance of evidence is in favor of their Devonian age."

The first workers who followed Peale recognized the age problem but were not troubled by it. Raymond (1907, 1909) clarified and delimited the unspecified "mingling" of Devonian and Carboniferous forms noted by Peale. Raymond demonstrated that the "mingling" is practically confined to a few tens of feet of poorly exposed yellow sandy and silty beds at the top of the Three Forks, and interpreted the relations to mean simply a transition from Devonian to Mississippian high within the established Three Forks shale. Haynes (1916a, p. 27–28) confirmed this stratigraphic and faunal transition.

The age of the upper part of Peale's Three Forks became a "problem" when Berry (1943, p. 14–16) raised the "yellow sandstone" at the top to formation rank, on the ground that it "is both lithologically and faunally a recognizable stratigraphic unit" of Early Mississippian, probably Kinderhook, age, and named it the Sappington sandstone, with a thickness of 60 feet at the type locality (NW¼ sec. 36, T. 2 N., R. 1 W., according to Berry, but the outcrops he described are actually in NE¼ SW¼) in the Three Forks quadrangle. Berry did not describe a type section. The fossil evidence cited is a Mississippian *Syringothyris* fauna, confined to the yellow sandstone, which has no species of the Devonian *Cyrtospirifer* fauna so abundant in the green shale below. He noted the possibility of a transitional fauna in 30 feet of "argillaceous limestone and carbonaceous shale" (his fig. 4, p. 17) between the base of the *Syringothyris* zone and the top of the *Cyrtospirifer* zone, but assigned these intervening beds to the Three Forks. Although it was not possible to do so without gross exaggeration, Berry separated the Sappington on his geologic map (pl. 1, scale about 1:125,000).

Several other local studies in and near the Three Forks area (Holland, 1952, p. 1704–1790; McMannis, 1955, p. 1396–1399; Verrall[11]) followed Berry in recognizing a separate Sappington sandstone of Mississippian age, though each introduced modifications, and the formation and its implications were accepted in the regional synthesis of Crickmay (1952, p. 588–589).

Holland (1952) interpreted the Sappington to be about 100 feet thick in the type locality (fig. 3, column 1). He not only found *Syringothyris* in orange siltstone more than 30 feet lower than any found by Berry, and below the 30 feet of "argillaceous limestone and carbonaceous shale" assigned to the Three Forks by Berry, but also discovered some decorticated specimens of *Rhipidomella* "believed to be *R. missouriensis* (Swallow)" and of Mississippian age in his opinion, near the base of a sequence of nodular rubbly yellowish-gray calcareous sandstone about 10 feet lower than the lowest occurrence of *Syringothyris*. Below the presumed Mississippian *Rhipidomella*-bearing sandstone is a thick covered interval with scattered outcrops of orange siltstone; rarely exposed within this interval is brownish black shale, a few feet thick. Holland (1952, p. 1705 and figs. 5 and 6) placed the base of the Sappington and the base of the Mississippian as well below this rarely seen shale, a choice of minimum usefulness for mapping, whatever its paleontologic virtues. Holland (1952, p. 1707) reported the yellow and orange sandstone at the top of the Sappington to be overlain with erosional unconformity by the gray crinoidal limestone of the Lodgepole.

If Holland's identifications and age assignments are accepted, he found Mississippian forms below some of the assemblages reported here as Devonian (collections 56, Mil–1, and 57MDu35, on fig. 3). This applies even if the decorticated *Rhipidomellas* are disregarded. Holland himself found no *Cyrtospirifer* faunas above *Syringothyris* faunas, but would not change his selection of a boundary if such were found, on the premise (paraphrasing Caster) that "it is the appearance of the new elements, not the last vestige of the old, that dates the strata" (Holland, 1952, p. 1709–1710).

McMannis (1955) accepted the brownish-black shale as the base of the Sappington, but extended the Sappington of Holland upward to include a few feet of conodont-bearing black shale which is widespread in the region (Sloss and Laird, 1946; 1947, p. 1411–1418). Insisting on the persistent presence, not only in the Bridger Range but throughout the "type Three Forks region" of black shale between beds typical of the Sappington and the "typical Three Forks of Haynes," and claiming a sharp faunal break marked also by this black shale, McMannis denied either a lithologic or faunal transition between the Three Forks formation and his Sappington sandstone. Somewhat surprisingly, however, he did not insist on a Mississippian age for the Sappington, recognizing the possibility that "these beds transcend the systemic boundary."

The idea of a sharp faunal and lithologic break between the siltstone of the Sappington and the underlying shale of the main body of the Three Forks, independently advanced by McMannis, was first proposed on a regional basis by Crickmay (1952). He concluded that the Three Forks shale near Three

[11] See footnote, p. 11.

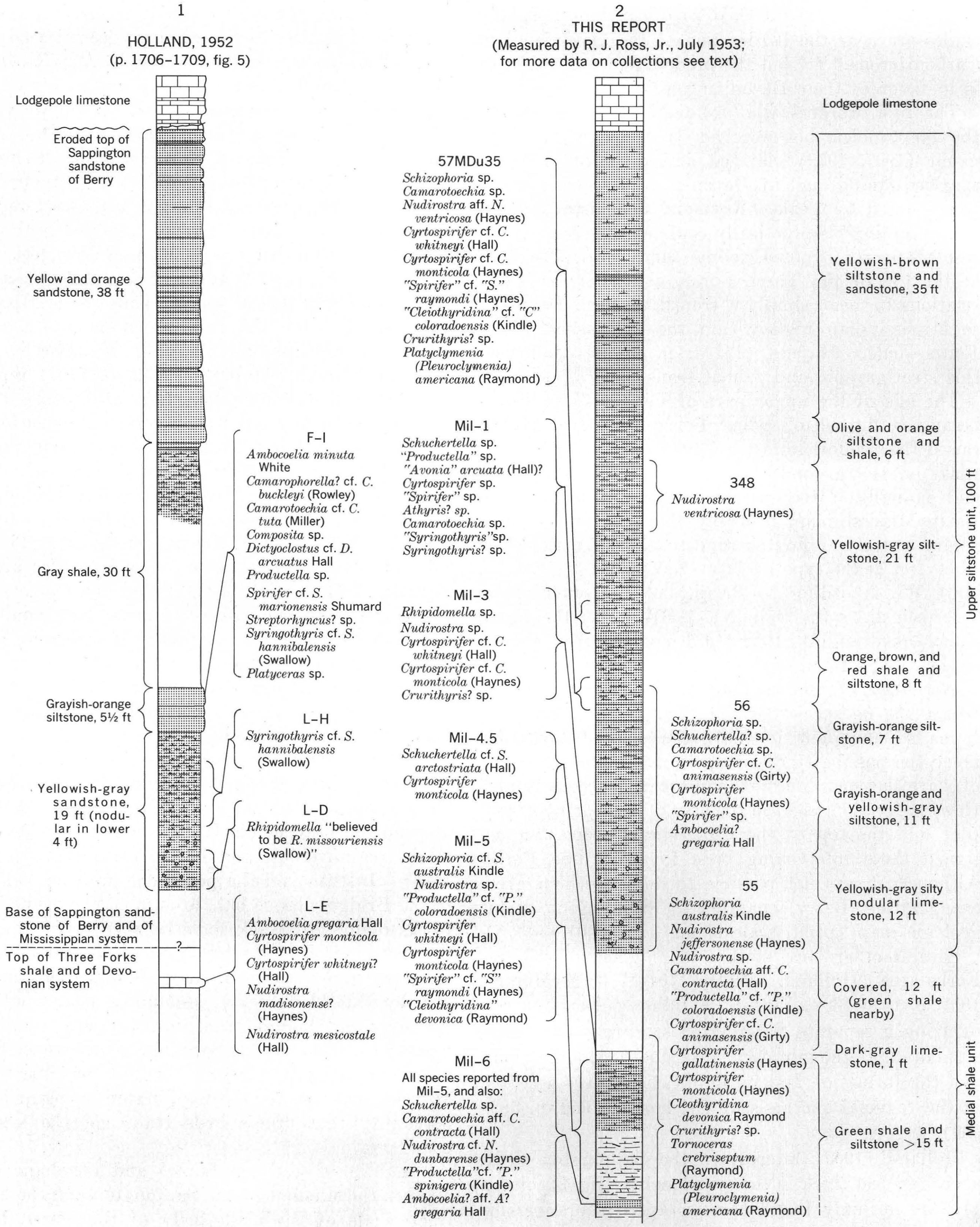

FIGURE 3.—Measured sections and fauna in upper part of Three Forks shale, NE cor. SW¼ sec. 36, T. 2 N., R. 1 E. (type locality of Sappington sandstone of Berry).

Forks does not carry a *Cyrtospirifer* fauna at all, but that the forms commonly identified as such are species of the superficially similar genus *Cyrtiopsis*, distinctly younger than the *Cyrtospirifer* fauna but wholly older than the *Syringothyris* fauna of the Sappington, from which it is separated by 1 to 19 feet of black shale without megafossils (Crickmay, 1952, p. 588). Crickmay assigned to the Mississippian all the rocks above the black shale which are similar to the Sappington, and implied that the black shale and possibly even some of the beds below it are also Mississippian. Later (Crickmay, 1956, p. 57–58) he specifically correlated this black shale with the Exshaw shale of Mississippian age in Alberta, but favored a Devonian age for the strata below.

A different interpretation is that of Sloss and Laird (1947, p. 1411), who said that "it can be demonstrated that the normal shale facies of the Three Forks, bearing the typical *Cyrtospirifer* fauna, is interbedded with the sandy facies bearing *Syringothyris*," and that the sandy facies grades laterally into typical shale of the Three Forks formation. These writers considered the "Sappington sandstone" to be a "local member of the Three Forks formation" and assigned it an unequivocal Late Devonian age, despite the Mississippian aspects that had impressed all previous workers. Such an assignment had, however, been foreshadowed by the hesitations of the committee that prepared the Devonian correlation chart (Cooper and others, 1942): they assigned the "Three Forks with *Syringothyris*" an equivocal Devonian or Mississippian age and remarked (p. 1785) that this zone may ultimately be assigned definitely to the Devonian. The opinions of Sloss and Laird as to the age of the Sappington have been reiterated by Sloss and Moritz (1951, p. 1252), Andrichuk (1951), Mann (1954, p. 11), Alexander (1955, p. 45–49), and others.

In the Three Forks quadrangle, an alternation of orange siltstone and sandstone beds of "Sappington type" with green shale and dark limestone beds of "Three Forks type" is unquestionably present, as Berry and Holland also recognized. Black shale intervenes locally between the orange siltstone and green shale but its persistence has not been demonstrated, despite the contrary claim of McMannis and Crickmay. Although alternation of lithofacies is proved, accompanying faunal alternation of the sort described by Sloss and Laird has not been confirmed. The anomalous relations west of Willow Creek lend support, but the fact that "*Cyrtospirifer* faunas" seem to have tolerated both muddy and silty environments does not. Whether or not a faunal alternation occurs and regardless of the exact ages of the faunas, the sequence of clastic orange siltstone, too thin to map by itself at the scale of this report, is better mapped with the unresistant, dominantly clastic Three Forks shale, with which it grades, than with the ledge-forming Lodgepole limestone.

The physical stratigraphy of the Three Forks shale in the Three Forks quadrangle thus supports the descriptions of Sloss and Laird. Whether or not they are correct about alternation of faunas closely controlled by lithofacies, however, their conclusion seems untenable that the entire formation is of Devonian age. In this matter the present work confirms the majority opinion, dating from Raymond and Haynes, that the Devonian-Mississippian boundary is high within these rocks, though just where is uncertain.

The Devonian-Mississippian transition at the type locality of the Sappington sandstone of Berry is within or possibly even at the top of the sequence of yellow siltstone, to judge by Dutro's interpretation (quoted earlier) of the Geological Survey collections reported on table 2 and figure 3. He does not accept any of Holland's or Berry's Mississippian age assignments. The crux of the matter is the time significance of *Syringothyris*. Dutro rejects the idea that "*Syringothyris* means Mississippian." In taking this position he is not merely considering the recorded range of the genus, which of course has long been recognized as Late Devonian to Mississippian. Rather, he opposes the view that any single species or genus can be meaningfully chosen as a systemic index, and bases his ideas of age on an evaluation of the whole fauna. The Devonian-Mississippian transition in this locality is placed by Dutro 60 to 100 feet above the boundary selected by Holland, and 20 to 60 feet above that chosen by Berry. The top of his zone of transition coincides with the boundary proposed by Sloss but only because fossils are lacking at this locality, for, on evidence from other nearby collections, he would expect the transition to be somewhat lower.

Despite all the attention given it, the inconsequential "Sappington problem" is about the same as when Haynes left it in 1916.

Cartographically, the concept of the Sappington sandstone, whether as a Mississippian formation or a member of a Devonian formation, is of little value in the Three Forks quadrangle. This is not to minimize the potential usefulness of the unit for mapping in other areas, however, particularly where the orange siltstone is thick and is set off by black shale, for beds lithologically similar to those of the Sappington near the Devonian-Mississippian boundary are now known from a rather large area: they appear as far west as the middle of the Jefferson Island quadrangle (shown me by W. R. Lowell), though not as far west as Philipsburg, where the entire interval of the Three Forks is absent (Emmons and Calkins, 1913, p. 65), nor, apparently, as far southwest as the Blacktail and Snowcrest Ranges (Klepper, 1950, p. 60); at least as far south as the Gravelly Range (Mann, 1954, p. 11) and perhaps even into northern Utah (Holland, 1952, p. 1719–1723), but the rocks of this interval have not been traced across intervening Idaho; as far east as the Bridger Range but seemingly not as far southeast as Yellowstone Park (Hague, Weed, and Iddings, 1896, p. 4); at least as far northwest as the Helena quadrangle (Adolph Knopf, oral communication, 1957) but not as far due north as the Canyon Ferry quadrangle (Mertie, Fischer, and Hobbs, 1951; p. 26). The impression is that rocks of Sappington lithology are somewhat less extensive than are the green shales, and that where limestone dominates the formation the upper siltstone is likely to be absent.

The thin black shale that generally intervenes between the Three Forks and the Lodgepole in central Montana and has been designated the Little Chief Canyon member of the Lodgepole limestone by Knechtel and others (1954) does not appear at any place where the contact is exposed in the Three Forks quadrangle. The shale is not reported by Berry (1943) or Alexander (1955) in the adjoining Jefferson Island quadrangle, or by Klepper and others (1957) in the Elkhorn Mountains to the northwest, or by Mann (1954) in the Gravelly Range to the south, and I have not seen it in the Toston quadrangle to the northeast. It crops out as near by as Logan, however, according to Verrall[12] and in the Bridger Range (McMannis, 1955, p. 1397), though only a few feet thick. This shale may have been stripped off the orange siltstone by pre-Lodgepole erosion, or it may merely wedge out near the Missouri River.

ORIGIN

Most of the Three Forks seems to be of shallow, nearshore marine origin. The dark shale with pyrite, in the form of crystals and fossil replacements, suggest sharply restricted communication with the ocean. The basal crumpled and brecciated thin-bedded mudstone and limestone, with their bright colors, appear to be residual from originally much thicker evaporite sequences, deeply leached, as recognized by Sloss and Laird (1946) and discussed by Andrichuk (1951). Though superfically similar to the basal evaporites, the upper orange and yellow siltstone and sandstone probably formed under much different conditions, presumably close to shore in shallow, well-aerated waters in free communication with the ocean. Very likely part at least of this nearshore facies was deposited contemporaneously with and grades into offshore dark shale, as Sloss and Laird (1947) concluded.

The dominance of free-swimming cephalopods among pyritized fossils in the dark shale beds, whereas brachiopods are the overwhelmingly dominant nonpyritized forms elsewhere in the formation, suggests that the cephalopod faunule is a thanatacoenose or death assemblage, for living mobile cephalopods would scarcely inhabit sulfide-rich muds, whereas the brachiopod faunule may approach a biocoenose or life assemblage (Wasmund, 1926).

The source of the clastic components of the Three Forks is hard to visualize. It may have been a southern extension of the uplift postulated by Deiss (1933) in parts of the Swan, Flathead, and Lewis Ranges, to account for a hiatus between the upper part of the Jefferson and lower part of the Madison. An eastern source is not out of the question, as Devonian rocks are absent in southeastern Montana and adjacent parts of Wyoming, and seem never to have been deposited there (Sloss, 1950, p. 439, fig. 7). Low relief of a possible eastern source is suggested by the dominance of nonclastic rocks in the Devonian sequence to the shoreline (Sloss, 1950).

CARBONIFEROUS

Rocks mapped as Carboniferous, divided into four map units, are 1,600 to 3,500 feet thick. They comprise about 80 percent limestone, 10 percent siltstone and sandstone, 5 percent dolomite, and 5 percent shale; the clastic ratio is less than 0.2. The boundary between the Mississippian and the Pennsylvanian is not yet clear in western Montana. The two great limestone formations, Lodgepole and Mission Canyon, of the Madison group near the base of the Carboniferous can safely be assigned to the Mississippian, the Quadrant formation at its top is assigned to the Pennsylvanian (though it may in part be as young as Early Permian) and the Mississippian-Pennsylvanian transition comes within the rocks here mapped together as the Big Snowy and Amsden formations.

The Carboniferous of western Montana, though not as thoroughly studied as the early Paleozoic, has been approached on a regional scale in three papers: the Madison group is treated by Sloss and Hamblin (1942); the Big Snowy group by Perry and Sloss (1943); and the Big Snowy, Amsden, and Quadrant by Scott (1935). None of these studies is both detailed and comprehensive, and none has much to say about western Montana, as their emphasis is on

[12] See footnote, p. 11.

oil-productive central Montana. Despite evident possibilities for useful contributions in this part of the geologic column, the Carboniferous rocks are cursorily treated here: the bulk of strata in the interval is limestone of the Madison group with monotonously uniform texture and composition, that discourages detailed scrutiny by ordinary methods; and the remaining interbedded clastic and carbonate rocks are on the whole too poorly exposed to justify elaborate treatment. The data at hand, though meager, lead to some thoughts that may be of interest on certain aspects of correlation and origin.

NAMES AND GENERAL CORRELATION OF CARBONIFEROUS FORMATIONS

Peale (1893, p. 32–33) assigned to two formations the rocks he recognized as Carboniferous, which at that time included the Permian: the Madison formation, consisting wholly of limestone, 1,250 feet thick; and the overlying Quadrant formation, consisting of interbedded limestone and quartzite, "the latter predominating at the top," 300–400 feet thick. "A prominent bed of quartzitic sandstone," 40–60 feet thick, capping his Quadrant he assigned to the Juratrias Ellis formation, though recognizing its possible Carboniferous age (Peale, 1896, p. 2). From his brief but clear descriptions, it is plain that he recognized all the rock types found by later workers, though the thicknesses he reported are puzzlingly small, but the scale of his mapping did not allow further subdivision.

The Madison was raised to group rank and divided into two formations, a lower thin-bedded unit named the Lodgepole limestone and an upper thick-bedded one named the Mission Canyon limestone by Collier and Cathcart (1922, p. 173), the usage now commonly followed throughout the region and in this report. The Lodgepole was further subdivided into a lower shaly member, the Paine, and an upper wholly limestone member, the Woodhurst, by Sloss and Hamblin (1942, p. 313–318); these members can be identified locally in the Three Forks quadrangle but not in enough places to permit mapping.

Although Peale was the first to use Quadrant formation in print (1893), the term originated with Weed contemporaneously working in Yellowstone Park. Weed restricted the name to post-Madison, pre-Permian rocks. Not recognizing any Permian in his area, Peale included the entire post-Madison pre-Mesozoic sequence in his Quadrant. Condit (1918) traced the Permian Phosphoria formation (Richards and Mansfield, 1912, p. 683–689) into the Three Forks region, and restricted the Quadrant there to rocks fitting Weed's original definition.

Later, Scott (1935) proposed radical changes of nomenclature for the beds included in Weed's Quadrant. He limited the Quadrant to the quartzitic sandstone beds just below the Phosphoria, correlating the restricted Quadrant with the Tensleep sandstone (Darton, 1904, p. 394–401) of Pennsylvanian age. The beds below the restricted Quadrant, and above the Madison, he subdivided into four conformable formations (in ascending order): Kibbey sandstone (Weed, 1899b), Otter shale (Weed, 1892, p. 307), Heath shale (new name), and Amsden formation (Darton, 1904, p. 394–401), all of late Mississippian (Chester) age. The three lower units were placed in the Big Snowy group, confined to central Montana, with the Amsden separately classified, presumably because of its differing, much wider, distribution. The stratigraphic proposals of Scott represent a distinct advance. His Big Snowy, Amsden, and Quadrant have been widely accepted and are used in this report, although it could be argued that the name Quadrant has lost its value through modifications, and should be abandoned.

The "red limestone" in the lower part of Peale's Quadrant (1896, p. 2) is clearly the Amsden of Scott. In most of the area studied by Peale, including much of the Three Forks quadrangle, the Amsden formation lies directly on the Madison group, and the Big Snowy is absent. In parts of the Three Forks quadrangle, however, rocks typical of the Amsden are underlain by as much as 400 feet of strata of the Big Snowy group. The beds contain fossils characteristic of the upper part of the Big Snowy, but lithologically they are only vaguely similar to type Big Snowy units and are not subdivided into established formations; consequently, they are designated Big Snowy formation rather than group. Because the Big Snowy is generally thin, and its contact with the Amsden formation hard to locate in most places, the two units are combined on the geologic map.

MADISON GROUP

LODGEPOLE LIMESTONE

Assigned to the Lodgepole limestone of the Madison group is a thick succession of thin-bedded limestone and silty limestone. Characteristically, the Lodgepole outcrops begin with one or two ledges, 10 to 20 feet high, succeeded by a long grassy valley slope, with only scattered outcrops; the slope ends in a great ledge, as much as 100 feet high—the base of the very thick bedded Mission Canyon limestone. Best exposures of the formation as a whole are on the north side of Dry Hollow in secs. 2 and 3, T. 1 N., R. 1 W., and on the east side of Milligan Creek, just below the canyon in sec. 36, T. 2 N., R. 1 E.

In contrast to the older limestones, the Lodgepole has rather constant thickness, as well as it can be determined, and constant gross lithology throughout the quadrangle. Thickness in the Milligan Creek and Mud Spring areas is in the range 600–700 feet; widely differing apparent thicknesses, from as little as 200 feet to as much as 700 feet, measured at close intervals in the Hossfeldt Hills and Willow Creek vicinities suggest unrecognized near-bedding faults and make estimates of average thickness hazardous, but it is perhaps not far from 600 feet in each sector. The main rock type is dark-gray microcrystalline limestone, in beds 4 inches to 2 feet thick, that weathers yellowish gray to light olive gray. More coarsely crystalline beds or zones are rare near the base but are progressively more prominent upsection; many of these coarser phases are medium gray and olive gray, have sugary textures suggestive of clastic fabric, and a fresh petroliferous odor. Good casts and molds of brachiopods and solitary corals are widespread in the microcrystalline limestone, and fragments of brachiopods, corals, crinoids, and bryozoans are abundant in the sugary zones. Partings of yellowish-brown and grayish-orange calcareous mudstone, having sharp contacts with the limestone, are abundant in the lower 200 feet or so, becoming fewer but thicker upsection (though nowhere more than a few inches thick). Mudstone seems to be somewhat more common in the Milligan Creek and Mud Spring areas than in the other two pre-Tertiary sectors. In the upper 100 feet, a few very thick-bedded sequences, 10–20 feet thick, appear. For crudely measured sections, see pages 127, 130, 132, and 134–135.

In the northeastern part of the Mud Spring area (secs. 24 and 25, T. 3 N., A. 1 E.), the Lodgepole has been altered as a result of the intrusion of a nearby monzonitic pluton and its apophyses. Most of the altered rocks are merely bleached nearly white and recrystallized to a coarse fabric approaching marble. Rarely exposed in shallow gulches near the north edge of sec. 25, however, are variably coarse calcsilicate rocks of varied composition, permeated with ramifying hornblende diorite dikelets. The commonest type is very coarse, with crystals as much as 2 inches across of white and pale-pink calcite intimately intergrown with grayish-olive to dark-yellowish-brown idocrase and subordinate similarly colored garnet. On casual inspection, the idocrase looks much like garnet but it has many envelope-shaped crystals, with well-developed bipyramids separated by narrow four-sided prisms (resembling a common zircon form); identity of the mineral with idocrase has been confirmed in thin-section and by oil immersion methods. In a few places these rocks have green stains suggesting copper salts, but have no readily recognized copper minerals.

A rare, and confusing, calcsilicate type is a dense hard fine-grained yellowish-gray rock studded with equant bluish-black crystals 1/10 to 2/10 inch across. The large crystals look like spinel, and the general association leads to the idea that the rock is an altered dioritic dikelet. The rock, however, actually consists of large zoned garnets, brown in thin section, in a fine groundmass of colorless garnet-diopside, and minor amounts of wollastonite, allanite(?), and a micaceous mineral; it presumably is a contact-altered sedimentary rock.

These rocks, from close to the base of the Lodgepole, probably represent original silty strata that have formed new mineral combinations without additions of anything but heat and perhaps water from the intrusions.

Much of the Lodgepole resembles thin-bedded parts of the Meagher and Pilgrim formations. Distinction is readily made on the basis of fossils, which can usually be found even in small exposures of the Lodgepole and recognized even by novices in paleontology: cup corals, segments of crinoid stems, and a varied brachiopod community conclusively indicate that a questionable outcrop is Lodgepole rather than a Cambrian limestone.

The Lodgepole, as noted earlier, overlies the Three Forks shale with seeming conformity, though the absence of correlatives of the Little Chief Canyon shale member suggest a brief hiatus. Its upper contact with the Mission Canyon limestone is conformable and transitional, being placed where very thick bedded sequences begin to dominate the section. In a few places, such as the mouth of Milligan Canyon, the transition is accomplished within a few feet, but in most places the combination of partial exposures and minor thick-bedded sequences within the upper part of the Lodgepole makes it difficult to place the contact closely. Consequently a gradational contact is drawn throughout. Some of the difficulties with the thickness of the Lodgepole may be due, not to structure as suggested above, but, to unconscious changes in the choice of the upper boundary, or real original variations in the stratigraphic level at which thinner bedded strata give way to thicker. This point could perhaps be resolved if the thickness of the Mission Canyon limestone could be measured with fair accuracy and its changes compared with those of the Lodgepole but this, as shown below, is not feasible.

AGE AND ORIGIN

The Lodgepole limestone is a typical marine carbonate, laid down in rather shallow water (sublittoral to shallow neritic). Its thin but even stratification suggests a gently oscillating sea. It is largely of chemical origin, but includes some clastic layers, probably reworked from nearby parts of the Lodgepole itself. The fact that mudstone interbeds are fewer but thicker upsection, whereas thick-bedded limestone is increasingly common, suggests a gradual slowing of the rate of oscillation, no doubt accompanied by slow reduction of borderlands, rather than shoreline retreat. The same data, considered along with the decrease in proportion of clastic limestone and in numbers of fossils upsection, suggest increasing average water depth during Lodgepole time.

The Lodgepole is highly fossiliferous, so that systematic collecting could have produced fossil lists of staggering length. However, the fauna is thoroughly known, and the age of the formation is established as Early Mississippian. Five collections, taken from various levels merely to confirm the stratigraphic assignment and age, and examined by J. E. Smedley (written communication, May 28, 1954) are listed below:

1. From uppermost 25 feet of the formation in Milligan Canyon [16430–PC]:

   Camarotoechia sp. indet.
   Spirifer centronatus Winchell
   Schuchertella cf. *S. chemungensis* (Conrad)
   Chonetes loganensis Hall and Whitfield
   Productus gallatinensis Girty
   Avonia? sp. indet.
   Spiriferina? sp. indet.
   Dielasma sp. indet.
   Platyceras sp. indet.

2. From 15–20 feet below top, south of Jefferson River in SW¼NE¼ sec. 3, T. 1 S., R. 1 W. [16434–PC]:

   Fenestella sp. undet.
   Spirifer centronatus Winchell
   Schuchertella sp. indet.
   Leptaena? sp. indet.
   Chonetes cf. *C. loganensis* Hall and Whitfield
   Schizophoria compacta Girty
   Spiriferina sp. indet.
   Platyceras sp. undet.
   Low spired gastropod frag. gen. indet.

3. From the middle Lodgepole in Milligan Canyon [16432–PC]:

   Camarotoechia metallica White
   Axiodeaneia sp. indet.
   Spirifer centronatus Winchell
   Torynifer cooperensis (Swallow)
   Composita madisonensis (Girty)
   Chonetes loganensis Hall and Whitfield
   Linoproductus sp. indet.
   Platyceras sp. undet.
   Proetid trilobite frag. gen. undet.

4. From the middle Lodgepole in center, sec. 36, T. 3 N., R. 1 E. [16433–PC]:

   Fenestella sp. indet.
   Camarotoechia metallica White
   Spirifer centronatus Winchell
   Composita madisonesis (Girty)
   Chonetes loganensis Hall and Whitfield
   Cleiothyridina aff. *C. incrassata* (Hall)
   Leptaena analoga (Phillips)
   Spiriferina cf. *S. solidirostris* White
   Orthotetes sp. indet.
   Platyceras sp. indet.
   Straparolid gastropod gen. indet.
   Proetid trilobite gen. indet.

5. From 20–25 feet above the base of the formation in Dry Hollow, center S½NW¼ sec. 3, T. 1 N., R. 1 W. [16431–PC]:

   Fenestella sp. indet.
   Schizophoria sp. indet.
   Schuchertella? sp. indet.
   Chonetes loganensis Hall and Whitfield
   Chonetes logani Norwood and Pratten
   Spirifer centronatus Winchell
   Spirifer striatus var. *madisonensis* of Girty 1899
   Straparollus sp. indet.
   Platyceras sp. indet.
   Paracyclus sp. indet.

MISSION CANYON LIMESTONE

The Mission Canyon is the last of the mountain-forming Paleozoic limestones. Great ledges carved from its basal beds form the highest and boldest ridge crests throughout the quadrangle, though they are only slightly higher than many ridges formed by basal beds of the Jefferson dolomite. Where dips are moderate, most of the Mission Canyon, like the Jefferson, is poorly exposed, and forms long grassy dip slopes broken by a few low ledges. The formation is magnificently exposed where dips are very steep, as in the Willow Creek area east of Ingleside Quarry and the eastern part of the Hossfeldt Hills. The most accessible continuous exposures are in Milligan Canyon and in the southeast-draining canyon in SW cor. sec. 29, T. 3 N., R. 2 E.

Throughout the quadrangle the Mission Canyon, like the Lodgepole, is lithologically constant. The main rock type is gray microcrystalline limestone much like that in the Lodgepole, but lighter in color. The Mission Canyon contains a few beds of medium-grained crystalline sugary limestone, like that in the Lodgepole, mostly near the base, but lacks mudstone layers and is but rarely fossiliferous.

The only striking compositional difference between the two formations of the Madison group is the abundance of chert in false-nodules, nodules and lentils in the Mission Canyon, especially the upper half. Most of the chert is grayish black to olive black, but some is much lighter, in the range of pale yellowish browns and grays. The nodules tend to be almond shaped and an inch or two long. Their long axes are usually parallel to bedding, and some grade into lentils as much as 5 feet long. The structures called false-nodules look like ordinary nodules, but consist of a thin skin of chert surrounding lumps of limestone indistinguishable from the country limestone. Gradations from false-nodules to true do not occur, and the two types seem to be of different origin. The general richness of the formation in silica is apparent on weathered surfaces, which in many places have thin rough crusts of finely crystalline quartz, easily mistaken for chert.

The distinguishing feature of the Mission Canyon is its thick, commonly indistinct, bedding. Individual beds 3–5 feet thick are the rule, and beds more than 10 feet thick are numerous; sporadically, however, thin-bedded sequences appear.

The apparent thickness of the Mission Canyon varies widely within individual pre-Tertiary areas. In the Milligan Creek area west of Milligan Canyon the formation is rather consistently 600–700 feet thick, whereas in seemingly undisturbed sections only 2 miles east of the Canyon measured thicknesses are 1,000–1,100 feet. In Milligan Canyon itself and for more than a mile to the northeast, the formation is so thickened by a combination of cross faults and unmappable near-bedding faults that measurements are precluded.

In the 3-mile strip of continuous Mission Canyon on the east flank of the Hossfeldt anticline, the apparent thickness varies rapidly, ranging between 700 and 1,000 feet. In the Willow Creek vicinity, the apparent thickness also varies widely, from 1,000 feet to more than 1,400 feet, within another 3-mile continuous strip without mapped faults. The problem of thickness variation does not arise in the Mud Spring sector for the full Mission Canyon is nowhere exposed; it is at least 700 feet thick in the trough of the syncline east of Dunbar Creek.

For crudely measured sections see pages 127 and 132.

The great local variations in thickness may be ascribed largely to erosion or to known but unmappable deformation. Rocks of the Big Snowy formation are missing west of Milligan Creek, where the Amsden lies on deeply eroded Mission Canyon, but the full Mission Canyon section is preserved beneath the Big Snowy east of Milligan Creek. No such explanation will serve for the Hossfeldt Hills and Willow Creek sectors, where the Big Snowy formation overlies the Mission Canyon throughout; in these areas, near-bedding faults, which are visible in many transverse canyons but of unknown displacement and untraceable beyond the canyon walls, are a possible explanation.

In the upper two-thirds of Milligan Canyon, where the canyon trends north-northeast, a thick zone of limestone breccia within the Mission Canyon limestone is spectacularly exposed (fig. 4). The breccia is an unresistant interval that here controls the course of Milligan Creek. The stratigraphic thickness of the zone changes abruptly from less than 10 feet to more than 40 feet; these changes largely reflect irregularities in the top, for the base seems to stay within narrow stratigraphic limits about 200 feet below the top of the formation. The zone is loosely filled with unsorted angular to subangular fragments of limestone, which range from granule to giant-boulder size but are mostly in the cobble to small boulder size. Most of the breccia is cemented by calcareous siltstone and claystone, colored orange or, rarely, brownish red; in some places, spaces between stones are open, or lined with coarsely crystalline calcite. Rarely, spaces that seem to have been tube-like, of unknown, possibly great length along the dip but only a few feet wide and high, are filled with laminated to very thin bedded calcareous siltstone like the matrix material of the breccia.

This breccia zone continues east of the canyon, where it dominates the long arm of Mission Canyon rocks east of Milligan Creek (NW¼ sec. 25, T. 2 N., R. 1 W.), but few signs of it appear farther to the east, though exposures are fairly good. To the west, the breccia zone can be followed to about the center

FIGURE 4.—Limestone breccia in upper Mission Canyon limestone, Milligan Canyon. View at right angle to strike of formation; dip 35° away from viewer. Note partly filled triangular cavity above man, and laminated siltstone layers above him to his right.

sec. 35, T. 2 N., R. 1 W., where it seems to disappear against the Amsden contact. Breccia zones, similar lithologically but much smaller in all dimensions, occur somewhat lower in the Mission Canyon in this vicinity, and in the upper few hundred feet of the Mission Canyon in all the other pre-Tertiary sectors.

A thin limestone conglomerate, 5–10 feet thick, is exposed high in the formation in one canyon (SW cor. sec. 29, T. 3 N., R. 2 E.). This facies, consisting of pebble-to-cobble size subrounded clasts of typical Mission Canyon limestone, solidly cemented by gray clayey, sandy limestone, could not be traced beyond the canyon walls.

In the Mud Spring area, Mission Canyon rocks form the west and north margins of the monzonitic pluton that intrudes Lodgepole rocks at its east margin. Curiously, the pluton despite its size and the vigorous effects it exerted on the Lodgepole limestone, had virtually no effect on the chemically similar Mission Canyon. The Mission Canyon at the north border gives no hint of an intrusive contact nearby, but the contact itself is not exposed; at the west border, the contact is almost continuously visible for nearly a mile, and marked by a discontinuous zone, in places several feet thick, of siliceous iron and manganese oxide deposits, but the limestone is hardly more than bleached, and that only for a few feet.

The Mission Canyon where fully preserved is easy to identify on the basis of its great thickness alone. In faulted terranes, it can be confused with the thick and indistinctly bedded parts of the Meagher; just such a situation appears in the Lombard thrust zone in sec. 19, T. 3 N., R. 2 E., where Meagher is thrust over Mission Canyon. In the absence of fossils, the bluish tinge of the weathered Meagher distinguishes it from the yellowish-gray surface of the Mission Canyon; in the instance cited, the assignment is confirmed by the presence of characteristic Meagher blue-and-gold strata nearby.

The Mission Canyon, as noted in the discussion of thickness, is succeeded in most parts of the quadrangle by the Big Snowy formation. Pre-Big Snowy disconformity is suggested by signs of weathering, such as discoloration, pitting, and slight channeling in the Mission Canyon just below the Big Snowy contact, by chips of Mission Canyon limestone just above the contact, and by the great variety of Big Snowy rock types in contact with the Mission Canyon. In the adjoining Toston quadrangle, disconformity is dramatically shown in fossil karst topography on the Mission Canyon, where local relief is as much as 100 feet along 300 feet of contact with basal Big Snowy rocks.

West of Milligan Creek and in the Mud Spring sector, Big Snowy rocks are missing and the Amsden overlies the Mission Canyon. Thickness measurements and truncation of the limestone breccia zone demonstrate deep pre-Amsden erosion of the Mission Canyon west of Milligan Creek. The contact itself is nowhere well enough exposed to give added evidence of disconformity; this is true also of the Mud Spring area.

The uppermost part of the Mission Canyon is more resistant than either the basal Amsden or basal Big Snowy, so that the top of the Mission Canyon is a dip slope into a depression or valley. Commonly, the succeeding rocks, whether Amsden or Big Snowy, are red and yield red soil. Where exposures are lacking, therefore, the contact is drawn at the change of slope between identified Mission Canyon and younger Carboniferous strata, just downsection from red soil.

AGE AND ORIGIN

Systematic fossil collections were not made from the Mission Canyon. Its Mississippian age, however, is established by collections of Early Mississippian fossils from the underlying Lodgepole limestone and Late Mississippian forms from the overlying Amsden and Big Snowy formations. Closer dating has been attempted by several regional workers, who generally assign it to the Early Mississippian. It may be, however, that the uppermost beds are middle Mississippian (Gardner and others, 1945).

The fairly deep neritic conditions that prevailed at the end of Lodgepole deposition continued while the Mission Canyon was being deposited, but with far fewer and fainter oscillations. Despite the extreme stability thus suggested, on the order of 1,000 feet of highly uniform limestone was laid down in water that was never more than a few hundred feet deep.

Limestone breccia in the upper part of the Mission Canyon is not merely a local phenomenon. It is a prominent feature as far north as the Limestone Hills, near Townsend (shown me by E. T. Ruppel); as far west as the southern Elkhorn Mountains (Klepper and others, 1957, p. 19–20); as far south as the Gravelly Range (Mann, 1954, p. 12); and as far east as the Bridger Range (McMannis, 1955, p. 1400–1401). The thickest and most persistent breccia in the region around Three Forks is in places where the Big Snowy is spotty or lacking. This observation certainly does not apply beyond the Three Forks region, for neither Big Snowy rocks nor breccia zones are present in south-central Montana (Gardner and others, 1945, fig. 2), in the Willis quadrangle in southwesternmost Montana (W. B. Myers, oral communication, 1957), or, by inference, in the Philipsburg quadrangle (Emmons and Calkins, 1913, p. 67–71).

Berry (1943, p. 16), apparently the first to publish a description of a persistent fragmental zone in the upper part of the Madison of western Montana, interpreted it to be a marine conglomerate "indicating a significant stratigraphic break." It is hard, however, to visualize a way in which ordinary submarine processes, acting far from known shores, could produce thin but areally extensive sheets of coarse unsorted angular limestone detritus in the midst of a completely conformable flat-lying limestone sequence. It is possible that Berry mistakenly included with the Madison red-cemented limestone conglomerate of Tertiary age that crops out at both ends of Milligan Canyon (see fig. 6) and was thus misled as to the nature of the process that formed the fragmental Madison rocks. Later workers in western Montana agree in general with the interpretation of Thom and others (1935, p. 27), made about similar fragmental rocks in the Madison of south-central Montana, that these rocks are solution or collapse breccia developed by cave-making processes above sea level. In this explanation, the rare lenses of conglomerate and well-bedded siltstone represent subterranean stream deposits.

If this is the correct mode of origin, the time of origin must be post-Mission Canyon, rather than intra-Mission Canyon as Berry concluded. The breccias are evidently older than the oldest folding, which is Late Cretaceous, for they are confined to a rather narrow stratigraphic range without regard for kind or degree of deformation, or for present topography. If due to underground solution, they must have been formed during one or more of the several known episodes of post-Mission Canyon pre-Laramide uplift and erosion. In the southern Elkhorn Mountains, where the Big Snowy is absent, Klepper and others (1957, p. 20) refer formation of the main breccia zone to a post-Mission Canyon pre-Amsden erosion interval. In the Three Forks quadrangle, the presence of some Big Snowy rocks leaves only part of Big Snowy time for breccia development. Farther east, in the Bridger Range, the basal Kibbey sandstone of the Big Snowy group lies on the Mission Canyon, leading McMannis (1955, p. 1401) to state that "the breccias developed prior to Big Snowy deposition, rather than prior to Amsden deposition." A reasonable inference is that the emergence which produced the restricted water body in which the main mass of the Big Snowy group was deposited also started solution breccia forming to the west. Where emergence was short-lived, breccia developed poorly; westward, longer emergence encouraged growth of thicker and more persistent solution breccia. Thus, both Klepper's and McMannis' interpretations would seem literally correct for each area.

BIG SNOWY FORMATION

Assigned to the Big Snowy formation is a bewildering variety of rock types whose unity lies largely in their stratigraphic position, between the relatively persistent lithologies of the Mission Canyon and Amsden formations. Mostly fine-grained clastic rocks and impure thin-bedded carbonate rocks, the Big Snowy is unresistant, and forms grassy topographic sags of irregular width, broken by a few low ledges.

Best, and least equivocal, exposures are in the canyon across the east limb of the Hossfeldt anticline in SW cor. sec. 29, T. 3 N., R. 2 E. There, 400 feet of strata of the Big Snowy consist of:

	Thickness (feet)
Yellowish-gray siltstone and dark-gray shale, largely covered	25
Thick-bedded olive-gray limestone, forming ledge	50
Thick beds of alternating yellowish-brown sandstone and fragmental limestone	25
Largely covered interval, probably dark shale, with a few thin fossiliferous dark-gray limestone beds near base	150
Yellowish-brown thin-bedded siltstone and mudstone	150

From this canyon the Big Snowy seems to swell and pinch irregularly both northeast and southwest, with rapid facies changes. The most persistent strata of the Big Snowy throughout the Hossfeldt Hills sector are the yellowish-brown siltstones at the base.

In the adjoining Mud Spring area, not more than 100 feet of rocks of the Big Snowy formation are present. They consist of a few ledges of thick-bedded dark-gray microcrystalline limestone, aggregating about 20 feet, separated by rarely exposed grayish-olive shale and dark clayey limestone.

In the Milligan Creek sector, the Big Snowy is confined to a strip about 6,000 feet long, east of Milligan Creek. The beds of the Big Snowy and lower beds of the Amsden are contorted through much of this strip into tight folds with amplitudes of a few hundred feet, and exposures are spotty, so that it is hard to get a clear picture of the stratigraphy. In a Big Snowy-Amsden section apparently 800 feet thick, a maximum of 200 feet seems assignable to the Big Snowy, and of this, float in the covered upper half includes some red sandstone that may be Amsden. Including the questionable strata, the Big Snowy of this sector comprises:

	Thickness (feet)
Largely covered. Rare outcrops are of thin-bedded light-gray dolomite and limestone; float includes some red sandstone	100
Yellowish- and greenish-gray sandstone, with a little thin-bedded light-colored mudstone and carbonate rock	50
Grayish-yellow siltstone and yellow-to-orange dolomite, poorly exposed	50

In the Willow Creek area, rocks of the Big Snowy are only about 50 feet thick but persistent, assuming that the base of the Amsden is correctly located below a thin sequence of red hematite-rich sandstone. Less than half the interval yields outcrops; these are of olive-gray and pale-yellowish-brown dolomite. The concealed part is probably underlain by dark shale.

For detailed measured sections of the Big Snowy formation, see pages 126–127, 130, 132, and 133.

Disconformity between the Big Snowy and the Amsden is suggested by rapid local changes in the thickness of both, by the great variety of rock types overlain by the basal part of the Amsden, and by the basal hematitic sandstones of the Amsden themselves, which seemingly represent worked regolith. All these relations, however, can be interpreted to indicate offlap in the Big Snowy and overlap of the basal part of the Amsden. The point is moot within the Three Forks quadrangle. In the neighboring Toston quadrangle, however, rocks of the Big Snowy group are thick in the southwestern part, but absent from the eastern part where they should be even thicker as the main basin of deposition of the Big Snowy is approached; this strongly suggests pre-Amsden erosion rather than offlap-overlap.

RELATION TO ESTABLISHED FORMATIONS OF THE BIG SNOWY GROUP

It seems futile to attempt correlations of these highly varied rocks from sector to sector, and even more so to force correlations at the formation level with other areas. Berry (1943, p. 18) handled the problem by recognizing no rocks of the Big Snowy group in his map area "except in the hills north of Three Forks where 3 feet of green shale, which may be Otter, separates the Madison and Amsden." Correlation of the Big Snowy of the Hossfeldt Hills with Scott's formations has, however, been attempted by Scott himself (1935, fig. 1, columnar section 7) and by Hendricks and Gardner (*in* Gardner and others, 1945, p. 20–21). Neither of these reports locates the section precisely, but there is no doubt that the "Eustis" section of both deals with the same outcrops described on page 44 and in section C (p. 132). The location of the section of Big Snowy measured by Hendricks and Gardner cannot be as described. They place it "about 1 mile west of Eustis [a flag station] * * * in secs. 7 and 8, T. 2 N., R. 2 E. in the Lombard Hills about 4 miles northeast of Three Forks," but Eustis is nearly 8 miles northeast of Three Forks, and secs. 7 and 8 are wholly covered by Quaternary deposits. However, their description of the terrain near the measured section, their statement that "the locality is within 1 mile of the Lombard overthrust," and details of the stratigraphy make it clear that the correct location is near that of the Hossfeldt Hills section of this paper, and would better be given as something like 2 miles southwest of Eustis and 6 miles northeast of Three Forks, in secs. 29 and 32, T. 3 N., R. 2 E.

A comparison of their interpretations with each other and with mine is illuminating. (See fig. 5.) The entire Big Snowy recognized by Hendricks and Gardner (see column D, units 1 and 2, fig. 5) includes about the same rocks as the lower three-fourths of the Big Snowy of this report (column C, units 3–6, fig. 5) therefore, they place in the Amsden about 100 feet of strata that are here regarded as Big Snowy. The strata in question are foreign to the basal part of the Amsden of the quadrangle and the region, which consistently is dominated by red clastic rocks, so that this part at least of their interpretation is untenable.

Scott's Big Snowy, on the other hand, while thinner than that of Hendricks and Gardner, paradoxically embraces the same span of strata as does the Big Snowy of this report, as shown by the similarity of Scott's Heath and Amsden formations (column B, units 3, 4, and 5, fig. 5) to the upper part of the Big Snowy formation and the Amsden formation of this paper (column C, units 7b–10, fig. 5). The differences in thicknesses among all three measurements does not mean significant operator error, but rather reflects measurements at different places of rocks with much initial lateral variation, and much disturbed by minor strike faults.

Comparison between Scott's interpretation and that of Hendricks and Gardner indicates that the Otter formation as used by Scott (column B, unit 2) comprises the same rocks as the Heath as used by Hendricks and Gardner (column D, unit 2); and the Kibbey formation as used by Scott (column B, unit 1) is the "Otter formation and perhaps Kibbey sandstone" as used by Hendricks (column D, unit 1).

If a choice had to be made, Scott's interpretation seems the better. On lithologic grounds his usage of Heath seems acceptable enough. But even if no importance is attached to the fauna identified by Smedley (see below) as similar to that of the Heath formation from about 150 feet above the base of the Big Snowy, and therefore far down in the Otter as used by Scott, the lithologies of Scott's Otter and Kibbey here are little like those of well-developed Otter and Kibbey northeast of Lombard, only 10 miles away (Scott, 1935, fig. 1, and p. 1026), that in turn resemble Scott's type sections (p. 1024–1025) generalized in column A, figure 5. Consequently, it seems best to treat the rocks of the Big Snowy group of the Three Forks quadrangle as a single unit of formation rank.

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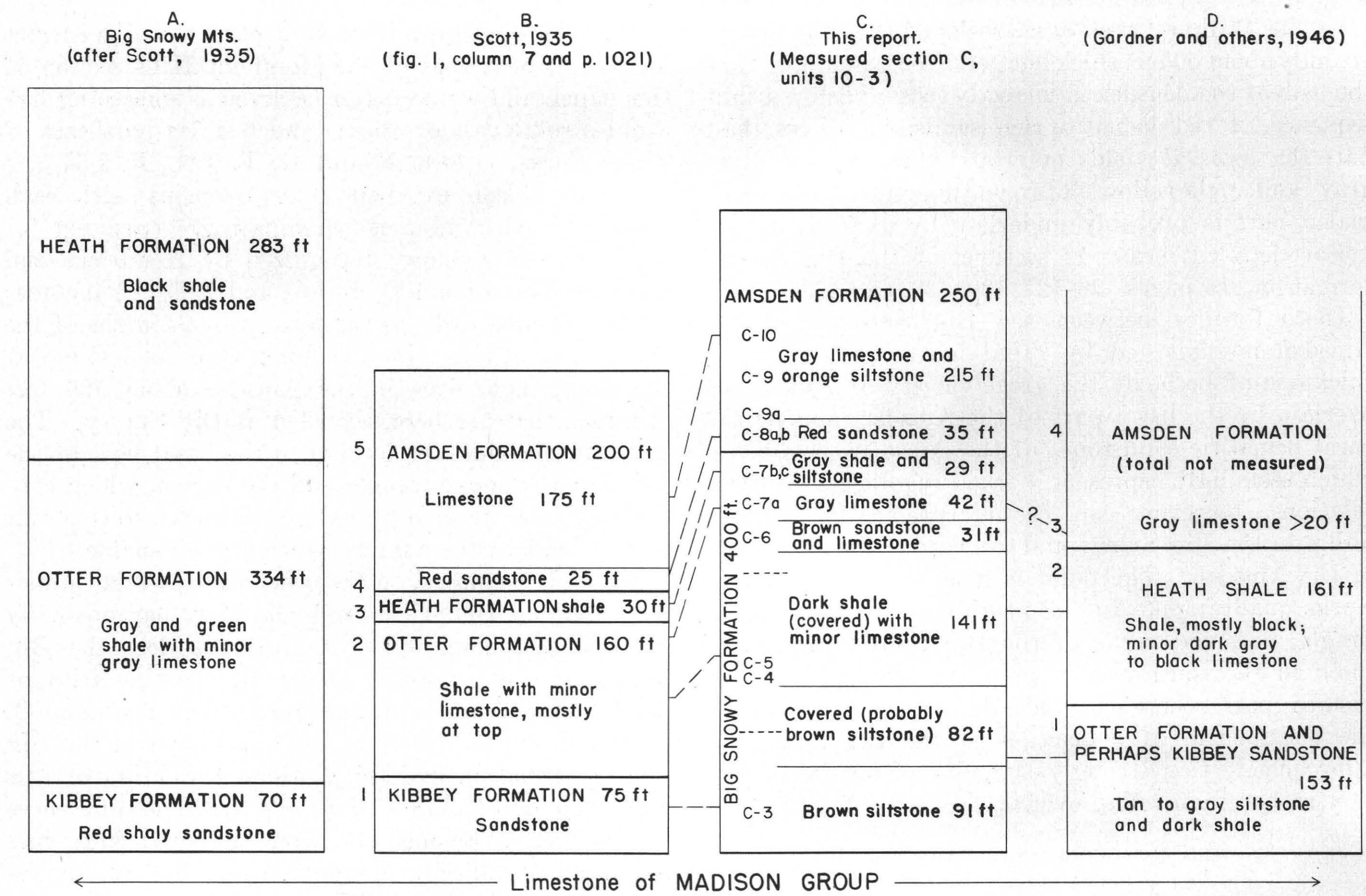

FIGURE 5.—Comparison of correlations of the Big Snowy west of Eustis Station, Gallatin County, Mont.

AGE AND ORIGIN

The age of the Big Snowy formation in the Three Forks quadrangle is only roughly determinate. Fossils are few, fragmentary, and extensively replaced by coarsely crystalline silica, making precise identification difficult. Only one locality, 150 feet above the base of the sequence in the Hossfeldt Hills sector (loc. 236 [16440–PC], pl. 2; center SE¼SW¼ sec. 29, T. 3 N., R. 2 E.) has yielded enough material, with adequate stratigraphic control, to be regarded as definitive. Smedley (written communication, May 28, 1954), who has also collected from this locality, lists the following fauna:

- Crinoid columnals
- *Chonetes* sp. undet.
- *Dictyoclostus inflatus* (McChesney)
- *Buxtonia* sp. undet.
- *Linoproductus* sp. indet.
- *Neospirifer* aff. n. sp. of Easton
- *Composita ozarkana* Mather
- *Punctospirifer* sp.
- Proetid trilobite frags., gen. indet.

and says:

> The species identified in this small faunule are similar to forms identified by Easton from the Heath formation in its type area. The *Neospirifer* corresponds to a new species being described by Easton from the Heath and the *D. inflatus* is similar to a new sub-species being described by him from the same unit. This faunule is of Late Mississippian age.

Crudely silicified corals from the same strata, studied by Helen Duncan (written communication, June 29, 1956), included bits of a favositid, tentatively identified as *Michelinia*?, and horn corals.

Two collections from beds probably of the Big Snowy formation in the Milligan Creek sector (loc. 43, SE¼ SW¼, sec. 24, and loc. 102, center E½NE¼ sec. 25, T. 2 N., R. 1 W., pl. 2) contain the following forms suggestive of Late Mississippian age, according to Smedley:

locality 43 [16436–PC]:
- *Schuchertella* sp. indet.
- *Stenocisma*? sp. indet.
- *Spirifer* aff. *S. brazerianus* Girty
- *Cleiothyridina hirsuta* Girty
- *Girtyella* cf. *G. woodsworthi* Clark

locality 102 [16437–PC]:
- *Fenestella* sp. indet.
- *Camarotoechia* aff. *C. herrickana* Girty
- *Eumetria* cf. *E. verneuiliana* Hall

The lowermost beds of the Big Snowy in this area may be of late Early Mississippian age, if fossiliferous thin-bedded gray limestone and orange siltstone, just below the Lombard thrust in the NW¼ sec. 12, T. 2 N., R. 1 E., (loc. 254 [16441–PC], pl. 2) are correctly assigned to the Big Snowy, but it is possible that these rocks are from the Madison group, in an undetected fault sliver. Of fragmental fossil material from these rocks, Smedley said: "It is reasonable to believe that the collections are of Mississippian age, and very likely Early Mississippian (Madison)". Forms include:

Crinoid columnals
Rhipidomella sp. indet.
Chonetes sp. indet.
Productus cf. *P. gallatinensis* Girty
Linoproductus? sp. indet.
Productid, gen. indet.
Spirifer centronatus Winchell
Punctospiriferoid frag., gen. indet.

The uppermost beds recognized as Big Snowy may be of Early Pennsylvanian age. They have yielded no diagnostic fossils, but the lowest fossiliferous strata of the Amsden cannot be dated more closely than Late Mississippian or Early Pennsylvanian (see later discussion), and the strata of the Big Snowy could cross the systemic boundary.

To summarize, rocks of the Big Snowy formation in the Three Forks quadrangle are mostly of Late Mississippian age, but may include strata as young as Early Pennsylvanian and as old as late Early Mississippian.

The clastic rocks of the Big Snowy formation are mostly if not entirely nearshore marine deposits. Possible exceptions are the basal siltstones which could be flood-plain deposits. Much of the rapid lateral variation in the Big Snowy may be due to the effects of the irregular weathered surface of Mission Canyon limestone on which it was deposited. Plainly suggested is a nearby land source of varied lithology, but the nearest such source seems to have been far to the southwest. In Big Snowy time, limestones of the Mission Canyon or its equivalents must have been the only exposed rocks for not less than a hundred miles southwest of Three Forks and for much greater distances in other directions (see, for instance, Sloss, 1950, fig. 8). Conscious of the distances involved, Sloss (1951, p. 56) nonetheless favored a source to the southeast on the "Wyoming Shelf" or even farther south in the "Ancestral Rockies." Part of the clastic components, especially near the base, were doubtless derived locally by accumulation of insoluble residues from the erosion of the Mission Canyon.

AMSDEN FORMATION

The Amsden formation, which crops out in all four pre-Tertiary areas, is a heterogeneous assemblage of thin-bedded vividly colored clastic and carbonate rocks. Rather poorly sorted mudstone and sandstone, dominant at the base, give way to limestone and dolomite upsection. The clastic phases change character as well as volume upsection, shifting to progressively better sorted siltstone and quartzitic sandstone. The unifying lithologic trait of the Amsden is its redness, which ranges from blackish red in the basal sandstones to delicate pink tints in the upper carbonate beds; where outcrops are poor, residual soils are typically reddish brown. Many beds are other colors, with yellow and orange especially common in the siltstone, and gray in the carbonate rocks, but the formation as a whole has a distinctive reddish cast. The base of the formation is invariably in a soil-covered saddle or valley; upsection, exposures are progressively better and relief is accentuated as resistant carbonate beds become more numerous.

In the Milligan Creek area, the Amsden is 500–600 feet thick, both west of the creek, where it lies on Mission Canyon limestone, and east of the creek, where it succeeds the Big Snowy formation. East of the creek, the sequence may be summarized as:

	Thickness (feet)
Interbedded gray dolomitic limestone and orange to reddish-brown sandstone	100
Dark-gray limestone, in a few thick ledges, separated by orange siltstone	50
Thin-bedded sandstone, mudstone, limestone, and dolomite, varicolored red, purple, pink, orange yellow; poorly exposed	450

West of the creek, exposures are too poor for detailed measurements, but it is plain that the basal unit is much thinner and the upper units concomitantly thicker. Almost every sandstone and mudstone has abundant carbonatic cement. An interesting feature of some of the brown sandstone beds in the upper unit is the presence of several percent of chlorite, determined in immersion liquids and checked by X-ray.

The Amsden appears in the Mud Spring area only in a short strip cut by Mud Spring Gulch. The full section is present for only half a mile north of the gulch, and much of that is covered. Here the Amsden is 300 feet thick, comprising:

	Thickness (feet)
Thick-bedded gray limestone and dolomite, forms ledges	200
Thin-bedded red and yellow limestone and siltstone, poorly exposed	100

On the east flank of the Hossfeldt Hills anticline the Amsden is even thinner, being rather close to 250 feet thick, and likewise divisible into two convenient units:

	Thickness (feet)
Thin-bedded gray limestone and orange siltstone	200
Thin-bedded red and greenish-gray mudstone and sandstone, poorly exposed	50

In the Willow Creek area the Amsden is generally around 400 feet thick, though in places it is more than 500 feet thick. Typical sequences include:

	Thickness (feet)
Thin-bedded yellowish-gray dolomite and limestone, with subordinate orange and red siltstone	100
Thin-bedded pinkish-gray quartzitic sandstone and yellowish siltstone	200
Thick-bedded gray, orange, and brown sandstone, with reddish-black hematitic sandstone at base	100

For detailed measured sections see pages 126, 130, 132, and 133.

Thickness of the middle unit varies erratically, whereas the upper and lower units are persistent.

The Amsden strata of the Milligan Creek and Willow Creek sectors are rather alike as are the Amsden of the Mud Spring and Hossfeldt Hills sectors, but the Amsden of the two southwestern sectors contrasts markedly with that of the two northeastern sectors. The contrasts are mostly with respect to clastic components. The carbonate rocks are rather uniform, being mostly microcrystalline and even textured, with considerable silica in the form of nodules of black or olive chert, scattered grains of quartz, and, in the limestone, thoroughly silicified fossils. The few fossiliferous beds have sugary coarser textures, and coarse sugary dolomite beds are rather common near the top of the formation. The Amsden and Big Snowy are unique among the Paleozoic rocks in containing many silicified brachiopods and corals; in older strata, these forms are replaced by carbonate or, rarely, iron oxides.

Despite the contrasts within the respective sections, each begins with basal hematite-rich sandstone or siltstone, and ends by grading into the overlying Quadrant formation through an alternation of thick beds of gray limestone or dolomite and pale quartzitic sandstone. In the Willow Creek and Milligan Creek areas, the passage from dominant limestone to dominant sandstone involves 50 feet or more of transitional beds in which the two rock types are about equal, so that the contact is mapped as gradational. In the Mud Spring and Hossfeldt Hills sectors, the transition is generally accomplished within 20 to 30 feet stratigraphically and the symbols for ordinary contacts are used. Where outcrops are poor, the lowest ledge (2 feet or more thick) of sandstone is arbitrarily used as the base of the Quadrant; the same stratum is surely not being chosen throughout but the errors are small. This criterion has been widely used in western Montana (for example, Gardner and others, 1946, p. 4–5; Klepper and others, 1957, p. 20). An obvious alternative, to choose the top of the Amsden as the highest thick limestone or dolomite bed, is not practicable, as thick carbonate strata appear throughout the Quadrant.

AGE AND ORIGIN

Fossils from the Amsden are few, poorly preserved, and confined to the middle of the formation. They indicate either Late Mississippian or Early Pennsylvanian age. Only the Milligan Creek and Willow Creek areas have yielded fossil material from the Amsden. The six collections cited below, from the 3-mile strip of Amsden at the north side of the Willow Creek area (see pl. 2), represent no strata lower than 150 feet above the base or higher than 100 feet below the top.

1. Collection 323 [16366–PC]from gray silty limestone 150–175 feet above the base (J. E. Smedley, written communication, Dec. 6, 1954):

   Cleiothyridina sp.
   Dictyoclostus sp. indet.
   Echinoconchus sp. indet.
   Linoproductus sp.

2. Collection 315 [16444–PC] from olive-gray limestone 350–400 feet above the base (J. E. Smedley, written communication, May 28, 1954):

   Composita cf. *C. subtilita* (Hall)
   Antiquatonia sp.
   Rhipidomella? sp. indet.
   Terebratuloid brachiopod n. gen.?

3. Collection 316 [16445–PC] from orange siltstone 20–30 feet below 315 (J. E. Smedley, written communication, May 28, 1958):

   Spirifer cf. *S. increbescens* Hall
   Composita sp. indet.
   Orthotetid brachiopod gen. indet.
   Productid gen. indet.
   Punctospiriferoid gen. indet.
   Cypricardella sp. indet.

4. Collection 321 [16446–PC] from 25 feet of yellowish gray and pale red siltstone, 275–300 feet above the base (J. E. Smedley, written communication, May 28, 1958):

   Fenestellid bryozoan gen. undet.
   Spirifer aff. *S. increbescens* Hall
   Composita aff. *C. sulcata* Weller
   Punctospirifer sp. indet.

5. Collection 322 [16447–PC] from 25 feet of strata like those of 321 and directly below 321 (J. E. Smedley, written communication, May 28, 1954):

   Fenestrate bryozoan cast indet.
   Fistuliporoid bryozoan gen. indet.
   Camarotoechia? frag. sp. indet.
   Spirifer increbescens Hall
   Composita cf. *C. subtilita* (Hall)
   Orthotetid brachiopod frag. gen. indet.
   Chonetes sp. indet.
   Linoproductus sp. indet.
   Productid frag. gen. indet.
   Rhipidomella sp.

6. Collection 309 [16443–PC] from light-colored, mostly yellowish gray silty limestone, dolomitic limestone and siltstone 150–250 feet above base (J. E. Smedley, written communication, May 28, 1954):

Fenestellid bryozoan gen. indet.
Spirifer cf. *S. increbescens* Hall
Chonetes sp. indet.
Derbya sp. indet.
Dictyoclostus cf. *D. inflatus* (McChesney)
Linoproductus sp. indet.
Myalina? sp. indet.

The Milligan Creek area furnished two Amsden fossil localities, both from east of the creek and from the middle limestone unit. One of these (loc. 106 [16439–PC], pl. 2; NE¼NW¼ sec. 25, T. 2 N., R. 1 W.) contains a faunal assemblage much like that in the Willow Creek localities (J. E. Smedley, written communication, May 28, 1954):

Echinoid plate gen. indet.
Spirifer cf. *S. increbescens* Hall
Composita cf. *C. subquadrata* Hall
Dictyoclostus cf. *D. inflatus* (McChesney)
Straparollus (*Euomphalus*) sp. indet.

and fragments of punctospiriferoid and productid brachiopods.

The other locality, 4,500 feet farther east (loc. 103 [16438–PC], pl. 2; SW cor. sec. 19, T. 2 N., R. 1 E.) has yielded, according to Smedley, two of the forms present in the other localities listed:

Spirifer cf. *S. increbescens* Hall
Composita sulcata Weller

and also several forms known from the upper part of the Big Snowy group in its type area:

Antiquatonia cf. n. sp. of Easton
Spirifer aff. n. sp. of Easton
Cleiothyridina hirsuta Girty
Eumetria aff. E. *acuticosta* Weller;

and also, according to Duncan, fragments of a coral of a "manuscript genus" known from the Otter formation. Like the fauna from the bulk of the underlying Big Snowy, the Amsden faunas can be dated no closer than Late Mississippian and (or) Early Pennsylvanian. This means that the systemic assignment of 300 to 800 feet of fossiliferous beds is in question. Furthermore, no fossils have been recovered from the overlying, supposedly Pennsylvanian Quadrant formation, leaving in question the age of 300 feet more of strata. Plainly, subdivision of the Carboniferous systems into Mississippian and Pennsylvanian in this area must await further work. The apparent occurrence of typical Big Snowy fossils high in the Amsden is puzzling, but can hardly be profitably discussed now in view of the meager local and regional evidence. One safe conclusion is that the erosional unconformity, if any, between the Amsden and Big Snowy represents a short time.

The Amsden is mostly a transgressive shallow-water marine formation, no doubt under climatic and orographic conditions much like those which led to deposition of the Flathead sandstone and Wolsey shale. The basal hematitic sandstones are mineralogically like the Flathead, though less well sorted and rounded, suggesting a shorter period of beach reworking. The differences between the finer grained parts of the Amsden and the Wolsey are mainly mineralogic, and reflect differences in the mineralogy of source rocks of the clastic components. As the source of the clastic rocks of the Flathead and Wolsey was Precambrian crystalline rocks like those to the south, and Belt rocks derived from such crystalline rocks, the source rocks of the bulk of the Amsden must be sought in other terranes. Part of the clastic rocks of the Amsden were no doubt derived from the erosion of nearby Carboniferous rocks. The chlorite in the upper sandstone of the Amsden, however, cannot have had a local source. Perhaps it and, of course, other detrital components as well were supplied by the erosion of Paleozoic, Belt, and Precambrian crystalline rocks exposed far to the west (but not necessarily farther than the vicinity of the Idaho border where the previously mentioned Cambrian land barrier recognized by Willis (1907), Walcott (1915), and others, could have persisted into, or been revived in, late Carboniferous time.)

The depositional environment of the interbedded carbonate rock and quartzitic sandstone in the upper part of the Amsden resembled that of the overlying Quadrant formation (p. 51–52).

QUADRANT FORMATION

The Quadrant formation, mostly variably cemented sandstone, crops out in all four areas of pre-Tertiary rocks. Where the sandstone is well cemented by silica, and is appropriately called quartzitic, it tends to be resistant to erosion, and forms prominent ledges and ridge crests. Where the rock is cemented by calcite, it crops poorly and yields a thin rubbly soil. The kind and degree of cementation varies greatly in individual beds, so that ledges are discontinuous, though the strata that form them may not be. Best exposures of the Quadrant are east of Milligan Creek, along the section line between secs. 24 and 25, T. 2 N., R. 1 W.

Although its thickness varies considerably, the gross lithology of the Quadrant is rather uniform throughout the quadrangle. The main rock type is thick-bedded light-colored medium-grained rounded quartz sandstone. Cementation is mainly by quartz overgrowths, but a little carbonatic cement is present in nearly every stratum, and in many carbonate is the principal cement. Rarely both fine-sand and coarse-sand phases occur. Cross-stratification occurs, but is not common. Thick interbeds of light-colored even-grained sandy dolomitic limestone, abundant in the lower part, are progressively rarer upsection.

In the Milligan Creek area, the Quadrant averages about 300 feet thick, though it is 400 feet thick where detailed measurement, summarized below, was made east of Milligan Creek:

	Thickness (feet)
Thick-bedded quartzitic sandstone, yellowish-gray, pinkish-gray, very pale orange; forms ledge	100
Sandstone like that above but with calcareous cement, poorly exposed	150
Alternating thick beds of light-colored calcareous sandstone (2/3) and of sandy and dolomitic limestone (1/3), poorly exposed	150

Gardner and others (1945, p. 18–20) measured the same section described here, but reported only 72 feet of Quadrant. At the base of their measured Quadrant is 10+ feet of resistant light-gray limestone that they assigned to the Amsden, so there is no way of knowing how they interpreted the 300 feet of quartzitic sandstone and limestone below. The impression is that they abandoned their own criterion, and used the top of the highest thick limestone rather than the base of the lowest thick sandstone for the Amsden and Quadrant contact. Their Quadrant, thin as it is, nevertheless includes 32 feet of strata assigned to the Phosphoria formation in this study, based on the presence of bedded chert and of early Permian fossils (see p. 52–53).

The comparatively great thickness of the Quadrant east of Milligan Creek may be initial, or may be due to unrecognized strike faults like the fault which duplicates the Phosphoria and Ellis just upsection in the same vicinity, or may reflect varying amounts of pre-Phosphoria erosion. The last of these possibilities seems remote; evidence is inadequate for choosing between the other two.

The Quadrant crops out extensively on both sides of Mud Spring Gulch, but due to folding and accidents of exposure there is no section suitable for measuring. Minimum thickness is 250 feet; average thickness may be much greater, as several map measurements, not worth much consideration singly, consistently are close to 500 feet. The basal 100 feet or so is thick-bedded yellowish-gray and yellowish-orange quartzitic sandstone, without carbonate interbeds. At least 100 feet of similar beds are at the top of the formation. Some covered zones within the formation are probably underlain by sandstone with calcareous cement.

In the Hossfeldt anticline, the Quadrant is comparatively thin, rarely exceeding 150 feet:

	Thickness (feet)
Quartzitic sandstone like that below, but lacking limestone, brecciated and sheared	50
Thick-bedded white to yellowish-gray quartzitic sandstone; a few beds 1–3 feet thick of grayish-yellow limestone	100

As the description suggests, the formation may be thin due, in part at least, to faulting. Scott (1935, p. 1017) reported a thickness of 160 feet here ("west of Eustis").

In the Willow Creek sector, the Quadrant is 200 feet thick, and consists of:

	Thickness (feet)
Thin- to thick-bedded yellowish-gray quartzitic sandstone, and a few thin beds of white limestone	150
Alternating thick beds of light-gray dolomitic limestone and quartzitic sandstone	50

The Quadrant is conformably overlain by the Phosphoria formation. Gradational relations are suggested by the appearance throughout the Phosphoria of quartzitic sandstone and sandy limestone strata identical with those in the Quadrant. Overlap of the Phosphoria is suggested by variations in lithology of the basal beds of the Phosphoria, but there is no hint of pre-Phosphoria subaerial erosion, except possibly in the erratic thickness of the Quadrant formation. In the Milligan Creek and Hossfeldt Hills areas, where the basal part of the Phosphoria is resistant thick-bedded carbonate rock, the contact has little topographic expression; elsewhere, the basal strata of the Phosphoria are unresistant chippy siltstone and bedded chert, and the contact is marked by a shallow and narrow depression. Where the contact is not exposed, it is arbitrarily placed at the downsection base of the lowest slope bearing outcrops or float of any distinctive Phosphoria rock type: chippy yellow siltstone, brown or gray chert, or phosphatic sandstone.

Detailed measured sections are given on pages 126, 130, 132, and 133.

AGE AND ORIGIN

No fossils were found in the Quadrant. On local evidence the Quadrant may be as young as early Permian, for Foraminifera from the base of the Phosphoria (see p. 53) are no older than middle Wolfcamp; it may be as old as early Pennsylvanian, for no fossils from the underlying local Amsden are dated as younger than early Pennsylvanian, if that young. Scott (1935, p. 1031) regards the Quadrant, redefined, as "the westward equivalent of the Tensleep sandstone" and of "basal Pennsylvanian" age. Early Pennsylvanian (Des Moines) fusulinids have been reported from near the top of the Tensleep as far west as the Pryor Mountains, 150 miles southeast of Three Forks (Richards, 1955, p. 31). But if the Quadrant of southwestern Montana is truly gradational with the Phosphoria, as suggested not only in this paper but also by Klepper and others (1957, p. 22) and by Scholten and others (1955, p. 366) then at least part of the Quadrant is late Pennsylvanian or even early Permian as Sloss and Moritz (1951, p. 2165), and later Scholten and others (1955), recognized. Furthermore, Verville (1957) presented evidence that the Tensleep in the Big Horn Mountains, Wyoming, is of Permian (Wolfcamp) age. On the opposite side, Mann (1954, p. 18) found the Phosphoria disconformable on the Quadrant in the Gravelly Range.

If the Quadrant near Three Forks is indeed the strict lithologic equivalent of the Tensleep of eastern Montana, it too may be of Pennsylvanian and Permian age. Sandstones like the Quadrant are deposited slowly, and the Quadrant lithofacies may well rise in time westward. Until definitive faunas are discovered in the western Quadrant, however, it seems prudent to assign the Quadrant near Three Forks the customary Pennsylvanian age.

The thick-bedded, locally cross-stratified sandstone of the Quadrant, almost monomineralic and thoroughly worked, suggests shallow neritic or beach conditions on a stable shelf, dominated by slow marine transgression. The sandstones are much like those of the Flathead, and conditions of deposition were presumably similar. The interbeds of limestone, much of it clastic, but not bioclastic, and with a persistent fraction of quartz sand grains, suggest the same depositional environment.

A nearby land source seems required. It is not necessary, of course, for the landmass to have been reduced to base level to account for each limestone bed, or even for the shoreline to have retreated. A sufficient inference is that local shifts of current caused sand to bypass a particular part of the basin long enough for limestone to accumulate, partly by direct chemical deposition, and partly by current reworking of poorly consolidated chemical limestone.

While it is not hard to account for the limestones, absence of intermediate fine-grained clastic rocks is a real problem, and one of more than local dimensions, for similar accumulations of interbedded quartzitic sandstone and limestone are widespread, particularly in the Pennsylvanian and Permian of western United States; for example, Oquirrh formation of central northern Utah (Gilluly, 1932), Wells formation of northeastern Utah, eastern Idaho, and southwestern Wyoming (Richards and Mansfield, 1912).

The coarse clastic rocks of the Quadrant were derived from a source either of varied rock types, or of rocks much like the Quadrant in composition and grain size. If the source was a normally varied assemblage of older rocks, then accumulation of nearly pure quartz sand in one place must inevitably have been attended by accumulation of clay rocks elsewhere. If the Quadrant is genetically like the Flathead sandstone and the offshore time-equivalent of the Flathead is the Wolsey shale, what rocks of an age overlapping that of the Quadrant are like the Wolsey? If the age of the Quadrant throughout is limited to Early Pennsylvanian as it seems to be in eastern Montana, there are no shaly rocks in the region which can play the part. But if the western Quadrant is younger, as suggested earlier, then a possible candidate for this role is the thick sequence of fine-grained clastic rocks in the lower part of the Phosphoria formation, far to the south.

Such interpretation implies a northern source. One possible source is in central Montana, roughly on the site of the old Big Snowy basin, as reconstructed by Scott, (1935, fig. 3). The Quadrant thins abruptly northeastward from Three Forks, being less than 50 feet thick in much of the Toston quadrangle, and absent from central Montana (Scott, 1935, p. 1015). It may never have been laid down there, as Scott implied; its absence can, however, also be interpreted as due to pre-Ellis (Jurassic) erosion, rather than to nondeposition (see, for example, Vine, 1956, p. 433). But although the former Big Snowy basin may have been an island or peninsula contributing detritus to the Quadrant sea, its dimensions do not seem to have been large enough to provide material for all the Quadrant of southwestern Montana, which blanketed many thousands of square miles to average depths of 200 or 300 feet, and is 1,000 feet thick in places (Condit, 1918, p. 111).

The other possibility, namely that the sandstone of the Quadrant was derived from rocks consisting almost wholly of coarse quartz sandstone, poses equal problems. The only rocks of this sort in the regional geologic column are the Flathead sandstone and the quartzites in the Belt series. The Flathead is plainly too thin to have been a major contributor to the Quadrant. In most places, the Belt series has too high a proportion of clay-rich strata, and its quartzite is too fine grained, to qualify. In northern Beaverhead County, a thick sequence of coarse-grained quartzite (W. B. Myers, oral communication, 1957) would seem lithologically suitable, but it is apparently a local phase of inadequate volume to supply a significant part of the Quadrant. The thick and extensive Hoodoo quartzite and its equivalents in south-central Idaho (Ross, 1935, p. 15–19) would appear suitable in nearly every respect, but are rather distant.

Derivation of the local Quadrant clastic rocks by winnowing of sand from a northerly landmass of varied lithology seems the better interpretation at present, but there are many uncertainties. The Quadrant of southwestern Montana clearly invites detailed study.

PERMIAN

Rocks of demonstrably Permian age are 75 to 200 feet thick, and are assigned to a single formation, the Phosphoria. They are made up of about 40 percent chert, 40 percent sandstone and siltstone, and 20 percent carbonate rock, mostly dolomitic limestone; the clastic ratio, assuming that chert is nonclastic, is near 0.7. The Permian rocks are unique in this region in containing substantial amounts of bedded chert and of phosphatic material.

Because of the economic importance of its phosphatic rocks, the Permian of the Cordilleran region has been thoroughly studied, mostly during the past decade by the Geological Survey. McKelvey and others (1956) summarized the regional stratigraphic relations and proposed an interesting nomenclatural scheme. Cressman (1955) described the physical stratigraphy of the Permian rocks of southwestern Montana.

PHOSPHORIA FORMATION

The Phosphoria formation, which crops out in all four areas of pre-Tertiary rocks, is a heterogeneous assemblage of clastic and chemical rocks, displaying almost as much variety as the Big Snowy formation. Best exposures are east of Milligan Creek in sec. 24, T. 2 N., R. 1 W. Limestone like that of the Quadrant and quartzitic sandstone dominate the section in some places; in other places, dark phosphatic sandstone and yellow siltstone prevail. The main unifying factor is the presence of bedded chert throughout. Considering the great internal variety and the presence of an important unconformity above, the variations in thickness of the Phosphoria are surprisingly small.

Topographic expression of the Phosphoria varies with the dominant lithology. Where rocks similar to those of the Quadrant are the main component, they yield low discontinuous ledges, and there is little topographic evidence of change from the Quadrant. In some places the contact appears inconspicuously on a back-slope valley wall; in others, on a steplike dip slope. Where unresistant rock types dominate, the basal Phosphoria forms a low grassy depression and rises upsection toward the overlying resistant basal rocks of the Ellis formation.

In the Milligan Creek area, the Phosphoria is at its thinnest; it is nowhere more than 150 feet thick and in places is as thin as 75 feet. A typical section includes:

	Thickness (feet)
Thick-bedded brown chert, and thinly interbedded chert, yellow very fine sandstone, and greenish-gray phosphatic sandstone	50
Dolomitic limestone, like basal unit; forms white ledge	25
Orange quartzitic sandstone and brownish-gray chert, poorly exposed	25
Yellowish-gray dolomitic limestone, with many pods of olive chert	25

Most of the chert is very thin bedded to laminated, in sequences a few inches thick. The brittle chert, intricately fractured, rarely yields good outcrops, but is readily traced by its abundant angular float. Near the top of the formation are a few thick-bedded sequences of chert that form bold outcrops. There, chert is invariably somber, mostly in muted tones of green and brown. A little of it is phosphatic, as shown by a bluish-white weathering bloom.

Although the quartzitic sandstone and some of the limestones are indistinguishable from those of the Quadrant, many of the limestones differ somewhat. They have considerable clay that produces a dull earthy surface, even when freshly broken, and distinctive knobby pods, commonly 4 to 8 inches long and a quarter as thick, of dark chert. Discontinuously present at the very base of the formation, east of Milligan Creek, is a bed, rarely as much as 3 feet thick, of gray limestone like that of the Quadrant that is riddled with lentils and partings of chert, and contains many silicified fusulinids; a few similar lenses are scattered in the lower unit of the formation.

The greenish-gray phosphatic sandstones in the upper part of the formation, though of small volume, are unusual enough to warrant description. They are rarely as much as 1 foot thick or traceable for more than a few hundred feet laterally; the thickest lens exposed, 3½ feet above the base of the upper unit in SW1/4SW1/4 sec. 24, T. 2 N., R. 1 W., is nowhere more than 1½ feet thick. Three or four similar, but much thinner, lenses typically appear higher in the unit. The rock is medium to coarse grained, rather poorly sorted, and well cemented, in places by carbonate, in places by silica. The sand grains are of rounded to subangular chert, phosphatic oolites and pellets, shell fragments (some of which seem to be phosphatic), and iron-stained clay pellets, in widely varying proportions. The phosphatic grains weather bluish white, making a distinctive speckled surface.

Gardner and others (1945, p. 19–20) describing the same section, reported a thickness of only 84 feet. As noted earlier, they included with the Quadrant 32 feet of chert-rich clayey carbonate strata that are here classed as basal Phosphoria.

In the Mud Springs sector, the Phosphoria is exposed only along two short stretches at the west edge, and there poorly. The section is about 150 feet thick and comprises:

	Thickness (feet)
Thick-bedded yellowish-gray quartzitic sandstone	50
Olive-gray thin-bedded chert	50
Thinly interbedded chert, carbonatic siltstone and phosphatic sandstone, largely concealed	50

Limestone, so prominent near Milligan Creek, is absent here, and phosphatic sandstone is near the base rather than the top.

In the Hossfeldt Hills, the Phosphoria is about 200 feet thick, and fairly well exposed. It consists of:

	Thickness (feet)
Thick-bedded yellowish-gray quartzitic sandstone	50
Thinly interbedded dolomitic limestone, siltstone, and chert, poorly exposed	25
Thick-bedded light-olive chert, forms ledge	50
Quartzitic sandstone, limestone, and chert, largely covered	50
Yellowish-gray limestone with large pods of chert	25

This section is much more like that near Milligan Creek, 10 miles away, than that near Mud Spring Gulch, 6 miles away, except for the upper unit. It differs notably from both in lacking phosphatic sandstone.

In the Willow Creek area, the full Phosphoria section is present only along a 1,000-foot strip in the NW¼SE¼ sec. 34, T. 1 N., R. 1 W.; elsewhere, it is covered by Cenozoic deposits. Exposures in this strip are good, though possibly not typical. They comprise:

	Thickness (feet)
Thick-bedded yellow quartzitic sandstone	25
Chippy yellow and orange siltstone; minor chert and phosphatic sandstone, poorly exposed	50
Olive-gray chert, poorly exposed, with three 3-foot beds of greenish-gray phosphatic sandstone near top	50
Orange siltstone and dolomite, and olive-gray chert, poorly exposed	25

The Phosphoria of this area largely lacks carbonate rock, and in this respect resembles that of the Mud Spring area; in general, it is much more like the two distant sectors than the nearby Milligan Canyon sector.

Measured sections of the Phosphoria are given on pages 125–126, 130, 131, and 133.

The Phosphoria is disconformably overlain by rocks of the Jurassic Ellis formation. In most places the basal beds of the Ellis are pebble and cobble conglomerate composed of chert of the Phosphoria, plain evidence of erosional unconformity. In the few exposures of the contact there is no hint of angular unconformity. Although the exact contact is rarely exposed, definitive exposures of both formations are numerous enough to fix the contact everywhere within a few hundred feet horizontally. (In the recently published Jurassic folio (McKee and others, 1956) plate 2, showing pre-Jurassic geology, is in minor error in showing the Phosphoria absent below the Ellis in the northeastern part of the Three Forks quadrangle. In fact, Phosphoria rocks, though thin, persist northeastward throughout all but the northeasternmost part of the Toston quadrangle, and the Phosphoria-Quadrant boundary of pre-Ellis time was thus on the order of 20 to 25 miles farther northeast than drawn.)

NAME, AGE, AND CORRELATION

Peale did not recognize Permian rocks in any part of the Three Forks 1° sheet. Most of the Permian rocks make poor outcrops and have few fossils; those that crop out well are unfossiliferous strata, like the Quadrant, and it is not surprising that Peale included them with the Quadrant. Condit (1918) was the first to recognize that lithologic equivalents of the Permian Phosphoria formation, with type area in southeastern Idaho and northeastern Utah, occur throughout the Three Forks quadrangle. Nearly all subsequent workers in southwestern Montana have applied that name to all the rocks, of whatever lithology, between the Quadrant formation and known Mesozoic rocks. Berry (1943, pl. 1, p. 20–21) included in the Ellis formation rocks regarded by Condit (1918), by Cressman (1955, pl. 4), and in this paper as Phosphoria equivalents. Faunal evidence that some of these rocks in the Three Forks quadrangle are of early Permian age and therefore at least rough time-equivalents of the Phosphoria was presented by Frenzel and Mundorf (1942).

Fossils collected during this study from the base of the Phosphoria in the strip in S1/2 sec. 24, T. 2 N., R. 1 W. (loc. 115 [f97901], pl. 2), apparently also the locality of Frenzel and Mundorf, have been examined by L. G. Henbest (written communication, May 19, 1954), who identified the following forms:

Pseudoschwagerina sp. cf. *P. montanaensis* Frenzel and Mundorf
Pseudoschwagerina or *Schwagerina* sp.
Pseudofusulina (or *Parafusulina*?)

indicative of middle or late Wolfcamp (or possibly early Leonard) age; that is, very early but not earliest Permian. The unfossiliferous upper part of the formation may, of course, be younger by an epoch or more but there is no reason to suspect that any part of the formation is younger than Permian, for lower Triassic rocks, though absent from the Three Forks quadrangle itself and from the region to the west, north, and east, overlie the Phosphoria equivalents not far to the south (see, for example, Klepper, 1950, p. 66; Moritz, 1951, fig. 11; Mann, 1954, p. 20–22).

The name Phosphoria is assigned to the entire Permian sequence even though phosphatic shale and chert, the characteristic lithofacies of the type Phosphoria, do not dominate the local sequence. There is much merit in the proposal of McKelvey and his associates (1956) that the name should be limited to rocks of a single lithologic community, the chert-mudstone-phosphorite facies, and that different names should be given to sequences of Phosphoria age but different lithofacies: Park City formation (of Boutwell, 1907) for carbonate facies, and Shedhorn sandstone (new name) for quartz sandstone facies.

The Three Forks quadrangle contains rocks that might be assigned to all three of McKelvey's formations, and in roughly equal proportions, but the scale of mapping prevents their separation, if all three formations are really present. The rocks must be mapped as a unit, preferably with a convenient unit name. This might as well be the time-honored Phosphoria formation, rather than such unwieldy, though more precise, constructions as "rocks of the Phosphoria interval," "rocks of Phosphoria age," or, in the usage of Williams (McKelvey and others, 1956, p. 2856–2857), "rocks of Park City age."

ORIGIN

The light-colored quartzitic sandstone and sandy limestone of the Phosphoria no doubt formed like similar rocks in the Quadrant; that is, not far from shore under conditions of slow marine transgression. The dark phosphatic sandstone and yellow siltstone bear the stigmata of shallow-water deposition also, but their poor sorting and high organic content suggest regression, and at a comparatively rapid rate. Evidence in this area is insufficient to indicate whether transgression and regression were cyclic, as concluded on regional evidence by McKelvey and others (1956, p. 2827–2829) for the western field as a whole, and by other authors, and if so, how many cycles are represented. The regional studies by Cressman (1955, p. 22–25) and McKelvey and others (1956, especially their fig. 1) suggest a land source of varied rocks northeast of Three Forks—perhaps about the same landmass which has been visualized as supplying the sand of the Quadrant.

As for the bedded chert nothing can be said, on the basis of this study, about the source of its silica, and little about the conditions of its deposition. Its uniform anisotropic texture indicates chemical deposition. Its intimate association with all the other Phosphoria rock types, including pebbly phosphatic sandstones, strongly suggests that it too was deposited in shallow water, but in quiet places bypassed by clastic material. This is the same environment sketched for the limestones of the Phosphoria and the Quadrant as well. The variables which controlled the proportions of limestone and chert in such environments offer an interesting field for investigation.

MESOZOIC

The exposed Mesozoic sedimentary rocks, subdivided into three map units, are 500 to 1,100 feet thick, but occupy less than 1 square mile. Comprising but a small part of the known Mesozoic column in adjoining areas, they are not a representative sample of the region. Indeed, exposures are so poor and fragmentary that they may not even be representative of the quadrangle. If they are representative, sandstone and siltstone make up the great bulk, perhaps three-quarters, of the deposits of the era with the remainder comprised of small volumes of limestone, mudstone, and conglomerate.

Rocks of Triassic age are absent. The oldest Mesozoic rocks are nearshore marine deposits of Late Jurassic age; these are the youngest mappable marine deposits in the quadrangle. Rocks of latest Jurassic and Early Cretaceous age are largely stream deposits.

The Mesozoic rocks are structurally conformable with the Paleozoic rocks, sharing their intense deformation.

Of probable Mesozoic (Late Cretaceous) age are a thick sequence of mainly andesitic extrusive rocks, that can readily be placed in the geologic column and could be treated as stratigraphic units. They are, however, discussed separately along with the intrusive rocks. Their stratigraphic position is shown in table 1.

JURASSIC

The Jurassic system is represented by 300–500 feet of highly varied rocks, separated into two map units, the marine Ellis formation, and the overlying continental Morrison formation. The Jurassic-Cretaceous time boundary, following regional custom (for example, see Imlay and others, 1948; Klepper, 1950, p. 67; Scholten and others, 1955, p. 367; McMannis, 1955, p. 1406), is arbitrarily placed at the lithologic boundary between the rather similar Morrison and Kootenai formations. As diagnostic fossils have been recovered from neither formation in the quadrangle, and only from the top of the thick Kootenai in neighboring areas, the system boundary may not coincide with the rock boundary.

The Jurassic rocks of the quadrangle are confined to eight scattered, mostly covered strips aggregating little more than 3 miles in length. With such information, only a minimum of discussion is in order. Fortunately, added information on the Jurassic of the Three Forks quadrangle is not urgently needed, as adequate statements concerning most elements of Jurassic geology have recently appeared at every thinkable level from the regional (Moritz, 1951), to the State (Imlay and others, 1948), to the provincial (Schmitt, 1953) to the national (McKee and others, 1956) and, finally, to the world (Arkell, 1956). On all significant points, the Three Forks quadrangle has been properly treated, considering scale.

ELLIS FORMATION

Conglomerate, limestone, and sandstone assigned to the Ellis formation crop out in all four pre-Tertiary areas. Reasonably good exposures of the full sequence are confined to the Milligan Creek and Hossfeldt Hills sectors; best exposures are in sec. 24, T. 2 N., R. 1 W.

The basal beds of the Ellis are about as resistant as the upper ones of the Phosphoria, and there is little topographic expression of the contact. In most places, where the Phosphoria appears in a steplike dip slope below a Quadrant-capped ridge, the Ellis continues the steplike terrain, ending in a sag that marks the base of the unresistant Morrison formation; in a few places, as east of Milligan Creek, the lower part of the Ellis caps a ridge, and only the upper part forms a dip slope.

The Ellis of the Milligan Creek area is 75 to 150 feet thick. A typical section includes:

	Thickness (feet)
Very thin bedded yellowish-gray limestone, with minor brown sandstone, poorly exposed	50
Thick-bedded brown sandstone, forms ledge	25
Brown, orange, and gray thick-bedded limestone	10
Brown pebble conglomerate, forms ledge	10

The conglomerate is made up of subangular to rounded pebbles of dark chert and subordinate orange siltstone, derived from the underlying Phosphoria formation, in a brown matrix of quartz and chert sand. The rock is well cemented, partly by silica, partly by calcite. Although its lithology is rather constant, its thickness is not. In some places, it is more than 30 feet thick; in others, it is absent and limestone is at the base.

In contrast to the conglomerate, the limestone in the lower part of the Ellis is rather persistent in thickness but highly variable in lithology. Within a few hundred feet laterally it grades from a brown or orange fragmental variety—containing many grains of chert and quartz sand, a few chert pebbles, and bits of pecten shells—to a dense finely crystalline medium-gray variety.

The sandstone weathers brown to a depth of several inches. The unweathered rock is gray. Subrounded medium quartz sand is the main component; subangular dark chert sand is abundant, and grains of limonite are common. Imlay and others (1948, loc. 42) describe the sandstone of the Ellis east of Milligan Creek as glauconitic, but glauconite was not mentioned by Hendricks and Gardner (*in* Gardner and others, 1946, p. 19) who measured the same section, nor recognized in this study. The rock is well cemented, partly by silica, partly by calcite. The sandstones thus are much like the matrix of the conglomerates. The sandstone strata are broken into blocks by persistent rectangular joints spaced several feet apart. On steep dip slopes, as in SE1/4 sec. 24, T. 2 N., R. 1 W., the blocks tend to slide downhill, forming spectacular talus jumbles at the base. Alexander (1955, p. 59) published excellent photographs of such piles.

The limestone at the top of the Ellis is variably microcrystalline to finely crystalline, and has a high proportion of sand and clay. The weathered rock is flaggy, with an earthy appearance and grayish-yellow color more like siltstones than typical limestones. This led Hendricks and Gardner to characterize this interval as 49 feet of "brittle calcareous sandstone * * * with some calcareous siltstone." The sequence includes a few thin beds of brown sandstone, much like that below.

In the Mud Spring sector, the Ellis formation is exposed only in two short strips in secs. 20 and 29, T. 3 N., R. 1 E. Though only half a mile apart, they are very different. In sec. 29 the upper and lower limits of the formation are easily located and its thickness is on the order of 200 feet; in sec. 20 the top of the Ellis is covered by Quaternary gravel, but the partial sequence exposed is close to 350 feet thick. The reasons for this discrepancy are not known. Exposures in sec. 29 are poor, so that knowledge of the Ellis in the Mud Spring sector rests, shakily, on observations in sec. 20, where the formation seems to comprise:

	Thickness (feet)
Thinly interbedded varicolored clayey limestone and calcareous siltstone, poorly exposed	225
Hard brown cherty sandstone, poorly exposed	35
Orange fragmental limestone	20
Mostly covered; rare outcrops are of chert pebble conglomerate	30
Thick-bedded olive-gray limestone, with chert pebbles	30

The lower half of this section seems to correlate fairly well with the entire Milligan Creek section, leaving the upper half with no correlatives at Milligan Creek.

The Ellis formation in the Hossfeldt Hills is 100–200 feet thick. The typical section comprises:

	Thickness (feet)
Covered interval. Rare outcrops are yellowish-gray siltstone and limestone	75
Thick-bedded brown chert pebble conglomerate	25
Thick-bedded yellowish-gray limestone, forms ledge	25

Locally, as much as 30 feet of uncemented breccia composed of boulder-size angular masses of chert and quartzitic sandstone separate the limestone from the underlying Phosphoria. This breccia may be tectonic but more likely is a rubble reflecting pre-Ellis weathering. The Ellis rock types in the Hossfeldt Hills all have their counterparts in the Milligan Creek and Mud Spring sectors, but the distinctive thick-bedded brown-weathering sandstones of those areas are lacking here.

In the Willow Creek area, Ellis rocks are exposed only in N1/2SE1/4 sec. 34, T. 1 N., R. 1 W. Topographic relations suggest that at least 100 feet of Ellis rocks are present, but detritus from the Phosphoria formation masks all but 3 feet of chert pebble conglomerate at the base.

Measured sections of the Ellis formation are given on pages 125, 130, 131, and 133.

Because the upper part of the Ellis and the lower part of the Morrison are alike unresistant, exposures of the contact are rare, and their relations are in doubt. The two formations are presumed conformable here, being conformable in nearby areas (Klepper and others, 1957, p. 23; McMannis, 1955, p. 1406) and regionally (Imlay and others, 1948).

NAME AND CORRELATION

The name Ellis formation was first used by Peale in 1893 (1893, map) for the rocks above his Quadrant formation and below the Cretaceous, but the rocks themselves were not described. The formation is briefly described later in his Three Forks folio (1896, p. 2–3) as being 300–400 feet thick, with thick upper and lower quartzitic sandstone separated by fossiliferous argillaceous limestone; he recognized that the basal sandstone may be Carboniferous. This basal sandstone, as noted earlier, is included in this report with the Quadrant formation, and part of the middle argillaceous limestone unit of Peale is the Phosphoria formation of this report. It is difficult to compare the rest of the Ellis as recognized here with Peale's generalized descriptions.

Berry (1943, p. 20–21) described and figured the Ellis formation from near Milligan Creek as being 170 feet thick, and including all the strata from the "white sandstone, mostly silicified" at the top of the Quadrant to a thick "brown calcareous sandstone" at the base of the Kootenai formation. I assign all but the upper 50 feet to the Phosphoria formation, and the brown calcareous sandstone that is Berry's basal Kootenai is here considered part of the Ellis.

In the Sweetgrass Arch, Cobban (1945, p. 1263) raised the Ellis to group rank and divided it into three formations, roughly comparable to the three informal divisions of Peale: sandstone-and-siltstone units at the base (Sawtooth formation) and at the top (Swift formation) separated by a limestone-and-shale unit (Rierdon formation). Imlay and others (1948, loc. 42) assigned the entire Ellis interval near Milligan Creek to the Swift formation, with the concurrence of Cobban (R. W. Imlay, written communication, Jan. 6, 1954). Thus there are grounds for applying the term Swift to all the rocks of the Ellis in the Three Forks quadrangle, but I hesitate nevertheless, because the lithologic succession can be otherwise interpreted (see, for example, Moritz, 1951, p. 1803, fig. 4, who figures an Ellis section from the "Three Forks area" as about 125 feet thick and consisting about equally of all three of Cobban's formations) and because needed faunal evidence is lacking. Consequently, the name Ellis formation is used in this report.

AGE AND ORIGIN

The only fossils recovered from the Ellis are fragments of the pelecypod *Camptonectes* studied by Imlay (written communication, Jan. 6, 1954), and of oysters, from limestone near the base. According to Imlay, "*Camptonectes* is evidence for Jurassic age." If only the Swift formation is present, the rocks are presumably of Late Jurassic age; if equivalents of the Rierdon and Sawtooth also occur, then part of the Ellis is of Middle Jurassic age (McKee and others, 1956, table 2).

The rocks of the Ellis formation are evidently of shallow marine origin. The flaggy yellowish rocks of mixed composition and fine grain in the upper part may be brackish water or estuarine deposits, heralding the retreat of marine waters from the quadrangle and the deposition of the landlaid Morrison formation.

As first recognized by Cobban (1945), accepted generally by recent workers, and shown in the maps of the Jurassic folio (McKee and others, 1956, especially pl. 8) a low island persisted in central Montana during Ellis time. The western limit of the Ellis sea was, according to the folio, close to the Idaho-Montana state line. The Three Forks quadrangle is near the southwest margin of the island, and probably received at least part of its Ellis clastic rocks, particularly the chert-pebble conglomerates, from the northeast. A westerly source for much or all the clastic rocks of the Ellis cannot, however, be ruled out.

MORRISON FORMATION

Fine-grained clastic rocks assigned to the Morrison formation crop out north of the Jefferson River and form an irregular but subdued topography. Good exposures are restricted to the Milligan Creek sector. Elsewhere, the Morrison is marked by a distinctive reddish-brown soil.

The Morrison seems to be rather constant in thickness and lithology wherever it is exposed. In the Milligan Creek sector it is 200–250 feet thick. It seems about as thick in the Mud Spring sector, where it is largely covered by Quaternary gravel, but somewhat thinner, 100–150 feet thick, in the Hossfeldt anticline.

Dominantly reddish-brown mudstone near the base, the formation becomes better sorted and coarser grained upsection, with siltstones and fine sandstones prevailing near the top, and the red-brown tone gives way to yellows and yellowish browns. This general trend is interrupted by sporadic yellowish-gray siltstones near the base and a few reddish-brown mudstone strata near the top; near the base, too, are lenses of dark-gray shale that weather bright bluish gray. Many mudstone beds, mostly reddish brown, are mottled or varicolored locally, in somber shades of red, brown, orange, and yellow in many combinations. Lenses of olive-gray microcrystalline limestone, a few feet thick and a few tens or hundreds of feet long, are scattered throughout the formation. The sandstone beds near the top are generally a few feet thick and made up largely of quartz sand, with a little clay, chert, and limonite. All the clastic rocks have calcareous cement. Carbonaceous shale and lignitic coal, widespread in the uppermost Morrison in much of Montana (Moritz, 1951, p. 1810–1811) and present in the adjoining Elkhorn Mountains (Klepper and others, 1957, p. 24) and Toston quadrangle, are not exposed in the Three Forks quadrangle, but may underlie covered areas. Two measured sections are described on pages 125 and 131.

The contact with the overlying Kootenai formation is conformable, perhaps gradational. The base of the Kootenai in this region is generally taken as the base of the lowest thick stratum of coarse sandstone rich in grains of dark chert and widely called "salt-and-pepper" sandstone (see for example, Gardner and others, 1946, p. 2; Klepper and others, 1957, p. 24; McMannis 1955, p. 1406; McKee and others, 1956, p. 3). Where the basal salt-and-pepper sandstone is well cemented, the contact with the Morrison is easy to locate. In some places, however, this sandstone is absent, or unresistant and thin, and choice of a contact is difficult, as the Kootenai contains many fine sandstone and siltstone beds much like those in the Morrison. Along such stretches the contact is merely guessed.

NAME AND CORRELATION

The Morrison and Kootenai formations of this report were placed by Peale (1896, p. 3) in the Dakota formation, of Cretaceous age. Since then, the term "Dakota" has been restricted to Upper Cretaceous rocks that apparently do not occur in southwestern Montana. Berry (1943, p. 20–21) placed the same rocks wholly in the Kootenai formation, of Early Cretaceous age, but noted that the "lower part of the section * * * may be equivalent to the Morrison and Cloverly formations" recognized nearby. Imlay (1948) measured the Jurassic section east of Milligan Creek in the course of a state-wide Jurassic study, and assigned "231 feet, siltstone and claystone, varicolored" above the Ellis to the Morrison formation of Eldridge (Eldridge and Emmons, 1896). As noted above, other workers have done the same with nonmarine dominantly red strata above the Ellis and below the lowest thick salt-and-pepper sandstone.

AGE AND ORIGIN

Although these rocks seem appropriately named Morrison formation, it does not necessarily follow that they are wholly of Late Jurassic age, as is the Morrison of the type area in faraway central Colorado (Simpson, 1926). Diagnostic fossils have not been found in the Morrison or in the lower several hundred feet of the Kootenai in the Three Forks quadrangle, or elsewhere in southwestern Montana. In the Sweetgrass Arch, more than 100 miles north of Three Forks, Cobban (1945, p. 1269–1270, 1281) found an erosional unconformity between the two formations, and Brown (1946, p. 246–247) found that the unconformity separates Jurassic plant remains from Cretaceous ones. At the southern end of the Big Belt Range,[13] and also in the Gravelly Range (Mann, 1954, p. 30–31) the Morrison is overlain with angular unconformity by rocks of Kootenai lithology (called Cloverly formation by Mann). McMannis (1955, p. 1406) also recognized a disconformity in the nearby Bridger Range but found no fossils to aid in determining whether or not it is sensibly contemporaneous with that in the Sweetgrass Arch country. It is thus possible that in and beyond the Three Forks quadrangle the arbitrary lithologic boundary between Morrison and Kootenai formations does not coincide with the equally arbitrary time-line between Jurassic and Cretaceous. In the absence of definitive new data,

[13] See footnote, p. 13.

however, the age of the Morrison formation is taken as Late Jurassic.

The Morrison exposures reveal little about their depositional environment, but the prevalence of poorly sorted red-brown mudstones and of siltstones, and the lenticularity of many layers, combine to suggest a nonmarine, largely fluvial, environment. The extremely fine grained limestones may well have been deposited in fresh-water overflow ponds or lakes.

Convincing reconstructions of the paleogeography of Morrison time (Kimeridgian according to Arkell (1956, p. 20) but commonly spelled Kimmeridgian in this country) by Imlay (McKee and others, 1956, pl. 8) indicate that the Three Forks quadrangle was then near the west edge of a great basin of continental deposition, that extended southward into Arizona and New Mexico, eastward into the Dakotas and Nebraska, and narrowed northward near the Canadian border. The provenance of the Morrison clastic rocks is therefore reasonably sought to the west, probably beyond Montana, as 250 feet of Morrison formation is present as far west as Drummond (Imlay and others, 1948, loc. 17). It should be noted that this view conflicts with the earlier interpretation of Schmitt (1953, p. 389), based on far less data: "* * * cessation of marine Jurassic deposition was accomplished by gradual withdrawal of the marine waters to the west into the Utah trough. As the waters withdrew, continental deposition occurred over a very extensive area as represented by the Morrison formation * * *" inescapably inferring an eastern source for the Morrison.

CRETACEOUS

In the Three Forks quadrangle, Cretaceous sedimentary rocks are exposed only north of the Jefferson River where they reach a maximum thickness of 650 feet. They are confined to a single landlaid formation, the Kootenai of Early Cretaceous age. This is in striking contrast to the Cretaceous sedimentary column of nearby areas, which is very thick and mostly of Late Cretaceous age: as much as 3,000 feet thick in the Elkhorn Mountains to the northwest (Klepper and others, 1957, p. 24); more than 12,000 feet thick near Whitehall to the west (Alexander, 1951, fig. 3); more than 2,500 feet thick in the Gravelly Range to the south (Mann, 1954, p. 31–32); not less than 10,000 feet thick in the Bridger Range to the east (McMannis, 1955, p. 1406–1409); and more than 3,000 feet just across the Missouri River in the Horseshoe Hills[14] (Verrall, 1951).

Studies of the Cretaceous, comparable to those for most of the older systems, have not been published on any regional scale that includes southwestern Montana.

[14] See footnote, p. 11.

KOOTENAI FORMATION

Sandstone and siltstone, with subordinate mudstone and limestone, assigned to the Kootenai formation, crops out in the three more northerly sectors of pre-Tertiary rocks. A complete section of the Kootenai is exposed at the east edge of the quadrangle in the Hossfeldt Hills area (secs. 29 and 32, T. 3 N., R. 2 E.). In the Mud Spring and Milligan Creek areas, only the lower Kootenai is exposed, in six scattered strips with an aggregate length of about 1½ miles, and in five of these the rocks have been altered, at least in color, due to igneous intrusions. Within these strips, the rocks are fairly well exposed, the resistant sandstones forming ledges and low ridges, separated by shallow depressions thinly mantled with reddish-brown soil. The fine-grained clastic rocks of the Kootenai, like those of the Wolsey shale and Three Forks shale, are preferred hosts for tabular intrusive rocks.

The Kootenai formation at the east edge of the Hossfeldt Hills area is 650 feet thick, but only 500 feet is present within the quadrangle. The formation includes:

	Thickness (feet)
Thick-bedded mottled orange to gray coarse gastropod limestone, forms ledge	10
Covered	75
Thick-bedded yellowish-brown sandstone and light-olive-gray quartzitic sandstone, forms ledge	10
Largely covered. Scattered outcrops are of gray and red limestone; main float is red and orange mudstone	200
Thick-bedded orange sandstone, with lenses of chert pebble conglomerate	50
Covered	50
Interbedded orange sandstone and olive-gray quartzitic sandstone	50
Largely covered. Scattered outcrops are brown limestone, gray claystone, red siltstone and sandstone	175
Thick-bedded orange sandstone, with much chert, forms ledge	25

For more detailed sections see pages 124–135.

Although sandstone is only about 20 percent of the formation, it provides most of the outcrops. The main variety is grayish orange and is composed of medium to coarse subangular quartz sand that contains many chert grains and chips of orange siltstone and is firmly cemented by iron-stained clay and calcite. Less abundant is an olive-gray variety that is similar in grain size and in rounding but lacks siltstone chips and is cemented largely by silica overgrowths; this is the common salt-and-pepper sandstone of the Kootenai throughout the region. Because this variety does not appear in the basal sandstone, but is 200 feet higher and above a thick covered interval that produces reddish soil, it might be argued that the two lower units of this section belong in the Morrison formation. Most of the other Kootenai exposures in the quadrangle, however, have no gray chert-bearing sandstone whatever, but all have basal sequences of orange or yellowish-gray chert-rich coarse sandstone. Orange or yellowish-gray chert-bearing sandstone must be accepted as diagnostic of the Kootenai in the Three Forks quadrangle, or the formation cannot be mapped except in the Hossfeldt Hills.

The limestone at the top of the formation, though thin, is persistent in the low hills just east of the quadrangle; similar gastropod-bearing limestone is at or near the top of the Kootenai in most of the region (Peale, 1896, p. 3, under heading Dakota formation; Klepper and others, 1957, p. 25; Freeman, 1954, p. 10–14; Scholten and others, 1955, p. 356) but not in the Bridger Range (McMannis, 1955, p. 1406).

In even the best exposures, at least 70 percent of the formation is covered. Float and scattered small outcrops suggest that the covered intervals are underlain mostly by thin, impersistent brown and red calcareous mudstone and siltstone, and subordinate red fine calcareous sandstone, brown coarsely crystalline sandy limestone, and gray claystone. Gray claystone is associated with lignitic coal in the upper part of the Kootenai in the Toston quadrangle, but no coal is exposed in this interval in the Hossfeldt Hills.

In the Milligan Creek sector, the lower 100 to 200 feet of the Kootenai is fairly well exposed. The lower 100 feet is largely thick-bedded faintly laminated yellowish-gray chert-rich medium-grained sandstone, with calcareous cement, much like the basal sandstone in the Hossfeldt Hills. The succeeding 100 feet, exposed only in secs. 24 and 13, T. 2 N., R. 1 W., includes similar sandstone, but in thin beds, intercalated with thin lenses of yellow- and reddish-brown siltstone and fine sandstone. Much of the sequence is reddened and hornfelsed by a large andesitic intrusive. The entire Kootenai is missing for about 500 feet in SE1/4 sec. 24, where the andesite is in contact with the upper part of the Morrison.

Several hundred feet of Kootenai appear to be present at the west edge of the Mud Spring area (SW¼ sec. 20, T. 3 N., R. 1 E.) but exposures are poor, and the rocks are disturbed and altered by a nearby intrusive mass. Principal outcrops and float are greenish-gray hornfels, presumably altered mudstone or siltstone. If these were originally red or brown, the common color of such rocks in the Kootenai of the Hossfeldt Hills, their colorants, presumably hydrous iron oxides, have evidently been reduced owing to heating by the intrusion, just the opposite of the oxidizing (= reddening) effect of the intrusion near Milligan Creek.

Just east of the quadrangle, the Kootenai is overlain conformably by varicolored sandstone and siltstone of the Upper Cretaceous Colorado shale. (Peale's geologic map (1896) shows large areas of "Colorado and Montana formations" of Cretaceous age along upper Milligan Creek, lower Mud Spring Gulch, and in the hills in the northwest corner of the quadrangle. The supposed Cretaceous rocks in the two valleys are fossiliferous Tertiary deposits, and those in the hills are igneous rocks assigned to the Elkhorn Mountains volcanics.) Within the quadrangle, the Kootenai is overlain with angular unconformity by lower Tertiary rocks of the Bozeman group.

NAME, AGE, AND ORIGIN

As noted earlier, the rocks divided into Morrison and Kootenai formations in this report were classed together as Dakota formation by Peale but as Kootenai formation by Berry. In this, Berry followed Fisher (1908) who extended the term from Canada into northern Montana. Fisher (1908, p. 18–19) recognized the presence of probable Morrison formation below his Kootenai in northern Montana, and Berry suspected it near Three Forks. Restriction of the term as made in this report is consonant with current regional usage.

No diagnostic fossils were recovered from the Kootenai. The gastropods in the uppermost limestone are numerous but invariably replaced by coarsely crystalline calcite and by silica in a manner that destroys their characterizing internal structure. Samples from uppermost limestone of the Kootenai in the Toston quadrangle were studied by Reeside (written communication, Nov. 23, 1955) who was able to find only a few roughly determinable shells:

Unio sp.
Reesidella montanensis (Stanton)?
Gyraulus sp.

These are not definitive as to age, but serve to confirm the formation as Kootenai, which is demonstrably of Early Cretaceous age as near by as the Great Falls region (Fisher, 1908, p. 20), the Harlowton area (Yen, 1951, p. 1–3), and the vicinity of Dillon (Yen, 1951, p. 1–3). While at least part of the Kootenai near Three Forks is assuredly of Early Cretaceous age, the possibility remains, as discussed earlier, that the lower part of it may be of Late Jurassic age; or, conversely, that part or all the Morrison formation may be of Early Cretaceous age. In the absence of significant new data, however, the Kootenai is tentatively assigned the customary Early Cretaceous age.

The rocks of the Kootenai are much like those of the Morrison, and seem to have been deposited under similar nonmarine conditions. The thick and extensive coarse sandstones in the lower part of the Kootenai do not have their counterpart in the Morrison, and indicate a distinct but mild orogenic interruption in a general history of fluvial lowering of a region of low relief during Morrison and Kootenai time.

CENOZOIC

Cenozoic sedimentary rocks, subdivided into nine map units, dominate the quadrangle, underlying 150 square miles. The Cenozoic units, all of nonmarine origin, have a maximum aggregate thickness of 3,000 feet, but the total local thickness is probably nowhere greater than 2,500 feet, and is generally less than 1,000 feet. The Cenozoic column is made up mostly of lower Tertiary rocks, but the surface exposures are largely of thin unconsolidated Quaternary deposits.

Long ago, Peale (1896, p. 3) recognized the broad geologic similarity among the Cenozoic intermontane basins of western Montana:

> The movements that began during the latter part of the Cretaceous period ultimately resulted not only in the folding of the strata but also in the formation of the numerous basins or valleys found throughout the mountainous portions of Montana.

The early paleontologic work of Douglass (especially his papers of 1899, 1902, and 1909) and of Matthew (1903) confirmed and strengthened this concept. But no systematic study of the Cenozoic geology of any part of southwestern Montana has yet appeared. The nearest approach to a systematic local study has been in the Canyon Ferry quadrangle, where Mertie (1951, p. 5–6, 29–43, 55–56, 84–85) subdivided the Cenozoic into six mapped units—four Tertiary and two Quaternary. Extensive fossil collections have been made by members of the Carnegie Museum of Pittsburgh, the Frick Laboratory, Princeton University, the University of Michigan, University of Chicago, and other organizations, but little publication has yet resulted. Some fairly detailed work on local Tertiary deposits and history appears in recent unpublished doctoral theses.[15] This report attempts to treat the little-known Tertiary rocks in somewhat more detail than the older rocks. The Quaternary rocks, though mapped in considerable detail, are otherwise summarily treated.

Large-scale syntheses of Cenozoic history in western Montana have been ventured by Atwood (1916), Pardee (1950, based on work done before 1947), and Alden (1953, based on work before 1948), and briefer summaries have been given by Pardee and Schrader (1933, p. 3–6, 24–28) and Lowell (1956a).

[15] See footnotes, p. 11, 13, and 111.

TERTIARY—BOZEMAN GROUP

The Three Forks basin is carved largely in soft, little-deformed Tertiary rocks that were laid down in a basin of dimensions similar to the present one. The Three Forks quadrangle, in the northwest part of the modern basin, is also near the northwest edge of the Tertiary basin, as shown first by Peale (1893, pl. 1). In the quadrangle, continental sedimentary rocks of known or probable Tertiary age are exposed over 35 square miles and are thinly mantled by unconsolidated Quaternary deposits over 40 square miles more. Subdivided into four formations, they have a maximum aggregate thickness near 2,500 feet, but such a figure means little, as the pre-Tertiary surface had much relief.

Principal rock types are siltstone, 30 percent; sandstone, 25 percent; conglomerate, 20 percent; clay, 20 percent; limestone and marlstone, 5 percent. About half of the Tertiary deposits is derived from erosion of pre-Tertiary rocks, and half is penecontemporaneous pyroclastic debris. Few beds are free of Tertiary volcanic ash, and many strata are made up wholly of such material. The Tertiary rocks are of Eocene, Oligocene, late Oligocene or early Miocene, late Miocene, and early Pliocene age, but only the Eocene and lower Oligocene rocks are shown on the map (pl. 1); the outcrops of upper Oligocene–lower Miocene, upper Miocene, and Pliocene rocks are too small to map.

NAMES AND CORRELATION OF UNITS

The Tertiary rocks of the Three Forks quadrangle were designated Bozeman lake beds by Peale (1893, pl. 1; 1896, p. 3), who regarded them as deposits in one of a series of coexisting lakes widespread in western Montana in "Neocene" (that is, post-Eocene, pre-Pleistocene) time. He concluded that some of the detritus was derived by erosion of the pre-Neocene rocks bordering each lake but that the bulk of the deposits is "fine volcanic dust" blown into the basin by winds; that a little fell on the water and settled to form "pure white sediments;" but that most of it fell on land and was washed into the lake along with a small proportion of foreign fragments, to form layers of "rusty color." In ascribing a lacustrine origin to these deposits he was following the ideas of his colleague Hayden, who, as early as 1869 (p. 114–115), stated that "from the dawn of the Tertiary period, even up to the commencement of the present, there was a continuous series of fresh-water lakes all over the continent west of the Mississippi River," and, more specifically of the headwaters of the Missouri: "These * * * broad valleys have all been lake-basins during the last portion of the Tertiary period"

(Hayden, 1872, p. 147). Peale's lacustrine thesis was accepted by his contemporaries Weed and Iddings (1894) who used the name Bozeman lake beds, and Douglass (1899), who did not use the name. Beginning with Matthew (1899), however, all other writers have opposed the idea that the basin deposits are largely lacustrine, favoring instead various combinations of fluviatile and eolian processes (see Davis 1900, and Osborn, 1909, p. 24–28 for summaries of early views), and the established name Bozeman lake beds has been avoided by the use of informal terms such as Tertiary deposits, or "basin beds" (Scholten and others 1955, p. 368) or used reluctantly, with actual or implied deprecatory quotation marks.

In the Three Forks quadrangle, rocks that Peale assigned to the Bozeman lake beds have been divided in this report into three gradational map units, designated as new formations. In ascending order, they are Milligan Creek formation, mainly limestone, of Eocene age; Climbing Arrow formation, mainly bentonitic clay and sand, of late (and middle?) Eocene and early Oligocene age; and Dunbar Creek formation, mainly tuffaceous siltstone, of Oligocene age. The Thompson Creek beds of Douglass (1902) and the Titanotherium beds of Matthew (1903) are part of the Climbing Arrow formation. The Oreodon beds of Matthew are difficult to locate from his description, but probably are from the lower part of the Dunbar Creek formation. The Milligan Creek and Dunbar Creek formations are largely lake beds with high proportions of volcanic ash; the Climbing Arrow is mostly stream laid; a very small fraction of the Climbing Arrow and possibly a larger fraction of the Dunbar Creek may be tuff, windborne into the basin and deposited on the dry basin floor. The three formations are here regarded as local, though it may eventually be feasible to trace them into adjoining basins.

Sharing many features, it is fitting that these rocks be designated a group; in deference to Peale's prior usage but to avoid debatable genetic implications, the name Bozeman group is proposed. Also regarded as part of the Bozeman group are small bodies of limestone conglomerate, with distinctive orange matrix, that gradationally underlie the Milligan Creek formation, are probably of Eocene age, and are assigned to the Sphinx conglomerate (Peale, 1896, p. 3); and unmappably small patches of Miocene and early Pliocene gravel. The limestone conglomerate is lithologically like parts of the "Laramide" (largely Paleocene) Beaverhead conglomerate of Lowell and Klepper (1953). The Miocene and early Pliocene gravel patches are apparently outliers of the thick units of similar age exposed in the east bluffs of the Madison River and called Madison Valley formation by Douglass (1907); they have been recently discussed by Dorr (1956).

In the Three Forks quadrangle, the Bozeman group overlaps an erosion surface that developed during and soon after Laramide deformation, which cannot be more closely dated than Late Cretaceous-early Eocene. The youngest rocks below this surface are the Elkhorn Mountains volcanics, of Late Cretaceous age.

At the top of the Bozeman group, upper Tertiary gravels lie with great erosional, and possibly some angular, unconformity on the Climbing Arrow formation. This unconformity, involving many hundred feet of Climbing Arrow and Dunbar Creek rocks, and unknown thicknesses of any other units which may once have intervened, might at first glance seem a natural top for the Bozeman group. But because basin deposits of Miocene and Pliocene age are thick and extensive just across the Madison River and were likely once abundant in the Three Forks quadrangle too, before Quaternary uplift and erosion, their remnants should be included with the Bozeman group. To omit them would be to bisect a useful genetic unit in both time and space, and place an undesirable restriction on Peale's original concept.

To recognize that Tertiary deposits are thick and have a wide range in age is not to imply that the Tertiary history of the Three Forks basin was largely depositional. All the mappable deposits in the western part may well have been laid down in a brief episode that extended from late Eocene or at most middle Eocene to early Oligocene time. The interval from middle Oligocene through middle Miocene time is represented by virtually no deposits anywhere in the basin and was likely a time of slow erosion. Upper Miocene deposits in the eastern part are thick and extensive but are largely tuffaceous sediments and may have been deposited fast. Pliocene deposits are thin and apparently very local. Thus the Tertiary was more a time of erosion than of deposition.

The rocks of the Bozeman group, laid down in a basin, are confined to the low parts of an existing basin, and have obviously not been deformed on a scale remotely approaching that of the pre-Tertiary rocks. They have, nevertheless, been considerably deformed. In fine-grained, even-bedded strata, presumed to have been horizontal, common dips are 5°–10°; rarely such rocks dip as steeply as 25°, and dips as steep as 60° are recorded. Many conglomerates and sandstones dip 15°–35°, but a large fraction of such dips is probably initial. No faults have been mapped in the Bozeman group, though at least one is suggested by relations near Dunbar Creek, and others may be concealed. At any rate, there are no faults profound enough visibly to affect the surrounding pre-Tertiary rocks.

The Bozeman group is on the whole poorly consolidated, and unresistant to erosion. Broad benches have been cut across it, and thin but continuous soils have been developed on these benches, where they have not been covered by Quaternary deposits. Consequently, although exposures are generally fair on valley walls, they are poor elsewhere.

Owing to the discontinuous exposure and inconstant dips, only the grossest units and structures can be traced.

It is hoped that the Bozeman group—defined as the Tertiary fluvial, eolian and lacustrine rocks which accumulated in the basins of western Montana after the Laramide orogeny—may become a useful regional term. If the concept is to be useful it must have rather elastic age limits, for it is not only reasonable but known that the earliest and latest deposits from basin to basin are not of identical age. Upper Eocene basin deposits have long been known from Sage Creek, in southwesternmost Montana (Douglass, 1903, p. 145–146; recently restudied by Hough, 1955) and now from the Three Forks quadrangle, but in most places the oldest basin fills seem to be lowest Oligocene, as indicated in the most recently issued Tertiary correlation chart (Wood and others, 1941). The youngest fossiliferous basin deposits in the Three Forks vicinity are of Pliocene age, but in basins farther west the youngest dated deposits of the basin-filling sequence seem generally of late Miocene age, though middle Pliocene basin deposits occur in Deer Lodge Valley (Konizeski, 1957). Both the earliest and latest time of filling may in many places not be datable due to lack of fossils or exposures.

In several places in southwestern Montana, the oldest Tertiary sedimentary rocks are coarse conglomerate whose potential correlation with the Bozeman group is debatable. In Beaverhead County such rocks are named Beaverhead formation and are shown to have formed in Late Cretaceous to early Eocene time during the late stages of the Laramide orogeny (Lowell and Klepper, 1953). The Beaverhead and similar rocks could be considered the earliest deposits in the Tertiary basins, and thus classified as basal units of the Bozeman group, as I have done with the Sphinx conglomerate. A different but tenable view is that they are synorogenic rather than postorogenic and are better kept separate.

Certain other rocks associated with the intermontane basins are plainly not referable to the Bozeman group as defined above. Glacial deposits, even if of Tertiary age, like the supposed Eocene Black Butte till (Scott, 1938) should be excluded, as should mappable igneous flow rocks. In Quaternary time, low-level lakes existed at several places in valleys cut into the Tertiary basin deposits and older rocks. Such a lake occupied part of the Madison River valley near Ennis, according to R. E. Swanson (oral communication, 1955). The deposits in and on the margins of this and similar lakes naturally resemble the Tertiary basin deposits (so much so that Peale, 1896, mapped those near Ennis with the Neocene Bozeman lake beds) but should, of course, be distinguished from the Bozeman group wherever possible.

The high-level gravel that overlies the Bozeman group cannot be closely dated. It may be of very late Tertiary age, but in this report it is arbitrarily assigned to the Quaternary, as are certain other somewhat younger fossiliferous deposits.

SPHINX CONGLOMERATE

At the base of the Bozeman group and assigned to the Sphinx conglomerate is limestone conglomerate with a reddish-orange calcareous matrix that appears in small outcrops at many places in the quadrangle. Every such exposure is near the present basin floor at the foot of a limestone ridge. Most of the bodies are too small to map separately. As seen in valley walls, most of the masses are less than 10 feet thick and less than 100 feet long; their third dimension is unknown but probably small. Four large patches are, however, shown on the geologic map: two at the ends of Milligan Canyon (fig. 6) and two on the south side of Jefferson River valley above Willow Creek. The largest masses have exposed thicknesses of around 100 feet. The subsurface extent of one, the mass above

FIGURE 6.—Sphinx conglomerate of Eocene age, above Milligan Canyon. East side of valley, SE¼SW¼ sec. 25, T. 2 N., R. 1 W. Thin dark layers are reddish brown bentonitic mudstone. Compare with figure 4, showing limestone breccia of cave origin in Mission Canyon limestone, half a mile downstream.

Milligan Canyon, is partly known as a result of drilling by the Bureau of Reclamation (Harold C. Elliott, oral communication, Aug. 5, 1957). This body is 30–50 feet thick below about 30 feet of alluvial cover for not less than 600 feet upstream from the northern end of the exposure; and it lies on pre-Tertiary rocks.

Just east of the quadrangle, in NW¼NW¼NW¼ sec. 28, T. 1 N., R. 2 E., a test well (Ryan, Tom Tice no. 1) penetrated a thick mass of conglomerate that is probably referable to this unit. The well penetrated a sequence of more than 300 feet of "colored" conglomerate, mostly orange, beneath a succession 2,000 feet thick of "sand" and "shale" with streaks of coal, presumably strata equivalent to the Dunbar Creek and Climbing Arrow formations.

The conglomerate is composed mainly of subangular pebbles and cobbles of dark fine-grained Paleozoic limestone. Larger clasts are rare but boulders as much as 3 feet long occur. A few of the stones are light-colored Paleozoic limestone, siltstone, and quartzitic sandstone. In addition, in the conglomerate above Willow Creek, gneiss and schist pebbles are common, and rarely present are hematite pebbles derived either from hematitic phyllonite of Precambrian age or hematitic layers in such Paleozoic formations as the Wolsey, Big Snowy, or Amsden. In places the stones are angular, and the rock might better be called breccia. In others, many of the stones are subrounded. Whether rounded or angular, the conglomerate is everywhere poorly sorted and obscurely bedded. The matrix, which makes up 20–40 percent of the rock, is earthy, porous, and generally reddish orange, but in places is reddish brown or yellowish gray. The sand-sized components are mostly quartz, but include feldspar, altered mica, and bits of noncarbonate rocks, firmly cemented by clay minerals and by powdery calcite. The orange color is confined to the matrix, the gravel presenting only the normal colors of the parent formations. The matrix color seems due to staining by finely disseminated ferric oxides and not to the body colors of component grains as in the somewhat similar Beaverhead conglomerate (Lowell and Klepper, 1953, p. 239). Conglomerate in the main mass above Milligan Canyon (trenched by the Bureau of Reclamation) contains layers a few inches thick of dark-reddish-brown bentonitic sandy siltstone and clay.

The conglomerate is variably deformed. Near Milligan Canyon the unit has low dips, conforming to those of neighboring Tertiary rocks. In the small mass in SE cor., sec. 25, T. 1 N., R. 1 W., the strata dip 25° northward, apparently passing below Milligan Creek formation to the north, but the contact area is concealed by alluvium. Bedding elsewhere is so poor that attitudes are unknown.

The conglomerate lies with angular unconformity on the Mission Canyon limestone. Below Milligan Canyon, the upper part of the limestone conglomerate is interbedded with tuffaceous limestone and coarse channel sandstone of the Milligan Creek formation, demonstrating gradational relations.

AGE AND ORIGIN

The conglomerate is younger than the Carboniferous Mission Canyon limestone it lies on. It is older than the Milligan Creek formation but not greatly. As the Milligan Creek formation, discussed later, is of demonstrated middle or late Eocene age, the conglomerate is also of Eocene age.

Similar rocks are widespread though not abundant in the Quaternary and Tertiary deposits of this region, and they are probably common throughout the Rocky Mountains. Composed mainly of limestone, and with a calcareous matrix, they differ greatly from the widespread and thick red-banded, but carbonate-poor, rocks of the Rocky Mountain region that have been much discussed (for summary and analysis see Van Houten, 1948). The fact that the conglomerates are mostly limestone pebbles, little rounded, demonstrates that they were not deposited in standing water or handled by perennial streams. Their mixed composition and slight rounding show that they are not merely talus, though they have surely not moved far. Their topographic position and texture suggest deposition by intermittent streams or by mudflows in fanlike forms. Except for the matrix, they are like ordinary alluvial cones or fans.

But the reddish-orange calcareous matrix offers problems, for redness and high carbonate content seem generally antithetic in land-laid sediments, though common enough in marine ones. Very likely the two properties originated separately. Perhaps the orange clayey fraction of the matrix represents lateritic soil developed on local uplands of low relief during warm and wet episodes in earliest Tertiary time. Such an origin would be in keeping with the manner and conditions of formation postulated by Van Houten (1948) for the red material in the early Cenozoic deposits of the region generally. Later, accelerated erosion might have stripped the clayey soil and mixed it as it moved downslope with limestone fragments shed from steep valley walls free of residual soil. The mixture might then have been deposited at lower altitudes, orange in color, but with little interstitial carbonate; the carbonate matrix might have been deposited later by ground water. Such a process would be easier to accept if any remnants of earliest Tertiary upland paleosols were known. Considering the subsequent history of the region it would be astonishing if any were ever identified, however widespread they may once have been. Suggestive of an upland paleosol of suitable age nearby is the "pebbly red mantle as much as 30 or 40 feet thick" at the base of the Tertiary tuffaceous sequence along the Boulder River in the Devils Fence quadrangle (Klepper and others, 1957, p. 42).

Worth considering is a reverse possibility—that the orange color developed later, after deposition under semiarid conditions of gray fan gravel with a lime-rich matrix. Perhaps the orange color was developed by ground water, partly by oxidation of ferrous iron from the carbonate rocks, and partly by physical redistribution of ferric iron washed from hematitic pebbles derived from a variety of iron-rich pre-Quaternary rocks. This concept, while not requiring vanished upland paleosols, is opposed by the sharp contacts of gray limestone pebbles with the orange matrix. If the color is postdepositional, the surfaces of the pebbles might be expected to be discolored as well as the matrix.

The Sphinx conglomerate may well have formed a widespread apron at the foot of the mountains that were the surface expression of Laramide uplift and accelerated erosion. Deposits formed under such conditions are likely to be stripped as erosion proceeds so that only a few fortuitously covered and preserved fragments tend to remain in the geologic column. Detailed study of such remnants on a regional basis would surely be rewarding.

MILLIGAN CREEK FORMATION

NAME AND DISTRIBUTION

The Milligan Creek formation crops out only in the southwest part of the quadrangle, on both sides of the Jefferson River valley. It is here named for Milligan Creek, a southward-flowing tributary of the Jefferson, which near its mouth cuts through and exposes rocks typical of the unit. The Milligan Creek formation is made up of light-colored fine-grained tuffaceous lake deposits, mainly limestone but ranging from limestone through marlstone to calcareous mudstone, and interfingering stream-channel sandstone and conglomerate.

The formation is exposed over an area of 3 square miles and underlies a thin covering of Quaternary deposits in perhaps 2 square miles more. Its original extent was less than 100 square miles, perhaps less than 50. West of Three Forks, distribution and topographic setting of exposures indicate that the formation was never more than 6 miles broad and suggest that its western margin may have been within the quadrangle, though it may have extended a mile or two into the Jefferson Island quadrangle. The formation probably does not extend more than a few miles east of its most easterly outcrop (sec. 16, T. 1 N., R. 1 E.) as several borings near the east edge of the quadrangle penetrated little or no limestone, the characterizing rock type of the Milligan Creek; there is no reason to think it ever extended farther east. The formation may have extended south a few miles beyond the southwest corner of the quadrangle to the base of high hills in the Precambrian crystalline rocks. In the north half of the quadrangle the Milligan Creek formation does not appear between the younger Tertiary units and the pre-Tertiary rocks. It might have been eroded before deposition of the younger units, or been laid down only locally and not yet exposed, but the simplest interpretation is that no Milligan Creek rocks were ever deposited.

A very small patch of Milligan Creek formation may, however, be represented by limestone (called "Cambrian" on the driller's log) in the bottom of a test well (Dunbar Oil Co., Dunbar no. 3) 534 feet deep in NE¼NE¼ sec. 11, T. 2 N. R. 1 W., on the west flank of an anticline that exposes the lowest recognized Climbing Arrow strata; similar limestones were not met in much deeper tests a few miles to the west and east.

It would be unsafe to predict the presence of much Milligan Creek formation beneath the Jefferson River flood plain. In the few places where attitudes can be measured near the riverward edges of the formation, the rocks are horizontal or dip away from the river. Possibly, the Jefferson River has completely removed the Tertiary rocks in this stretch, and filled the cavity with Quaternary alluvial deposits.

TOPOGRAPHIC EXPRESSION

The Milligan Creek formation appears in near-white dissected benchlands, thinly mantled with light-colored sandy soil, between the rugged hills of pre-Tertiary rocks and the Jefferson River flood plain. The limestone of the formation, unlike that of the pre-Tertiary rocks, is not very resistant to weathering, owing to its high proportion of altered volcanic glass and clay, and the fine-grained lake clastic rocks are even less resistant. Where the fine-grained rocks dominate, as east of Willow Creek, exposures are poor. Where the formation has a high proportion of conglomerate, as near Milligan Creek and west of Willow

Creek, it is more resistant, producing fairly prominent outcrops. Throughout, the fine-grained strata wear back to produce wide steps capped by sandstone and conglomerate.

LITHOLOGY AND THICKNESS

The bottom of the Milligan Creek basin is nowhere exposed, being covered by Quaternary alluvium or perhaps never reached by prealluvium downcutting. The exposed gradational contact with underlying Sphinx conglomerate, mentioned earlier, is at the edge of the basin and the strata there are probably unlike the lowest beds farther out. The original top of the formation is visible only in the NW¼ sec. 16, T. 1 N., R. 1 E., and there poorly; elsewhere, it has been removed by erosion, and in many places the formation is unconformably covered by Quarternary gravel or silt. The best exposures of the formation are in the low hills north of Highway 10S and west of Milligan Creek. This vicinity (specifically, E½ sec. 11, NW¼ sec. 12, and SW¼ sec. 1, T. 1 N., R. 1 W.) is therefore designated the type area of the formation.

A composite partial section, about 300 feet thick, from the type area, includes:

	Thickness (feet)
Top of hill.	
Limestone: thick-bedded massive, yellowish-gray to very pale orange; glassy, hard, sugary, with many tiny irregular vugs; scattered earthy yellowish-gray rounded sand and granules of limestone and clay	25
Limestone: thin- to thick-bedded, platy to flaggy, yellowish-gray to white; earthy, hard, microcrystalline; much clay, mostly in silt-size fragments; small irregular silicified patches. A few lenses, as much as 3 feet thick, of subangular pebble conglomerate and nearly white subangular sandstone	60
Marlstone and limestone: thick-bedded light-olive-gray earthy marlstone, weathering white, interbedded with subordinate amounts of dusky-yellow and yellowish-gray finely crystalline clayey limestone. All beds contain from traces to several percent of clay pellets mostly in silt and sand sizes	100
Limestone: thin-bedded, white, clayey, hard, microcrystalline, with many tiny vugs; small irregular silicified patches. Many sand- to granule-size pellets of white clay; a few rounded grains of glassy quartz sand. A few layers, as much as 3 feet thick, of light-olive-gray marlstone and of white coarse subangular sandstone	60
Limestone and marlstone: thick- to thin-bedded, light-yellowish-gray and yellowish-brown fine-grained strata ranging from calcareous mudstone to clayey limestone. Many sand-size pellets of clay	50
Bottom of gulch.	

Some of the strata here called marlstone may contain enough clay and silt to be called calcareous mudstone. Locally, the limestone is replaced by laminae of white, red, or green chert.

A typical earthy tuffaceous limestone from the upper part of the formation west of Willow Creek (sample 13, pl. 2; NE¼NW¼ sec. 6, T. 1 S., R. 1 E.) consists mostly of large fragments in a very fine matrix. As figure 7 shows, more than a quarter of the rock is coarser than coarse sand and more than half is silt and clay; the four intervening sand classes comprise less than a fifth.

The large fragments, rounded to subangular, are mostly clay with a little coarse vitric ash (these and subsequent terms for pyroclastic material are used as defined by Wentworth and Williams, 1932) variably altered to clay; the sand and silt fraction consists mostly of aggregates of tiny calcite crystals with subordinate clay aggregates; and the clay fraction is mostly clay minerals but with a heavy sprinkling of calcite crystals. Minor sand and silt components are angular fragments of light-green hornblende, brown biotite, chlorite, quartz and feldspar, both twinned and untwinned. The large clay fragments resemble the ash fragments in shape and color and there are all gradations between fairly fresh ash and pure clay. All the large clay fragments are probably thoroughly altered ash, and at least some of the clay of the matrix is also altered ash. Glass in the ash has a refractive index near 1.50.

Two X-ray analyses (one [Lab No. 141369] by C. J. Parker; the other [Serial No. 226415] by A. J. Gude, III) also show that the clay and silt fraction of the rock contains mostly calcite and montmorillonite with very small amounts of feldspar (labradorite, according to Gude), chlorite, and quartz.

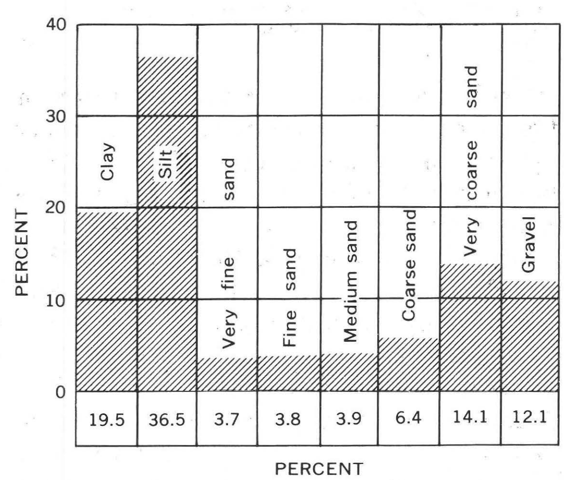

FIGURE 7.—Mechanical analysis of typical tuffaceous limestone from the Milligan Creek formation. Sample from locality 13 (see pl. 2); laboratory no. 141369. Grain-size classification according to Wentworth scale. Analysis by P. R. Blackmon.

A chemical analysis[16] (column *A* in table 3) is not easy to interpret in terms of mineral composition, even with the information given by the other methods. Assuming that essentially all the CO_2 in the analysis represents calcite, the rock has about 66 percent of this mineral, which is somewhat more than the visual estimate. But neither the microscope nor X-ray analysis reveals any mineral phase that would account for the rather high phosphate content: presumably the phosphate is present submicroscopically in organic matter amorphous to X-rays. If the remaining material is wholly volcanic ash, it should be possible to approximate its chemical and thence its normative composition by subtracting calcite, phosphate in the form of normative apatite, and excess water, and then recalculating. This operation gives the compositon shown in columns *B* and *C* of table 3.

TABLE 3.—*Chemical analysis of typical tuffaceous limestone from the Milligan Creek formation, locality 13 (see pl. 2)*

	A	B	C
SiO_2	19. 8	57. 5	64. 7
Al_2O_3	4. 5	13. 0	14. 6
Fe_2O_3	1. 8	5. 2	5. 7
FeO	. 10	. 3	. 3
MgO	. 52	1. 5	1. 7
CaO	39. 9	7. 8	8. 7
Na_2O	. 41	1. 2	1. 3
K_2O	. 50	1. 4	1. 5
TiO_2	. 18	. 5	. 6
P_2O_5	. 30	----------	----------
MnO	. 06	. 8	. 9
H_2O	3. 9	11. 3	----------
CO_2	28. 9	----------	----------
Sum	101	100. 5	100

A. Analysts, P. L. D. Elmore, and K. E. White (rapid method).
B. Recalculated free of $CaCO_3$ and $Ca_3(PO_4)^2$.
C. Recalculated free of water.

Even with allowances for reasonable variation this rock fits none of the average rocks of the common igneous groups as calculated by Nockolds (1954), nor the glass-rich rocks of the Climbing Arrow formation (columns 7 and 8 in table 6) nor those of the Dunbar Creek formation (columns 9, 10, 11, table 8*B*). The principal difficulty is in the CaO content, far too high for rocks of comparable SiO_2 and Al_2O_3 content; part of the calcium may be present as sulfate, for many similar rocks contain visible gypsum, but this point cannot now be established as SO_4 was not determined. The total iron is also too high, and Na_2O and K_2O too low. The noncarbonate part of this rock thus seems to be polygenetic. It is nevertheless likely that the ash fraction is chemically similar to the ash component of the Climbing Arrow formation, as the refractive index of their glass is similar. The ash of the Climbing Arrow is in the range of calc-alkali rhyolite in Nockolds' classification.

Some limestone and marlstone strata in the formation contain tests and casts of small gastropods and clams, as well as charophyte oogonia, and, rarely, ostracodes.

The conglomerate interbeds, typically 1–3 feet thick and a few tens or hundreds of feet long, are made up of quartzitic sandstone, vein quartz, and much lesser amounts of intermediate coarse-grained igneous rocks. Rare components are chert and mud balls. The matrix is mostly subangular quartz sand, with a little feldspar, chert, and clay. Induration varies from poor, where the rock is loosely bound by clay and a little carbonate, to excellent, where the rock is solidly cemented with silica, usually clear and crystalline but in places milky and chertlike. The sandstone layers are similar in composition to the matrix of the conglomerate, but are rarely well indurated, being loosely cemented by clay and carbonate.

The outcrops on the east side of Milligan Creek are much like those of the type area, but with a higher proportion of platy rather than massive limestone and of conglomerate and sandstone lenses.

Between these two masses of mostly fine grained beds is an area underlain by a thick body of cross-stratified conglomerate and sandstone, best exposed on the north side of the eastward-draining gulch in S½NW¼ sec. 1, T. 1 N., R. 1 W. The strata, commonly a few feet thick and dipping 5°–10° SE., are alternating coarse pebbly sandstone and subround to round pebble conglomerate, well cemented with silica. Fragments of many of the pre-Tertiary sedimentary formations above the Madison group and of a variety of igneous rocks are present. The principal rock types are quartzitic sandstone, andesitic volcanic, and monzonite. Limestone is rare; shale, siltstone, and chert are absent. Because of its position and its lithologic similarity to the sandstone ond conglomerate interbedded with fine-grained Milligan Creek facies, this mass is regarded as part of the Milligan Creek formation. Actual interbedding with known Milligan Creek rocks is, however, not visible, and the conglomerate-sandstone body may be a part not of the Milligan Creek formation but of the Climbing Arrow formation, deposited in a deep channel cut into the Milligan Creek formation.

The formation west of Willow Creek differs from the type area in several respects. Fine-grained rocks,

[16] This analysis and all subsequent chemical analyses of rocks cited were made by rapid methods similar to those described by Shapiro and Brannock (1956). The results, summed to the nearest percent, are entirely adequate for the purposes of this report. Calculations made on the basis of such analyses are, of course, correspondingly inelegant but serviceable.

including a few beds of bentonitic siltstone, mudstone, and shale, in addition to the dominant limestone and marlstone, make up only about 40 percent of the sequence, the remainder being beds a few feet thick of poorly rounded rudely cross-stratified coarse sandstone and pebble conglomerate, locally cemented with silica and very hard. The sandstone and conglomerate are much like their counterparts across the river, dominated by quartz and quartzitic sandstone, but differ in having considerable chert and a few fragments of gneiss and amphibolite, and in lacking andesitic and monzonitic debris.

East of Willow Creek, coarse layers are rare, and the principal rock types are thin-bedded yellowish-gray earthy tuffaceous calcareous mudstone and light-brownish-gray clayey limestone. Present here are a few lenses, less than a foot thick, of rounded limestone-pebble conglomerate, with yellowish-gray muddy matrix.

The maximum exposed thickness of the Milligan Creek formation is 300 feet. Although the base is not exposed, it probably is not far below the lowest exposures, and the total thickness is not much greater than the exposed thickness. In the type area and east of Willow Creek the exposed thickness is about 300 feet; west of Willow Creek and east of Milligan Creek less of the formation is present, perhaps no more than 200 feet. The conglomerate-sandstone body west of Milligan Creek is as much as 150 feet thick.

The Milligan Creek formation seems to be conformably overlain by the Climbing Arrow formation but exposures of the contact, limited to a few places southwest of the Buttleman Ranch buildings, are too poor to establish this with certainty. Both formations in this vicinity are nearly horizontal and their lithologies appear gradational, for the Milligan Creek has some bentonitic strata much like the prevalent rock type of the Climbing Arrow, and white limestones like those of the Milligan Creek are abundant in the lower part of the Climbing Arrow.

A possibility, hard to evaluate, is that part of the Climbing Arrow formation in the north half of the quadrangle was formed contemporaneously with all or part of the Milligan Creek formation. Such a relation is favored by the fact that the Climbing Arrow formation overlies limestone conglomerate above Milligan Canyon, whereas the Milligan Creek formation overlies limestone conglomerate below the canyon.

FOSSILS AND AGE

The Milligan Creek formation has yielded a variety of fossils but none are diagnostic of age. Because the Milligan Creek formation is seemingly gradational (and partly contemporaneous?) with the Climbing Arrow formation, and because the lower part of the Climbing Arrow is definitely of rather late Eocene age, it is assumed that the Milligan Creek formation is also of Eocene age.

The formation contains many snails and a few clams and ostracodes. The only floral remains recognized are charophyte oogonia. The gastropods and pelecypods were studied by D. W. Taylor (written communication, May 29, 1957), the ostracodes by I. G. Sohn (written communications, Jan. 5 and March 24, 1955); the charophytes, examined by R. E. Peck of the University of Missouri, proved to be too poorly preserved to permit study.

Snails are the most abundant fossils. Unfortunately, they are poorly preserved; external structures are obscured by the coarse texture of the replacing calcite, and internal structures are practically obliterated by calcite fillings, so that close identification is rarely possible. Nevertheless, because little is known of North American continental Tertiary gastropods, the scant data on those from Three Forks seem worth giving. Gastropod forms identified from the Milligan Creek and the other formations of the Bozeman group by Taylor are listed by him in table 4 for convenient reference. They include Lymnaea of indeterminate species ranging from Triassic on; Planorbidae of the subfamily Segmentininae indeterminate as to genus and similar to forms described from both the late Eocene Tepee Trail formation of Wyoming and the Oligocene Florissant formation of Colorado; representatives of four genera of the family Pupillidae known throughout the Tertiary in Europe but little studied in the pre-Pliocene of the United States; and a new species of *Oreohelix* (*Radiocentrum*).

The pelecypods are all of the wide-ranging genus *Pisidium.*

The ostracodes include two small smooth forms, one of which is questionably assigned to *Cypridopsis;* the other is indeterminate even as to genus.

According to Taylor, all snails identified from the Milligan Creek outcrops north of the Jefferson River and most of those from the south side are fresh-water forms of pond or lake habitat. Those from south of the river include both fresh-water and land snails. Most of the land snails are of genera that seem to adapt to a wide variety of environments. *Oreohelix*, however, is characteristically a mountain snail.

ORIGIN

Peale's concept of eolian-lacustrine origin for the entire Tertiary basin sequence applies well to the Milligan Creek formation.

TABLE 4.—*Check list of gastropods in Tertiary rocks of Three Forks quadrangle, Montana*

[Localities shown on data map, pl. 2]

	Milligan Creek formation						Climbing Arrow formation			Dunbar Creek formation								
	20002	20003	20004	20005	20006	20007	20008	20009	20010	20011	20012	20013	20014	20015	20016	20017	20018	20019
Fresh-water snails:																		
Lymnaea	×	×	×	×	×	×			×								×	×
Australorbis pseudoammonius (Schlotheim)									×									
Planorbidae, Segmentininae		×			×													
Planorbidae indet	×		×						×									×
Physidae								×										
Land snails:																		
Pupillidae A						×												
B						×						×						
C						×					×	×		×		×		
D						×												
E								×										
F											×	×		×	×			
G											×							
H												×						
I														×	×	×		
indet														×				
Polygyrella							×	×										
Oreohelix n. sp						×						×	×	×	×	×		
Helminthoglyptidae n. gen								?		×				×	×			
Indeterminate		×																

The number given for each locality is that of the U.S. Geological Survey Cenozoic series. Altitudes are correct to ±10 ft. All localities are in the Three Forks quadrangle (1950) 1:62500.
20002. NW¼SE¼ sec. 11, T. 1 N., R. 1 W. Altitude 4,350 feet.
20003. E½NE¼ sec. 11, T. 1 N., R. 1 W. Altitude 4,430 feet.
20004. Same locality as 20003, but 5–10 feet topographically and stratigraphically lower.
20005. SW¼SW¼ sec. 1, T. 1 N., R. 1 W. Altitude 4,470 feet.
20006. NW¼NW¼ sec. 7, T. 1 N., R. 1 E. Altitude 4,185 feet.
20007. NW¼SW¼ sec. 36, T. 1 N., R. 1 W. Altitude 4,260 feet.
20008. SE¼NE¼NE¼ sec. 10, T. 2 N., R. 1 W. Altitude 4,600 feet.
20009. NE¼NE¼NE¼ sec. 11, T. 2 N., R. 1 W. Altitude 4,580 feet.
20010. Center sec. 12, T. 2 N., R. 1 W. Altitude 4,430 feet.
20011. SE¼NW¼ sec. 6, T. 2 N., R. 2 E. Altitude 4,360 feet.
20012. NW¼NW¼ sec. 5, T. 2 N., R. 2 E. Altitude 4,350 feet.
20013. NW¼SW¼ sec. 3, T. 2 N., R. 1 E. Altitude 4,230 feet.
20014. Same location as 20013 but 18 feet stratigraphically higher.
20015. NW¼SW¼ sec. 3, T. 2 N., R. 1 E. Altitude 4,235 feet.
20016. S½NE¼SE¼ sec. 19, T. 3 N., R. 1 E. Altitude 4,450 feet.
20017. NW¼SW¼ sec. 3, T. 2 N., R. 1 E. Altitude 4,235 feet.
20018. SW¼NE¼ sec. 6, T. 1 N., R. 2 E. Altitude 4,225 feet.
20019. Sec. 29, T. 1 N., R. 2 E. 600 feet E., 0 feet N. of SW corner. Altitude 4,230 feet.

The limestone and other fine-grained rocks of the Milligan Creek formation, as attested by their lithology and fossil content and distribution, were deposited in a lake. The thick bedding suggests that the lake was perennial rather than seasonal. The lake basin seems to have existed at about the position of the present Jefferson Valley. The coarse marginal deposits and the presence of the mountain snail *Oreohelix* suggest that the basin lay in terrain not much different from the present.

The volcanic ash fragments no doubt were partly airborne into the lake and partly washed in, after brief periods of subaerial weathering. The alteration to clay seems to have occurred mainly after the fragments reached the lake. Favoring sublacustrine alteration is the common presence of patches and streaks of silica, presumably released in the devitrification of nearby glass. Further, the coarse ash must have been mainly glass when it entered the lake: that which was airborne into the lake could scarcely have been much altered, and that washed into the lake could not have floated out and been deposited eventually in a very fine matrix if its pumiceous character has been destroyed by argillization.

Perhaps repeated lapilli-ash showers, by interfering with normal soil formation on the lake margins, were responsible for the restriction of life to the sturdy Charophyta and rare ostracodes.

The vents from which the pyroclastic debris in the Milligan Creek and later Tertiary formations were erupted are unknown and likely long to remain so, mainly because early Tertiary volcanism was so widespread in southwestern Montana and adjoining regions.

Source areas for the coarse interbeds are more readily identified. The composition of the conglomerate and sandstone indicates that those on the north side of the Jefferson River, lacking metamorphic rock fragments, came from farther north and those south of the river, rich in gneiss, came from farther south. Their degree of rounding and sorting, their common cross-stratification, and their lack of limestone pebbles show that they were not shed directly off the limestone-dominated slopes on which they lie but were deposited by streams after considerable transport. None have the sorting and rounding nor the steep foreset bedding expectable in deltas. The simplest interpretation is that these masses are deposits made in stream channels scoured in the lake deposits when the lake was low or dry. But surely the streams also flowed when the lake was high, and a large fraction of the sand and gravel deposits must have been made where streams entered the lake. The absence of well-developed standing-water features in any of the sandstone and conglomerate may be due to rapid deposition, resulting from steep gradients and perhaps from significant transport only during floods.

The climate of Milligan Creek time is hard to visualize from the data at hand. The thick bedding in the formation suggests that the lake was perennial and this in turn favors a fairly humid climate. Conversely, the coarseness of conglomerate at the margins, suggesting torrential deposition, and the absence of fossil wood faintly hint at semiarid conditions.

CLIMBING ARROW FORMATION

NAME AND DISTRIBUTION

The Climbing Arrow formation is made up of olive thick-bedded sandy bentonitic clay and coarse sand with subordinate light-colored siltstone, sandstone, conglomerate, and limestone. The formation is here named for the Climbing Arrow Ranch, of which the principal buildings (SW cor. sec. 8, T. 1 N., R. 2 E.) are on fairly typical outcrops of the formation though rather remote from the main areas of exposure. The Climbing Arrow, except for Quaternary alluvium, is the most extensively exposed unit in the quadrangle; it crops out over 27 square miles and is obscured only by a thin mantle of Quaternary detritus over 20 square miles more. Rocks of similar lithology underlie large areas in the adjoining Radersburg quadrangle (V. L. Freeman, written communication, Mar. 4, 1954) and some northward extension of the formation may ultimately be feasible.

TOPOGRAPHIC EXPRESSION

The formation in general is unresistant, producing subdued landscapes. In the northwest part of the quadrangle it tends to form low but distinct rounded hills, banded in tones of light gray and white, and separated by broad smooth valleys. These hills are held up by resistant limestone, sandstone, and conglomerate strata in the lower part of the formation.

The bulk of the formation, poorly consolidated and rich in clay, tends to develop somber badland topography (fig. 8) with bare clay-capped hills, rounded on top but with steep sides, separated by grassy poorly drained lowlands. In the lowlands, the local drainage pattern has been disrupted by many small-scale earth flows.

FIGURE 8.—Climbing Arrow formation in type area. View southeast from road in NW cor. sec. 12, T. 1 N., R. 1 E.

LITHOLOGY AND THICKNESS

The base of the Climbing Arrow is definitely known only from the poor exposures south of the Buttleman Ranch. Exposures in the minor anticline in secs. 1 and 12, T. 2 N., R. 1 W., are presumably close to the base. The top is much better known from good exposures east of the Mud Spring Gulch road near Highway 10N and along the Madison Bluffs in sec. 6, T. 1 N., R. 2 E. Within the formation it is hard to establish even the general stratigraphic position of many exposed segments. The composite section below, though surely containing gross errors, gives the general tenor north of the Jefferson River where the formation is not less than 750 feet thick, and may be considerably more than 1,000 feet thick. (Total thicknesses greater than 1,500 feet can be interpreted if use is made of driller's logs of test wells in secs. 7 and 8, T. 2 N., R. 1 E., but such interpretations are of little value as the logs are terse, incomplete, and unsupported by cuttings.)

	Thickness (feet)
Upper dark unit: thick-bedded olive bentonitic sandy clay with many lenses of dark coarse sand and a few layers of white coarse tuffaceous siltstone and sandstone; near basin edges, thick tongues of dark coarse pebbly sandstone and conglomerate	200 to 300
White unit: thick-bedded white tuffaceous siltstone and finely crystalline porous limestone with a few thick interbeds of olive bentonitic sandy clay and poorly consolidated dark sandstone	100 to 200
Lower dark unit: like upper dark unit but with thin lenses of white limestone and thick lenses of poorly consolidated sandstone and pebble conglomerate throughout	>500
Base concealed.	

The three units are gradational and contacts between them can only locally be drawn.

The lower dark unit is especially well exposed at the west edge of the quadrangle near the common corner of secs. 8, 9, 16, and 17, T. 2 N., R. 1 W. The white unit is best seen in the syncline south of the Silver Sage Ranch in secs. 3 and 10, T. 2 N., R. 1. W.; it is also well exposed along the county road at the south edge of SW1/4 sec. 8, T. 2 N., R. 1 E.; and the isolated areas of Climbing Arrow rocks in secs. 22 and 27, T. 3 N., R. 1 W., and centering in NE1/4 sec. 34, T. 2 N., R. 1 W. are also probably part of this unit. The upper dark unit is well exposed in NW1/4 sec. 16 and NE1/4 sec. 17, T. 2 N., R. 1 E., and along Highway 10N, about a mile south of its junction with Mud Spring Gulch road. Basin-edge tongues of coarse detritus in the upper units are well seen near the mutual corner of secs. 17, 18, 19, and 20, T. 2 N., R. 1 E.

The three units, only broadly recognizable north of the Jefferson River, cannot be distinguished south of the river. There, the formation, about 700 feet thick, is dominated throughout by olive clay and dark sand, with limestone prominent only near the base and white tuffaceous siltstone numerous only near the top. Excellent continuous exposures of nearly 350 feet of the section appear in the hills east of the county road in SE1/4 sec. 11, and W1/2 sec. 12, T. 1 N., R. 1 E. This area (shown in fig. 8) is designated the type area of the formation, as it offers thick and continuous exposures of the dominant clay-and-sand lithology. For a full view of the formation, however, it is necessary to examine the subordinate limestone, siltstone, sandstone, and conglomerate north of the Jefferson River in the places listed previously. A sequence about 100 feet thick exposed below the summit bench in center W1/2 sec. 12 (measured section F, pl. 2) is representative of the type area:

	Thickness (feet)
Gravel-covered bench, top of hill.	
Clay, thick-bedded, grayish-olive, bentonitic, and a few streaks of clayey fine subangular sand	12
Sand, white to yellowish-gray, clayey, medium to coarse, subangular; one bed	1
Clay, thick-bedded, grayish-olive, bentonitic	5
Clay and sand; alternating lenses and beds, 4 inches to 1 foot thick, of olive-gray clayey sand and pale-olive sandy clay	25
Sand, thick-bedded, yellowish-gray, fine to coarse, subangular	6
Clay, thin-bedded, pale-olive, sandy; includes several reddish-brown and grayish-olive bentonitic zones a few inches thick	4
Sand, thin-bedded, white to yellowish-gray, clayey, coarse, subangular	8
Clay, thick-bedded, olive; grades into underlying sandstone	5
Sandstone, grayish-olive, medium to fine, subrounded, cemented with clay; one bed	3
Sand, yellowish-gray, coarse, feldspathic and biotitic, subrounded; one bed	3
Clay, grayish-olive to dusky-yellow, markedly bentonitic; one bed	3
Clay, thick-bedded, grayish-olive, sandy, bentonitic	7
Sand, thick-bedded, pale-olive to yellowish-gray, coarse, feldspathic and biotitic, subangular to rounded	7
Clay, grayish-olive, very sandy, bentonitic; one bed	1
Sand, thick-bedded, pale-olive to yellowish-gray, medium to coarse, feldspathic, biotitic, subangular; grades into overlying clay	6
(Sequence starts at bench at 4,330 ft alt.)	

The relative volumes of rock types in the Climbing Arrow, based on several crudely measured partial sections and with some assumptions as to subsurface thinning and thickening of strata with distance from the hill fronts, are estimated at bentonitic clay, 45 percent; quartzose sand, 25 percent; quartzose sandstone and conglomerate, 10 percent; tuffaceous siltstone and sandstone, 10 percent; limestone, 10 percent.

The bentonitic clays that dominate the formation are somber rocks, mostly colored olive tones of gray, but some are olive brown and reddish brown. Most of them swell appreciably when wetted, but rarely more than double their original volume. Unlike nonswelling clays, they tend to maintain outcrops, producing steep backslopes and gentle dipslopes, alike barren of vegetation and mantled with loose popcornlike lumps of dark clay. (See figs. 8 and 9.) Although clay is the dominant material, few beds are pure. Most contain from traces to several percent of subrounded sand and angular silt grains of quartz and clotted clay minerals, and, less commonly, of biotite, feldspar, dark aphanitic volcanic rocks, and volcanic glass. Crystals of gypsum, evidently authigenic, are scattered through many clay strata, and, rarely, gypsum is a major component, forming plates, as much as an inch thick and a foot across, of large intricately intergrown crystals.

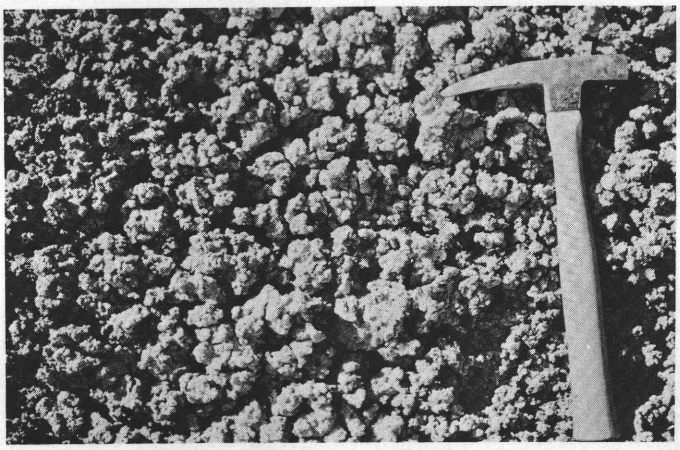

FIGURE 9.—Popcorn surface on weathered bentonitic siltstone of Climbing Arrow formation.

The foregoing field observations have been supplemented by laboratory study of six representative samples. The obvious presence of small amounts of coarse material is quantitatively shown by mechanical analysis of the six samples (fig. 10).

In thin section, the six clays are similar. By reflected light they are varicolored pale red, yellowish gray and light gray, rather than the olive tones of the fresh outcrop. This is presumably a dusty film for, by transmitted light, the clays are largely dark colored, mostly greenish brown and translucent but locally dark brown, grayish red, and pale red. They consist mostly of closely packed unoriented splotchy masses of dark translucent clay, generally in the clay-silt range but locally as coarse as fine sand, feathering out into lighter colored clear clay at the edges. In places dark translucent and light transparent clay are in alternating bands. Some grains are separated by reticulating veinlets of clear opal, and a few contraction fractures are lined with opal. Under crossed nicols, the dark clay masses appear as groups of cloudy anhedral grains with yellowish-brown (anomalous low second order?) polarization colors; the clear clay is microcrystalline, with myriad unoriented straight-extinguishing length-slow fibers giving gray and white first-order colors. The dark clay is of slightly lower maximum refractive index than the clear clay which in turn is slightly lower than balsam. Apparently two clay species are represented but they are here both regarded simply as montmorillonite. Within and between the clay masses are many angular bits of quartz, ranging in size from clay to fine sand, and in places making up as much as 10 percent of the rock. Much less abundant, and not present in all specimens, are slightly bleached greenish-brown biotite, clayey muscovite, sodic plagioclase, bubble-rich clear glass in both curved and equant shapes, epidote, sphene, hydrated iron oxides, and black opaque minerals. Carefully sought but not found were relict pyroclastic textures of the sorts reported from literally hundreds of occurrences of montmorillonite rocks and summarized by Ross and Hendricks (1945, p. 64–65, pls. 1–5).

The identification of the clay minerals as montmorillonite is based on X-ray analyses of the six samples. The X-ray studies, tabulated in rows 1–6 of table 5, show that the rocks consist almost wholly of montmorillonite, with minor amounts of quartz, feldspar, mica, and, in one only, kaolinite. Not only is the clay fraction (5*A*) dominated by montmorillonite but the silt fraction as well (5*B*).

Chemical analyses of the combined clay and silt fractions are presented in table 6*A*. The chemical compositions of the six clays, from widely separated parts of the formation, are remarkably similar. They are well within the chemical ranges of the sedimentary montmorillonites studied by Ross and Hendricks (1945); 34 analyses of reasonably comparable material have been selected from the group presented by Ross and Hendricks and summarized as to median and range of oxides, for comparison with the median and range of the analyzed Climbing Arrow samples.

TABLE 5.—*X-ray mineralogy of clay and silt fractions of some rocks of the Climbing Arrow formation*

[Analysts: J. C. Hathaway, samples 1–5 and 7; C. J. Parker, samples 6 and 8. Localities shown on pl. 2. For laboratory numbers, see table 6]

Sample	Percent [a]	Montmorillonite group	Kaolinite group	Feldspar	Quartz and cristobalite	Chlorite and mica
			***A*, Clay fraction**			
1	69.1	△	------	------	×	×
2	86.4	△	------	------	×	×
3	95.2	△	×	------	×	------
4	26.8	△	------	×	------	×
5	60.7	△	------	------	×	×
6	77.9	◬	------	------	------	------
7	27.0	△	------	------	------	×
8	22.3	◬	------	------	------	------
			***B*, Silt fraction**			
1	27.6	△	------	------	------	×
2	10.1	△	------	×	------	------
3	4.2	△	------	×	------	×
4	70.5	△	------	------	------	------
5	30.3	△	------	------	------	------
6	10.2	×	------	------	------	×
7 [b]	66.8	------	------	------	------	×
8	27.9	×	------	------	------	------

[a] Percent of rock in size class, from figure 10 for samples 1–6; from figure 11 for samples 7 and 8.
[b] Glass is major component. (Glass not reported but probably present in most other silt and clay samples.)
◬—only component reported.
△—major component (>20 percent).
×—minor component, generally only a trace.

The lenses of yellowish-gray incoherent sand that alternate with the clay strata are typically several feet thick, several hundred feet broad, and, to judge by low ridges they produce on the surface, a few thousand feet long; their trends, though diverse, are consistently easterly. The sand is mostly coarse, subangular to subrounded, and composed of quartz with subordinate feldspar (mostly plagioclase but partly alkali feldspar), biotite, and dark fine-grained volcanic rocks. A pecularity of the biotite is that it occurs not only as flakes but also in equidimensional booklike aggregates of sand and granule size. The thick tongues of sandstone and conglomerate, near basin edges, are probably lateral equivalents of the sand lenses, cemented mostly by clay but partly by silica with the aid of ground water.

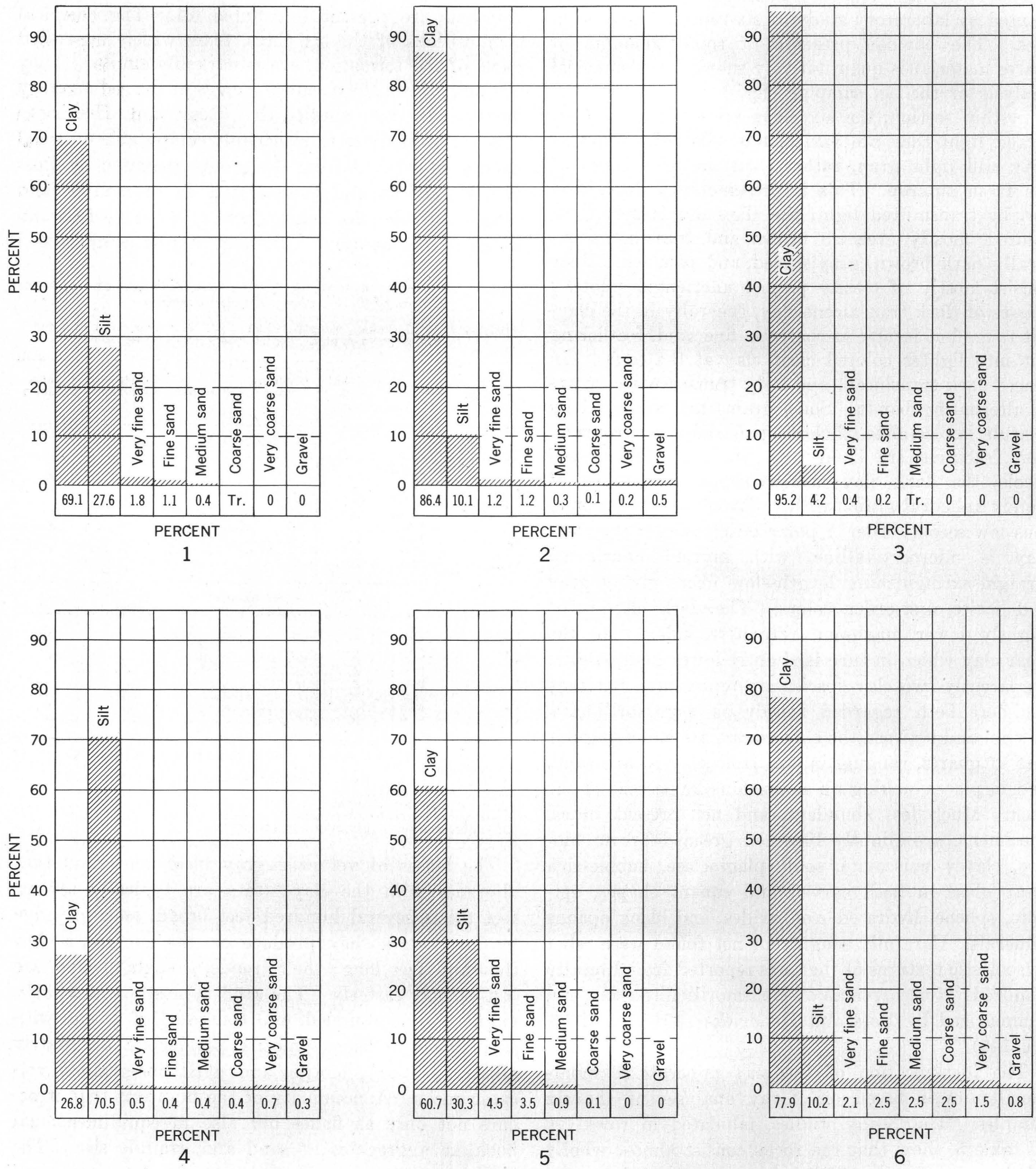

FIGURE 10.—Mechanical analyses of clay-rich rocks from the Climbing Arrow formation. Sample numbers correspond to those in table 6 where laboratory numbers and chemical analyses are given. Grain-size classification according to Wentworth scale. Analyses by P. R. Blackmon.

TABLE 6.—*Chemical data on representative clay-rich and glass-rich rocks, Climbing Arrow formation*

[Rapid-method analysis by P. L. D. Elmore and K. E. White. Localities from which samples where collected are shown on pl. 2. Laboratory numbers are last 3 digits of 6-digit numbers of which the first 3 digits are 141. Rc, range of samples 1-6; Mc, median of samples 1-6; Mr, median of 34 analyses of sedimentary montmorillonites, and Rr, range of 34 analyses of sedimentary montmorillonites, published by Ross and Hendricks, 1945, p. 34-35, analyses 1-5, 7-12, 14-16, 18-20, 23, 25-27, 30-32, 34-35, 37, 38, 46, 48, 50-53]

	Clay-rich rocks						Glass-rich rocks		Rc	Mc	Mr	Rr
Sample	1	2	3	4	5	6	7	8				
Laboratory No	360	361	362	363	364	366	365	368				
					A. **Chemical composition**							
SiO_2	57.7	56.1	55.1	57.5	59.4	56.0	68.6	67.8	55.1 -59.4	56.8	50.4	44.0 -55.4
Al_2O_3	15.6	16.4	19.0	15.0	16.6	14.7	14.0	13.6	14.7 -19.0	16.0	17.8	11.7 -28.9
Fe_2O_3	5.6	5.9	2.9	5.9	5.2	5.2	1.5	2.0	2.9 - 5.9	5.4	2.8	.8 - 7.3
FeO	.16	.06	.17	.22	.22	.20	.19	.14	.06- .22	.18	0	0 - 1.4
MgO	2.4	2.8	2.1	2.8	2.0	3.0	.88	.64	2.0 - 3.0	2.6	4.1	.50- 8.6
CaO	2.2	1.8	1.8	1.9	1.3	1.9	.81	.80	1.3 - 2.2	1.85	1.2	0 - 3.2
Na_2O	1.5	.92	.96	1.0	.72	1.2	2.8	2.0	.72- 1.5	.98	.17	0 - 3.8
K_2O	2.0	1.4	1.5	1.2	2.8	1.5	4.4	4.6	1.2 - 2.8	1.5	.35	0 - 2.3
TiO_2	.62	.54	.60	.68	.74	.66	.18	.10	.54- .74	.64	.11	0 - .80
P_2O_5	.19	.09	.05	.06	.08	.12	.04	.02	.05- .19	.085	0	0 - .24
MnO	.04	.16	.03	.18	.05	.06	.04	.06	.03- .18	.055	0	0 - .18
H_2O	12.3	14.1	15.4	13.0	11.1	15.2	6.7	8.0	11.1 -15.4	13.6	20.9	12.4 -24.1
CO_2	<.05	<.05	<.05	.08	<.05	.18	<.05	<.05	<.05- .18	<.05	0	0 - .37
Sum	100	100	100	100	100	100	100	100	-----	-----	-----	-----
			B. **Chemical composition, recalculated free of water and CO_2**									
SiO_2	65.8	65.3	65.5	66.4	66.7	66.2	73.4	73.7	-----	-----	-----	-----
Al_2O_3	17.7	19.1	22.5	17.4	18.7	17.3	15.0	14.8	-----	-----	-----	-----
Fe_2O_3	6.4	6.9	3.4	6.8	5.8	6.1	1.6	2.2	-----	-----	-----	-----
FeO	.2	.1	.2	.3	.3	.2	.2	.2	-----	-----	-----	-----
MgO	2.7	3.2	2.5	3.2	2.2	3.5	.9	.7	-----	-----	-----	-----
CaO	2.5	2.1	2.1	2.2	1.4	2.2	.9	.9	-----	-----	-----	-----
Na_2O	1.6	1.1	1.1	1.1	.8	1.4	3.0	2.2	-----	-----	-----	-----
K_2O	2.3	1.6	1.8	1.3	3.2	1.8	4.7	5.0	-----	-----	-----	-----
TiO_2	.7	.6	.7	.8	.8	.8	.2	.1	-----	-----	-----	-----
P_2O_5	.2	.1	.1	.1	.1	.2	-----	-----	-----	-----	-----	-----
MnO	-----	.2	-----	.2	.1	.1	-----	-----	-----	-----	-----	-----
Sum	100.1	100.3	99.9	99.8	100.1	99.8	99.9	99.8	-----	-----	-----	-----

The white siltstone and sandstone are composed essentially of poorly sorted fragments of volcanic glass. The refractive index of the glass (9 samples studied) ranges from near 1.500 to 1.505; most of it seems to be close to 1.502. A little is in curved bubble-rich shards but most of it is in equant fragments, angular to rounded, riddled with cracks in a rudely rectangular pattern. From 10 to 20 percent of the equant fragments are altered to montmorillonite, mostly the pale-green or greenish-brown semiopaque variety described earlier from the Milligan Creek formation, but including lesser amounts of the clear fibrous type. The curved shards are typically unaltered. Minor components are angular to subrounded grains of quartz, alkali feldspar (mostly orthoclase), plagioclase (albite to andesine), brown biotite, light-green hornblende, angular opaque minerals, and dark aphanitic volcanic rocks. Some of these white rocks contain interstitial calcite; a few contain large quantities of it, and grade into tuffaceous limestone. Mechanical, X-ray, and chemical analyses of a typical carbonate-free white tuffaceous siltstone, from the white unit in SE¼NE¼NE¼ sec. 10, T. 2 N., R. 1 W., appear as sample 7 in tables 5 and 6 and figure 11.

Several thin-sections and many grain mounts of white siltstone that seems to be nearly pure volcanic ash contain intimately admixed, beautifully preserved organisms (identified by Estella Leopold and Richard Rezak) including pollen, pelagic diatoms, nonpelagic diatoms, filamentous algae, and blue-green algae.

The white limestone of the formation is similar in all important respects to that of the Milligan Creek formation and needs no further description.

Very rare are thick lenses of white conglomerate made up of unsorted subrounded fragments of pumice as much as 2 inches long, cemented by white silt and clay. One of these lenses, which furnished sample 8 (tables 5 and 6 and fig. 11), is well exposed in a prospect pit in NW¼NW¼ sec. 32, T. 3 N., R. 1 E.

In addition to the micro-organisms in the siltstones, the formation contains fossil evidence of a rich fauna and flora that included—in decreasing size—brontotheres, tapiroids, oreodonts, turtles, rodents, lagomorphs, insectivores, marsupials, gastropods, and ostracodes; and coniferous and deciduous trees, flowering plants, and charophytes. The vertebrate remains are generally in the bentonitic clays. The gastropods and ostracodes occur in both the white siltstone and the limestone. The fossil wood is largely in the sandstone and conglomerate lenses. Tree and plant leaves have been found only in clays, and charophyte remains only in limestone. Algae, pollen, and diatoms occur widely in white limestone, as well as in siltstone.

By far the most abundant fossil material is silicified wood. Some fragments are rather large: a stump 5 feet across and a single log 26 feet long and 9 feet across have been observed. These and many other

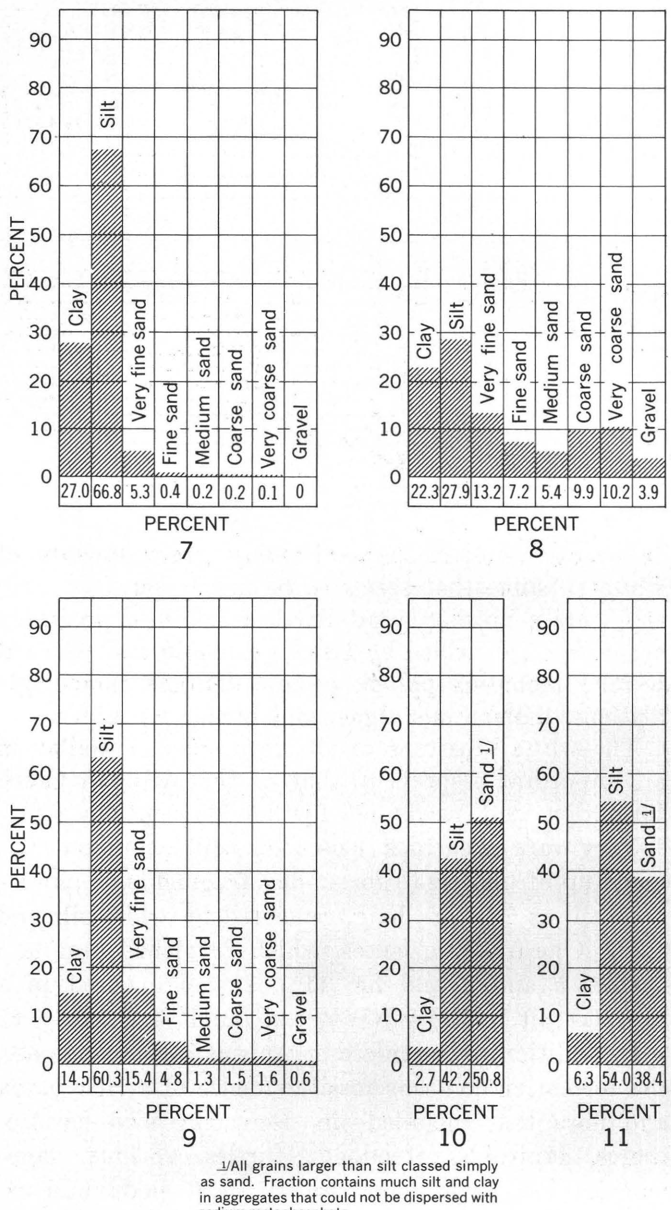

Figure 11.—Mechanical analyses of tuffaceous siltstone from Climbing Arrow (7, 8) and Dunbar Creek (9, 10, 11) formations. Sample numbers correspond to those in tables 6 and 8 where laboratory numbers and chemical analyses are given. Grain-size classification according to Wentworth scale. Analyses by P. R. Blackmon (7, 8, 9) and J. C. Hathaway (10, 11).

smaller fragments are largely yellowish-brown chalcedony and bluish-black opal. Many cavities and fractures are lined with earthy greenish-yellow jarosite that looks disturbingly like carnotite. Jarosite is also common in the cement of the associated sandstone and conglomerate. In a different type of fossilization the wood is replaced by translucent to opaque yellowish-gray microcrystalline quartz that has well-developed prismatic jointing. Small pencil-like fragments of this replaced wood are easily mistaken for crystals of some hard authigenic mineral.

The Climbing Arrow grades upward into the Dunbar Creek formation over a stratigraphic interval of perhaps 50 feet by an alternation of dark bentonitic clay with the white and yellow tuffaceous siltstones that characterize the basal Dunbar Creek.

PEBBLY SAND MEMBER

Included with the Climbing Arrow formation, though not without hesitation, is a large body, not less than 150 feet thick, of cross-stratified yellowish-orange pebbly subrounded sand and lesser amounts of pebble gravel, that is exposed above the east bank of Willow Creek. A small mass of similar sand is well exposed above the opposite bank. The cross strata in the sand have northerly or northeasterly dips of a few degrees. The sand, mostly quartz, also contains much feldspar and a little biotite, hornblende, and dacitic rock.

The sand seems to have been locally derived from crystalline rocks to the south and deposited by a predecessor of Willow Creek. It is entirely possible that the predecessor was a late Quaternary stream rather than an early Tertiary one, for the sand lacks fossils, and its obscure stratigraphic relations permit either interpretation.

FOSSILS AND AGE

The age of the Climbing Arrow formation ranges from middle or late Eocene to early Oligocene. The age of the lower part of the Climbing Arrow formation has been established as no younger than late Eocene on the basis of small but diagnostic vertebrate and invertebrate faunas. These Eocene faunas are of more than routine interest as only one other Eocene locality has heretofore been known in southwestern Montana—near Sage Creek, 100 miles to the southwest. The vertebrate fauna is from a single locality in SE¼NW¼SE¼ sec. 3, T. 2 N., R. 1 W., from an area scraped smooth by bulldozers (in 1954) at the south abutment of the small dam downstream from the Silver Sage Ranch buildings (loc. 443, pl. 2 [(red) D111]). The fauna, studied by G. E. Lewis (written communication, Jan. 19, 1956), includes the following forms:

Hyopsodus cf. *H. uintensis* Osborn; right maxilla, P^4–M^3 [(red) D111–1]
Dilophodon sp. or *Isectolophus* sp.; right ramus, P_3–M_2 [red) D111–2]
Protoreodon sp.; two upper molars [(red) D111–3]

Lewis concludes that this assemblage is probably of Uinta age. The vertebrate locality is within 100 feet of the top of the lower dark unit, so that on this evidence at least 400 feet of the formation is of late Eocene age or older.

The stratigraphically lowest Oligocene fossils were recovered from near the top of the white unit, just north of the county road at the south edge of sec. 8, T. 2 N., R. 1 E., 1,600 feet east of southwest corner (loc. 158, pl. 2 [(red) D104]). These include two rami and a cheek tooth of *Paleolagus temnodon* Douglass, of earliest Oligocene (Pipestone Springs) age, according to Lewis (written communication, Jan. 19, 1956). If the stratigraphy is correctly understood, therefore, the transition from Eocene to Oligocene is now located within less than 300 feet of strata. (This conclusion would be more convincing if both Oligocene and Eocene fossils had been recovered from a single continuous section.)

A single locality in center sec. 12, T. 2 N., R. 1 W. (20010, pl. 2) has yielded fresh-water gastropods of rather definite Eocene age, along with other snails of wide or indeterminate range in age. The snails are in strata far down in the lower dark unit, exposed in a small anticline, and therefore are several hundred feet stratigraphically below the Eocene vertebrate locality. The fauna, studied by D. W. Taylor (written communication, May 29, 1957) and summarized in table 3, includes *Lymnaea* and *Planorbidae* of indeterminate species, that occur in both the Milligan Creek and Dunbar Creek formations as well and are of little chronologic value, and the fresh-water planorbid *Australorbis pseudoammonius* (Schlotheim). According to Taylor, this species is widely known in Europe where it is restricted to the Eocene. In the United States it is more widely distributed in middle Eocene than in younger deposits, but is known from at least three deposits dated as late Eocene on vertebrate evidence (Kishenehn formation of northernmost Montana and southeastern British Columbia; Tepee Trail formation, Wind River basin, Wyo.; and an unnamed unit at Beaver Divide, Fremont County, Wyo. The species is also known from one locality (Beaver Divide conglomerate member of Nace, 1939, Fremont County, Wyo.) dated as very early Oligocene on sparse vertebrate evidence. On its own merits, then, the occurrence of this species is strong but not indisputable evidence of Eocene age.

The upper dark unit north of the Jefferson River contains the following forms identified by Lewis (written communication, Jan. 19, 1956):

brontotheres cf. *Allops* sp. or *Brontotherium* sp. left metacarpal IV [(red) D107]; right radius [(red) D108–1]
Merycoidodon cf. *M. gracilis* Leidy; right ramus [(red) D106]; right M_3 [red) D108–2]

These collections came from the NW¼ sec. 16, T. 2 N., R. 1 E. (locs. 344–346, 362, pl. 2).

Merycoidodon remains have been found only in the upper 200 feet of the formation north of the Jefferson River, and in the succeeding basal 200 feet of the Dunbar Creek formation (loc. 409, pl. 2). Based only on oreodont evidence, this part of the Climbing Arrow formation might be regarded as of middle Oligocene (early Brule) age. Entire large brontothere bones have, however, been recovered from within 100 feet of the top of the Climbing Arrow and these are regarded, by both Hough and Lewis, as diagnostic of early Oligocene (Chadron) age. Fragments of brontothere bones have also been recovered from the Dunbar Creek formation (NE¼ sec. 6, T. 2 N., R. 2 E., loc. 250, pl. 2) so that the upper part of the Climbing Arrow, and at least part of the Dunbar Creek as well, is of early Oligocene age.

The only vertebrate remains recovered from the Climbing Arrow formation south of the Jefferson River are fragments of brontothere teeth, identified in the field by Lewis, from about 350 feet below the top, in NE¼SE¼ sec. 11, T. 1 N., R. 1 E. They suffice, however, to date this sequence as of early Oligocene age and to confirm its unity with the strata of similar lithology across the river.

The fossil wood and leaf remains offer little aid in age determination but are of some paleoecologic interest. Ten collections of fossil wood were studied by Elso S. Barghoorn of Harvard University (written communication, Jan. 14, 1955). Most of the reasonably well preserved material he found to be coniferous wood assignable with more or less definiteness to the family Cupressaceae. For most of the material, generic determination is not feasible but at any rate the wood is not that of *Juniperus* and is probably either *Thuja plicata* (Western Red Cedar), *Chamaecyparis lawsoniana* (Port Orford Cedar), or *Libocedrus decurrens* (Incense Cedar). The very large stump, mentioned previously, is that of a dicotyledonous tree which is "in all reasonable probability that of *Acer* (Maple)." "Growth rings are normal and well developed in all specimens, indicating seasonal climatic conditions." Most of the woods identified are from the lower dark unit of the Climbing Arrow and are thus probably of Eocene age, but the age evidence from the woods themselves is not more closely determinable than Tertiary, according to Barghoorn.

A collection, made by Harold Masursky, of plant material from the lower dark unit of the Climbing Arrow formation (NE¼NE¼ sec. 12, T. 2 N., R. 1 W.) was studied by R. W. Brown (written communication, Oct. 12, 1956). He identified the following forms, to which he was unable to assign an age:

Equisetum sp. (horsetail)
Typha sp. (cattail)
Fragment of a dictoyledonous leaf
Part of a fruit, undetermined

Of great promise for future age and ecologic interpretations are the abundant well-preserved microorganisms in the white rocks of the formation. The obviously varied assemblage of pollen, characeous remains, algae, and diatoms invites systematic collecting and study.

ORIGIN

The Climbing Arrow formation is largely the product of an aggrading stream system. The lenses of yellowish-gray quartzose sand, sandstone, and conglomerate spread throughout the formation, and the thick mass of pebbly sand flanking Willow Creek are clearly stream-channel deposits. Most of the streams were probably small, as suggested by the rather low degree of rounding and sorting, and the rather poor bedding. The gravels are consistently finer grained than those on the present flood plains of tributary streams, suggesting that the relief of the time was more subdued than the present.

The rudely east-west orientation of the channels, the increase in proportion and coarseness of conglomerate westward, and the general distribution of the Climbing Arrow formation suggests that the streams flowed eastward. The suggestion is strengthened by the lithology of the coarse deposits, which are dominated by debris from biotite-rich intermediate plutonic rocks, and dark aphanitic volcanics, a combination characteristic of the southern Elkhorn Mountains that form the west edge of the Three Forks basin. More distant transport cannot, of course, be ruled out.

The dark bentonitic clay strata, with their persistent fraction of silt and rounded sand and their intimate association with channel deposits are evidently overflow deposits that accumulated on the flood plains in short-lived ponds and lakes. If the coarse sediments were shed from the ancestral Elkhorn Mountains, it is reasonable to expect that the associated fine sediments were similarly derived. The situation is complicated by the mineralogy of the clay and by the known contemporaneous pyroclastic volcanism, for most large masses of montmorillonite rock are derived from the alteration of volcanic glass, as Hewitt (1917) long ago recognized and as subsequent work (summarized by Ross and Hendricks, 1945) abundantly confirmed. The possibility must be considered that some, perhaps most or even all, of the montmorillonite in the clays and tuffaceous siltstones is altered contemporaneous pyroclastic debric like the clay pellets in the Milligan Creek formation. If so, textural confirmation is lacking. As Ross and Hendricks noted (1945, p. 64) "thin sections show that bentonites [their term for montmorillonite or beidellite rock] commonly retain the structure of volcanic ash with surprising perfection." None of the six thin sections of clays of the Climbing Arrow show any relict ash textures, suggesting that whatever the origin of the clay, it was already montmorillonite before deposition.

The general absence of diagenetic silica in the Climbing Arrow is strong negative evidence that alteration to montmorillonite with concomitant release of silica occurred before deposition.

Very likely, the montmorillonite is polygenetic: some is residual clay from normal weathering of the monzonitic (and andesitic?) rocks of the clastic source area and some is altered latitic volcanic ash, perhaps ash first deposited in headwater areas during the earlier eruptions of Milligan Creek time and altered in place and in transit. Contemporaneous ash, whether falling on surrounding slopes and washed into the clay ponds, or falling directly into them from the air, may not have had time or appropriate conditions to alter much and may be represented by the glass fraction of the clay-rich strata.

The white siltstone, sandstone, and conglomerate made up largely of fresh vitric, crystal, and lithic ash fragments appear at first glance to be tuffs deposited on dry land. The tuff was calcalkaline rhyolite as indicated by comparison of the chemical composition of two analyzed specimens (recalculated free of water and CO_2 in columns 7 and 8 of table 6*B*) with Nockolds' tabulation (1954, p. 1012). A few of these strata may be tuff in the strictest sense, especially those composed only of angular fragments, but most of them were water laid as indicated by their lenticularity, their small but persistent content of rounded sand grains and their varied population of well-preserved pollen, diatoms, and algae. These strata probably represent rhyolitic ash that fell on or near the flood plains and was only slightly reworked by the meandering streams. That their glass is scarcely altered does not weaken the case for water deposition, even if the clay-rich strata should be shown to be direct deposits of latitic ash in standing water, for the difference in composition of the ash is apparently an important element in determining the course and rate of alteration. As Ross and Hendricks (1945, p. 65) point out: "It is not uncommon to find geologic sections of volcanic rocks where the tuffs of feldspathic composition have altered to bentonite, whereas the rhyolitic rocks in the same section are almost unaltered."

Even those few strata composed wholly of volcanic detritus may be flood-plain deposits. Sufficiently thick and extensive ashfalls could have provided streams and ponds with an exclusively pyroclastic load for a time, leading to flood-plain deposits of volcanic siltstone and clay and with channel fills of water-rounded lapilli and coarse ash.

The limestone of the Climbing Arrow presumably formed in ponds isolated from frequent flooding. The rather inhomogeneous and porous texture and the presence of characeous remains suggest that they are at least partly organic.

The abundant and varied flora and fauna show that the climate was warm and humid; according to Taylor (written communication, May 29, 1957) *Australorbis pseudoammonius* indicates tropical climate. The climate inferred from flimsy local evidence is consonant with the climate inferred from large-scale studies. As summarized by Barghoorn (1953, p. 244):

> During the early Cenozoic the northern mid-latitudes were covered by a vegetation, the botanical equivalent of which is now largely confined to subtropical and even tropical climates * * *. It seems quite unlikely that truly tropical conditions occurred in mid-latitudes during the Eocene. It is more probable that the absence of pronounced winter freezing allowed an unusual extension poleward of tropical plants * * *. Hence the absence of cold * * * rather than extreme warmth is indicated.

DUNBAR CREEK FORMATION

NAME, DISTRIBUTION, AND TOPOGRAPHIC EXPRESSION

The Dunbar Creek formation crops out in the northeast and southeast parts of the quadrangle. The spectacular bluffs that rise on the west bank of the Madison River are carved largely in this unit. The formation is here named for Dunbar Creek, a southward-flowing tributary of Mud Spring Gulch, that flows, occasionally, along the east edge of the wedge of Dunbar Creek rocks east of Highway 10N. The Dunbar Creek formation is mainly white to grayish-yellow thick-bedded tuffaceous siltstone, partly lacustrine and partly eolian, intricately laced with fluviatile sandstone and conglomerate. Rare components are dark bentonitic clay and white limestone. In color and general appearance, the Dunbar Creek resembles the Milligan Creek formation, but the Dunbar Creek contains little limestone, the main rock type in the Milligan Creek.

The formation is exposed over an area of 6 square miles and is thinly mantled by Quaternary deposits over more than 20 square miles more. The northern and southern parts, widely separated, are correlated on the basis of lithology, stratigraphic position, and the ages of contained fossils. Easterly dips carry the formation below the Madison River; similar strata cropping out on the east bluffs are mostly, if not wholly, of much younger age (late Miocene and early Pliocene according to most workers).

Rocks very much like the Dunbar Creek formation crop out extensively in the Townsend Valley and Clarkston Valley portions of the Radersburg and Toston quadrangles. Detailed study may eventually show that the Dunbar Creek formation is widely mappable in these valleys.

The Dunbar Creek formation develops topography much like that of the Milligan Creek—white dissected benchlands rising steplike from the flood plain, gashed with many steep-walled canyons. Because the rocks are well consolidated and yield little soil, outcrops are plentiful throughout.

LITHOLOGY AND THICKNESS

The base of the Dunbar Creek is well exposed in secs. 3 and 4, T. 2 N., R. 1 E., flanking Highway 10N. It is also visible at the north end of the Madison Bluffs in sec. 6, T. 1 N., R. 2 E. The top is nowhere exposed in the quadrangle. The stratigraphically highest exposures are in the bluffs at the southeast corner. On the flanks of the Hossfeldt Hills no more than 250 feet of the formation are preserved but in the southeast corner of the quadrangle not less than 800 feet and perhaps more than 1,000 feet are present.

Near Dunbar Creek the formation is mostly a monotonous succession of thick-bedded white to grayish-yellow tuffaceous siltstone. The sequence includes a few beds, several feet thick, of white tuffaceous limestone, many tongues of poorly sorted, poorly rounded carbonate-rich gravel that thicken hillward, and a few thin lenses of slightly bentonitic gray clay.

South of Three Forks, the lower 300 feet of the formation is much like the section near the Hossfeldt Hills, though with fewer limestone beds and more clay beds; conglomerate beds are better sorted than those near Dunbar Creek, and lack carbonate pebbles.

The upper part of the formation has little yellow siltstone and no limestone; it consists largely of thick-bedded white tuffaceous siltstone and fine sandstone, and many thin lenses and a few very thick ones of well indurated, rudely cross-stratified, yellowish-gray and yellowish-brown coarse to fine pebbly sandstone. It also contains a few thick sequences of dark clay-rich siltstone and bentonitic clay.

Chosen as the type area of the formation is the E½ sec. 7, T. 1 S., R. 2 E., for its exceptionally good and continuous exposures of about 315 feet of strata in the upper part of the formation. The lower 180 feet of the section in the type area is at the base of the bluffs in the southeast corner of the section; the upper 135 feet is exposed in the most northerly gulch in sec. 7 (measured section G, pl. 2). This partial section comprises:

	Thickness (feet)
Quaternary gravel:	
Gravel, pebble to cobble, rounded, largely quartzose rocks but with some volcanic and gneissic rocks, heavily cemented by calcium carbonate	4
Dunbar Creek formation:	
16. Sandstone, thick-bedded, yellowish-gray, pebbly, fine, subangular, tuffaceous, porous	18
15. Siltstone, thick-bedded, very pale orange, pebbly, micaceous, tuffaceous	29
14. Siltstone like that in unit 4	20
13. Sandstone, thick-bedded, faintly cross-stratified, grayish-yellow, subangular, medium	5
12. Claystone, thick bedded; like unit 9	12
11. Sandstone; like unit 1	2
10. Sandstone; like that in unit 4	36
9. Claystone, yellowish-gray with distinctive subconchoidal fracture; one bed	2
8. Siltstone; like that in unit 4	12
(break in section at gravel covered bench)	
7. Siltstone; like that in unit 4	36
6. Sandstone; like unit 1	10
5. Covered (probably sand)	55
4. Siltstone and fine sandstone; thick-bedded, yellowish-gray to white, subangular, tuffaceous	38
3. Sandstone; like unit 1	12
2. Sandstone, thick-bedded, white, fine, micaceous, tuffaceous	6
1. Sandstone, interbedded, moderate - yellowish - brown, pebbly, coarse, subangular cross-stratified sandstone and yellowish-gray, medium subangular micaceous sandstone	22
Bottom of gulch.	
Thickness of Dunbar Creek formation	315

Not exposed in the type are are any limestone or carbonate-cemented clastic rocks, fairly common nearer the base. These rocks, and the basal part of the formation, are best seen in SW¼ sec. 3, T. 2 N., R. 1 E.

The formation as a whole is made up of about 80 percent tuffaceous siltstone and sandstone; 15 percent quartzose sandstone, sand, and conglomerate; and 5 percent limestone, clay, and claystone.

The tuffaceous rocks are composed almost wholly of fine volcanic ash. The chemical composition of three representative samples of tuffaceous siltstone is reported on table 7*A*; recalculated to standard oxides 7*B*, these rocks are nearest to adamellite (quartz monzonite) or dellenite (quartz latite) in Nockolds' classification (1954, p. 1014). That they are lower in total alkalis and higher in alumina than their presumed source rocks is an expectable consequence of the presence of a large clay fraction. The clay and silt fractions of these same rocks are comprised overwhelmingly of montmorillonite (table 8). They are virtually identical with those of the Climbing Arrow formation in appearance and similar in chemical composition (compare the recalculated analyses of table 7*B* with analyses 7 and 8, table 6*B*). The refractive index of the glass (21 samples studied) seems to have a slightly wider range than that of the Climbing Arrow, from 1.498 to 1.506; most of it is close to 1.501. The younger tuffaceous rocks also seem to have a wider range of common accessory minerals, including labradorite, yellow-brown chlorite, and sphene, as well as all the minor components of siltstone of the Climbing Arrow, but less clay. The greater variety in the siltstone of the Dunbar Creek may merely reflect the fact that many more samples of the Dunbar Creek have been examined. The white sandstone seems iden-

TABLE 7.—*Chemical analyses and norms of representative samples of tuffaceous siltstone, Dunbar Creek formation*

[Rapid-method analysis of sample 9 by P. L. D. Elmore and K. S. White; samples 10 and 11 by P. L. D. Elmore, S. D. Botts, and M. D. Mack]

	A. Chemical composition			*B.* Chemical composition, recalculated free of water and CO_2			*C.* Norms			
Sample	9	10	11	9	10	11		9	10	11
SiO_2	63.2	65.2	66.4	69.9	71.7	72.7	Q	37.6	40.2	43.3
Al_2O_3	14.8	13.5	14.0	16.3	14.9	15.4	or	17.2	22.2	20.00
Fe_2O_3	2.6	3.3	1.9	2.9	3.6	2.1	ab	16.8	15.2	14.2
FeO	.55	.07	.81	.6	.1	.9	an	17.5	10.6	11.4
MgO	1.2	1.3	1.1	1.4	1.4	1.2	di: wo	.0	.0	.0
CaO	3.4	1.9	2.1	3.7	2.1	2.3	di: en	.0	.0	.0
Na_2O	1.8	1.6	1.6	2.0	1.8	1.7	di: fs	.0	.0	.0
K_2O	2.6	3.5	3.1	2.9	3.8	3.4	hy: en	3.5	3.5	3.0
P_2O_5	.35	.44	.28	.4	.5	.3	hy: fs	.0	.0	.0
MnO	.08	.00	.00	.1	.0	.0	mt	.7	.0	2.1
TinO	.04	.06	.04	----------	.1	----------	il	.8	.3	.6
H_2O	9.1	8.8	8.6	----------	----------	----------	ap	.3	.0	.0
CO_2	.39	<.05	<.05	----------	----------	----------	hm	2.4	7.7	.8
							C	3.5	4.0	4.8
Sum	100	100	100	100.2	100	100	ru	.0	.3	.0
							plagioclase	$Ab_{49}An_{51}$	$Ab_{59}An_{41}$	$Ab_{55}An_{45}$

TABLE 8.—*X-ray mineralogy of clay and silt fractions of representative tuffaceous siltstones, Dunbar Creek formation*

[Analysts: J. C. Hathaway, samples 10 and 11; C. J. Parker, sample 9. Localities shown on pl. 2. Laboratory Nos. 141367, 152106, 152107]

Sample	Percent [a]	Montmorillonite group	Kaolinite group	Feldspar	Quartz and cristobalite	Chlorite and mica
A. Clay fraction						
9	14.5	◬	--	--	--	--
10[b]	2.7	◬	--	--	--	--
11[b]	6.3	◬	--	--	--	--
B. Silt fraction						
9	60.3	×	--	△	--	--
10[b]	42.2	×	--	△	△	--
11[b]	54.0	△	--	△	--	--

[a]Percent of rock in size class, from figure 11.
[b]Glass is major component. (Glass not reported but probably present in most other silt and clay samples.)
◬—only component reported.
△—major component (>20 percent).
×—minor component, generally only a trace.

tical mineralogically with the white siltstone. The sandstone not only is interbedded with the siltstone but intergrades with it, both vertically and laterally. Like their counterparts in the Climbing Arrow formation, the tuffaceous rocks locally contain a rich and varied assemblage of micro-organisms.

In the tuffaceous rocks calcite generally occurs as scattered interstitial aggregates of small clear crystals. In a few places, however, calcite is very abundant in rocks that look like carbonate-poor siltstone, but are harder and contain many tiny vertical tubelike openings. In such beds, calcite forms as much as half the rock, surrounding the ash fragments and replacing and embaying their margins. Such calcite-rich rocks are rarely more than 3 feet thick, with gradational lower contacts into calcite-poor siltstone or sandstone but sharp upper contacts. Similar calcite-cemented tuffaceous sandstone, from much younger rocks on the east side of Madison Valley, has been described by Howard (1932, p. 11–12).

The other rock types closely resemble their lithologic counterparts in the older formations of the Bozeman group, except that the rocks of the Dunbar Creek formation despite their lesser age are somewhat better indurated, so that the sand of the Climbing Arrow corresponds to the yellow sandstone of the Dunbar Creek, and some of the Dunbar Creek clay is indurated enough to have subconchoidal fracture and be classed as claystone.

Fossils are not uncommon in the formation north of the Jefferson River. In addition to micro-organisms, the limestone and calcite-rich siltstone locally contain many gastropods and a few ostracodes. The yellowish siltstone and sandstone contain scattered brontothere, turtle, oreodont, and rodent remains. South of the Jefferson River macrofossils of all kinds are rare, but parts of an anthracothere have been recovered from quartzose sandstone, and some of the white siltstone contains gastropods.

The Dunbar Creek formation is overlain only by Quaternary deposits, with marked angular unconformity, except in sec. 3, T. 2 N., R. 1 E., where a small mass of gravel that has yielded the remains of a late Miocene or early Pliocene mastodon lies on the Dunbar Creek. This gravel, discussed later, is no more than 250 feet above the base of the Dunbar Creek formation, demonstrating deep middle Tertiary erosion.

AGE

The Dunbar Creek formation is at least in part of early Oligocene age. Early Oligocene brontothere and oreodont remains, mentioned earlier in connection with the age of the Climbing Arrow formation, have been recovered from all parts of the Dunbar Creek formation flanking the Hossfeldt Hills. The only vertebrate materials from the formation south of the Jefferson River are many articulated trunk and limb bones of an anthracothere (identified by G. E. Lewis, written communication, Apr. 12, 1957) from hard pebbly sandstone about 400 feet above the base of the formation in the bluffs facing the Madison River about 18 feet below the rim in NW¼SW¼ sec. 29, T. 1 N., R. 2 E. (loc. 384, pl. 2). Unfortunately the bones, though numerous, include none sufficiently diagnostic to permit assigning a more specific name, or a more precise age than Oligocene.

Poorly preserved gastropods are widespread in the formation. According to D. W. Taylor (written communication, May 29, 1957), who studied nine collections, none are diagnostic as to age. Of some interest, however, in connection with depositional environment, the gastropods are discussed later under the heading of origin.

Thus, no more than 400 feet of the formation can be dated as specifically as Oligocene and only the lower 250 feet as early Oligocene. Possibly, part of the formation may eventually prove to be of middle Oligocene age but not younger. Deposits with late Oligocene or early Miocene fossils, discussed later, are known to lie on the eroded surface of the Climbing Arrow formation and it is unlikely that the Dunbar Creek continued to be deposited during the pre-late Oligocene or early Miocene hiatus. For the present the Dunbar Creek is assigned simply an Oligocene age.

ORIGIN

The rock types in the Dunbar Creek are nearly all the same as those in the Climbing Arrow and presumably are of like mixed origin: partly fluviatile, partly eolian, partly lacustrine. Their very different proportions, however, indicate a marked shift in depositional environment. In Dunbar Creek time, drainage rarely passed out of this part of the Three Forks basin, but was largely interior, to judge by the nature of the dominant rock type of the formation. The well-sorted thick-bedded laterally persistent white sandstone and siltstone, formed largely of poorly rounded, little altered rhyolitic ash, are plainly not deposits made in stream channels or flood plains, but rather are indicative of lake or bolson deposition.

If the ash was not brought in to the basin by streams, it must have blown in. But whether the ash of a particular stratum fell on dry land or in standing water is not always easy to determine. The ash in many strata certainly fell in standing water, as demonstrated by the presence of intimately mixed, abundant, and varied quiet-water micro-organisms. Ash-rich strata in the Madison Bluffs have also yielded fresh-water snails (loc. 20018–20019, table 3 and pl. 2). On the other hand, samples from many other ash-rich beds lack micro-organisms and most of the gastropods from the formation north of the Jefferson River are land snails (locs. 20011–20017, table 3 and pl. 2). This, however, is purely negative evidence and of little value.

Another bit of negative evidence is the absence of upside-down graded bedding. Theoretically, this should develop, with finer ash grading upward into coarser, when pumiceous ash of varying grain size falls into standing water, because larger pieces of pumice, with more air holes, should float longer than smaller ones. At least one such occurrence is known (Bateman, 1953). Possibly, this type of gradation is actually present in some beds, but was not detected in casual examination owing to the small total range in ash size.

Better evidence, favoring at least a modest proportion of bolson deposition for the white strata, comes from the fact that they are everywhere channeled by sands and gravels derived from the neighboring highlands and containing little contemporaneous pyroclastic debris. Such channels appear wherever the formation is exposed for as much as 50 feet vertically or a few hundred feet horizontally. If the bulk of the white rocks were deposited in a lake, it must have been shallow and often dry.

The paucity of identifiable pyroclastic material in the coarse channel clastic rocks is puzzling. Rhyolitic ash surely fell on the surrounding hills at the same time it fell on the basin floor, and much of the stream load must have been contemporaneous pyroclastic debris. The crystal fraction of the ash is of minerals common in the pre-basin rocks, and obvious signs of their pyroclastic origin disappeared when they entered the stream system. The rhyolitic glass that fell on valley sides, presumably as slow to alter to clay as glass of similar composition in a lake, was probably transported as glass, rapidly reduced to silt-size in transit and eventually deposited with the fines outside the channels. Glass that fell on interfluves and long remained there may have altered to clay before entering a stream and in transit lost all overt signs of its pyroclastic origin.

The rare thin porous limestones that grade upward from tuffaceous siltstones, but have sharp tops, may well be evaporitic rocks reflecting periodic drying; or they may be fossil caliche horizons, even more suggestive of general bolson rather than lake conditions. The fabric, however, might have been produced by circulating ground water (Howard, 1932, p. 21).

The climate of Dunbar Creek time is as uncertainly known as the mode of deposition of the tuffaceous rocks, and for much the same reasons. The idea that conditions were less humid than in Climbing Arrow time naturally follows from the suggestion of dry-land deposition. If there was a major shift in climate from Climbing Arrow to Dunbar Creek time it was not sudden, as the two units are gradational and the lower part of the Dunbar Creek contains the same sort of vertebrate population as the Climbing Arrow.

MIDDLE AND UPPER TERTIARY GRAVEL

Fossil evidence indicates that deposits of late Oligocene—early Miocene, late Miocene, and possibly early Pliocene age occur in the quadrangle, but the rocks containing the fossils are not mappable.

Articulated remains of a late Oligocene or early Miocene rhinoceros, *Diceratherium armatum* Marsh, from a few miles south of Three Forks, have been described by Wood (1933, 1938). Wood did not himself collect the material, and the locality data furnished are contradictory (2 miles southwest of Three Forks in the 1933 report; 4 miles south of Three Forks in the 1938 abstract) but it seems that the locality was in coarse sandstone overlying the bentonitic clays of the Climbing Arrow formation (called Thompson Creek beds by Wood), and probably therefore from between the Climbing Arrow formation and Quaternary bench deposits. Other remnants of Miocene deposits may be preserved similarly, appearing only at bench rims. In this category may be channel deposits, consisting of 10–25 feet of cross-stratified pebble conglomerate and fine sandstone, exposed for a few hundred feet in the bluffs in the NW¼ sec. 29, T. 1 N., R. 2 E., below a thin capping of Quaternary cobble gravel. These rocks are stained dark yellowish orange, are variably cemented with clay and silica, and contain many fragments of opalized wood but no other fossils. They resemble some of the channel deposits in the Climbing Arrow formation, and the resemblance is strengthened by the fact that they overlie grayish-olive bentonitic clay and grayish-yellow bentonitic siltstone. The bentonitic rocks here, however, are less than 10 feet thick, and overlie a thick succession of white glass-rich siltstones and fine sandstones, typical of the Dunbar Creek formation. The orange conglomerate and sandstone are much younger than Climbing Arrow and could be Miocene. Too small to map separately, the exposures are included with the Dunbar Creek formation.

Fragments of a horse tooth, identified as *Merychippus* sp. of late Miocene age by Lewis (written communication, Apr. 12, 1957) were collected from rounded clay-matrix pebble gravel capping a low hill in SE¼ NW¼SE¼ sec. 3, T. 2 N., R. 1 W., 400 feet southwest of the Eocene vertebrate locality in the Climbing Arrow formation, and 15 feet higher (loc. 444, pl. 2 [(red) D276]. The gravel is much like that designated old Quaternary alluvium, widespread in this vicinity, and it well may be that some of the alluvium mapped as Quaternary is really of late Tertiary age; or, conversely, the gravel may indeed be Quaternary and the tooth reworked from older rocks.

About 1947, most of the skull, several teeth, and some other bones of a mastodon were found west of the junction of Highway 10N and Mud Spring Gulch road by Dr. Nick Helburn, Professor of Geography at Montana State College, and a group of students. The remains were taken to the college campus where I examined them. The exact locality is not known, but seems to have been within NE1/4SE1/4 sec. 3, T. 2 N., R. 1 E., just below the rim of the bench, capped with angular gravel, that occupies the northeast part of sec. 3. Lenses of subrounded gravel crop out at several places around the rim, between the angular-gravel cap and the Dunbar Creek formation, and the mastodon remains presumably came from one of these, for many subrounded pebbles were packed in the skull cavity when I saw it. Through the generosity of Professor Helburn, the remains of two cheek teeth were made available for study. Jean Hough (written communication, Mar. 4, 1954) found the material to be too incomplete for certain generic identification, but definitely "to belong to a primitive mastodon and not to the more common upper Pliocene or Pleistocene species. More complete material is needed to determine to which of the Miocene or Pliocene genera they should be referred." G. E. Lewis also examined this collection (written communication, Jan. 19, 1956); he described the animal as "a gomphotheriid mastodon [(red) D109], possibly referable to a *Rhynchotherium* such as that from the Deep River fauna of Montana. The age is certainly not older than late Miocene, and might be early Pliocene."

The distribution and stratigraphic relations of these later Tertiary gravels plainly indicates great disconformity with the Eocene and some Oligocene rocks. In terms of time, the shortest interval separates the late Oligocene or early Miocene deposits with *Diceratherium* from the early Oligocene part of the Climbing Arrow formation; longer is the interval between the mastodon-bearing latest Miocene or early Pliocene gravel and the underlying Oligocene Dunbar Creek formation; longer still the interval between the *Merychippus*-bearing late Miocene gravel and the underlying late Eocene part of the Climbing Arrow formation. In terms of structure, the younger Tertiary gravels channel the older formations; the scale of channeling, however, seems no greater than that of the sandstone and conglomerate lenses within the older formations, so that significant angular unconformity is not suggested. In terms of thickness, it is reasonable to assume that many hundreds of feet of Climbing Arrow and Dunbar Creek rocks have been removed by intra-Tertiary erosion, but due to the inconstant initial thicknesses and lateral variation of the units, and their uncertain structural history, the assumption is hard to fortify with even crude measurements.

Little can be said about the origin of these elusive later Tertiary deposits. The gravels are much like those in the older rocks of the Bozeman group, suggesting similar sources and regimens. The meager faunal evidence suggests a more humid climate with a narrower temperature range than that of the present.

QUATERNARY

Rocks of known or probable Quaternary age, subdivided into six map units, are thick and continuous enough to map over more than 110 square miles, and over a vertical range of more than 1,000 feet. The thickest and most extensive Quaternary deposits are those made by the present streams and their predecessors. The stream deposits flooring valleys, whatever their textural variations, are combined in a single map unit designated young alluvium. Older stream deposits, graded to surfaces above the present flood plains, constitute a second unit of formational rank, named old alluvium. It is subdivided over large areas into two members, rounded gravel, on valley plains, and subrounded gravel, in tributary valleys. Also mapped are thin but extensive deposits of angular and subangular fan gravel and sheets of eolian silt.

The old alluvium, the fan gravel and the eolian silt have intricately overlapping relations and cannot readily be discussed in chronological order. They are described, therefore, in order of abundance, followed by a brief description of the young alluvium and a discussion of the ages and stratigraphic relations of all the Quaternary deposits.

Although the limits of the Quaternary deposits were mapped with care, the rocks within each unit have been but casually studied. If the preceding discussion of the Tertiary rocks may fairly be called a start toward their understanding, the ensuing discussion of Quaternary deposits must be regarded as merely a sympathetic gesture in their direction.

OLD ALLUVIUM

Old alluvium, graded to levels above the flood plains of the Jefferson and Madison Rivers and actively being eroded, is exposed over 30 square miles. Remnants at low altitudes, on the sides of Jefferson Valley, are as much as 400 feet thick; at higher altitudes they are much thinner, generally less than 15 feet thick on high benches.

The formation lies mostly on soft, easily eroded Tertiary rocks. In some places, however, particularly in the southwest and northwest parts of the quadrangle, it lies in shallow valleys cut on all sorts of hard bedrocks ranging from Precambrian gneiss and arkose to Paleozoic limestone and quartzitic sandstone, Upper Cretaceous volcanics and Cenozoic basalt.

The old alluvium almost everywhere dips at angles of less than 2° toward the junction of the Madison and Jefferson Rivers.

Most of the old alluvium is gravel. The higher an alluvial remnant, and the farther from the present rivers, the greater its proportion of gravel: the highest old alluvium, capping the 5,050-foot bench at the south edge of the quadrangle, is all gravel; the lowest, just above the flood plain of the Jefferson River, contains large volumes of sand and silt; the old alluvium on the sides of the valleys of the present tributaries has increasingly large proportions of sand and silt downstream.

The main size-grades in the gravel are large pebbles and small cobbles, though boulders as much as 2 feet long are present. The gravel tends to coarsen somewhat away from the present master streams, but the rate of coarsening seems much less per unit distance than in the young alluvium. The average degree of rounding in the stones varies markedly among gravel bodies. Large volumes in the southern part of the quadrangle are dominated by rounded gravel, whereas other large volumes, mostly north of the Jefferson River, are mainly subrounded gravel. This is an overall aspect quite apart from the downstream increase in degree of rounding to be expected in any stream gravel, and is the basis for recognizing two gravel members, rounded gravel and subrounded gravel.

The gravel is dominated by noncarbonate rocks of low porosity: quartzitic sandstone, medium- and coarse-grained quartzite, vein quartz, quartz-rich gneiss, and dark volcanic rocks. All the abundant rock types are common in the surrounding hills.

ROUNDED GRAVEL

The rounded gravel member has been mapped over an area of 16 square miles south of the Jefferson River. Rounded, well-sorted pebble and cobble gravel mantles most of the broad smooth benches south of the river and drapes discontinuously over the intrabench slopes. Benches are numerous north of the river but, remarkably, no similar deposits cap them, unless some are concealed by soil on the benches in the large area of undifferentiated old alluvium across the river from Three Forks. It is unlikely that rounded gravels were once extensive north of the river, but have been utterly stripped, while their counterparts to the south remained. Rather, rounded gravel seems never to have been deposited on the northern benches. The gravel capping evidently retards erosion of the southern benches while the northern ones, unprotected, are being rapidly removed.

Many gravel patches are large, covering more than a square mile, and the gravel cover on the highest (5,050-foot) bench extends continuously for 7 miles south of the Three Forks quadrangle. The gravel is by no means the continuous and presumably thick mass it appears to be, however, in the 9-mile stretch south of Fairview Cemetery, where it blankets benches and slopes alike. Actually, the gravel is thick only on the benches. On the slopes Tertiary strata make innumerable small outcrops and have been encountered at many places by digging through a gravel film a stone or two thick, but the Tertiary outcrops are rarely continuous enough to map. The gravel film, no doubt let down from the bench above, is not distinguishable from gravel in place and is thus mapped on many slopes. The original relation between gravel and terrain is well shown at the base of the highest, oldest bench where streams have become integrated enough to sap the colluvial cover and reveal the Tertiary rocks continuously along the deeply dissected bench front, with gravel confined to the bench surface.

Although the gravel protects the underlying bench, even the youngest lowest bench is dissected enough to demonstrate that the gravel is thin in comparison with its areal extent, although close measurement of thickness is made difficult by the tendency of the uncemented round stones to roll downhill even on low slopes. Where gullies intersect locally cemented gravel, however, the thickness can readily be measured: in such places it is generally on the order of 5 feet and rarely exceeds 15 feet. Probably, it averages around 10 feet thick and has a maximum thickness of less than 50 feet. The gravel at the head of the highest bench, 17 miles south of Three Forks, is thicker, about 100 feet thick according to Alden (1953, p. 35).

Judging by the surface layer only, the gravel at each bench level differs markedly in coarseness, sorting, and lithology from other bench levels. The surface of the highest gravel body is smoothly paved with rounded egg-shaped or discoidal cobbles, mostly 4 to 8 inches long, of quartzite or quartzitic sandstone, colored pale grayed tones of orange, brown, or red. Nearly every cobble has a yellowish-brown skin, whatever the color of the rock, and many are studded with neatly curved percussion marks. The surface layer on each successively lower bench exhibits similar large cobbles but has successively higher proportions of small cobbles and of pebbles, so that the average grain size is progressively smaller and the sorting poorer; also, the prevailing quartzite lithology is progressively diluted with other rock types, particularly gneisses and volcanic rocks.

A foot or two below the surface, however, these apparent differences disappear. If there are any systematic lithologic or textural differences from bench to bench, that might be demonstrated by detailed counts or other measurements, they are small. Viewed in three dimensions, the rounded gravels from highest to lowest can be categorized as indistinctly bedded sheetlike pebble and cobble strata, with a little interstitial sand, dominated by quartzite rocks but with large proportions of other crystalline rocks.

Though generally unconsolidated, the rounded gravel in several places is thoroughly cemented by white chalky calcite. Lime-cented zones occur in the highest rounded gravel but are most abundant at intermediate levels. None have been found at altitudes below 4,440 feet. As exposed in stream cuts, the cemented zones consist of strips, a few tens or hundreds of feet wide and the entire thickness of the gravel, that are thoroughly impregnated with calcite, in places so much so that the pebbles are separated an inch or more (fig. 12). The distribution of lime cement seems to be channellike rather than blanketlike but the trends of the channels are unknown as the cemented zones cannot readily be traced away from the outcrop. Transitions to uncemented gravel are abrupt, occurring within a foot or two laterally. The calcite is apparently not the product of evaporation but rather appears to have been deposited from laterally moving channeled ground water or perhaps from hot-spring water rising along fractures.

FIGURE 12.—Lime-armored Quaternary gravel, capping bench in NW¼ NW¼ sec. 18, T. 1 S., R. 2 E.

SUBROUNDED GRAVEL

Subrounded pebble gravel is mapped separately from the other deposits of old alluvium in many large lobate areas north of the Jefferson River, and in the drainage of Willow Creek. Covering about 15 square miles, the bodies of subangular gravel differ from the bodies of rounded bench gravel in being relatively thick in comparison with their width, and in having rudely triangular rather than sheetlike cross sections. Only locally are they distinctly benched. In many places, they reach almost or quite to the top of the valleys they occupy but do not spread over the bordering benches, suggesting that their original extent was never much greater than their present extent. The base of the subangular gravel is rarely visible so its thickness can only be estimated. On the unsafe assumption that the valleys holding these gravels had cross sections and long profiles shaped like those of present valleys, the wedges of subangular gravel appear to be more than 50 feet thick over long stretches and exceptionally as thick as 100 or even 150 feet. A thickness of at least 60 feet of old alluvium, in a patch too small to map, is penetrated by a pumping water well at the head of the valley in Three Forks shale in NW¼SE¼ sec. 19, T. 3 N., R. 2 E.

Most of the gravel is in the range of pebbles and small cobbles, but small boulders are common. Surface layers contain little fine material but excavation reveals a large sand and silt fraction. Although most stones are subrounded, many are subangular.

The lithology of the subrounded gravel varies widely. In the northwestern part of the quadrangle, it is dominated by dark andesitic volcanic rocks. On the north side of Mud Spring Gulch, it contains large proportions of coarse monzonitic rocks. Near the southwest corner of the quadrangle it is heavily charged with fragments of gneiss and of vein quartz.

ORIGIN OF OLD ALLUVIUM

The form and distribution of the deposits show that the rounded gravel and most of the undifferentiated old alluvium near the present Jefferson River are valley-plain deposits and that the subrounded gravel is fill in tributary valleys.

That the rounded gravel was deposited by a major perennial river is indicated by its rounding, its sorting, and its considerable volume and extent. All the gravel could have been derived from the basin margins, to judge by lithology. More distant transport for much of it is suggested by the fact that the gravels are about as well rounded and sorted at the remote edges of the basin as they are near the outlet at the head of the Missouri. The gravels are little different from highest to lowest except for the surface layer, indicating that the regimen was about the same during the cutting and filling of the entire flight of terraces capped by rounded gravel. The differences between surface layers are regarded as lag phenomena.

Most of the rounded gravel, as noted earlier, is in the southern part of the quadrangle: the patches descending from the Madison Range are broad and long; those descending to the Jefferson River from the north are narrow and short. This asymmetry of distribution, apparently a consequence of northerly or northwesterly tilting, is discussed later.

The distribution and texture of the subrounded gravel show that it was deposited by tributary streams. That the tributaries were short is suggested by the lithology of the gravel, everywhere directly reflecting the kind of resistant rocks that crop out upstream.

Presumably, deposition of subrounded gravel in tributaries accompanied deposition of all the rounded gravel. The visible subrounded gravel, however, seems to represent only the later part of the long history indicated by the size and distribution of the rounded gravel. No subrounded gravel is preserved above 4,700 feet. None is strikingly benched, and it is recognized only in valleys that are the sites of present streams. Possibly, older remnants exist, but were not recognized because they are not dissected. More likely, the older subrounded gravel has been eroded. Lacking discharge for effective sidecutting, the tributaries became fixed at an early stage; the older fills were swept out during episodes of downcutting and only the younger valley fills are preserved.

Certainly the landscape and probably the climate were greatly different from the present when the highest gravel was laid down. The relief within the quadrangle must have been far more subdued than now. Only 10 square miles of very resistant bedrocks are now topographically higher than the highest rounded gravel on the 5,050-foot bench, and less than 1 square mile is as much as 500 feet higher. Relief between ridge tops and the flood plains matching that of today—800 to 1,300 feet—would require removal of a volume of resistant rocks inconceivable in the brief time available, even if the oldest old alluvium is of late Pliocene age. Lower relief than at present, or at any rate lower average stream gradient, is directly suggested by the fact that the average and maximum sizes of clasts in the old alluvium are distinctly smaller than in the young alluvium. In turn the young alluvium was formed under geomorphic conditions much like those of the present if not necessarily under the present climate. Successively lower deposits of the old alluvium were of course deposited under physiographic conditions that successively approached those of the present. Their distribution indicates increasing relief and narrowing flood plains during old-alluvium time.

As for climate during that time, little positive evidence exists. Ancient soil horizons are lacking, or at any rate not notably developed; lacking, too, are comparable features such as rotted pebbles or authigenic growths of such climate-sensitive minerals as iron and manganese oxides. The local lime-cementation has little climatic implication if correctly interpreted as of ground water or spring rather than evaporative-capillary origin. The varying lag concentrations of quartzite cobbles seem to be a function of time rather than of climate.

The textures and structures of the subrounded gravel compared with those of young alluvium in the same valleys imply a much more humid climate than now. Not only are the older gravels much better sorted and rounded than adjacent young alluvium but they virtually lack torrential false-bedding, common in the youngest fill, implying that they were the products of streams with more and steadier discharge than the present intermittent tributaries.

FAN GRAVEL

Widespread north of the Jefferson River, but rare to the south, are extensive fan-shaped deposits of angular and subangular gravel. Only the largest are mapped. They consist mostly of unsorted slabs of limestone heavily encrusted with caliche on the base, and rarely on other sides. Most of the deposits are sheetlike, commonly no more than 2 or 3 stones thick; a few are wedge-shaped and much thicker. The clasts are generally in the pebble and small boulder range, but rarely are as much as 3 feet long. In places the stones are in a sand or mud matrix but usually they are loosely strewn about the surface. The older the deposit, as indicated by its degree of dissection and of topographic isolation, the smaller the proportion of interstitial fines and the thicker and more uniform the caliche crusts.

Most of the fan gravel mantles benchlands, below altitudes of 4,900 feet, between the limestone ridges and the flood plains. In plan, the shapes of the benchland deposits vary somewhat but they are all rudely fanlike, broadening downhill, or consist of scattered patches of gravel that seem to be the dissected remnants of an originally fanlike gravel film. The larger deposits are draped over several benches and the intervening steeper slopes. Blanketing the base of the Hossfeldt Hills is a virtually continuous apron of subangular gravel, with a copious sand matrix, that extends to the Jefferson River flood plain on the southeast and to the flood plain of Mud Spring Gulch on the southwest. North of Highway 10S near the west edge of the quadrangle are scattered patches, with an aggregate fan shape, of angular limestone gravel that lies on a variety of rocks ranging from Belt arkose to old alluvium; the distal end of this tattered fan is more than 100 feet above the Jefferson River flood plain. A large area of angular gravel east of Willow Creek is mostly fragments of dacite. Smaller strips or caps of angular limestone gravel lie on benches in the center of the quadrangle, west of Three Forks Junction; about 1½ miles west of Silver Sage Ranch at the edge of the quadrangle; and above Fairview Cemetery a mile southeast of Three Forks.

Thicker deposits of subangular gravel occupy the valleys of a few short tributaries. Such deposits are at the north edge of the quadrangle in secs. 20, 21, 28 and 29, T. 3 N., R. 1 E.; between the monzonite pluton and Copper City in secs. 25 and 26, T. 3 N., R. 1 E.; east of the mouth of Milligan Canyon; and on tributaries of Sand Creek in sec. 33, T. 1 N., R. 1 W., and sec. 4, T. 1 S., R. 1 W. These valley-filling gravels seem slightly better rounded than those on the benchlands and have relatively little caliche. They tend to be composed of whatever well-indurated rock crops upslope. This is generally limestone but in places is quartzitic sandstone or coarse-grained igneous rock. Exposures reveal only the upper few feet of these deposits; in places they are no doubt many tens of feet thick.

ORIGIN

The texture, lithology, and topographic position of the unsorted angular limestone gravel on the benchlands suggest colluvial rather than stream origin. The rudely fanlike ground plans, and the plain evidence of miles of movement of large fragments over low slopes, especially well shown in the fan remnants lying on North Boulder formation north of Jefferson School, favor transportation by sheetflow or, better, mudflow rather than by unlubricated creep or mass flowage. It well may be that such deposits are not the results of torrential rains or thaws but instead may require the gentle and protracted wetting of material of appropriate texture lying at the base of steep slopes. One trouble with a torrential origin is that there are too few mudflow deposits, given the many hundreds of torrential rains that must have occurred in the many thousands of years during which the benchlands have been exposed. Perhaps the operative process was akin to fluidization processes in gas-solid systems, now widely used in industry (summarized by Reynolds, 1954). The essential element in connection with gravity-driven water-solid systems like mudflows is that optimum small amounts of water may act to reduce the coefficient of friction of the mud, to permit flow on low slopes, but not reduce its viscosity to a point below which it cannot move large fragments. The water requirements for inducing and maintaining flow in a gravel-soil mixture may have so narrow a range that exactly the right rate and duration of precipitation (or thaw) on properly prepared materials and slopes is statistically improbable, thus accounting for the paucity of colluvial gravels. However transported, the angular gravels are plainly the products of an arid or semiarid climate like the present.

The subangular gravels in valleys and mantling the Hossfeldt Hills are probably partly colluvial, but mostly alluvial. That their stones are subangular and include much limestone reflects minimal transport of the materials in running water, a simple consequence of the shortness of the depositing streams. These deposits appear to be genetically intermediate between the angular colluvial benchland gravels and the subrounded valley fills of old alluvium.

SILT

Light-colored silt of probable eolian origin is widespread in the benchlands. Such silt is distributed over perhaps 30 square miles, of which about 10 square miles are covered continuously enough to map. There are three large general areas of silt; one at the north edge of the quadrangle, one at the west-central edge, and one south of Three Forks. The deposits along the north edge are at the south margin of a great silt sheet that cloaks the broad divide between the Townsend and Three Forks valleys. Those at the west edge seem to be a part of a long strip of similar material that extends 3 or 4 miles farther west into the Jefferson Island quadrangle. Thickness of the silt, hard to measure without much digging, is estimated to be less than 5 feet over most of the mapped areas and only exceptionally more than 15 feet. It weathers readily to soil, and most of the silt areas are in active cultivation.

The silt lies at varying altitudes and on a variety of rocks and surfaces. A unifying relation is that all the mapped silt bodies are south of extensive bare exposures of poorly consolidated Tertiary rocks rich in clay and volcanic glass.

The unit consists mostly of yellowish-gray to very pale orange angular silt and clay with scattered grains and streaks of rounded fine sand. Several inches at the top are commonly altered to a yellowish-brown clayey soil. No stratification is visible, even in continuous vertical exposures as much as 10 feet high, as in the silage pit in SW¼NW¼ sec. 2, T. 1 S., R. 1 E., and along the former course of Highway 10N (west of the present highway) in NW¼ sec. 7, T. 3 N., R. 1 E. The dominant materials are volcanic glass, quartz, and clay minerals with minor amounts of mica, feldspar, and calcite. The silt is riddled in many places with thin vertical tubular openings, presumably root scars. Despite its lack of cement the silt, presumably because of the angularity of its grains, is able to maintain vertical walls for many years in artificial cuts, a feature well shown at the localities mentioned.

The deposits have the classic features of loess. Their eolian origin seems certain. Distribution of the loess suggests that northerly winds swept the grains off bare outcrops of the Climbing Arrow formation and similar units, eventually dropping them either at the windward base of high obstructions, like the Madison limestone ridge above Dry Hollow, or in the lee of low obstructions. Part of the loess, especially that south of Three Forks, may have been blown off aggrading flood plains.

YOUNG ALLUVIUM

Stream deposits graded to or below the present flood plains occupy 60 square miles. Regardless of their texture, they are indiscriminately shown as young alluvium. Exposures are abundant, owing to a very recent, possibly still continuing, episode of rejuvenation that has led to as much as 15 feet of trenching by tributaries. (See fig. 15.) The deposits exposed along present channels of the Jefferson and Madison Rivers are dominantly subrounded to rounded, clean, fairly well sorted gravel and sand. The gravel is but rarely coarser than cobble size. Large areas of swampy or soil-covered ground are probably underlain by silt and mud but little of such finer material is exposed. Along the tributaries the dominant visible material is poorly sorted subangular silty and sandy gravel containing many boulders. Rarely, lenses a few feet thick of better sorted material—sandy cobble gravel, pebble sand, and reddish-brown clayey silt—appear. Judging by a few tributaries followed to their mouths, this poorly sorted, poorly rounded alluvium extends along the tributary channels almost if not quite to their junctions with the through rivers. With more observations it would have been feasible to subdivide the young alluvium in the same way as the old, into two gradational lithologic units in characteristic settings: poorly sorted, poorly rounded material, in tributary valleys; and much better sorted and better rounded detritus on the flood plains of the through rivers.

The thickness of the young alluvium is unknown but is probably less than 50 feet throughout. In the tributaries the thickness is greater than the depth of recent trenching—15 feet maximum—and may be several times this thickness; above Milligan Canyon, drilling by the Bureau of Reclamation (H. C. Elliott, oral communication, 1957) has shown that the valley fill is rarely more than 30 feet thick. Presumably it is not greatly thicker under the flood plains of the Jefferson and Madison Rivers, for the tributaries are accordant. Many water wells have been drilled in the flood plains but it is hard to tell from driller's log where the wells leave young alluvium and enter poorly consolidated Tertiary deposits.

Very little of the young alluvium was deposited under present climatic conditions. Under the present climate, the streams are primarily intrenching themselves in their earlier deposits; such deposits as they make are scanty and confined to mere patchy films on valley floors or sides. The bulk of the young alluvium was surely laid down under more humid climate than now prevails, and most of it is of Pleistocene rather than Recent age.

AGE AND STRATIGRAPHIC RELATIONS OF QUATERNARY DEPOSITS

Fossils have been recovered only from the young alluvium so that the age of the other Quaternary deposits must be approached indirectly.

The age of the young alluvium is late Pleistocene and Recent. The right scapula of a *Bison antiquus?*

of Mankato or Cary age (identified by the Upper Cenozoic Research Group, over the signature of Edward Lewis, written communication, Nov. 18, 1955) was excavated from alluvium graded to the Jefferson River flood plain 2 miles southeast of the Buttleman Ranch in center NW¼ sec. 23, T. 1 N., R. 1 E. (loc. 394, pl. 2). The alluvium here is deeply trenched by the present stream. The scapula, showing little sign of erosion, was found embedded in the east bank of the trench 3½ feet below the surface. Because no other bones of the animal were attached or nearby, there is some possibility that the bone was reworked from Quaternary gravel that crops out upstream and uphill.

Remains of modern cow (identified in the field by Jean Hough) were recovered from the upper foot or two of alluvium or associated colluvium in several places. The presumption is that modern cow did not appear in the area until the first settlement by white people around 1860, and thus that significant flooding and deposition has occurred in the area within the past century (the trenching is not necessarily younger than burial of the cows but may be contemporaneous).

Neglecting the scant paleontological evidence, it seems reasonable to assume that the flood plains of the Jefferson and Madison Rivers, 2 to 4 miles broad, and the great volume of deposits beneath them required far longer than the Recent to form. In developing the present flood plains something on the order of 400 to 600 feet of deposits, mostly old alluvium, had to be removed from the Jefferson Valley. This is indicated not only by the height of paired bench gravels above the rivers but is also required to achieve superposition of the present Jefferson channel upon at least 450 feet of cover between the Madison limestone ridges 4½ miles southwest of Willow Creek (see fig. 16). It also must have taken far longer than the Recent to cut the towering cliffs below Milligan Creek. Of course both of these latter erosional feats may have been performed by early Tertiary rivers, and the Jefferson's task may have been the somewhat easier one of exhuming the evidence by removing soft earlier Quaternary deposits.

The old alluvium represents a long time interval, as implied by its vertical distribution of 1,000 feet, and by the large dimensions of individual gravel masses. Downcutting to each successively lower bench level may have happened quickly, but the extensive lateral shifting necessary to cut and fill terraces miles wide must have required much time. Of course, the entire present width of each terrace need not have been developed before incision of the next lower one; part of the widening has doubtless been by mass-wasting processes long after the rivers had begun working at lower levels and could only affect higher terraces by narrowing them.

The old alluvium is no younger than late or middle Wisconsin for it is disconformably succeeded by young alluvium. Part of the old alluvium could be of Pliocene age because the latest fossil—the gomphotheriid mastodon—locally found in the preceding deposits is no younger than early Pliocene or possibly latest Miocene, and no demonstrably younger rocks have been reported in the vicinity. A Pliocene age would, of course, give more time to develop the long flight of terraces and also the quartzite lag gravel at the surface of the higher deposits, which closely resembles the widespread Flaxville gravel of Collier and Thom (1918) of Miocene and Pliocene age.

Probably the old alluvium is entirely of Pleistocene age. The Madison Valley formation of Douglass (1903) forms the upper part of the east Madison bluffs and dips gently northeastward (mostly in Manhattan quadrangle; see fig. 1B). The plain that slopes northeastward at about 150 feet per mile from the top of the bluffs approximates a dip slope on the Madison Valley formation, according to Dorr (1956, p. 72). The Madison Valley formation is generally regarded as of Miocene and Pliocene age (Wood and others, 1941; Charles H. Falkenbach, oral communication, 1954) although there is some doubt as to the location of the time boundary. Dorr (1956, p. 73) concluded that "Most if not all of the Madison Valley formation is Miocene rather than early Pliocene in age" but Falkenbach cited recovery of Pliocene oreodonts from the uppermost gravels on the bluffs as convincing evidence of Pliocene age for part of the formation. As Dorr presented no fossil evidence that the uppermost bluff-making rocks are pre-Pliocene it seems premature to abandon the long accepted view that some at least are Pliocene. The dip of the Madison Valley formation projects it far above the Three Forks quadrangle. The formation has evidently been truncated by the erosion which preceded the earliest deposits of old alluvium, and removed practically all post-Oligocene Tertiary rocks from the quadrangle. The oldest old alluvium, with its much lower and more northerly dip, appears to be structurally unconformable with the Madison Valley formation. This interpretation of unconformity, highly questionable on local evidence because the rocks involved are not in contact, is supported by direct evidence from the Manhattan quadrangle, sec. 26 and 35, T. 1 N., R. 2 E. (shown me by O. M. Hackett). There, the northeast-dipping Madison Valley formation is unconformably overlain by north-dipping gravel, identical lithologically with the oldest old alluvium and at the same altitude, about 5,050 feet.

If sufficient time is allowed for the great amount of erosion represented by the unconformity between the Pliocene part of Douglass' Madison Valley formation and the rounded gravel, and at the opposite end enough time is allowed to develop the quartzite lag-gravel surface on the old alluvium, it is a reasonable working hypothesis that none of the old alluvium is older than Pleistocene, but that some of it may be earliest Pleistocene.

The fan gravels, to judge by their greatly varying degrees of dissection, have a rather wide age range. All, however, extend to such low altitudes as to show that they are younger than the bulk of the old alluvium. The youngest fan gravel is in the great blanket swathing the Hossfeldt Hills. This blanket, extending to the flood plain of the Jefferson and Madison Rivers and scarcely touched by erosion, is probably younger than all but a small fraction of the young alluvium. The oldest fan gravel would appear to be the most tattered fans of angular gravel at the highest altitudes, but because the deposits are colluvial and not graded to any stream, this sort of evidence is weak. The valley-filling fans seem younger than most of the old alluvium but older than most of the young alluvium.

If the fan gravels are the products of semiarid climate, they were not laid down simultaneously with any of the humid-climate old alluvium. Probably one or more of the episodes of cutting that produced the successively lower terraces on which old alluvium was deposited was accompanied by climatic changes (interstadial intervals?), so that fan gravels were deposited alternately with old alluvium as the climate oscillated in later Quaternary time. (It does not necessarily follow, however, that climatic change was responsible for any of the terraces involving old alluvium.)

The stratigraphic relations of the small area of angular fan gravel above Fairview Cemetery are unclear. This angular gravel is older than rounded gravel on the low-level bench to the east but its low altitude suggests that it is not among the oldest Quaternary deposits. The nearest bedrock source for its limestone blocks, and the nearest deposits of similar material, are 4 miles to the west across the Jefferson Valley. Perhaps this gravel is the remnant of a mudflow that started on the northwest valley wall at a time when the valley was filled with alluvium to the level of the 4,400-foot bench. The present downhill northward extension of the gravel to below 4,200 feet is probably due to postdepositional movement in response to erosion of the northern base of the deposit.

The eolian silt may represent more than one depositional event. The silt along the north edge of the quadrangle may be older than the subrounded gravel, for the silt is cut by the valleys in which the subrounded gravel is exposed. Furthermore, considerable antiquity is suggested by the development of extensive arable soil on the silt. But although the silt is older than the present valley incisions, it could be younger than the subrounded gravel, now being exhumed simultaneously with the silt. The other areas of silt all seem to be younger than most of the old alluvium and older than most of the young alluvium. The silt south of the Jefferson River overlies and is thus younger than the rounded gravel on the 4,400-foot bench, but is older than the young alluvium in the valley below. The windblown silt south of upper Milligan Creek overlies and is younger than the old alluvium there; it is probably a little older than the nearby young alluvium.

There is no direct evidence of the time relations between eolian silt and fan gravel. Presumably windblown deposits are more likely to form in a dry climate, so that the silts were probably deposited concurrently with some of the fan gravels.

IGNEOUS ROCKS

Igneous rocks in the quadrangle occupy 14 square miles and are divided into five main map units. They are mostly of Late Cretaceous to early Tertiary (pre-Bozeman group) age and fall into two categories: extrusive rocks—the Elkhorn Mountains volcanics—of much deformed andesitic lava flows and volcanic breccia that were emplaced before or in the early stages of regional folding, and an array of monzonitic, dacitic, latitic, and andesitic intrusives, emplaced during or after folding in the form of sills, laccoliths, and one semiconcordant pluton. Some of the intrusive rocks, particularly the andesite and latite, may be contemporaneous with the volcanic rocks; most of them, however, seem to be a little younger. Igneous rocks that do not fit these categories are pockets of Precambrian pegmatite in the Precambrian crystalline rocks, noted in the section on the crystalline rocks, and two small masses of olivine basalt, intrusive into rocks near the base of the Bozeman group and of probable early Tertiary age. Not classed as igneous rocks, though they might be, are the large volumes of Tertiary sedimentary rocks made up mainly of volcanic ash but here included with the Bozeman group.

The igneous rocks are much less resistant to erosion than the pre-Tertiary carbonate rocks and quartzitic sandstone but more resistant than the pre-Tertiary clay-rich rocks or the poorly consolidated Cenozoic sediments. Where extensive enough, they develop low rolling somber-hued topography with even less vegetation than the surrounding rocks, but with a thick carapace of weathered material that in most places effectively conceals the nature of the rock and its relations.

About 80 thin sections were studied by H. Frank Barnett who wrote the petrographic descriptions that follow. The locality and rock type represented by each thin section studied is identified by a number and symbol on the data map, plate 2. The thin sections were examined by routine petrographic microscope methods. Mineral volume percentages were estimated against comparison charts; they are least accurate in heavily altered and in very fine grained rocks. Plagioclase compositions were mostly inferred from measurements of extinction angle compared with the high-temperature curves of Tröger (1952); a few untwinned feldspars were studied in immersion liquids and further checked by staining with sodium cobaltinitrite.

Rock names follow the classifications of Kemp and Louderback as modified by R. C. Mielenz (1948). Sufficient information is given for the common rocks so that other classifications could readily be applied. Quartz is considered essential if it amounts to more than 5 percent of the total rock volume. Rocks are described as porphyritic if phenocrysts aggregate from 2 percent to 50 percent, or porphyries if phenocrysts exceed 50 percent.

Eleven typical igneous rocks have been analyzed chemically. The localities represented by analyzed rocks are shown on plate 2. Results of these analyses, and the norms calculated from them, are presented in table 9; the analyses are the basis for a silica-variation diagram (fig. 13).

UPPER CRETACEOUS AND LOWER TERTIARY ROCKS

EXTRUSIVE ROCKS—ELKHORN MOUNTAINS VOLCANICS

Assigned to the Elkhorn Mountains volcanics of Klepper and others (1957, p. 31–41) are stratified, deformed andesitic volcanic rocks that crop out over more than 6 square miles and are the most extensive igneous rocks in the quadrangle. They appear in two areas: in the northwest corner of the quadrangle, and below the Jefferson Canyon thrust west of Jefferson School. In both areas, the formation yields low rolling barren topography that is a distinctive steely purplish gray when viewed from a distance. Close at hand, purplish tones are not so apparent and the weathered surfaces are mostly yellowish brown or olive brown. Freshly broken surfaces are mostly medium dark gray though a greenish or bluish cast is not uncommon.

The dominant rocks are porphyritic andesite with many somewhat indistinct stubby phenocrysts of feldspar and scattered long ones of pyroxene. Nearly all are compact, with few vesicles or amygdules, and their groundmasses are fine grained to medium grained. Many fractures are lined with coarsely crystalline yellowish-gray stilbite in stellate clusters. The andesite layers are generally a few tens of feet thick. Presumably, most of them are lava flows but a few may be shallow penecontemporaneous sills. Interbedded with them are a few thin layers of latite, most of which are no doubt lava flows also; some have a streaky banded structure suggesting welded tuff.

In the northwest corner, the formation includes much volcanic breccia in units as much as 500 feet thick; their lateral extent is unknown but may well be miles. The breccia is an interesting and confusing rock. It is polylithic, with crowded subangular to subrounded fragments, generally 3 to 15 inches long but in places several feet long, of a variety of extrusive and hypabyssal rock types including biotite diorite, andesite, latite, and quartz latite, in a holocrystalline andesitic matrix exactly like the dominant layered andesite of the formation. There are no obvious signs of interaction between the matrix and any of the clasts, and no marked change in grain size in the matrix near clasts. These features, combined with the range in texture, composition and degree of roundness of the fragments suggest mudflow origin. But the uniformly dense holocrystalline porphyritic texture of the matrix, an appearance maintained in thin section, seems unequivocally to indicate a fluid-poor melt rather than a watery mud. Tentatively, these rocks are interpreted to be lava flows, and the large clasts to be xenoliths, largely cognate and in equilibrium with the andesite matrix.

The sequence west of Jefferson School includes one especially interesting and unusual layer, apparently a flow but conceivably a welded tuff, of porphyritic perlite (loc. 16, pl. 2: NE¼NE¼NE¼ sec. 21, T. 1 N., R. 1 W.). It forms a layer 5–20 feet thick continuous for at least 400 feet between massive ledges of latite; it does not thin out but continues beneath cover at either extremity. The rock is very dark gray with a vitreous lustre and subconchoidal, almost hackly fracture. Scattered throughout the glassy groundmass are large grains of feldspar, golden biotite, and green mafic minerals exhibiting rude flow banding.

The total thickness of these layered volcanics is a few thousand feet. The main body of Elkhorn Mountains volcanics is probably more than 10,000 feet thick, according to Klepper and others (1957, p. 32). Near-

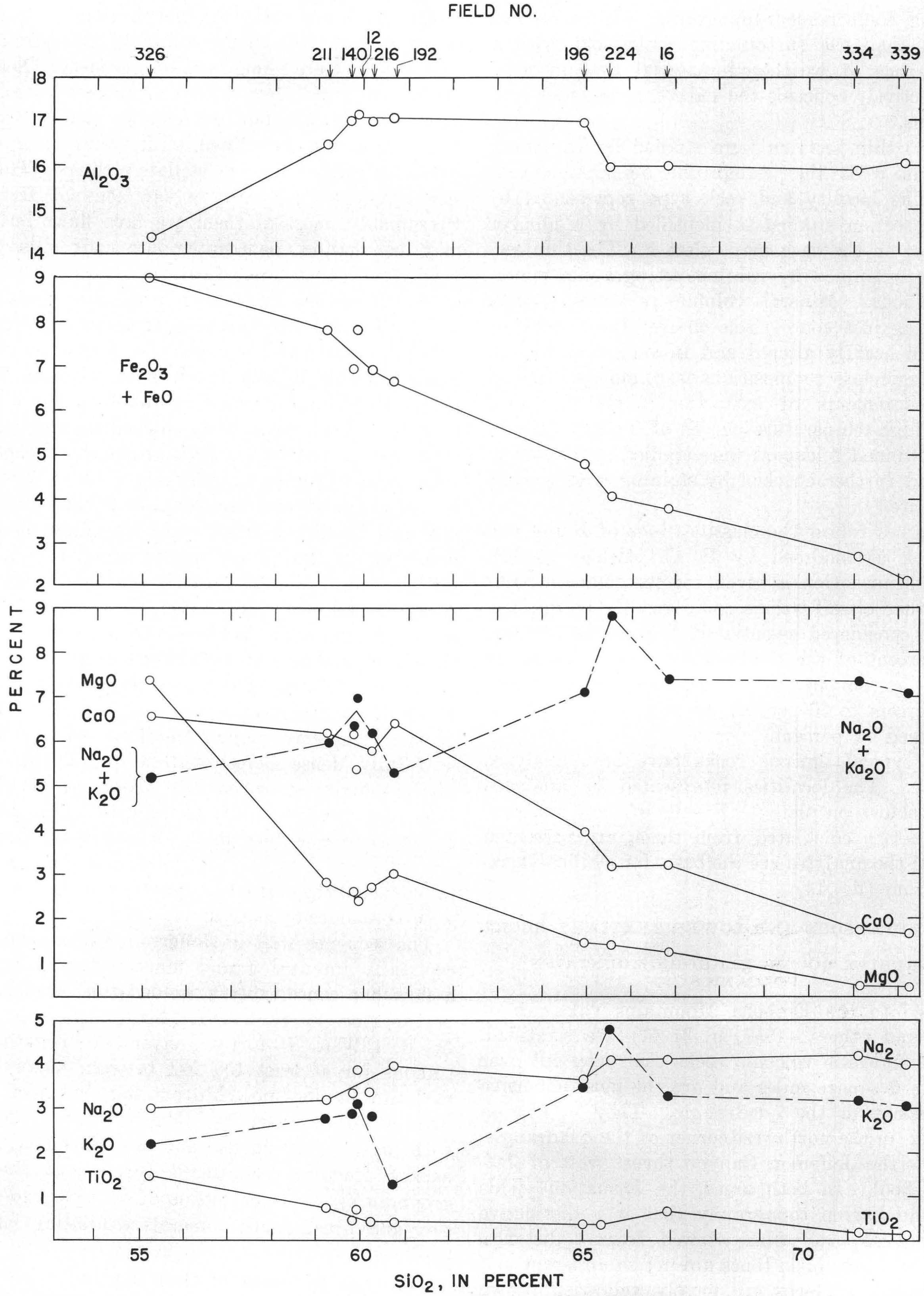

FIGURE 13.—Silica-variation diagram for chemically analyzed igneous rocks, Three Forks quadrangle, Montana.

TABLE 9.—*Chemical analyses and norms of some igneous rocks, Three Forks quadrangle, Montana*

[Localities and rock types shown on pl. 2]

Sample	12	16	140	192	196	211	216	224	326	334	339
Laboratory No. 152	095	096	097	098	099	100	101	102	103	104	105
Chemical composition [Rapid-method analysis by P. L. D. Elmore, S. D. Botts, and M. D. Mack]											
SiO_2	57.9	63.6	58.5	59.4	62.1	58.4	59.7	65.1	54.4	70.2	71.4
Al_2O_3	16.8	15.2	16.8	16.7	16.2	16.3	16.9	15.8	14.2	15.9	15.9
Fe_2O_3	5.9	1.8	2.7	2.0	2.7	3.8	3.4	2.9	1.4	1.5	1.0
FeO	1.7	1.8	4.1	4.5	1.8	3.9	3.5	1.3	7.5	1.1	1.0
MgO	2.3	1.2	2.5	2.9	1.4	2.8	2.7	1.4	7.3	.37	.42
CaO	5.2	3.0	6.1	6.2	4.2	6.2	5.8	3.2	6.5	1.7	1.6
Na_2O	3.7	3.9	3.3	3.9	3.5	3.2	3.4	4.1	3.0	4.2	4.0
K_2O	3.0	3.1	2.8	1.3	3.4	2.8	2.8	4.8	2.2	3.2	3.1
TiO_2	.64	.71	.50	.46	.38	.66	.53	.36	1.5	.20	.17
P_2O_5	.37	.10	.30	.27	.18	.31	.31	.16	.30	.14	.02
MnO	.14	.10	.20	.16	.12	.20	.18	.11	.16	.06	.08
H_2O	1.6	5.0	1.0	1.9	2.2	.88	.78	.42	1.4	.91	1.2
CO_2	$<.05$	$<.05$	.07	$<.05$	.76	$<.05$	$<.05$	$<.05$	$<.05$	$<.05$	$<.05$
Sum	99	100	99	100	99	99	100	100	100	100	100
Norms [Norms calculated by G. A. Macdonald, and Elizabeth Jaffe]											
Q	12.0	20.3	11.5	13.1	19.5	12.1	12.9	15.3	2.5	29.4	31.5
or	17.8	18.4	16.7	7.8	20.0	16.7	16.7	28.4	12.8	18.9	18.4
ab	31.5	33.0	27.8	33.0	29.9	27.3	28.8	34.6	25.2	35.7	34.1
an	20.3	14.7	22.8	24.2	15.3	21.7	22.5	10.6	18.9	7.5	8.1
di wo	1.2	.1	2.3	2.0	0	3.0	1.7	1.9	4.8	0	0
di en	1.0	.1	1.2	1.0	0	1.9	1.1	1.6	2.8	0	0
di fs	0	0	1.1	.9	0	.9	.5	0	1.7	0	0
hy en	4.7	2.9	5.0	6.2	3.5	5.0	5.6	1.9	15.4	.9	1.0
hy fs	0	1.0	4.1	5.1	.8	2.6	2.8	0	8.8	.7	1.1
mt	4.2	2.6	3.9	3.0	3.9	5.6	4.9	3.7	2.1	2.1	1.4
il	1.2	1.4	.9	.9	.8	1.2	1.1	.8	2.9	.5	.3
ap	1.0	0	.7	.7	.3	.7	.7	.3	.7	.3	.0
hm	3.0	0	0	0	0	0	0	0	0	0	0
C	0	0	0	0	1.1	0	0	0	0	2.8	3.0
cc	0	0	0	0	1.7	0	0	0	0	0	0
plagioclase	Ab_{61}	Ab_{69}	Ab_{55}	Ab_{58}	Ab_{66}	Ab_{56}	Ab_{56}	Ab_{77}	Ab_{57}	Ab_{83}	Ab_{81}

by to the west, Alexander (1955, p. 61–73) described a sequence of similar rocks, which he assigned to the Livingston formation, as being more than 9,600 feet thick. Here, near the southeastern limit of the formation, the maximum thickness is likely to be much less. No units seem to be repeated in the steeply dipping sequence west of Jefferson School; if so, the thickness there approaches 1,000 feet.

Within the quadrangle, no depositional contacts of the Elkhorn Mountains volcanics are exposed. In the Elkhorn Mountains (Klepper and others, 1957, p. 24, 40) the volcanics in places gradationally overlie the Slim Sam formation, of Late Cretaceous age, and in other places lie on an erosional unconformity that represents removal of the Slim Sam and other formations at least as far down section as the Morrison formation; a few miles west of the Three Forks quadrangle, volcanic rocks of the same sequence lie with angular unconformity on rocks as old as the Madison group (Alexander, 1955, p. 67). In the quadrangle, the volcanics are unconformably overlain by rocks as old as the Eocene part of the Climbing Arrow formation.

PETROGRAPHY

By H. FRANK BARNETT

The Elkhorn Mountains volcanics are represented by 13 thin sections, located on plate 2; 2 typical andesites (192, 196) are also represented by chemical analyses on table 9. The matrix of the breccias seems identical with the andesite flows, and the descriptions apply to both. The rare latite in the formation (represented by specimen 160, pl. 2) is similar to that of large masses mapped separately and described later; it will not be described here.

ANDESITE

All the 11 andesites represented by thin sections are porphyritic, with phenocrysts, commonly 0.5 to 1.5 mm long, averaging 30 percent. Of the phenocrysts, two-thirds, or 20 percent of the rocks, are zoned plagioclase, ranging from oligoclase-andesine to sodic labradorite. Locally, feldspar phenocrysts are clumped in crude "sunburst" patterns.

The remaining phenocrysts, 10 percent of the rock, are mafic minerals. Half are colorless to very pale green clinopyroxene, probably diopsidic augite. The clinopyroxene is generally in long prisms, randomly scattered, but in places it occurs as spheroidal clumps of stubby crystals accompanied by much magnetite. The other half is pale-green orthopyroxene (probably hypersthene), greenish-brown hornblende and brown biotite, in varying proportions.

Plagioclase microlites with weak fluidal texture comprise about half of the groundmass. The microlites appear to be more sodic than the phenocrysts but are too altered for sure determination. Alkali feldspar, about a quarter of the groundmass, is mostly anhedral, filling many interstices; less commonly, it assimilates plagioclase and encloses other minerals. Rarely, alkali feldspar appears in well-formed crystals showing Carlsbad twins. Mafic minerals, a quarter of the groundmass, are anhedral and indeterminate. Accessory minerals are apatite, magnetite, and, rarely, sphene.

All the rocks are somewhat altered, with sericite, clay, calcite, epidote, chlorite, and secondary magnetite present in amounts approaching 5 percent. Calcite commonly replaces pyroxene, amphibole, and the cores of plagioclase phenocrysts; rarely, it fills cracks. Clay, sericite, and epidote are secondary after both phenocrystic and groundmass feldspar. Chlorite and magnetite marginally replace the mafic minerals. Small clear spherical masses of low birefringent material are probably chalcedonic silica but possibly a zeolite.

PORPHYRITIC PERLITE

In thin section (no. 16, pl. 2), the porphyritic perlite is vividly perlitic with contorted fluidal structure. It is in part porphyritic and in part fragmental. Phenocrysts, generally 0.5 to 1.5 mm long, comprise 15 percent of the rock. Half are fresh plagioclase, zoned from andesine-labradorite to mid-andesine; locally, they are embayed by glass. Other phenocrysts are biotite and clinopyroxene, probably augite, each forming about 2 percent. Magnetite, in large euhedra, is nearly as abundant as the mafic minerals; large crystals of apatite are sporadically present.

The bulk of the rock, 70 percent, is light-brown completely isotropic glass with many perlitic cracks. The refractive index is near 1.512.

Fragments floating in the glass amount to 15 percent of the rock. Many are broken bits of plagioclase of the same composition as the phenocrysts. Others are chunks of glass that differ somewhat from the groundmass: some are darker brown with contorted banding and fillings of a zeolite or silica mineral in perlitic cracks and small vesicles; others are light brown, with crowded alined plagioclase microlites.

A chemical analysis (no. 16, table 9) is not as revealing as it might be, for no attempt was made to separate the fragments or phenocrysts from the glass. Nevertheless, the rock, with due allowance for its water content, is clearly of a composition close to Nockolds' (1954, p. 1014) average rhyodacite.

CORRELATION AND AGE

The volcanic works in the northwest corner of the quadrangle are continuous with the very large body of similar rocks mapped and described by Klepper and his associates (1957, p. 31–41) on the east flank of the Elkhorn Mountains and named by them Elkhorn Mountains volcanics. Klepper (oral communication, August 1957) believes that the volcanic rocks in the Three Forks quadrangle are in the basal part of the formation. The volcanic rocks west of Jefferson School are correlated with them, despite the considerable distance separating the two, not only because of their general petrographic and chemical similarity but also because they are similarly deformed.

On Peale's geologic map (1896) neither of the two areas involved is shown as containing mappable igneous rocks. The mass in the northwest corner is shown as sedimentary rocks of the Colorado and Montana formations. That west of Jefferson School is shown as Bozeman lake beds, but similar rocks on strike to the west are shown as part of the Livingston formation, described as conglomerate, sandstone, and andesitic tuff.

Berry (1943, p. 22–23, pl. 1) assigned both bodies to the Livingston formation but, anticipating Klepper's work, noted that they are mainly andesitic lava flows and "coarse conglomerates derived from them" and probably "should be correlated with similar rocks south of Helena," rocks that are part of the Elkhorn Mountains volcanic field of Klepper.

Klepper (1957, p. 38–40) decided against using the term "Livingston formation" because the Livingston is by definition a water-laid sedimentary unit whereas the volcanics, more than 10,000 feet thick, in the Elkhorn Mountains field contain little sedimentary material and therefore need a separate name, even though the two sequences are at least partial age equivalents. A distinction between them is especially desirable because much of the Livingston formation is derived from erosion of the Elkhorn Mountains volcanics.

The textures, structures, and mineralogy of the Elkhorn Mountains rocks are much like those of the very Late Cretaceous trachybasalts from the Adel Mountains about 80 miles north of Three Forks and midway between Helena and Great Falls (Lyons, 1944, especially p. 460–461). Especially striking is the association in both sequences of alkali feldspar (which Lyons designates simply as orthoclase) and calcic plagioclase. The Adel Mountain rocks differ, however, in having much more calcic plagioclase (calcic labradorite) and a little modal olivine.

The Elkhorn Mountains volcanics are older than the Climbing Arrow formation and are presumably younger than the Kootenai formation, which is the youngest exposed pre-Tertiary formation that lacks volcanic or igneous components. In the neighboring Elkhorn Mountains, Klepper and others (1957, p. 37–38) have been able to fix the age more closely. They cite varied, if scanty, faunal and floral evidence that the volcanics are wholly of Late Cretaceous age; their range cannot be delimited more closely than very late Niobrara or Telegraph Creek to Judith River or younger.

The porphyritic perlite, if a flow, welded tuff, or penecontemporaneous sill and thus also of Late Cretaceous age, is among the oldest glassy rocks on record. Obsidians and related siliceous rocks older than Miocene are rarely reported. There is only one other occurrence of glassy rocks as old as Late Cretaceous known to me: it happens to be in this general region, near Wolf Creek, 35 miles north of Helena (Barksdale, 1951, called to my attention by Prof. W. H. Mathews).

INTRUSIVE ROCKS

Intrusive igneous rocks that are contemporaneous with or younger than the Elkhorn Mountains volcanics and the regional folding, but older than the Bozeman group, occupy 8 square miles. They are divided into four compositional types: quartz monzonite and related, rather coarse-grained, rocks; latite; dacite; and andesite. The monzonitic and dacitic rocks tend to form large and thick masses: plutons or laccoliths. The latite and andesite appear both in extensive thin sills and in large thick bodies.

QUARTZ MONZONITE AND RELATED ROCKS

Monzonitic rocks are the most abundant intrusive rocks in the quadrangle. Most of them are light-colored equigranular medium- to coarse-grained quartz monzonite, but they grade into and include substantial volumes of monzonite, quartz diorite, and diorite. Mapped with them is a rich assortment, but small volume, of spatially associated textural and compositional variants. These rocks are largely confined to a single mass—the 10N pluton—at the north edge of the quadrangle, but include a large sill in the Willow Creek sector, and a few unmappably small sills east of Milligan Creek.

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10N PLUTON

Much the largest intrusive mass in the quadrangle is a quartz monzonitic pluton at the north edge of the quadrangle. Within the quadrangle it extends for more than 3 miles east-west, between Mud Spring Gulch and Dunbar Creek, and crops out for more than 2 miles north-south. It is exposed for at least 2 more miles into the Radersburg quadrangle and may extend several miles farther north and south beneath younger deposits. It is here named 10N pluton, for Highway 10N which cuts through its center. Merrill (1895, p. 663) described this body as diorite porphyrite. Berry (1943, p. 23) described it as "a small stock with average composition of a quartz monzonite" and noted that it includes monzonite, quartz monzonite porphyry, and, in the western part, quartz latite porphyry.

The pluton is made up of a wide variety of rock types but is dominated by light-colored pinkish quartz monzonite. Weathering usually darkens the rock only slightly, to pale grayish brown or rarely to moderate brown, and produces rolling terrain with scattered low rounded outcrops protruding from light-colored sandy soil. Exposures are too poor to reveal the shapes of any but the grossest petrographic units or to show whether contacts between rock types are sharp or gradational and thus whether it is a compound or composite body.

Within the pluton three subunits, based primarily on texture, have been roughly delineated: two eastern masses of roughly equigranular quartz monzonite rocks, one medium grained and one coarse grained; and a western mass of mixed rocks, slightly more mafic than the quartz monzonite and characterized by porphyritic texture.

Considering the size and siliceous composition of the pluton, it contains amazingly few xenoliths even near intrusive contacts, and has had little effect on the intruded rocks beyond bleaching and local recrystallization of some of the carbonates, effects discussed earlier in connection with the sedimentary rocks. One large xenolith or pendant of green thin-bedded medium-grained sandstone, most resembling Kootenai or Morrison rocks but conceivably from one of several other older units, is exposed in a small valley in the southeast corner of sec. 28, T. 3 N., R. 1 E.

The pluton is largely intrusive into the Madison group, where its original contacts are not concealed by Cenozoic cover; at its eastern edge, however, the pluton or its apophyses intrude the Jefferson and Three Forks formations.

To judge by the attitude of the bordering limestone, the pluton is a steep-sided but rather flat-topped structure that is roughly concordant on its upper surface but discordant on its sides. With little information in the vertical plane, it can only be guessed whether the pluton is stocklike, as it casually appears, or sheetlike, roughly parallel to bedding. On the weak ground that other intrusive rocks of the suite are sills, sheetlike structure seems slightly favored; the structure is so drawn in section *D–D′* (pl. 4).[17]

It might be argued that an intrusive structure of this form, intruded into a sedimentary column evidently no thicker than a few thousand feet (even if duplicated by deformation), is not well termed a pluton. In practice, however, pluton is often used for coarse-grained masses of large area, the third dimension being largely conjectural except for the upper few tens or hundreds of feet known due to erosion or subsurface exploration. Less committal than the terms stock or sill, it seems useful in just such equivocal situations as the present one. As so little is known about the depths at which large bodies of coarse-grained igneous rocks crystallize, the use of the name pluton surely does not necessarily imply much in the way of depth.

THICK SILL WEST OF SPRING CREEK

At the opposite corner of the quadrangle from the 10N pluton is another large monzonitic mass. There, a large sill, 2 miles long and as much as 200 feet thick, intrudes the basal Wolsey shale west of Spring Creek. The sill is mostly of monzonite, but locally is quartz monzonite. It resembles the main medium-grained rocks of the 10N pluton in color, texture, and visible mineralogy. The western part of the sill is intruded by several thin dikes of latite.

For much of its length the sill displaces essentially all the shale strata in the lower part of the Wolsey, between the Flathead sandstone and the sandstone member of the Wolsey. The sill steadily thickens from west to east, until it abruptly ends, but it is uncertain whether this reflects forcible spreading in the eastern part or permissive displacement of a shale interval that steadily thickens eastward. Permissive emplacement is favored by the fact that the sill continues without marked change in thickness eastward from the pinchout of the sandstone member, so the body is not regarded as a laccolith; that is, a concordant injected mass that lifted its roof by arching.

[17] Preliminary study of gravimetric and magnetic surveys of the pluton made in 1959 (Isidore Zietz, written communication, January 1960) indicate that at least the upper surface of the body plunges to the southeast, and suggest that the pluton may well be bottomed at a depth of several thousand feet.

Berry (1943, p. 23 and pl. 1) mapped this sill as syenite.

PETROGRAPHY

By H. Frank Barnett

Though separated on the geologic map, the medium- and coarse-grained nonporphyritic quartz monzonite (represented by six thin sections in plate 2 and by chemical analyses 216, 224 on table 9) are described together because they differ notably only in grain size. Grains in the medium variety are mostly in the range 0.5 to 1.5 mm. In the coarse variety grains are nearly twice as large.

The typical quartz monzonite is hypidiomorphic granular with faint but definite parallelism of plagioclase and mafic laths. Some is vaguely porphyritic. The plagioclase is zoned from sodic labradorite in the cores to mid-andesine, and makes up 30 to 50 percent of the rock. Quartz, 5 to 10 percent, is interstitial to the plagioclase, is anhedral, and is usually slightly strained. Fine-mosaic aggregates of quartz are developed along many grain boundaries and fractures. Twinned potash-rich alkali feldspar, comprising 30 to 50 percent of the rock, replaces plagioclase and is poikilitic in mafic minerals. The fine-mosaic quartz seems to have overlapping relations with alkali feldspar, here veining it, there apparently assimilated by it.

Green hornblende is the most abundant mafic mineral, averaging 5 percent. It commonly is fractured, and healed with mosaic quartz. Brown biotite and pale-green augite are about half as abundant. Biotite appears as plates and shreds generally within hornblende laths. Augite is widely replaced by hornblende. Accessory minerals include abundant apatite, lesser amounts of magnetite and sphene, and, rarely, zircon. Chlorite locally replaces biotite along cleavages. A little clay, sericite, and iron oxides appear in weathered zones.

Other rock types in the nonporphyritic part of the pluton have similar textures and minerals but the proportions of minerals differ. Two rocks (specimens 272 and 222, pl. 2) in the northern part of the mass, classed as granodiorites, have 20 percent subhedral alkali feldspar and 60 percent plagioclase. The outlier at the Copper City workings (specimen 269) has little quartz and is called monzonite. A thin apophysis (specimen 230) in limestone a quarter of a mile north of the workings contains strongly zoned augite as its only mafic mineral, lacks quartz and has less than 15 percent alkali feldspar; it is called diorite.

In the monzonitic sill west of Spring Creek (represented by specimens 298, 304, and 293, pl. 2) the dominant mafic mineral is biotite, deeply chloritized, rather than hornblende; the plagioclase grains commonly show little zoning; the alkali feldspar is in distinct anhedral grains rather than irregular ramifications; and quartz is rare.

The rocks in the western subunit of the 10N pluton, represented by six thin sections on plate 2 and by one chemical analysis (211) in table 9, range even more widely than indicated by the thin sections, no two of which are assigned the same name. In addition to the coarse-grained diorite of 211, monzonite, quartz diorite, and granodiorite, the group of thin sections includes fine-grained quartz latite and andesite. Present, too, are outcrops, not represented by thin sections, of granite aplite or quartz monzonite aplite, of alaskite, and of hornblende diorite porphyry.

In this perplexing mixture by far the most common types seem to be diorite and quartz diorite that are allied in containing persistent but subordinate amounts (3 to 10 percent) of large hornblende phenocrysts and very small amounts of alkali feldspar, in contrast to the plentiful alkali feldspar of the nonporphyritic phases. Their general mineralogy is similar to that of the nonporphyritic phases.

AGE

The youngest rocks intruded by the main mass of the 10N pluton are Mission Canyon limestone of early Carboniferous age. The oldest rocks in sedimentary contact with the pluton are those of the Dunbar Creek formation of Oligocene age. The range can be narrowed greatly, however, if the small mass of porphyritic diorite in SE¼SW¼ sec. 20, T. 3 N., R. 1 E., is accepted as part of the pluton, for it intrudes and alters rocks as young as the Lower Cretaceous Kootenai formation. The age of the pluton, and presumably of the monzonitic sill west of Spring Creek also, thus is post-Kootenai and pre-Dunbar Creek.

The range can be narrowed still further if the monzonitic rocks can be dated with respect to regional folding, which began slightly before but ended after the Elkhorn Mountains volcanics were erupted in late Late Cretaceous time. Dating the pluton in this way is difficult. On the one hand, it may be viewed as of prefolding age, with its steep western side and faulted eastern side giving evidence of crumpling and failure of the less competent Paleozoic strata against the monolithic pluton during compression. On the other hand, the main body of the pluton shows no signs of response to major compression, such as shearing or the development of zones of retrograde metamorphism, and this suggests postorogenic age.

The field relations of the monzonitic sill west of Spring Creek, however, strongly favor postfolding or intrafolding age. The sill persists for 2 miles across a gentle syncline and along the low-dipping west limb of an asymmetric anticline; in fact, it thickens toward the anticlinal axis. But it ends abruptly where the steeply dipping east limb begins. Not only does the monzonite fail to continue on the east limb, but that limb holds no sills of any kind, though latite sills are abundant in the low-dipping strata to east and west. The monzonite and the other sills as well seem to have been emplaced after asymmetric folding had progressed far. The low-dipping segments were intruded because these beds were easy to spread apart, while the steep-dipping segments acted as shoulders and remained free of sills. Intrusion might have come long after folding ceased, but it seems more likely that folding, differential separation of beds, and intrusion formed a continuous process. By this reasoning, the monzonitic intrusive rocks, assuming they are of a single age, are most likely post-Elkhorn Mountains and pre-Bozeman group; that is, latest Cretaceous to middle Eocene.

The rocks of the pluton are much like those of the Boulder batholith, 25 miles to the west; similar rocks, in similar settings, are widespread in the intervening area (Klepper and others, 1957, p. 48–52). Probably all these monzonitic and related calcalkaline intrusive rocks are of about the same age. Most modern workers (Klepper and others, 1957, p. 44, 60; Chapman and others, 1955, p. 607) have recognized that the possible age range of these rocks, based on stratigraphic evidence, is post-Niobrara (middle Late Cretaceous) to pre-early Oligocene. Evidence from this study suggests a somewhat earlier upper limit, for detritus from the batholith is a major component of the Eocene and Oligocene Climbing Arrow formation, and supports the conclusion of Klepper and others (p. 60) that "a very late Cretaceous or Paleocene date is most probable."

A still narrower range is suggested by recent age determinations of Boulder batholith rocks by the Larsen method (Chapman and others, 1955). Five determinations ranging between 61 and 72 million years, and averaging 68 million years, suggest emplacement "at or near the close of the Cretaceous" (Chapman and others, 1955, p. 609). The utility of the Larsen method is still, however, in debate.

LATITE

Latite, though of small volume, is the most widespread intrusive rock type. Thin sills of it occur in each of the areas dominated by other intrusive rocks, as well as in the Elkhorn Mountains volcanics, but only five bodies of intrusive latite are large enough to map. All the latite is medium gray to medium dark gray, weathering to steely-purplish tones like

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the Elkhorn Mountains volcanics, with sparse and small phenocrysts of feldspar and pyroxene in a dense, nonvesicular holocrystalline matrix. The forms and settings of the bodies, however, range widely. Two are pluglike masses, surrounded by Elkhorn Mountains volcanics; two are thick sills in the Wolsey and basal Meagher; and one consists of many closely spaced thin sills interleaved with basal strata of the Meagher.

PLUGLIKE BODIES

The largest body of latite is at the west edge, north of Highway 10S; it centers in NW¼ sec. 21, T. 1 N., R. 1 W. Exposed for a length of a mile, it is no doubt much larger beneath gravel cover to the south and the Jefferson Canyon thrust to the north. It is represented by specimen 12 on plate 2, and chemical analysis 12 in table 9. In places, the rock is coarse enough to be called monzonite, but most of it is not greatly coarser than the andesite and latite volcanic rocks to east and west. Its general field relations suggest that it is intrusive into the Elkhorn Mountains volcanics, but there is no positive evidence of intrusion.

The other pluglike mass, much smaller, is north of Cowan Spring in SW¼ sec. 33, T. 3 N., R. 1 W.; it is represented by specimen 168 on plate 2. Fine-grained throughout, it is separated with difficulty from the surrounding andesitic volcanic rocks, from which it visibly differs only in lacking signs of stratification and in being faintly lighter in color. Its intrusive origin, too, is not demonstrated but must be inferred from its rudely circular outline and seeming transgressive relations to the volcanic rocks.

THICK SILLS

Two thick sills of latite appear between Willow Creek and Spring Creek, in sec. 11 and 12, T. 1 S., R. 1 W. The larger, traced for nearly a mile in SE¼ sec. 11 and SW¼ sec. 12, is more than 300 feet thick; it is represented by specimen 300 on plate 2. The smaller, represented by specimen 301, is more than 50 feet thick and 1,700 feet long. Neither is well exposed.

The larger mass may be a laccolith. It not only displaces virtually all the shale between the top of the Flathead sandstone and the base of the quartzitic sandstone member of the Wolsey, but it seems to have increased greatly the apparent stratigraphic separation between the two sandstones. Scattered outcrops suggest that the intrusive is fairly homogeneous. It differs from the other latites in having many miarolitic cavities. These are lined with small quartz crystals that in turn are thinly coated with sooty black hematite(?).

The smaller sill seems to have wedged its way into the contact between Wolsey shale and Meagher limestone. The rock has a much higher proportion of feldspar phenocrysts than the other latites; in places it approaches latite porphyry.

THIN SILLS

Layers of latite, generally 1 to 4 feet thick, are abundant in the southwest corner of the quadrangle; they are represented by specimens 296, 306, and 289 on plate 2. Several appear in the western part of the thick monzonite sill that displaces the basal Wolsey, others are scattered in the adjacent remaining shale strata, and one is in gneiss (west bank of most southeasterly gulch in sec. 15, T. 1 S., R. 1 W.). In the monzonite, the latite bodies must be called dikes though they are rudely parallel to the outer walls. It was apparently one of these which Merrill (1895, p. 664) described as "brownish coarsely porphyritic andesite * * * extruded through the porphyrite sill" (here, monzonite). In the shale, gneiss and limestone, the latite bodies are sills.

The only latite in this area that is mappable, however, is a swarm of at least 15 latite layers more than a mile long in the basal 100 feet of the Meagher limestone. The silled zone is well exposed along the creek that flows west across the N½ sec. 9, T. 1 S., R. 1 W. The sills have thin glassy selvages and are separated by a few inches to several feet of slaty black limestone. Surprisingly, these blackened slaty beds can be traced along the strike into typical banded blue-and-gold beds.

PETROGRAPHY

By H. Frank Barnett

The latite varies considerably in texture, from rather coarsely fluidal to vitrophyric, and from poorly to richly porphyritic. The bulk mineral composition is fairly constant, however, and most of the variant textures seem to represent quantitatively insignificant phases. The main rock type has a fluidal texture with sparse phenocrysts, less than 5 percent, of mafic minerals. Most are stubby prisms of clinopyroxene, probably augite but too extensively altered for clear determination; the rest are green hornblende, in part as kelyphitic rims on pyroxene; brown biotite; and, rarely, phlogopite.

Microlites of plagioclase, ranging from calcic oligoclase to andesine, make up as much as 80 percent of the rock. Many grains are thinly rimmed by alkali feldspar, probably albite; late-formed alkali feldspar also occupies many interstices, but rarely totals as much as 5 percent. Quartz is rare, occurring only as rounded xenocrysts and in the lining of a few cavities. Subhedral clinopyroxene, hornblende and biotite appear in the groundmass to the extent of perhaps 10 percent. Apatite and fine-grained magnetite are persistent accessories; sphene is rare. The mafic minerals are extensively altered to chlorite and coarse-grained magnetite. Other common alteration products are calcite, sericite, limonite, and clay.

In a rather common variant, the rock has similar texture and groundmass but the phenocrysts are largely feldspar in the andesine range, and many have thin rims of clear albite.

AGE

Assuming that all the latite is of a single age, the unit is no older than very late Cretaceous, for it apparently cuts Elkhorn Mountains volcanics. It may, of course have been intruded while the Elkhorn Mountains volcanics were being erupted. In the Elkhorn Mountains there are many sills of quartz-poor intermediate rocks that were emplaced about at the same time as the volcanic rocks (Klepper and others, 1957, p. 44–45); such rocks, however, are largely diorite porphyry and andesite porphyry, little like the slightly porphyritic latite of the Three Forks quadrangle. Its upper age limit can with certainty be placed no lower than Quaternary, as latite is nowhere demonstrably overlain by Tertiary rocks. Latite, however, intrudes and is younger than the monzonitic sill west of Spring Creek. Like the monzonitic sill, the latite is probably younger than the folding but not greatly so.

DACITE

Dacite is mostly confined to a single large mass, the Buttleman laccolith, cut by Willow Creek. Unmappably small sills of dacite also appear in the large andesite mass above Milligan Canyon.

BUTTLEMAN LACCOLITH

The Buttleman laccolith, flanking Willow Creek near the south edge of the quadrangle, is more than 4,000 feet thick and 3 miles long. It is named for the Buttleman Ranch on which it crops out. Peale (1896, p. 4) and Merrill (1895, p. 664) classed the rock as hornblende porphyrite; Berry (1943, p. 23 and pl. 1) described the mass as "an intrusion of quartz latite and quartz latite porphyry."

The great bulk of the laccolith is dacite, represented by six thin sections on plate 2 and by chemical analyses 334 and 339 in table 9, but at its east edge it is somewhat more siliceous and alkalic, grading into quartz latite (represented by specimens 340, 341, pl. 2).

The laccolithic rocks are light colored, mostly medium light gray. Weathering generally bleaches them even lighter but locally they have yellowish-brown or yellowish-green splotches due to oxidation of iron-bearing minerals. They are uniformly fine grained and porphyritic, with many stubby prisms of feldspar, scattered needles of hornblende, and flakes of biotite.

The base of the laccolith is the Flathead sandstone throughout; its top has varying stratigraphic position, but rises no higher than the base of the Meagher limestone. At its western limit the laccolith has simply displaced the lower shales of the Wolsey between the Flathead and the quartzitic sandstone member of the lower part of the Wolsey, in the fashion of the sills farther west. The apparent stratigraphic interval between the two sandstones increases rapidly eastward, presumably due to forceful widening by the intrusion. Near the west wall of Willow Creek, the sill thickens abruptly where the sandstone member of the Wolsey seems to thin out, cutting across a thick interval of shale to bottom against another thick sandstone (or possibly the same one offset by an unrecognized pre-intrusion fault). East of the creek the outcrop of the laccolith widens tremendously to 1⅔ miles, but apparently the only rocks displaced are those of the Wolsey, for the northeast edge of the laccolith is near the base of the Meagher limestone. A thick mass of Wolsey remains as a giant xenolith or pendant within the intrusive east of Willow Creek (mostly in NW cor. sec. 8, T. 1 S., R. 1 E.).

Despite its size and siliceous composition the laccolith has had essentially no effect on its wall rocks. Where shales are preserved near contacts they are slightly hardened and reddened, but the bordering sandstones and limestones are not visibly changed. Xenoliths are few. The missing shale seems to have been bodily pushed aside by the intruding dacite, but not to have interacted with it.

PETROGRAPHY

By H. Frank Barnett

In thin section, the typical dacite is holocrystalline porphyritic, with 10 to 30 percent of small phenocrysts in a microcrystalline, distinctly fluidal groundmass. Of the phenocrysts, 95 percent are zoned plagioclase in the andesine range. Remaining phenocrysts are mafic minerals, largely needlelike prisms of green hornblende but including small amounts of biotite and traces of clinopyroxene. The mafic minerals are much altered, the hornblende to magnetite, sphene, and calcite; the biotite to chlorite. Some feldspar phenocrysts also are replaced by calcite.

The groundmass consists mostly—as much as 80 percent—of feldspar. In some places, the feldspar is wholly plagioclase, in the andesine range, in dense felted aggregates of microlites; in a few places it is late-formed alkali feldspar (no more calcic than albite-oligoclase) in cloudy anhedral masses; most commonly, both types appear, in varying proportions. The only other prominent mineral in the groundmass is quartz, in both anhedral interstitial grains and tiny well-formed crystals, averaging 10 percent.

Apatite, magnetite, and sphene are the common accessories. Calcite, in addition to replacing many phenocrysts, also replaces groundmass feldspar and quartz. Sericite, clay, and limonite are sparingly developed along fractures.

AGE

On direct stratigraphic evidence, the age limits of the Buttleman dacite are broad indeed. It is younger than the Wolsey shale but older than Quaternary gravel. Yellow sands assigned to the Climbing Arrow formation contain dacite pebbles but these strata are not firmly established as early Tertiary; they may be Quaternary.

A reasonable assumption is that the Buttleman dacite is temporally, if not genetically, related to the latite sills on strike with it, and shares their intra- or post-folding age. If so, the dacite is of late Late Cretaceous to pre-early or middle Eocene age.

ANDESITE

Within a radius of 2 miles of Milligan Canyon are a number of intrusive bodies mapped together in view of their generally andesitic composition and geographic position, though they are by no means similar in all other respects. They fall naturally into two subgroups: one, south of the canyon, made up mostly of thin andesitic sills in the Three Forks shale; the other, north of the canyon, consisting of large masses of andesite intruded into late Mesozoic rocks.

Berry (1943, p. 23 and pl. 1) designated the sills south of the canyon as syenite. Those north of the canyon were mapped as porphyrite by Peale (1896) and as andesite by Berry (1943).

SILLS SOUTH OF MILLIGAN CANYON

Many thin sills of andesite appear in a narrow stratigraphic interval east of Milligan Creek below Milligan Canyon. Where unaltered, the andesite is medium grayish green, fine grained, and very dense, with scattered needles and flakes of mafic minerals and many small feldspar phenocrysts. In most places, however, particularly near the creek, the andesite is deeply altered and has a thick porous dusky-yellow to olive crust of secondary minerals.

The andesite sills, represented by five specimens on plate 2, are nearly all intruded into the Three Forks shale; one is in the overlying Lodgepole limestone. They have an interesting relation to the host rocks and the pair of northward-trending faults that offset them. West of the faults, only a single, seemingly discontinuous sill, rarely as much as 30 feet thick, appears in the shale, just below the orange siltstone unit. Between the faults there is also only one sill; it, however, is continuous and thicker, averaging perhaps 40 feet. Just east of the faults, the upper half of the Three Forks contains at least three thick sills, and sills are progressively more numerous eastward, so that where the Three Forks passes under Quaternary gravel in center sec. 31, T. 2 N., R. 1 E., the shaly parts of the formation are ribbed with sills, totaling at least 31 in number and averaging 3 feet thick, with somewhat thicker septa of baked, reddened, hornfelsed, and locally slaty shale between. None of the Three Forks is missing; rather, the interval between the Jefferson and the Lodgepole formations has spread to accommodate the intrusive rocks. In this segment, too, there is a sill as much as 8 feet thick and at least 200 feet long in the Lodgepole limestone, 25 feet above the base.

Where the Three Forks interval emerges north of the gravel patch its character is much different: the Three Forks, all present, has been neatly split by a single concordant intrusive 300 feet thick and at least 3,500 feet long. This intrusive seems properly called a laccolith. It is recognizable for miles from the north and northeast because it forms a distinctive rounded ridge just below a huge letter T (for Three Forks High School) on the limestone mountain above. The main mass of the laccolith (represented by specimen 77 on pl. 2) is fine-grained diorite that differs from the andesite sills farther west only in texture; the borders are andesite (represented by specimen 69 on pl. 2).

Plainly, the sills are not offset by the faults, as they casually seem to be, but are controlled by them and therefore are younger than the faults.

INTRUSIVE ROCKS NORTH OF MILLIGAN CANYON

Above Milligan Canyon are two roughly elliptical masses of fine-grained medium-gray porphyritic andesite: a large one east of the creek, centering in N½ sec. 24, T. 2 N., R. 1 W., and a much smaller one west of the creek, straddling the line between secs. 22 and 27. The typical andesite contains many small phenocrysts of both feldspar and mafic minerals, only slightly larger than the groundmass, that give the rock a salt-and-pepper granularity. Weathering produces pitted dusky-yellow to yellowish-brown surfaces. Locally, the rock is more coarsely porphyritic latite, and in a few places it is so rich in mafic phenocrysts as to be classed as latite porphyry (represented by specimens 126, 134, 147 on pl. 2). A little dacite (represented by specimen 281 on pl. 2) occurs in the western mass. Exposures are too poor to tell whether the latite and dacite are gradational variants or dikes.

Lacking any signs of extrusive origin such as layering, oxidized surfaces or vesicularity, these masses are hesitantly interpreted to be shallow-seated intrusive rocks. Their only exposed contact with older rocks is transgressive, the southern border of the large mass east of Milligan Creek cutting at a small angle across bedding so that it is in contact with varying levels in the Kootenai formation in most places and for a short distance cuts across the Morrison formation. The Mesozoic strata are hardened and reddened for a few tens of feet from the contact. No apophyses crop out in the underlying strata and there are few sedimentary xenoliths in the andesite. All other margins of the large intrusive and the entire perimeter of the small one west of the creek are blanketed by Cenozoic sediments, the oldest of which are low in the Climbing Arrow formation. The two masses are probably plugs that remained as hills in the Tertiary basin after the less resistant sedimentary deposits, which they intruded, had been stripped.

PETROGRAPHY

By H. Frank Barnett

The andesite south of the canyon is described separately from that to the north, to bring out some differences between them.

The andesite south of the canyon (represented by five specimens on pl. 2) has a porphyritic faintly fluidal texture, generally veiled by alteration. The phenocryst content ranges widely but averages 30 percent. At least 80 percent of the phenocrysts are small laths, no longer than 1.5 millimeters, of plagioclase ranging from calcic oligoclase to calcic andesine. Some display oscillatory zoning and most have distinct rims of alkalic feldspar, probably albite. The remaining phenocrysts, aggregating less than 10 percent of the total rock, are yellow-brown biotite, pale-green hornblende, and colorless augitic clinopyroxene, in order of decreasing abundance. Present throughout are fresh euhedra of magnetite and apatite, mostly in association with mafic minerals or their altered remains.

The original groundmass was presumably largely feldspar; now it is mostly saussuritic material, of which some is sufficiently crystallized to yield determinable grains of clinozoisite and calcite. Recognizable in the groundmass are minute euhedra of untwinned alkali feldspar, probably albitic, that locally total as much as 15 percent of the rock, and a little interstitial quartz.

Plagioclase phenocrysts are extensively replaced by calcite and clinozoisite, and the mafic minerals are replaced by calcite and chlorite, in many places pseudomorphously. In the western sills, calcite and subordinate clinozoisite total as much as 65 percent of the rock. Elsewhere, these alteration products, and chlorite, rarely aggregate 20 percent, and clinozoisite is more abundant than calcite. The widespread development of well-crystallized clinozoisite and the abundance of unaltered magnetite combine to suggest that much of the alteration is deuteric rather than related to weathering. Ascribable to weathering, however, are widespread limonite and variable amounts of a cryptocrystalline yellowish brown interstitial material.

The northern andesite (represented by 7 specimens on pl. 2, and analysis 140 in table 9) averages 20 percent of small phenocrysts in a fluidal groundmass. Plagioclase, zoned normally from andesine-labradorite to sodic andesine, makes up 80 percent of the phenocrysts. Thin clear outer rims on many grains are even more sodic and may approach albite. The remainder of the phenocrysts is mostly colorless to yellowish-green augite in well-formed but usually altered crystals. Some specimens contain brownish-green hornblende and others reddish to yellowish-brown biotite. A few contain scattered grains of pale-green orthopyroxene.

The groundmass seems initially to have been almost wholly composed of microlites of plagioclase with mere traces of mafic minerals, but it is saussuritized throughout so that its fresh composition is conjectural. Alkali feldspar, much less altered, appears sporadically in the groundmass as roundish crystals and ramifying irregular masses. Magnetite and apatite are abundant accessories, commonly reaching 2 percent each.

The augite crystals are extensively replaced by mixtures of magnetite, hornblende, and chlorite. Hornblende and biotite are also somewhat altered to magnetite and chlorite. In places, the groundmass is recrystallized to determinable clinozoisite and epidote. Sericite, clay, and calcite are widely distributed and many fractures are filled with zeolites (mostly stilbite) or chalcedonic silica.

The northern andesite is distinctly more mafic than the southern, as the plagioclase is much more calcic, and augite is the main ferromagnesian mineral rather than a subordinate one. In addition, the southern andesite is characterized by intense deuteric alteration, whereas the northern andesite is but mildly modified.

AGE

On stratigraphic grounds only, the andesite sills south of Milligan Canyon can be dated no more closely than as younger than the Lodgepole limestone, which they locally intrude, and older than overlying Quaternary gravel. The andesite masses north of the canyon can be more closely dated: they are younger than the Early Cretaceous Kootenai formation which they intrude or unconformably overlie and older than the Climbing Arrow formation, which overlies them unconformably and to which they contribute detritus. Their petrographic similarity to the Elkhorn Mountains volcanics leads readily to the idea that some of the andesite bodies may have been intruded at the same time as the volcanic rocks were extruded.

Structural evidence permits somewhat narrower age assignment. The relation of the southern sills to the faulting in the Lodgepole and Three Forks vigorously suggests that these sills are younger than the faults, that in turn are younger than regional folding, if but slightly. The andesites, then, seem to be no older than late Late Cretaceous or younger than middle Eocene.

SUMMARY OF AGE RELATIONS

On direct stratigraphic evidence, and assuming only that all the intrusive rocks of a particular rock type are substantially contemporaneous, the monzonites are of post-Kootenai, pre-Dunbar Creek age; the latites are of post- or intra-Elkhorn Mountains pre-Quaternary age; the dacites are of post-Wolsey, pre-Climbing Arrow(?) age; and the andesites of post-Kootenai, pre-Climbing Arrow age.

Given an additional assumption, that intrusion did not begin until regional folding began, the lower limit of the period of intrusive activity can be raised to latest Cretaceous. The upper limit, of course, is not changed by this assumption.

With a third assumption, that all the intrusive rocks, having evident close structural relations, are temporally related too, and were emplaced in essentially a single irruptive episode, their aggregate age can be further narrowed to intrafolding and pre-Climbing Arrow, or latest Cretaceous to middle Eocene.

IGNEOUS ROCKS INTRUSIVE INTO TERTIARY ROCKS

OLIVINE BASALT

Olivine basalt crops out in two small areas near the Jefferson River in sec. 34, T. 1 N., R. 1 W. Much of the largest mass is on the north bank in NW¼ sec. 34 where it forms a conical hill mantled at the base by Quaternary gravel. The mass exhibits rude vertical flow banding striking N. 50° W. It is complexly jointed; one main joint set seems concentric to the rounded outcrop shape, another tangential, but this impression was not confirmed by systematic measurement. Most of the rock is compact but vesicular structure appears locally. The homogeneous flow bands are rarely interrupted by thin and impersistent zones of flow breccia.

Across the river in center E½ sec. 34, is a remnant of basalt less than 100 feet long and no more than 20 feet thick cutting Sphinx conglomerate. (The size of this body is exaggerated on the geologic map.) The basal few feet of the basalt is pock-marked with large calcite amygdules, and many pebbles from the conglomerate are incorporated in it. Away from the contact the rock is dense and only spottily vesicular.

The basalt in both masses is similar, with scattered small phenocrysts of mafic minerals set in a compact dark-gray microcrystalline to glassy groundmass. Weathered surfaces are grayish brown to brown and pitted where the phenocrysts have weathered out. The small mass is intrusive into the limestone conglomerate. The relation of the large mass to older rocks is unknown, but it is evidently related to the small mass and presumably contemporaneous with it. Both are probably plugs.

PETROGRAPHY

By H. Frank Barnett

A single thin section represents each mass (326 and 320, pl. 2). The larger mass is also represented by chemical analysis 326 in table 9. Both rocks are much alike microscopically, with around 10 percent of small phenocrysts, mostly euhedral olivine, in a fluidal groundmass of pale-greenish-brown glass studded with tiny grains of plagioclase and clinopyroxene. A few clinopyroxene phenocrysts, probably diopsidic augite, also occur in the smaller mass. Both have scattered xenocrysts of quartz, rounded and with thick reaction rims.

Groundmass plagioclase, which makes up 30 percent of the rock, is in grains too small for accurate determination by the methods used. Its indices are well above balsam and it is probably near calcic andesine or labradorite. Groundmass olivine and clinopyroxene each contribute perhaps 5 percent. Present also is a little microcrystalline pleochroic brown material that is probably biotite. Magnetite is abundant, in small euhedra and widely disseminated dust. The rest of the rock, near 50 percent, is glass with index well below that of balsam.

The olivine in both phenocrysts and groundmass is intricately cracked and much altered to bowlingite(?). Locally, the glass is slightly devitrified. Otherwise, the rock is unaltered.

AGE

The basalt is the youngest igneous rock in the quadrangle, for it is the only one that intrudes rocks of Cenozoic age. If the intruded limestone conglomerate is correctly dated as slightly older than the middle Eocene or late Eocene Milligan Creek formation, the basalt may be only a little younger than the other intrusive rocks. It may, however, be much younger, though older than the older Quaternary gravel that mantles it. For convenience, it is placed at the oldest possible position on the geologic map explanation.

SOME PETROLOGIC RELATIONS

The petrologic succession is obscure. Most of the intrusive rocks, including intrusive andesite, seem to be younger than the extrusive andesitic volcanics of the Elkhorn Mountains, and monzonite is intruded by latite west of Willow Creek; otherwise, no sequence is indicated.

The field evidence supports the view that all the intrusive rocks are magmatic and rose to the present level of exposure in a fluid but very viscous condition, with little superheat. Only the highly varied rocks in the western part of the 10N pluton suggest hybridization; if hybridization occurred, it must have been at far deeper levels than are now exposed, for the visible signs are of only trivial interaction between magma and wall rocks.

The inference is strong that all the igneous rocks are genetically related. The most direct, and perhaps the only convincing, evidence is the simple fact that the main rock types, though each is dominant in certain areas, are each widespread in the quadrangle, and were apparently emplaced in a single time span of no great geologic length. Some sort of genetic relation may, with reservations, be inferred from the close chemical relation of the igneous rocks, as shown by the shape and smoothness of the curves they yield in a silica-variation diagram (fig. 13), prepared from the analyses of table 9, recalculated free of H_2O and of $CaCO_3$, on the assumption that all the CO_2 reported is in carbonate form. (Incidentally, the fact that the porphyritic obsidian (no. 16) fits the curves may be taken as a faint confirmation of its contemporaneity with the Elkhorn Mountains volcanics; but the same sort of reasoning probably does not apply with equal force to the age of the basalt plugs (represented by no. 326), simply because basalts of similar composition are common throughout the Cenozoic column in the region.) Petrography and chemistry combine to demonstrate that all the igneous rocks involved are normal components of the calc-alkalic group (Peacock, 1931); the alkali-lime index of the analyzed group cannot be given as a single figure because the curve for $Na_2O + K_2O$ intertwines rather than intersects with the curve for CaO (see fig. 13); the silica range in the zone of intertwining is 59.5 to 61.8, which is within the range of the calc-alkalic group as defined by Peacock. To say that a group of rocks are genetically related is not to say much about their genesis—but is all that is warranted at present.

The Three Forks quadrangle is close geographically to the petrographic province of central Montana, as defined by Larsen (1940, p. 888–890). Chemically and mineralogically, however, the Three Forks igneous rocks are little like the orthoclase-rich subsilicic rocks that characterize the province, even from as close by as the northern Big Belt Mountains (Lyons, 1944). Instead of having a high potash-to-soda ratio, characteristic of all but the Crazy Mountains subprovince of the central Montana province, the Three Forks igneous rocks have potash-soda ratios close to those of average calcalkaline rocks of similar general composition, as tabulated by Nockolds (1954); that is, they have distinctly more soda than potash or, in the case of quartz monzonitic rocks (adamellites of Nockolds), slightly more potash. It should be noted that the Three Forks dacites (nos. 334, 339) are not appropriately compared with Nockolds' dacites (1954, p. 1015) from which they differ radically, but with his dellenites (p. 1014). The latites of the Three Forks quadrangle, in fact, have less potash than soda, if analysis 12 is representative, whereas Nockolds' average latite (p. 1017) has more potash. At the latitude of Three Forks, therefore, the western limit of the central Montana petrographic province is east of the Missouri River.

The Three Forks igneous rocks are, however, almost identical with certain seemingly contemporaneous rocks in the Elkhorn Mountains to the west and north. For example, virtually every chemical analysis in table 9 can be matched by analyses from the southern Elkhorn Mountains (Klepper and others, 1957, tables 1–3). As might be expected, however, from the far greater area covered in the Elkhorn Mountains report, the extrusive and intrusive rocks of that area are far more diverse.

STRUCTURE

Structural events are implicit in many of the facies changes and virtually all of the unconformities recorded in the bedded rocks, and in the locations and shapes of igneous masses. For present purposes, however, they are sufficiently noted in the discussions of the rocks they affect.

Discussed here, in chronologic order, are those structural features—folds, faults, tilted blocks—whose character and origin do not appear in the discussions of the rocks themselves. Unfortunately, only a fraction of the structures developed before deposition of Cenozoic sediments are visible. A wholly satisfactory local structural synthesis is thus out of the question, much less integration with the little-known regional structure.

Structural elements for which there is reasonably good evidence are sketched and named on plates 1 and 3. Plate 3 shows how few these are.

SUMMARY OF STRUCTURAL HISTORY

The earliest recorded structural event was the folding and metamorphism of the earlier Precambrian rocks. Much later, but still in Precambrian time, the southern limit of the embayment that received Belt sediments became established at about the latitude of Willow Creek village apparently through faulting. This discontinuity exerted a powerful if intermittent influence on subsequent deposition and deformation.

From late Precambrian to Late Cretaceous time, the quadrangle was structurally quiet, though far from dormant. Most of the folds and faults formed in response to regional compression applied recurrently during very Late Cretaceous to middle or late Eocene time. For convenience, this time interval is referred to as the Laramide interval or Laramide time, and the structural events within it as the Laramide orogeny, without any implication that deformation elsewhere assigned to the Laramide orogeny necessarily has the same time limits.

In the Laramide orogeny large-scale folding along roughly north-south axes occurred first. Major thrusting followed, perhaps after a period of quiescence. A broad belt of thrusting developed, but only the uppermost thrust is exposed in the Three Forks quadrangle. A cluster of tilted fault blocks above the thrust surface probably was formed during thrust advance. Some high-angle faults below the uppermost plate cannot be definitely dated in relation to thrusting but are likely also contemporaneous with the thrusting. Only two steep faults are known or inferred to be distinctly younger than the thrusting but older than the Tertiary basin deposits; possibly, however, many developed and controlled the general dimensions of the Tertiary basin, but if so they are masked by continental deposits. All but two of the many intrusive masses were emplaced during or shortly after folding and before deposition of the Bozeman group. Time relations between intrusions and thrusting are unknown but a reasonable guess is that the period of intrusion overlapped the period of thrusting.

The Tertiary basin is structurally controlled at least to the extent that it lies on a zone of repeated faulting, and its borders cut sharply across deformed older rocks. But there is no deciding to what extent it is due directly to diastrophism and to what extent indirectly, by erosion of unresistant rocks exposed by diastrophism. It is not, however, a product of dissection of an Eocene peneplane, as has been postulated.

Deformation during the Tertiary seems to have been mostly in the form of gentle easterly tilting, recurring or continuing from late Eocene until the close of the Tertiary, and punctuated by shallow local folding.

Quaternary deformation seems to have been limited to slight northward tilting.

STRUCTURE OF PRECAMBRIAN CRYSTALLINE ROCKS

The shape of the surface of the Precambrian crystalline mass below the younger rocks and the distribution and nature of mobile units within it have no doubt been basic elements in the structural development of the area, as Peale (1893, p. 14) early noted and as several later workers (see, for instance, Bucher, Thom, and Chamberlin, 1934) have hypothesized for the entire Cordilleran region. They must, however, be left to conjecture. All that is known of the Precambrian crystalline rocks in the quadrangle is based on a few poor exposures that do little to brighten the corner where they are. Banding and foliation strike N. 70°–90° W. and dip 50°–70° N. North-south compression older than the North Boulder formation is indicated if the crystalline rocks have acted only as buttresses during subsequent compressional deformation. Perhaps they have, but to judge by the patterns made by Paleozoic rocks folded within Precambrian crystalline masses in southwestern Montana, as shown on the current geologic map of the State (Ross and others, 1955), the Paleozoic rocks are molded smoothly around crystalline cores in neat anticlinal form, suggesting that the crystalline rocks have responded to some extent by folding. Thus, the structural trend of the small body of Precambrian crystallines here, which is parallel to that of overlying Paleozoic rocks, may be in part of post-Paleozoic age and cannot be confidently taken as the trend of pre-Belt structure.

STRUCTURAL SIGNIFICANCE OF PRECAMBRIAN SEDIMENTARY ROCKS

The North Boulder formation was deposited at a time of active tectonism, as indicated regionally by its coarse texture, great thickness, and linear distribution. The small volume of North Boulder rocks in the quadrangle and their position on the upper plate of a Laramide thrust zone of unknown but possibly great displacement make it difficult to appraise their structural significance on local evidence, but on a regional scale the meaning of the elongate wedge of coarse clastic rocks of which the North Boulder rocks are a part seems fairly clear. Peale (1893, p. 14) early interpreted the texture and distribution of these rocks to indicate an east-west Precambrian (Belt) shoreline at about the latitude of Bozeman. Most later workers agree that the shoreline probably is a fault or fault zone, downthrown on the north in pre- and intra-North Boulder time. (Deiss, 1935; Hinds, 1935; Berry, 1943, p. 7; Sloss, 1950; Alexander, 1951; McMannis, 1955, p. 1390–1392; and Verrall.[18]) The projected trace probably passes close to the village of Willow Creek, and the structure is therefore named the Willow Creek fault.

As Sloss and McMannis recognized, the influence of this ancient fault zone did not end with deposition of the North Boulder formation. In post-North Boulder pre-Flathead time, further uplift along this zone, by arching or tilting rather than faulting, is suggested by the fact that southward the Flathead overlaps lower and lower strata in the Belt series, and ultimately lies on pre-Belt crystalline rocks: in the northeastern part of Canyon Ferry quadrangle (see fig. 1B) the Flathead lies on Helena limestone, in the southwestern part on Empire shale (Mertie, Fischer, and Hobbs, 1951, pl. 1); in the northern part of the Toston quadrangle it lies on Spokane shale, farther south on Greyson shale; in the Manhattan quadrangle it rests on rocks as old as the Newland limestone.[19]

The fault zone seems to have had little effect on the deposition of Paleozoic or Mesozoic sedimentary rocks of the Three Forks area, which lie without exceptional differences in thickness or lithology on both sides of the zone. It seems, however, to have reappeared as an important influence in Late Cretaceous and early Tertiary time, as it apparently limited the Elkhorn Mountains volcanics on the south and provided the southern terminus of the great thrust system that includes the Lewis thrust. It is no coincidence that the Three Forks basin overlies this zone and is elongated parallel to it, thus being one of the very few intermontane basins in Montana that do not trend north-south.

Ironically, this important structure cannot be located closely except where it inferentially coincides with the Jefferson Canyon thrust.

LARAMIDE STRUCTURAL FEATURES

EARLY FOLDING

Folding on a large scale was the first major local event in the Laramide orogeny. The folding, presumably with concurrent erosion at the surface as sketched by Lowell (1956a), began before deposition of the Elkhorn Mountains volcanics, for beyond the quadrangle, as noted earlier, they rest on an angular unconformity representing the removal of thousands of feet of earlier Cretaceous, Mesozoic, and late Paleozoic rocks. But folding continued after the volcanic rocks were erupted, for the volcanics are folded to about the same degree as the older rocks in most places; this is true of the main mass of Elkhorn Mountains volcanics also (Klepper and others, 1957, p. 32, 55). Large-scale folding ended before thrusting began, as thrust faults that shear off several large folds are not themselves folded.

The largest fold for which there is evidence is the great crumpled Radersburg syncline or synclinorium that, plunging northward, occupies most of the northwest quarter of the quadrangle and extends across and beyond the Radersburg quadrangle to the north, between the elongate dome at the east edge of the southern Elkhorn Mountains (Klepper and others, 1957, p. 56 and pl. 2) and the northward-trending anticline whose west flank forms the Limestone Hills, a few miles southwest of Townsend (Ruppel, 1950). This fold, recognized but not named by Klepper and by Ruppel, seems appropriately called the Radersburg syncline. The horizontal distance between the trough of the syncline and the crest of each bordering anticline is 5 or 6 miles at the latitude of Radersburg and seems twice this at the latitude of Three Forks. The syncline involves the entire column of pre-Tertiary sedimentary and volcanic rocks present in the quadrangle, from the North Boulder formation to the Elkhorn Mountains volcanics. Most of the area

[18] See footnote, p. 11.
[19] See footnote, p. 11.

underlain by the fold is unconformably covered by younger deposits, and the fold structure in much of the rest has been complicated by faulting so that only its general outlines can be sketched, and it is impractical to show its axis on the geologic map. The syncline is abruptly cut off by the Jefferson Canyon and Highway thrusts at the south; the remnants of steeply dipping contorted Paleozoic and Mesozoic rocks east of Mud Spring Gulch road are all that is visible of its east flank; on the west the structure continues at least 7 miles into the Jefferson Island quadrangle. The average dip on the flank is about 30° and the plunge perhaps 10° northward, but local dips are in almost any amount and direction due to minor folds and to the influence of faults. The gross structure trends slightly east of north, as Klepper and Ruppel noted for the bordering anticlines, but the trend within the quadrangle seems to be a little west of north. Sections *B–B′* and *C–C′* illustrate the southern part of the syncline.

The pre-Tertiary sedimentary rocks south of the Jefferson River, below the thrusts that limit the Radersburg syncline, are regarded as separate structurally, though with similar general trends. Folds, though well formed in places, change trend and die out rapidly. Faulting seems to have been the main method of structural adjustment. The structure of this block is shown in section *A–A′*.

The most pronounced fold here is an asymmetric anticline. It forms the northerly bulge in the contact between Flathead sandstone and Precambrian crystalline rocks in NE¼ sec. 10, T. 1 S., R. 1 W. From the Flathead, it plunges first north and then northwest, disappearing under Quaternary gravel in NE¼ sec. 4, T. 1 S., R. 1 W. Its western limb, interrupted by a long strike fault, has low to moderate dips; its eastern limb, broken by two strike faults, is steep. A few thousand feet to the west is a poorly defined syncline of similar trend and of low amplitude. To the east, no large well defined folds are apparent.

The tight, locally overturned and broken anticline in N½ sec. 12, T. 1 S., R. 1 W., though small, is worth noting. In it are unusually good exposures of Park shale in complex relation to the Pilgrim limestone. The fold illustrates that thoroughly indurated rocks can pass from open to intensely deformed structure and back again, within very short distances.

These folds probably are related to the Radersburg syncline. Their axial trends diverge from that of the Radersburg syncline due partly to the influence of the underlying crystalline rocks and perhaps partly to refolding by thrusting, as hinted by the curving axes. Conceivably, the folds may be open in part because of "unfolding" (Kelley and Del Mar, 1957) related to thrusting.

The most spectacular fold, shown in sections *D–D′* and *E–E′*, is in the northeast corner, where all the pre-Tertiary rocks are involved in a great overturned plunging anticline, here named the Hossfeldt anticline. The lowest formation exposed in the core of the fold is the Pilgrim limestone. On the west limb, cut off by the Lombard thrust, the highest formation is the Amsden; on the east limb, the Kootenai. The structure continues eastward into the Manhattan quadrangle, as shown by Verrall,[20] ultimately involving rocks high in the Colorado shale of Late Cretaceous age. The horizontal distance from the crest of the Hossfeldt anticline to the trough of the parallel Eustis syncline of Verrall is about 6,500 feet. North-northeastward, the structure continues for at least 7 miles in the Toston quadrangle, where it disappears beneath the northern extension of the Lombard thrust. The anticline is overturned to the east, with moderate dips, mostly 30° to 40°, on the upright western limb, and steep dips, commonly 60° (that is 130°) to vertical, on the overturned limb. The axis plunges S. 30° W. at angles of 15° to 30°. Peale (1896) chose this structure for the west end of a section (*B–B′*) but did not recognize the thrust fault and was led to interpret the entire structure east of the large intrusive (10N pluton of this report) as an unfaulted syncline, overturned to the east. Haynes (1916b), in his original description of the Lombard thrust (or overthrust as he called it), recognized the fold structure to be an overturned anticline and illustrated it with a photograph (his fig. 9); his map (fig. 1) shows no rocks older than Jefferson dolomite or younger than Madison limestone on the flanks. Berry (1943, pl. 1) mapped the structure similarly and identified Pilgrim limestone in the core, but no rocks younger than Madison limestone on the flanks.

A smaller open synclinal fold, involving only rocks between the upper part of the Pilgrim limestone and the lower part of the Mission Canyon limestone, lies between the Lombard thrust and Dunbar Creek. The limbs are roughly symmetrical with average dips of about 40°, but the fold is complicated by many minor crenulations with dips ranging from 15° to vertical. It plunges S. 35° W. at about 20°. At the east edge of this syncline, against the Lombard thrust, is a small but striking asymmetric, in part overturned, anticline involving the uppermost Pilgrim, the Maywood and the basal Jefferson formations. Its axis, which can be traced for about 1,500 feet, plunges steeply southwestward. Possibly it is a large drag

[20] See footnote, p. 11.

fold and contemporaneous with thrusting rather than with the older folding. East of the Radersburg syncline, a broad shallow anticline, or dome, on the order of 3 miles broad, is the main structure in the pre-Tertiary rocks, mostly limestone of the Madison, intruded by the 10N pluton. On its west flank, where it passes into the Radersburg syncline, the structure develops steep dips and many complex minor folds. Section *D–D′* attempts to show part of this fold.

All of the folds could have been developed in a single episode of folding with roughly east-west compression. Most axial planes dip westward, suggesting a vertical couple with the western element moving relatively upward. This is part of a widely noted regional pattern, suggesting that large-scale tangential compression was the principal type of deformation, rather than local readjustments controlled by gravity.

The Madison group played a special role in folding (Berry, 1943, p. 24). It acted locally much the same as the Precambrian crystalline mass acted regionally, by serving as a buttress against which the thinly bedded, less competent overlying and underlying formations were crumpled. More distant units, both competent and incompetent, were more gently folded. North of Mud Spring Gulch the Madison is folded but elsewhere it is broken by closely spaced faults, both along and across the bedding. These faults individually are small but their aggregate displacement is great and causes large changes in apparent thickness. The tremendous apparent thickness of the Madison east of Milligan Canyon is largely due to such faulting, and in the valley on the northwest side of this thickened mass is an example of the buttress effect. There the Big Snowy and basal Amsden rocks have been intricately contorted. The folds have amplitudes up to 200 feet but are too small to show on the geologic map. The competent formations above and below the crumpled rocks maintain consistent strike and dip.

STEEP FAULTS OLDER THAN THRUSTING

Two small steep faults are overridden by a thrust, and are therefore of pre-thrust age. They are exposed under the Highway thrust in the low hills west of Highway 10S in SW cor. sec. 33, T. 2 N., R. 1 E., and in NE corner of adjoining sec. 3, T. 1 N., R. 1 E.

The fault in sec. 33 trends slightly west of north and repeats part of the Pilgrim limestone. The fault trace is marked by discordant attitudes of the strata on opposite walls. The relative movement is not known because banded blue-and-gold beds, present at several levels in the formation, appear on both walls; the throw cannot exceed a few hundred feet. Along the nearly parallel fault in sec. 3, lower Meagher limestone on the west has been dropped a few hundred feet against Pilgrim limestone on the east.

THRUST FAULTS—SIXTEENMILE THRUST ZONE

Thrust faults cut the masses of pre-Tertiary rocks in the northeastern, central, and west-central parts of the quadrangle. The thrust in the northeastern part —the Lombard thrust of Haynes (1916b)—trends north-south and dips west. That in the west-central part—the Jefferson Canyon fault of Peale (1896) and Berry (1943)—trends east-west and dips north. These faults are linked in the center of the quadrangle by the Highway thrust of intermediate attitude, and appear to be elements of a single thrust along which gently to moderately folded rocks as old as Precambrian moved relatively eastward over tightly folded rocks as young as Late Cretaceous or even very early Tertiary. In turn, this thrust is not isolated but is the uppermost plate in a broad zone of thrusting, named the Sixteenmile thrust zone (Robinson 1959), most of which is concealed by Cenozoic deposits in the Three Forks quadrangle. The concealment is not fortuitous but reflects controls on Cenozoic sedimentation exerted by the thrusts and related steep faults, both actively by providing gradient through the vertical component of displacement, and passively by providing fractured rocks for easy erosion.

A few miles to the northeast this broad zone of thrust faulting emerges from beneath the Cenozoic cover. In the Toston quadrangle, the Lombard thrust is the most westerly of three westward-dipping thrusts in a strip some 8 miles wide, each of which has similar maximum stratigraphic throw: Belt series rocks on Kootenai formation. Another, similar thrust probably extends the thrust zone a mile or two farther east into the Maudlow quadrangle (Klemme's work [21] in the same area does not agree in all particulars but gives the same general tectonic picture). All the thrusts east of the Lombard appear in the northern part of the Manhattan quadrangle, where, according to Verrall,[22] they swing into southwesterly trends. The easternmost thrusts die out in the sharp, steeply plunging folds of the Horseshoe Hills but two thrusts, the Trident and Green, 7,000 feet apart, continue throughout the hills and disappear beneath Cenozoic deposits above the mouth of the Gallatin River. With the Lombard thrust, they form a zone of known but largely covered thrusting more than 3½ miles wide at the east edge of the Three Forks quadrangle.

[21] See footnote, p. 13.
[22] See footnote, p. 11.

This thrust system was not named by previous workers, although among Peale, Berry, Klemme, and Verrall all the individual thrusts comprising it have been named. The probability is good that the system is the southern extension of the Lewis thrust zone, as Clapp (1932) suggested, but the linkage is far from complete and a local name is desirable. Because most of the system is well exposed along Sixteenmile Creek, where much of it was first mapped by Klemme, I have proposed the name Sixteenmile thrust zone (Robinson, 1959).

LOMBARD THRUST

In the northeastern part of the quadrangle, rocks as old as the Middle Cambrian Meagher limestone are thrust over rocks as young as the early Carboniferous Mission Canyon limestone. This is the southern end of the Lombard overthrust of Haynes (1916b). The trace of the thrust, curving but with an average strike near N. 20° E., is exposed continuously for 4½ miles, from the north edge of the quadrangle, in secs. 19 and 20, T. 3 N., R. 2 E., to sec. 12, T. 2 N., R. 1 E., where it disappears beneath Quaternary gravel. For much of its length the thrust places Pilgrim limestone in the western upper plate on Mission Canyon limestone in the eastern lower plate. Toward the south, formations in both plates are progressively younger so that where the fault disappears below gravel Jefferson dolomite is thrust on Amsden formation. For most of its length the thrust has a single trace parallel to the bedding above and below. At the north edge of the quadrangle, however, it splits into four branches with an aggregate width of more than half a mile. Here, splits introduce slivers of Meagher and of Jefferson formation below the Pilgrim on the uppermost plate. The lowermost trace cuts sharply across the west limb of the Hossfeldt anticline and down section, so that Jefferson is thrust on Jefferson at the quadrangle edge. Another smaller split, into two branches, appears along 2,000 feet of the thrust in the southern part, mostly in sec. 6, T. 2 N., R. 2 E. This split introduces a sliver of Jefferson dolomite between Mission Canyon limestone in the lower plate, and Pilgrim, Maywood, and Jefferson rocks in the upper plate.

The trace of the thrust is marked by a slight depression and can easily be followed, but exposures of the fault surface are rare and the dip of the fault has been measured only at one point, where it is 30° to the west. The average dip is probably closer to 45° judging by the bends in the trace across rugged topography. This is about the average dip of the formations bordering the thrust. Signs of deformation such as drag folds, breccia, gouge, or mylonite are virtually absent. In many places on and downslope from the sag made by the thrust trace are concentrations of fragments of orange chert, apparently related to the fault surface but not seen in place. The maximum stratigraphic throw, Meagher limestone on Mission Canyon limestone, is 4,000 feet.

Haynes (1916b, p. 271, fig. 1) traced the Lombard thrust for 13 miles, with an average strike of N. 30° E., from 3 miles north of Three Forks across the double-horseshoe bend of the Missouri River near Lombard and 2 miles beyond. The stratigraphic throw increases northward with the result that near Lombard Haynes reported that the fault, dipping about 40° to the west, "has brought strata of the Belt series over strata of Cretaceous age" so that the maximum displacement "is approximately 2 miles and strata which are stratigraphically about 6,800 feet apart are here in contact." The northern 10 miles of the thrust has been recently mapped in detail (Robinson, 1959, pl. 2). The thrust is overlapped by Tertiary deposits 3 miles north of the horseshoe bend; thus it has a continuous length of 14 miles. Detailed mapping confirms Haynes' interpretation as to scale of throw and displacement. More precisely, strata as old as the Greyson shale are thrust over strata as young as the uppermost part of the Kootenai formation, with a stratigraphic throw of more than 8,000 feet.

HIGHWAY THRUST

In the center of the quadrangle a thrust fault, offset by a high-angle oblique fault, places Wolsey shale and Meagher limestone on complexly faulted Meagher and Pilgrim formations east of the high-angle fault, and uppermost part of the Wolsey and Meagher on North Boulder formation west of the high-angle fault. The thrust, exposed for 2 miles at the base of the hills that rise above Highway 10S (fig. 14), strikes northeast and dips northwest; the dip is at angles of 10° to 20° to judge by the sinuosity of the thrust trace and by one exposure in the east branch of the gulch that drains south along the west edge of SW¼ sec. 33, T. 2 N., R. 1 E. The thrust, not previously recognized, is here named the Highway thrust.

The trace of this thrust, unlike that of the Lombard thrust, is marked by several striking, if small, disturbed features. For example, in SE cor. sec. 33, Meagher limestone 200 feet above the thrust trace is dragged into a tight asymmetric anticline with a vertical southeast limb. At several places, especially where red arkose of the North Boulder formation is at the top of the lower plate, the rocks for 2 to 3 feet

FIGURE 14.—Highway thrust, exposed on Highway 10S, 3¼ miles southwest of Three Forks junction. North side of highway at turnout in SW¼ sec. 5, T. 1 N., R. 1 E. Thrust fault indicated by T. Geologist stands on upper red arkose of North Boulder formation (p€n). Thrusting has reduced thickness of Wolsey shale (€w) normally 200–400 feet, to 20 feet. Part of Meagher limestone (€m) is also thrust out above upper thrust slice, as basal Meagher beds are missing. Thick ledge above geologist is glauconitic quartzite, low in the Wolsey shale.

on either side of the fault trace are microbrecciated, slightly recrystallized, and bleached almost white. Where the thrust is in Wolsey shale much of the chlorite is locally reoriented parallel to the thrust surface. The maximum stratigraphic throw is less than 500 feet.

JEFFERSON CANYON THRUST

From the west edge of the quadrangle a large thrust fault trends east parallel to and a mile north of Highway 10S. This thrust, which dips 40° N. where exposed in the creek valley near the south edge SE¼ sec. 16, T. 1 N., R. 1 W., places Precambrian North Boulder formation on steeply dipping Upper Cretaceous Elkhorn Mountains volcanics and on a latite mass intrusive into the volcanic rocks. The apparent stratigraphic throw is on the order of 10,000 feet. The only signs of deformation along the trace are a few small drag folds, with steeply dipping south limbs, in the conglomeratic North Boulder rocks at the very edge of the quadrangle.

This fault is the eastern end of a thrust system that was traced first by Peale (1896, p. 5) and later by Berry (1943, p. 24–25) for at least 9 miles eastward from the mouth of South Boulder River; on Peale's map the east end of the fault is placed a mile west of the Three Forks quadrangle. Both Peale (his section *A*) and Berry (his section *B–B′*) interpreted the fault zone as a thrust system dipping 45°–60° N. Berry (1943, p. 25) estimated its stratigraphic throw at 12,000 feet and named it the Jefferson Canyon fault. It seems more evocative, however, to call it the Jefferson Canyon thrust.

The thrust is on the trend of the buried Willow Creek fault of Belt age, and may be the same fault, revived with opposite throw. In the Bridger Range the presumed same Belt fault zone has been interpreted as the locus of a similar reversal of throw in Laramide time (Pass fault of McMannis, 1955, p. 1420–1421).

MINOR THRUSTS

Two very small thrust slices flank Milligan Creek above Milligan Canyon. East of Milligan Canyon in SE¼ sec. 24, T. 2 N., R. 1 W., a thrust dipping northwest more steeply than 40° repeats the Phosphoria formation and much of the Ellis formation. West of Milligan Canyon, in N½ sec. 35, T. 2 N., R. 1 W., a thrust that dips about 30° N. repeats most of the Amsden formation and the basal Quadrant formation. Each of these small thrusts has a few hundred feet of stratigraphic throw, is covered by Cenozoic deposits at one end, and disappears into bedding at the other.

THRUST DISPLACEMENT

A start toward estimating thrust displacement is to compare it with the measurable stratigraphic throw, 1½ to 2 miles. The dip is roughly parallel to bedding, and the displacement may be many times the throw. But, the rocks were folded before thrusting and possibly the displacement is less than the stratigraphic throw. Because folding is open in most of the upper plate, minimum displacement should much exceed the stratigraphic throw and 3 miles would seem a reasonable estimate of minimum displacement.

To estimate maximum displacement is harder. The absence of Paleozoic and Mesozoic windows west of the thrust system foredooms attempts to estimate maximum displacement by direct measurement. A laborious kind of stratigraphic approach through the construction of palinspastic maps (Kay, 1945) might, however, ultimately succeed in yielding rude estimates. Its basic element would be an attempt to restore the deformed rocks to their prefold and prethrust positions with the aid of contour or isopleth maps of such variables as thickness, clastic ratio, dolomite-calcite ratio, or coarse-to-fine ratio of formations or systems or other rock or time-rock groupings. The four sets of detailed measured sections (p. 124–135) were made with a reconstruction of this sort in mind, but it was abandoned when rough comparisons among the summaries of these measurements (spread through the discussions of the sedimentary rocks) indicated that there is about as much stratigraphic variation within each plate as there is across the thrust. The measurements reveal a surprising amount of lateral variation in supposedly uniform Paleozoic shelf rocks but help little with the problem of displacement. With the poor data at hand it seems prudent to conclude merely that displacement on the Lombard-Highway-Jefferson Canyon thrust is probably on the order of a few miles. Displacement on the whole Sixteenmile thrust system is presumably several times greater than that on the Lombard thrust alone. Casual measurements suggest that far more crustal shortening is due to the Laramide folds.

If the eastward-trending Jefferson Canyon and northward-trending Lombard thrusts are part of the same thrust fault, it is necessary to reconcile their great apparent difference in relative direction of movement. If only one fault is involved, and the different traces are due to accidents of erosion, was the relative movement of the upper plate southeasterly, or did one direction of relative transport, east-west or north-south, dominate so that the movement on one of the segments was largely strike slip?

Regional evidence suggests that the dominant relative transport was east-west. The Sixteenmile thrust zone seems to be at the southern end of a sinuous belt, more than 350 miles long, of thrust faults that dip westerly (Clapp, 1932). The zone can be traced with scarcely a break from southern Alberta to the Canyon Ferry quadrangle (see fig. 1*B*) east of Helena, but a stretch of about 25 miles from Canyon Ferry to the north end of the Toston quadrangle is covered by Cenozoic deposits so that structural continuity, though reasonably certain, cannot be proved. Assuming continuity, the zone near Three Forks curves abruptly to the southwest and then disappears beneath younger deposits. The overall movements on the great plates involved were surely about at right angles to the northerly regional trend of the thrust belt, which means a principal eastward rather than southward component. No thrusts that might be direct southern continuations of the thrust zone have been recognized in the Precambrian crystallines south of the curving end of the known thrust belt. It seems likely that the faulted northern margin of the Precambrian crystalline rocks was the southern limit of the thrust belt and that the crystalline body acted as a buttress that absorbed any southward components of movement in the production of minor folds and faults. Berry (1943, p. 23) and Alexander (1955, p. 99) made similar interpretations. If this reconstruction is correct, movement on the Jefferson Canyon thrust was largely strike slip.

The problem of the direction and amount of displacement should not be left without reference to a bolder concept than I have ventured. Certain geographically discordant stratigraphic sequences in a reconstruction of the initial distribution of facies in the Upper Cambrian formations, led Lochman and Duncan (1944, p. 16) to conclude that southwestern Montana has been subjected to post-Cambrian crustal shortening in a northeasterly direction of at least 50 miles and to imply that the shortening reflects Laramide thrusting on the Lewis-Sixteenmile system; merely to flatten Laramide folds would not suffice. In my opinion their evidence, supplied by a few widely spaced measured sections, is insufficient, but their conclusion may nevertheless be correct.

TIME RELATIONS OF FOLDING AND THRUSTING

The thrust system is younger than the folds confined to pre-Tertiary rocks, for the thrusts shear off several folds but are not significantly folded themselves. If these were thrusts of Alpine type, in which thrusting is the end result of folding continued beyond the rupture point (break or stretch thrusts), the thrusting followed directly on folding. But in such thrusts folds both above and below a thrust surface should be typically tight and asymmetrically steepened or overturned in the direction of thrusting. Although the Hossfeldt anticline in the lower plate is tight and overturned, the larger structures in the upper plate nearby are open and only moderately asymmetric. Consequently, the thrusts seem to be subsequent shear thrusts.

If so, questions arise as to the age relations among thrusting, Laramide igneous intrusions, and related hydrothermal ore deposits. The intrusive rocks and any ore deposits genetically related to them are of intra- or post-folding age. It is conceivable that some or all of the thrusting is younger than the intrusives and their attendant ore deposits, though generally considered older (see, for example, Klepper and others, 1957, p. 60). For example, the 10N pluton and any buried ore deposits that might be associated with it may be cut off a few thousand feet down by the Lombard thrust.

STEEP FAULTS IN UPPER PLATE OF HIGHWAY THRUST

From the roads that flank it, the hilly mass that lies above the Highway thrust appears to be a straightforward pile of little-deformed strata, nearly horizontal above Milligan Creek and with moderate to low northwesterly dips elsewhere. Actually, the mass is broken by steep faults into a mosaic of tilted blocks. The individual faults, most of which are followed faithfully by valleys or by marked topographic sags, can generally be traced for no more than a few thousand feet. They are cut off by or merge with other faults, or trend beneath Cenozoic deposits, or simply die out in the thick carbonate units. All the faults, even those with curving traces, seem to be vertical or nearly so. Maximum stratigraphic throw and net slip on most of them is on the order of 500 feet, rarely as much as 1,000 feet, so that Upper Cambrian rocks are thrown against Devonian ones, and Devonian against lower Carboniferous, but there are no larger stratigraphic throws. All the faults are probably oblique slip but with large vertical components. Significant horizontal offset is indicated only on the fault that, striking N. 15°W., cuts across the Jefferson-Broadwater County line in secs. 36, T. 2 N., R. 1 W., and 31, T. 2 N., R. 1 E., where drag and shearing in Three Forks shale indicate that the east wall moved south.

These faults are most readily interpreted as gravity faults. Little crumpling, gouge, or breccia is associated with them, except where three faults merge near the township corner half a mile east of the mouth of Dry Hollow. There, prominent ridges of Jefferson dolomite are intensely brecciated.

Most of the blocks bordered by these faults have been tilted perhaps 10° to 20° southward, opposing the general moderate northwesterly dip pattern. The small rectangular block that centers in the NW center sec. 6, T. 1 N., R. 1 E., however, seems to have popped up like a cork.

Included in this group of steep faults is the long, gently curving fault that bisects sec. 31 from northeast to southwest. The map pattern of this fault suggests that it is a minor thrust parallel to the Highway thrust and much like the two small slices farther upsection, but excellent exposures show conclusively that the fault is a trivial, near-vertical one, that for all but a small fraction of its length drops beds close to the base of the Three Forks shale on the west against strata at the top of the Jefferson dolomite, with a throw of less than 50 feet. Near its western end, it rises slightly in section and repeats basal beds of the Three Forks.

None of these minor faults are large enough to have been shown by Peale. Berry (1943, pl. 1, p. 24–25) generalized the pair of faults that affect the Madison limestone as a single "high-angle thrust" with horizontal offset of nearly a mile, but recognized none of the other faults nearby.

These faults presumably formed during development of the Highway thrust but their contacts with the thrust are concealed. Like the thrust, these faults are older than the Bozeman group for they are overlain by unfaulted rocks of the Milligan Creek formation. They are also older than the basin occupied by the Milligan Creek rocks for there was a reversal of topography after faulting and before the Tertiary basin existed. The upthrown sides of these faults along the east side of the valley now occupied by Milligan Creek had to be reduced by postfaulting erosion to produce this part of the Tertiary basin. Simple differential erosion contemporaneous with initial faulting will not suffice, for the carbonate rocks on the opposite sides of the faults are of equal resistance to weathering. Like the thrust faults, too, the steep faults are younger than the pre-Tertiary folding for they show no signs of being folded themselves. A reasonable inference is that these structures formed as blocks jostled on the shallow upper plate as thrusting proceeded. The absence of similar complications above the Jefferson Canyon and Lombard thrusts possibly reflects a greater thickness of cover. On the other hand, the two slices higher in the pile may simply not have moved far enough to develop jostled blocks.

OTHER STEEP FAULTS CLOSELY RELATED TO THRUSTING

Three long faults roughly parallel to bedding have been mapped in upper Paleozoic rocks south of the Jefferson River. The longest extends for more than 2½ miles from NW ¼ sec. 1, T. 1 S., R. 1 W., to SE¼ sec. 33, T. 1 N., R. 1 W.; it may be much longer, as it is concealed by Quaternary deposits at each end. For most of its length it raises rocks in the lower part of the Lodgepole limestone on the north against strata low in the Mission Canyon limestone on the south. Along two stretches, however, the north wall is raised enough to expose distinctive upper strata of the Three Forks. With the upthrown side downdip, the dip of the fault must be greater than that of the strata, which ranges widely along the fault but averages about 60°. The stratigraphic throw is perhaps 900 feet, a few hundred feet more than the thickness of the Lodgepole, presumed to average about 600 feet in this vicinity. The dip slip necessary to achieve this much stratigraphic throw is 2,000 feet. The amount of strike slip, though unknown, is probably small.

The parallel fault to the south has several similarities but is not identical. At its east end in center SE¼ sec. 2, T. 1 S., R. 1 W., the fault raises beds low in the Jefferson dolomite on the north wall against upper Lodgepole limestone on the south. Westward, the upthrown side gradually rises in section so that both walls are in Lodgepole limestone in SE¼ sec. 3, and the fault can no longer be recognized. The maximum stratigraphic throw at the east end is around 800 feet. Because the dips of beds along the fault average about 75° the theoretical dip slip is more than 3,000 feet. At first glance, this fault seems to be rotational, with the actual displacement declining westward to zero within a little more than a mile. But eastward the displacement seems to decline to zero rapidly, too, as the fault, covered by Quaternary valley deposits at the east end, does not reemerge in the hills only half a mile away.

The third fault mapped in this area apparently joins the first described one under Quaternary gravel in SE cor. sec. 33, T. 1 N., R. 1 W. Trending west-southwest, it crosses N½ sec. 5, T. 1 S., R. 1 W., and leaves the quadrangle to pass under Cenozoic deposits along upper Sand Creek. Whether or not it continues on the west side of the creek is not known. It is exposed at the south edge of Ingleside Quarry, where its dip is near vertical. Its curvature nicely reflects the syncline it crosses but it departs far from bedding, have Lodgepole in the south wall and Mission Canyon in the north wall at its east edge and the reverse situation where it leaves the quadrangle. It appears, therefore, to be a rotational fault, hinged somewhere in the NE¼NW¼ sec. 4 where both walls are in Mission Canyon limestone. The north wall has moved relatively up at the west and down at the east. Maximum throw at either end is a few hundred feet. This fault is not merely a curving continuation of the fault with which it merges for that fault must continue toward the mouth of Sand Creek to explain the relations between Mission Canyon and Lodgepole, north of the fault junction.

Like the block faults across the river, these faults are not only older than the Tertiary basin deposits, but are also older than the basin itself. The evidence is the same as for the block faults: the upthrown sides of the faults, with equally resistant units on both sides, face the present valley, which was also the Tertiary valley, and must have been reduced by erosion (or reversal of movement for which there is no evidence) in the development of the basin.

STEEP FAULTS YOUNGER THAN THE THRUSTING

Only one fault is demonstrably younger than the thrusting and a second is interpreted to be younger. Neither of these faults can be traced for as much as a mile before passing beneath Cenozoic deposits. Nevertheless, they may be symptomatic of a major stage in the tectonic history of the quadrangle: an episode of post-thrust steep faulting that did much to define the Cenozoic basins. That such faults, if present, are rarely exposed seems due to rapid retreat of initial scarps in the humid earliest Tertiary climate and to the mantling of the fault traces by the resulting debris.

The single certain post-thrust fault splits the Highway thrust in the center of the quadrangle. Curving gently in plan, with an average trend close to due north, the fault is in Meagher limestone for most of its exposed length of 4,000 feet but at its southern end it drops the upper(?) part of Meagher on the east wall against Wolsey shale on the west. Its stratigraphic throw and dip slip are both on the order of a few hundred feet. Comparable displacement is also indicated by the vertical separation in the projected dip of the thrust plane. Considerable strike slip, with the east side having moved relatively south perhaps more than 1,000 feet, is indicated by lateral offset of the Wolsey and by drag folds. This fault is marked by a bare ridge of contorted limestone trending athwart the normal consequent drainage to the Jefferson River. A remarkable box canyon, as much as 200 feet deep and not much wider at the top (but shown on the topographic map as simply a flaring gulch) follows the fault trace in SE¼SE¼ sec. 32, T. 2 N., R. 1 E., and NE¼NE¼ sec. 5, T. 1 N., R. 1 E.; making a right angle bend across the fault ridge, the canyon widens abruptly in Wolsey shale above the Highway thrust only 600 feet from the Jefferson River.

This fault is pre-Quaternary. It is older than old alluvium which covers it and it does not seem to offset any Quaternary landforms. There is no direct evidence of its age relations to Tertiary basin deposits but it seems reasonably regarded as of pre-basin age.

The second fault in this category is exposed for half a mile near the mine workings in sec. 25, T. 3 N., R. 1 E. This fault, striking N. 45° E., disappears to the north in a poorly exposed dip slope of Jefferson dolomite; to the south, it passes under Quaternary fan gravel. Downthrown on the northwest, it brings Lodgepole limestone against Jefferson dolomite on the southeast with a stratigraphic throw of perhaps 500 feet. The net slip may be much less because the fault is oblique to bedding. Exposures are inadequate to show whether monzonite, intrusive into the Lodgepole and bottoming on the fault, is offset by the fault or controlled by it. It is easy to imagine that this fault continues for several miles, more or less following Dunbar Creek, determining the contact between Lodgepole limestone and Tertiary deposits and, initially or by recurrent movement, being responsible for the abnormally high dips in the isolated mass of Dunbar Creek rocks in SE¼ sec. 3, T. 2 N., R. 1 E. The fault is regarded as of postthrusting age only because it resembles the one known postthrusting fault in being accordant with basin topography; it may even be post-Tertiary, but older than the Quaternary fan.

ORIGIN OF CENOZOIC BASIN

When deposition of the Bozeman group began, the topography of the quadrangle seems to have been much like that of the present, with steep ridges separated by broad valleys. How that terrain developed is largely conjectural. Strong diastrophic movements must in some way have been involved in the creation of the Three Forks basin, whose borders cut abruptly across faulted and folded older rocks, whose early record is of interior drainage in a humid climate, and whose shape closely follows known fault trends. It is surely no accident that the basin lies on and shares the trend of the east-west Precambrian Willow Creek fault and the Jefferson Canyon thrust, and that the adjoining Clarkston and Townsend basins trend at right angles and lie on the Sixteenmile thrust zone. Assuming this fundamental structural control, the basin might have developed by differential erosion of unresistant Precambrian rocks (and shaly Upper Cretaceous ones east of Three Forks) exposed as a result of structural events, or it might have originated as a direct expression of structure, from the downdropping or tilting of blocks on steep faults younger than the thrusting but genetically related to it.

If differential erosion were the dominant element, it should be possible to detect large-scale patterns in which the basin in earliest Tertiary time was integrated with drainage to the sea, for vast amounts of rocks had to be removed. Atwood (1916) tried to demonstrate an erosional origin of all the intermontane basins of southwestern Montana. He concluded that all the basins initially developed from the dissection of an Eocene peneplane by consequent streams that drained southward, from a Continental Divide at the present eastern front of the Rocky Mountains to the ancestral Snake River. The "lower Bozeman beds," in his nomenclature, were deposited in the valleys of this stream system which was closed in late Oligocene time by the "great outpouring of lava in the Snake River region." Subsequent drainage changes he ascribed in part to structural events, but stream erosion was named the primary agent in creating the intermontane troughs. Original drainage to the south was proposed to explain "exceedingly peculiar" drainage conditions off the present Continental Divide involving alternating basin and canyon stretches along major streams, barbed tributaries, abandoned valleys, great wind gaps, and the like.

Atwood's reconstruction, though ingenious, does not seem adequate, as Blackwelder (1917) early pointed out. A glance at Atwood's illustrations (pl. 27, fig. 52) shows that the explanation creates about as many peculiar physiographic conditions as it resolves. Further, the early Tertiary Continental Divide could not have been as far east as postulated by Atwood, as J. J. Tanner [23] and H. D. Klemme [24] both showed that early Tertiary drainage across the Big Belt Mountains and Maudlow Basin was eastward rather than westward, and Billingsley (1915, p. 32), McMannis (1955, p. 1412) and Klepper and others (1957, p. 38–40) have shown that much of the Late Cretaceous and Paleocene Livingston formation was derived from the eastward transport of detritus from the Elkhorn Mountains volcanics.

More significantly, there now appears to be little basis for an Eocene peneplane on which to develop integrated consequent drainage, to the south or in any direction. The principal evidence advanced by Atwood for the peneplane is the accordance of many summits in the Butte region. Because basins below those summits were then known to contain deposits as old as Oligocene(actually, Eocene, for the deposits near Lima had been described by Douglass in 1903), Atwood reasoned that the peneplane formed before Oligocene time but after Late Cretaceous deformation, and thus was Eocene. But the considerable Cenozoic folding and faulting now known in the region (see Pardee, 1950, for partial summary) make it most unlikely that remnants of a low surface dating back to the early Tertiary could be preserved in accordant summits. The summits may signal the existence of a former extensive surface of little relief, but if so, it is of much later age.

The foregoing objections to an Eocene peneplane are largely negative; in addition, there is some opposing positive evidence. The extensive and thick Beaverhead conglomerate (Lowell and Klepper, 1953), at least part of which is of Paleocene age and which is itself folded and overridden by pre-Oligocene thrusts, and the thick Eocene basin deposits near Lima and Three Forks, combine to indicate a degree of early Tertiary crustal unrest and surface irregularity hardly compatible with the idea of a peneplane.

There seems no reason to assume that the Three Forks basin is fundamentally the product of an integrated drainage system that flowed south or west, and was later blocked by deformation or lava damming. If the basin is mainly erosional, northerly or easterly

[23] Tanner, J. J., 1949, Geology of the Castle Mountain area, Montana: Princeton Univ. Ph.D. thesis (available on microfilm).

[24] See footnote, p. 13.

drainage is far more compatible with the apparent distribution and height of the pre-Tertiary ridges, much like the present ones. Easterly drainage during the early Tertiary is particularly favored by the pattern of gravel channels in the Climbing Arrow formation, and by the fact that an eastward-flowing stream system is known to have existed as late as Paleocene time, to transport debris from the Elkhorn Mountains volcanics to the Maudlow and Livingston basins, to provide the andesitic components of the Livingston formation.

Possibly, the earliest Tertiary drainage was not simply subsequent on uplifted unresistant strata but directly consequent on topographic irregularities produced by rapid Laramide faulting. The meager direct evidence, in the form of the Sphinx conglomerate apron and of the lake limestone of the Milligan Creek formation, supports the idea that the Three Forks basin was born closed. If so, its initial relief was primarily tectonic and not erosional.

It will probably never be possible to decide whether contemporaneous deformation or simple stream erosion was the main mechanism of basin formation, if either was. An unspectacular but reasonable view is that tectonism was generally slow enough and erosion fast enough so that both played important, complexly interrelated parts in the genesis of the basin.

POST-LARAMIDE STRUCTURAL FEATURES

However the Tertiary basin formed, recurrent diastrophism was probably responsible for its recurrent existence as the site of thick continental sedimentation during the dominantly warm and humid Tertiary climate. Often, exterior drainage systems to the east must have threatened capture of the basin; the long hiatus in the middle Tertiary record suggests that they sometimes succeeded. To maintain aggrading conditions in early and late Tertiary time, capture of the basin must have been defeated by diastrophism, for extensive and thick lava flows that might have dammed its natural eastern outlet are unknown. As suggested by McMannis (1955, p. 1426–1427), discontinuous relative uplift of the Bridger Range during the Tertiary offers a persuasive mechanism for maintaining a depositional environment. The barrier, whether this or some other, was apparently operative from very early Tertiary to middle Oligocene time, and again in late Miocene and at least part of Pliocene time, but ineffective in the long interval between.

Though the structures that influenced deposition of the basin deposits are obscure, some of the structures imposed on the deposits themselves are fairly plain.

The gross structure of the Tertiary rocks is that of an easterly tilted block that has been thrown into several broad gentle folds of diverse trend. Regional eastward dip has been noted by Peale (1896, p. 3), Pardee (1950, p. 365), McMannis (1955, p. 151), and others. Easterly tilt of a few degrees is shown not only by the dominance of dips in that direction, but also by the steady rise in stratigraphic position and age of the rocks eastward and southwestward without regard for intervening masses of older rocks. This asymmetry in the distribution of the Tertiary formations is apparent throughout the Three Forks basin. At the west edge of the basin, the exposed Tertiary rocks are tuffaceous sediments of early Oligocene age (Klepper and others, 1957, p. 42; Alexander 1955, p. 76). At the east edge, as noted earlier, the Tertiary rocks are of late Miocene and Pliocene age. A similar distribution prevails in the adjoining Townsend Valley. At the west side, the Tertiary rocks are eastward-dipping lower Oligocene strata (V. L. Freeman, written communication, Mar. 22, 1954). Just east of the Missouri River in the Toston quadrangle the Tertiary strata dip mainly east and are mostly of early(?) and middle Oligocene age (dated on the basis of vertebrate remains identified by G. E. Lewis, written communication, Mar. 12, 1957). At the east edge of the valley are Miocene (and Pliocene?) gravels (dated on the basis of vertebrate remains identified by G. E. Lewis, written communication, May 6, 1957) and dip eastward into the west frontal fault of the Big Belt Mountains.

It seems clear that the eastern edge of the Three Forks basin is a fault at the west edge of the Bridger Range; its throw has been estimated (Pardee, 1950, p. 379–381) as 3,000 feet or more. The same fault zone forms the east edge of Townsend Valley. But downfaulting does not explain the dips of the Tertiary strata in the Three Forks quadrangle into islands of pre-Tertiary rocks, because there is no repetition of beds on the east sides of the islands. Tilting after basin filling is also inadequate to account for the eastward rise in stratigraphic position, though it can account for the regional dip.

Eastward tilting, beginning after deposition of the Milligan Creek formation and continuing or recurring throughout deposition of the rest of the Bozeman group, with progressively shifting depositional troughs eastward, seems the most satisfactory interpretation. A consequence of such deformation is that the entire thickness of basin deposits at any point may be only a fraction of the apparent total thickness derived by adding the thicknesses of individual units at their outcrops. The tilted block may have been the eastern flank of a huge anticlinorium trending north-northeast, the core of which was the Boulder batholith, long since crystallized, and this rising fold might have provided both gradient and material for the alluvial part of the Bozeman group.

At the scale of the geologic map, the wrinkles on the tilted block are more apparent than the block itself. Perhaps the largest of these wrinkles is a northeasterly trending anticline underlying the Jefferson River valley and largely concealed by Quaternary alluvium. Its dimensions are vague but measurable in miles. The few dips visible on its flanks are mostly under 5° but in a few places are as high as 15°. The fold is doubtless partly responsible for the exposure of the early Tertiary rocks along the river, both by uplifting the rocks and by localizing the course of the river. How much of the persistent southeasterly dips in the Tertiary rocks south of Buttleman Ranch are to be ascribed to this anticline and how much to regional tilt is moot. North of this anticline the Milligan Creek formation is folded into an open syncline. To the east of the large anticline, another vaguely defined but apparently broad and low anticline trends east-west or northwest-southeast, with its crest emerging on the Madison bluffs about a mile north of Climbing Arrow Ranch. This anticline, with limbs dipping generally 5° to 10°, seems to plunge gently southeastward and disappear beneath the Madison Valley. Up plunge, it disappears under Quaternary alluvium.

Many smaller folds appear in the Tertiary rocks north of the Jefferson River, but only two have been traced for as much as a mile along the axis. These are the anticline and syncline previously mentioned in discussion of the Climbing Arrow formation. Most clearly defined is the syncline south of the Silver Sage Ranch, whose axis trends north-northwesterly across the W¼ sec. 10, T. 2 N., R. 1 W., disappearing to north and south under Quaternary alluvium. The fold is asymmetric, the west limb gentle with dips of 10° or less, the east limb steeper with common dips of 15° to 25°. The anticline, whose crest trends roughly parallel, across the W½ sec. 10, T. 2 N., R. 1 W., is a much gentler fold with opposing dips of 3° or less.

These folds are much too large to be slump structures or the result of differential compaction of underlying strata. Possibly they are the products of regional compression but their divergent trends invite the interpretation that they are shallow reflections of local readjustments of underlying Laramide fault blocks.

No faults profound enough visibly to affect older rocks transect the Bozeman group. Many minor faults are probably present, as suggested by linear trends athwart bedding visible on aerial photographs, but they could not be found on the ground. As noted earlier, the fault in Paleozoic rocks on upper Dunbar Creek may have initially or by recurrence deformed the Tertiary rocks, too, but the evidence is concealed by Quaternary gravel. If faulting occurred after the Dunbar Creek formation was deposited, the total displacement was too small to produce recognizable repetition of Tertiary beds on opposite sides of the Hossfeldt Hills.

The Quaternary rocks are nowhere markedly deformed except for scattered slump structures on steep valley walls. Slight Quaternary uplift with northerly tilt offers adequate explanation for at least two major geomorphic events: the capture of the basin, formerly closed or draining east, by the Missouri River; and the development of the series of Quaternary benches. Discussion of these events and their probable tectonic origin is deferred to the section on geomorphology.

GEOMORPHOLOGY

Near the close of Tertiary time, the Three Forks quadrangle seems to have been an area of low relief, with scattered hills protruding no more than a few hundred feet above a graveled plain. The development of the present topography is the history of the dissection of that surface under a climate that has alternated between humid and semiarid, and of the exhumation of a very early Tertiary surface of mature relief, that had been buried by the Tertiary basin deposits.

The main agent of dissection has been a stream system consisting of large perennial rivers, deriving their discharge from a persistently humid region far beyond the Three Forks basin, and their tributaries, varying from perennial to nearly dry as the basin climate varied. The former flood plains of the through streams remain as terrace remnants, capped by rounded gravels, the courses of their tributaries are marked by subrounded gravel fills. In the Mud Springs Gulch drainage, broad benchlands that are virtually bare-rock surfaces are identified as pediments; apparently they have long been isolated from integrated streams and have been slowly lowered by mainly colluvial processes. Some of the detritus used as tools in dissection remains behind, evidence alike of the cutting processes and of their inefficiency, but the volume of Quaternary deposits is small in comparison with the amount of material removed. The dissection was episodic, presumably with long periods of lateral cutting-and-filling by the streams alternating with short ones of vigorous downcutting.

Although uplift probably caused the dissection, discontinuities in uplift were not necessarily responsible for the discontinuities in dissection revealed by the steplike benches. Proximate cause for the development of most benches, and perhaps for all, may be differing resistance of the strata at the outlet canyon of the Missouri River from the Three Forks basin.

The lowest benches may be nonorogenic. They may be climatic terraces or the products of base level changes stemming from the growth and drainage of glacial Lake Great Falls. The present downcutting in tributary valleys is of probable climatic origin and has continued for only a short time; the flood plains above incised stream courses may therefore be regarded as climatic terraces.

HILLS

Most of the hills and ridges of the quadrangle, though neither high nor extensive, are rugged, rising abruptly from the benchlands to summits that are generally 800 to 1,300 feet above the valley floors, and thus to altitudes ranging from 5,000 to 5,400 feet; a few ridges rise several hundred feet higher. Most summit areas are flattish, though narrow. The prominent ridges of the quadrangle are without exception strike ridges of Paleozoic carbonate rocks. Lower ridges are developed on quartz-cemented sandstones of the Flathead, Quadrant, Ellis, and Kootenai formations.

Low hills of irregular shape are developed on large intrusive masses and on the Elkhorn Mountains volcanics.

The hills are exhumed from a late mature early Tertiary topography that was much like the present topography. The Tertiary deposits, however deformed, faithfully follow the details of the present hills and valleys. The surface below the Tertiary deposits, wherever it is visible in stream cuts or has been intersected by the drill, is a smooth continuation of the hill slope above. There is only a small notch where the subaerial and subsurface slopes meet, indicating that the modern hill fronts are not only parallel to the early Tertiary ones but not far from them. This slight amount of retreat further suggests that most of the present hill slopes were buried by Tertiary deposits until well into Quaternary time.

Added indication that the hills are exhumed is given by the Tertiary deposits themselves. They invariably coarsen toward the hills. The lithology and attitudes of many border sandstones and conglomerates demonstrate that they were shed from the hill directly above.

The coarseness of the very early Tertiary limestone conglomerate suggests mountainous relief. The later Tertiary conglomerates at hill edges are, however, distinctly finer grained than the modern gravels in similar positions, suggesting somewhat less relief than at present, and general lowering of relief during the Tertiary as the basin deposits grew thicker and the summits were worn lower. This general tendency may have been reversed for a time during the middle Tertiary hiatus. Low relief in late Tertiary or earliest Quaternary time is suggested by the fineness of the oldest Quaternary subrounded gravel, by the rude accordance of summits, and by the roundness of crests and divides.

How the hills initially formed is unknown. Most likely it was by a combination of Laramide faulting and differential stream erosion, during and soon after Laramide deformation.

BENCHES

Most of the quadrangle is a series of broad benches that slope toward the junction of the Madison and Jefferson Rivers. The benches are especially well defined south of Three Forks, just above the Madison River, where five distinct levels appear. The lowest is about 40 feet above the Jefferson flood plain three-fourths of a mile southeast of Three Forks; northeastward, it merges with the flood plain within a mile; southwestward, it continues unbroken for 3 miles. A second bench level is about 150 feet higher; it is prominent east of Fairview Cemetery and between Willow Creek and the Buttleman Ranch. The third level, about 100 feet higher and 300 feet above the flood plain, forms the extensive benchlands between Willow Creek and the Climbing Arrow Ranch, and is the only prominent level west of Willow Creek. About 180 feet higher is a fourth bench, conspicuous only near the Madison River. The fifth bench, more than 500 feet higher, and 1,000 feet above the flood plain, forms the highest surface at the southeast edge of the quadrangle; this is the north tip of an extensive bench that continues for many miles southward into the Norris quadrangle. The road that leads southeast from Three Forks to Fairview Cemetery, and then follows section lines for 6½ miles south, intersects the leading edges of three of these benches: the lowest at approximately 4,100 feet, the third at about 4,350 feet, and the fourth at about 4,530 feet. East of Fairview Cemetery the east edge of the second bench is close to 4,240 feet; between Willow Creek and the Buttleman Ranch its leading edge is slightly above 4,260 feet. The lower edge of the highest bench is not far from 5,050 feet.

Most interbench slopes are moderate, with grades perhaps two or three times that of the bordering benches. The highest bench, however, is bordered by steep slopes throughout and for long stretches the faces between other benches are steep also.

North of the river, benches are not nearly so well developed. Around the mouth of Mud Spring Gulch the two lower benches are extensively developed, and the third level is prominent west of Milligan Creek, but elsewhere in the Mud Spring Gulch drainage and along upper Milligan Creek benches are numerous but small, much dissected, and difficult to correlate with each other or with the extensive surfaces south of the river. These northern benches are irregularly rolling and poorly drained.

Most of the broad benches south of the Jefferson River are thickly capped by rounded gravel. The lowest bench north of the river is similarly capped, as is the distal end of the very broad bench above it. These surfaces were evidently cut by the same stream or streams that deposited the rounded gravels upon them. They are valley-plain terraces of the ancestral Jefferson and Madison rivers.

The highest gravel-capped terrace seems to be in the lowest part of a gently rolling surface elsewhere defined by the summits of the bedrock ridges. This surface apparently had less than 500 feet of relief, but the coarseness of the gravel on the terrace shows that mountainous terrain existed not far away.

The old valleys filled with subrounded gravel no doubt represent the primary consequent tributaries of the ancestral river system.

All but the lowest of the northern benches lack caps of rounded gravel, or any vestiges of such caps. Doubtless they are far more dissected than the southern benches owing to the lack of a protective cover of coarse gravel. Most of these benches are cut across the edges of folded strata of the Bozeman group, and therefore they are not stratum benches as defined by Rubey (1952, p. 113). All contain patches of fan gravel and windblown silt, and a few are mantled widely by such materials. Evidently they were never part of the valley plain of perennial rivers nor of the flood plain of perennial tributaries of the sort that deposited the subrounded gravel. Presumably they are pediments initially cut by sheetflow or by intermittent tributaries, left as benches after the master streams and their accordant tributaries were incised to lower levels, and subsequently lowered by colluvial processes. Details of their origin and history are, however, hard to visualize. If slow colluvial processes have indeed been major agents in their genesis and subsequent lowering, a necessary implication is that the surfaces have been effectively isolated from the much more rapid attack of integrated stream systems. This could have been achieved if the Jefferson River had flowed consistently south of its present course in earlier times, so that the northern part of the quadrangle remained a divide area. Later discussion of the asymmetry of terrace distribution leads to this very suggestion.

The origin of seemingly contemporaneous river terraces and pediments is one of the intriguing general problems not yet studied in those extensive parts of the arid West crossed now, or formerly, by perennial through rivers.

The manner of retreat of the fronts of both terraces and pediments seems fairly clear. The frontal scarps, steeper on terraces capped with resistant gravel than on unprotected pediments cut on soft rocks, were formed as the products of stream downcutting and maintained by stream sidecutting until a new episode of incision. Thereafter, their retreat has presumably been mostly by mass-wasting processes. Whatever the mechanism of retreat, it is evident from inspection of the frontal scarps that they have retreated parallel to their original slopes, for the scarps are of about equal average declivity from highest terrace to lowest and from highest pediment to lowest.

STREAMS

The Jefferson River meanders across the quadrangle in a broad flood plain interrupted by a short canyon near the west edge. The flood plain slopes from slightly below 4,200 feet altitude at the west edge to 4,040 feet at the east edge; the river has a channel gradient of 5.6 feet per mile. No measurements are available of width, depth, or velocity. For a stream meandering on a broad flood plain, however, the Jefferson flows remarkably fast.

The Madison River flows back and forth across the east edge of the quadrangle in a flood plain less broad. The slope of this flood plain is somewhat greater than that of the Jefferson, from 4,240 feet at the south edge to 4,040 feet at the river mouth. Unlike the Jefferson, the Madison has a comparatively straight but braided course, and its channel gradient is much steeper, 12.5 feet per mile. No measurements of width, depth, or velocity are at hand; the impression is that the bankfull Madison within the quadrangle is a little broader, much shallower, and about as swift as the Jefferson.

The two rivers join at the east edge of the quadrangle to form the Missouri River; above the junction is a long reach of distributaries. The Missouri leaves the quadrangle at an altitude of about 4,035 feet and is joined by the Gallatin River a few hundred yards east.

To reach the quadrangle, each river passes through alternating reaches of wide flood plain and of narrow canyon. The rivers head within a few miles of each other in the high humid Yellowstone Plateau and are perennial from head to mouth. The Jefferson, flowing in a broad arc convex to the west, drains about 9,200 square miles, and has average discharge near Three Forks of around 2,300 second-feet; the Madison, flowing pretty much straight north, drains only 2,500 square miles and has an average discharge of about 1,600 second-feet. (For detailed discharge data, see U.S. Geol. Survey, 1953, p. 19, 71, 76; 1954, p. 40, 43, 46.)

Only a few of the tributaries are perennial. Willow Creek has a small annual flow that is maintained by dams upstream from the quadrangle. Segments of Milligan Creek and of two or three other northern tributaries of the Jefferson are perennial, being fed from springs, but their dry season flow is rarely greater than a trickle. Most water courses are dry most of the year.

The primary and secondary tributary valleys are mostly flat-floored, even near their heads. Exceptions are the short tributaries of the Madison at the east edge: they are V-shaped canyons from mouth almost to head.

Along much of their courses the larger tributaries are markedly incised in their flood plains. The Madison and Jefferson are not strikingly entrenched nor are the tributaries at their mouths, but beginning a few hundred feet upstream the tributaries are progressively more deeply incised, to depths of as much as 15 feet (see fig. 15). The entrenchment commonly persists for a few miles, then gradually dies out in headward areas.

Drainage throughout the Quaternary seems to have had the same general pattern as now to judge by the distribution of older alluvial deposits, and by the detailed drainage relations of through streams, primary tributaries and secondary tributaries.

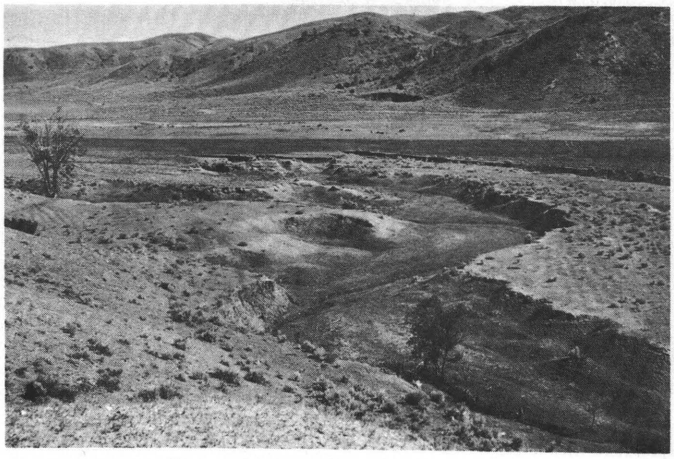

FIGURE 15.—Milligan Creek entrenched in its flood plain above Milligan Canyon.

In general, the present drainage pattern seems consistent enough with the interpretation that the through rivers and the primary tributaries developed as consequent streams with dendritic pattern in soft relatively undeformed Quaternary and Tertiary deposits with an initial northeastward slope, and that the secondary tributaries, including such as Dry Hollow and Dunbar Creek, are rudely trellislike subsequent streams controlled by exhumed Laramide structures.

But some stream features require more involved explanations. The Jefferson River and each of its main tributaries run for a small part of their courses within a comparatively narrow hardrock valley or steep-walled canyon that flares out into broader valleys upstream and downstream. At the present level of exposure, soft unresistant Cenozoic rocks nearby seemingly offer far easier paths. As there is no indication that these restricted stretches have been subjected to localized Quaternary uplift, the ready conclusion is that the streams have been superposed from a continuous cover of virtually unconsolidated, unresistant Cenozoic deposits. The position of Tertiary outcrops and of terrace gravels suggest that such a blanket may well have existed to altitudes of slightly above 5,000 feet, adequate to account by superposition for all the anomalous canyon reaches within the quadrangle.

At one or two of the canyons the tributary streams may at times have been unable to maintain their course across hard rocks. The valleys just above Milligan Canyon and above the Buttleman laccolith on Willow Creek flare broadly and some of the deposits preserved on their sides are unusually fine grained and well sorted. These relations suggest that the streams were dammed and built lake deposits, over which they ultimately flowed to complete cutting the present canyons. Though resembling the process of anteposition as visualized by Hunt (1956, p. 65–67) there is no evidence of active deformation during stream evolution, as required for anteposition.

The aggradation that produced the present flood plains is presumably due to the same tectonism or downstream control by differential hardness that was responsible for the higher benches. Why flood plains are absent from the Madison tributaries is unclear. Perhaps rapid westward sidecutting by the Madison is responsible. In the present climate, the stream mouths, under the influence of the perennial Madison, may be migrating westward faster than their heads. The overall gradient may, therefore, be increasing and any deposits that may be made after floods are quickly removed.

Why the lower Madison is straight and braids but the lower Jefferson meanders is not clear. The two rivers are much alike in many respects, including the important ones of source area and types of rocks crossed, but the Jefferson has much lower average gradient than does the Madison, and much greater discharge. To make any headway on this problem, far more information is necessary on such ~~variables~~ as bed load and channel shape.

ORIGIN OF BENCHED TERRAIN

Preceding discussion of the landscape in the benchlands and lowlands has touched mostly on the development of the near-horizontal surfaces—terraces, pediments, flood plains. The purpose now is to consider the origin of the intervening steeper surfaces. Involved are at least four widespread sets of steep surfaces below benches, with heights of as much as 500 feet, in a total vertical interval of 1,000 feet. The mere scale of these features suggest strongly that uplift was their prime cause. The nature of the uplift, however, is far from clear. Was it all at the start of the bench-forming episode or did it continue throughout bench development? If it continued, was it smoothly slow or spasmodic? Was it vertical or tilted?

As to the timing and duration of uplift, continuing but spasmodic upward movement offers, of course, the simplest answer: one bit of uplift for each paired bench. Certain simple regional evidence, however, opens the field to other explanations. The number and size of valley-plain terraces, and the height and continuity of interterrace scarps vary widely and abruptly from basin to basin in the Missouri headwaters region. In general the number, complexity, and vertical spacing of benches increases upstream. In the valley of the Missouri above the Gates of the Mountains and below Canyon Ferry only two broad low benches seem to be developed, judging by a rapid automobile reconnaissance. In the Townsend Valley (Pardee, 1925, p. 6–8) are three well-developed bench levels, in a vertical interval of about 300 feet. In the Three Forks basin are five main levels over a vertical range of 1,000 feet. In the middle Madison Valley above Ennis are at least six conspicuous benches over a range of more than 2,000 feet. This simple picture is complicated by the existence of many fragmentary benches that are slip-off slope terraces or local rock-defended terraces rather than basinwide valley-plain terraces; further, some of the benches noted may not be stream terraces at all, but pediments or structural benches. Nevertheless, the generality seems to hold. This indication of a measure of independence in bench development from basin to basin suggests that differential resistance of folded and faulted hard rocks encountered by the rivers in the canyons between basins may have been an important element in valley sculpture. Thus, steady smooth uplift or a single early upward movement might both have resulted in alternating benching and scarping in a basin as the master streams met rocks of varying resistance in the downstream canyon. In this view, even the most extensive paired bench might be regarded as a species of rock-defended terrace and the whole flight of terraces visualized as the products of "continuous valley excavation during [variably] restrained downcutting," to modify the description of Davis (1909, p. 514–586).

The uplift, whether smooth or jerky, seems to have been not vertical but tilted northward, on an axis somewhere north of the Gates of the Mountains, as suggested by the increase in number of benches from basin to basin upstream (southward) and the increasing total relief within the benched interval upstream. Several other lines of evidence of varying validity lead to the same idea. The course of the Jefferson River, to judge by the distribution of valley-plain terraces, has been shifting progressively northward throughout the Quaternary. Small remnants of paired benches show that the river has in the past flowed north of its present path but not far and not for long. Its present course is on its north bank across most of the quadrangle; presumably it would be there throughout if it were not superposed on limestone of the Madison group at the south side of the valley above Willow Creek. (See fig. 16.)

These observations are unimpressive alone, but begin to be more meaningful when taken in regional context, for the entire headwaters region of the Missouri shows strong signs of having developed initially

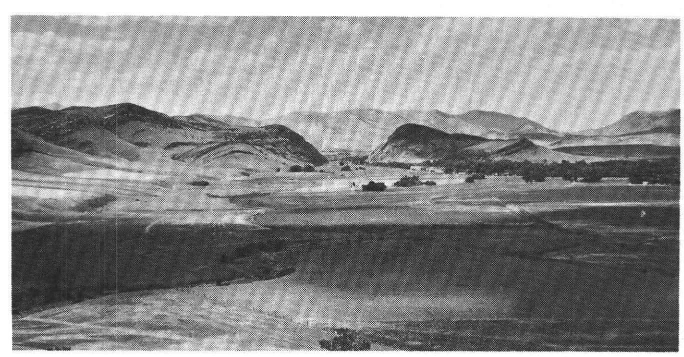

FIGURE 16.—Jefferson River superposed on limestone of the Madison group, 4½ miles above Willow Creek. View is due west from Pleistocene terraces, visible in center left. Course of river is marked by line of trees in middle right.

by consequent flow down a regional northward slope and of having experienced northward, or better northwestward, tilting late in its history. Initial northerly slopes, prior to superposition, are suggested by the overall north-pointing dendritic pattern of the master streams and primary tributaries. Northwesterly tilting is favored by the persistent tendency of the through rivers, where in broad valleys, to flow against their western and northern valley walls, clearly shown by examination of 15-minute and 7½-minute quadrangles in the region, but not very satisfactorily on smaller scale maps.

The dips of benches seem to support the idea of northerly tilt. The benches in the quadrangle are graded to the Jefferson and Madison Rivers and their initial surface dips were presumably about the same in amount though differing in direction. The impression is strong that the benches north of the Jefferson River now have lower dips, on the undissected parts of their surfaces, than do those south of the river. But measurements testing the impression have not been made and are probably not feasible, owing to the considerable dissection of the unprotected pediments that dominate the northern benches.

Perhaps the greatest virtue of the concept of continuing or repeated northerly tilt during the Quaternary is that it accounts for the capture of the eastward-draining Tertiary basin by the present northward-draining system.

Although all the terraces can be rationalized as the product of continuing or recurrent uplift, possibly complicated by differential resistance of rocks in outlet canyons, at least one other idea of origin should be considered. The terraces were formed coincidentally with some of the major climatic changes of the Quaternary and perhaps some of the terraces are climatically controlled. Given enough initial uplift, it is not inconceivable that climatic change dictated the entire sequence of alternate downcutting and sidecutting, with differential resistance at outlets accounting for the differences in terrace development from basin to basin. Climatic terraces (Bryan, 1925) result when flood plains built during periods of relatively humid climate are cut into as the climate shifts toward aridity, protective vegetation dies out, and gullying ensues. The concept has often been invoked to account for low terraces produced by the "Recent epicycle of erosion," first recognized by Bryan and thoroughly described by Hack (1942), in the arid Southwest. Whether the idea can be applied to features as large as the high terraces is conjectural.

Whether or not climate played an important part in forming the high terraces, it probably controlled incision of the flood plains of present tributary streams. A marked shift toward aridity during the past century is well documented in Western United States, presumably including Montana (see Bryan and Hack for data), and the present trenches have probably been cut entirely within that time. The perennial rivers, not much affected by the climatic change because they derive their flow from the persistently humid Yellowstone Plateau, have not become entrenched, and have limited the entrenchment of tributaries near their mouths.

Conceivably, some of the minor low terrace remnants may be neither tectonic nor climatic in the direct sense, but related to the growth and drainage of glacial Lake Great Falls (Calhoun, 1906). Glacial Lake Great Falls, ice-dammned at the north and penned by the Highwood, Big Belt, and Little Belt Ranges on the south, maintained itself with an upper surface on the order of 800 feet above the prelake channel of the Missouri long enough to produce recognizable shoreline features and, partly drained, had another steady stage at about 400 feet above the prelake channel. Such lakes certainly must have been effective temporary base levels for the stretch of the Missouri just upstream, but it is uncertain that even deep lakes existing for no more than a few hundreds or thousands of years could significantly affect the relation between downcutting and sidecutting almost 150 miles upstream and across several reaches of hard-rock canyons.

GEOLOGIC HISTORY

Whatever has seemed worth saying about the origin and history of each element of the geology of the Three Forks quadrangle has been said earlier in this report, but under topical headings. The object of this chapter is to place all the main known and inferred deposits and events in their chronologic order. Many uncertainties and gaps in the record, brought out previously, are here arbitrarily resolved and filled, to smooth the story.

The known geologic record of the area begins in early Precambrian time with the piling up of thick arkosic sand and gravel in shallow marine or continental waters. Mafic lava occasionally poured out on the clastic sediments and was injected into them as sills. This sequence was indurated, folded, and recrystallized to form gneiss from the sandstone and amphibolite from the igneous rocks. The metamorphic cycle ended with the formation of small masses of granite pegmatite and of vein quartz.

By late Precambrian time the area was part of a large low landmass contributing fine detritus to the shallow sea not far to the north and west in which the late Belt rocks of the Missoula group were accumulating. By Newland time, or perhaps as early as Chamberlain time, the crystalline rocks along the southern edge of the quadrangle began to rise rapidly along an east-west fault zone at about the latitude of Willow Creek village—the Willow Creek fault—and shed coarse feldspathic detritus northward to intertongue with the fine deposits that continued to thicken farther north. Repeated recurrence of movement on the fault zone produced coarse arkosic sand and gravel—the North Boulder formation—throughout Newland and probably well into Greyson time. As part of the orogenic disturbance, small masses of diabase were erupted near or at the surface, and fragments worn from the diabase were incorporated with the sand and gravel that accumulated to thicknesses of more than 4,000 feet. As faulting proceeded, hematitic phyllonite was produced in the gneiss along the fault zone.

Before Spokane time, movements on the Willow Creek fault zone apparently ceased. The landmass was reduced and only fine detritus was shed into the Spokane sea. Little or none was deposited south of the fault zone. The shallow sea may have persisted long after the end of Spokane time and the other formations of the Missoula group above the Spokane may have been deposited in this area.

In very latest Precambrian or Early Cambrian time, the sea retreated northward or westward and a long period of subaerial erosion followed. The retreat and erosion were probably due to another uplift on the Willow Creek fault zone. A low rolling surface was carved that transgressed progressively older Belt rocks southward and extended eventually across the fault zone and onto the gneiss and amphibolite mass.

Across this surface the Middle Cambrian sea advanced slowly from the west eroding Belt and crystalline rocks alike, dropping a veneer of coarse quartz sand—the Flathead sandstone—near shore and around the bases of low islands, and thickly spreading the clayey and silty waste—the Wolsey shale—farther from shore and in places atop surfaces too high to be mantled by sand. In areas bypassed by currents carrying clastic debris, thin layers of limestone were deposited.

As the shoreline gradually transgressed far beyond the quadrangle layers of limestone became thicker and more numerous, terrigenous strata thinner and fewer until a thick sequence of largely carbonate beds—the Meagher limestone—was formed.

The shoreline retreated briefly in latest Middle Cambrian time and fine-grained clastic sediments were again washed into the area to form the Park shale.

In Late Cambrian time eastward transgression of the shoreline was resumed. Conditions much like those of Meagher time were restored and the thick Pilgrim limestone was formed.

The region began to rise gently in very Late Cambrian time and the sea to withdraw. Thin regressive shale, siltstone, and clastic limestone—the Dry Creek shale and Pebbly limestones of Peale—were probably deposited at this time but were stripped or so deeply weathered that their original character disappeared in the ensuing vast interval of Ordovician, Silurian, and earlier Devonian time in which the area stood close to sea level, receiving no new deposits and losing very little of its old ones.

In Middle or Late Devonian time the sea again began to advance across the area from an unknown direction, but probably the west, reworking the weathered Dry Creek strata and redepositing the debris in hollows, as the basal mudstone and sandstone of the Maywood formation. As transgression continued, the waters cleared and quieted. Chemical carbonate deposition began to prevail over clastic deposition and the upper limestone of the Maywood formed.

Later, in the early part of Late Devonian time, the sea continued shallow, but persistently agitated, and the largely clastic petroliferous Jefferson dolomite was deposited.

Well into Late Devonian time, the body of marine water overlying the area, already in limited communication with the open sea, was cut off from the sea, and the basal muddy evaporitic beds of the Three Forks shale resulted.

When restricted communication with the sea was restored, the dark pyritic shale and thin muddy red limestone of the bulk of the Three Forks shale were laid down. Renewed transgression at the end of Devonian time ended these semieuxenic conditions. Along the advancing shoreline the upper orange siltstone and sandstone of the Three Forks shale were laid down contemporaneously with dark shale offshore.

As the sea advanced farther, clastic deposition in the area virtually ceased and limestone, mostly of chemical origin, was laid down. A time of gentle oscillations is represented by the thin-bedded Lodgepole limestone. Extremely stable conditions later led to formation of the thick-bedded Mission Canyon limestone.

Shortly after the Mission Canyon was deposited, the area emerged from the sea, probably due to easterly tilting in this region, leaving a small restricted depositional basin—the Big Snowy basin—to the northeast. The exposed Mission Canyon strata in the Three Forks area were deeply weathered. Cave systems developed in the upper part of the Mission Canyon and a red regolith formed at the surface. Solution- and collapse-breccias eventually filled the caves, probably not long after they were formed. Very near to the quadrangle, perhaps locally within it, the solution networks rose to the surface, producing karst topography. Later, the Big Snowy sea began to spread westward, eventually filling and leveling the irregularities on the eroded Mission Canyon surface with a heterogeneous assemblage of nearshore mudstone, sandstone, and carbonate rocks—the Big Snowy formation.

Another episode of uplift, late in Big Snowy time, led to deep erosion of the Big Snowy rocks in most of the quadrangle, and their complete removal from the west-central part.

In middle or late Carboniferous time, transgressive marine conditions were reestablished. The Amsden formation was deposited, beginning with muddy red strata derived from the pre-Amsden erosional mantle and later, as the water cleared, passing into carbonate rocks and subordinate pale quartzitic sandstone. In late Carboniferous and possibly early Permian time, sandstone deposition prevailed over limestone, and the Quadrant formation resulted. Later, but still in the early Permian, dominantly regressive shallow marine conditions became the rule, and the cherty and phosphatic sandstone, siltstone, and dolomitic limestone of the Phosphoria formation developed. Regression was probably southward due to gentle uplift on the north, perhaps on the Sweetgrass Arch.

Before the start of Triassic time, the sea had receded south of the quadrangle and did not return again, apparently, until Middle or Late Jurassic time. The area was slightly emergent during at least part of this interval and chert fragments worn from the Phosphoria became incorporated in the base of the succeeding Ellis formation, a heterogeneous complex of clastic limestone, sandstone, and conglomerate laid down under unstable shoreline conditions. The Ellis sea had advanced from the east; it retreated toward the east, still within Jurassic time. In its wake, streams eroded lowlands rising slowly to the west and laid down the continental deposits of the Morrison formation in latest Jurassic time. A surge of more rapid uplift to the west in Early Cretaceous time produced the basal coarse continental sandstone of the Kootenai formation. Sluggish streams deposited fine-grained strata during most of Kootenai time; occasionally, they were rejuvenated and laid down coarse sandstone.

In Late Cretaceous time the sea again invaded the area, depositing thick shale and sandstone of the Colorado formation and probably younger units. There were probably several advances and retreats of the sea during very late Cretaceous time, but their record is not preserved in the quadrangle.

The deformation that drove the Late Cretaceous sea from the area for the last time marked the local start of the Laramide orogeny. This deformation was probably in the form of gentle folding along north-south axes. The rising folds were deeply eroded, and on the irregular surface thus formed the Elkhorn Mountains volcanics were erupted in Late Cretaceous time. Compression continued and the folds became asymmetrically steepened or overturned toward the east.

As the folds developed they provided space for a host of igneous intrusions of calcalkaline composition: mostly andesite, dacite, latite, monzonite, and quartz monzonite. Most of the intrusions came in as thin sills in the shales of the Wolsey, Three Forks, Amsden, and Kootenai formations. Some made larger masses: dacite and andesite formed large laccoliths in shale; monzonite formed one complex semiaccordant pluton that mostly displaces Paleozoic carbonate rocks; and large pluglike bodies of andesite and latite were intruded into late Mesozoic rocks.

After folding, and probably during the long interval of intrusion, a broad system of thrust faults—the Sixteenmile thrust zone—developed diagonally across the quadrangle from northeast to southwest. On the thrust system, dipping 30°–50° northwesterly, rocks as old as the Belt series were pushed a few miles relatively eastward over rocks as young as the Kootenai formation. Thrusting was accompanied by complex block faulting in one segment of the upper plate and by steep faulting in a lower plate crushed against the Precambrian crystalline rocks. In late Paleocene or early Eocene time, many block faults possibly provided the framework of the Tertiary basin, but only two minor faults of the sort required are known. Others, if they exist, are concealed by the Cenozoic debris they gave rise to.

While deformation and intrusion were proceeding at depth, an eastward drainage system developed on the surface. The Elkhorn Mountains volcanics were deeply eroded and the fragments carried to the Maudlow and Livingston basins where they were deposited as part of the Livingston formation.

Diastrophism, probably the first uplift of the Bridger Range, interrupted eastward drainage in middle or late Eocene time, and a brief episode of interior drainage ensued. At an early stage the basin floor was dry, and limestone gravel that became the Sphinx conglomerate was deposited as an alluvial apron around many hills. At this time, or perhaps somewhat later, olivine basalt plugs were intruded. A lake developed in the southwestern part of the quadrangle and the Milligan Creek formation accumulated in it. The lake received not only limy muds offshore and coarse hill wash nearshore but also much volcanic ash, blown into the area from distant vents.

Elsewhere in the quadrangle, streams continued to carry material eastward, but perhaps slowed or blocked by the rising Bridger Range began to deposit gravel and sand in their channels and montmorillonitic mud in overflow ponds on the flood plains. These deposits—the Climbing Arrow formation—were largely detritus from the ancestral Elkhorn, Madison, and Tobacco Root Mountains, but were partly contemporaneous volcanic ash. After the Milligan Creek lake was filled or breached, fluviatile Climbing Arrow rocks began to accumulate thickly in the southern part of the quadrangle, too.

Climbing Arrow sediments continued to pile up until well into early Oligocene time. Then by a combination of eastward tilting in the area and further uplifts to the east, another basin like that of Milligan Creek time was created but with its center farther east, outside the quadrangle. In this basin, occupied only intermittently by lakes, grew the Dunbar Creek formation, formed mostly of airborne volcanic ash with only small amounts of evaporitic limestone and of streamlaid ash and mountain waste.

Continued eastward tilting shifted the area of deposition out of the quadrangle, and by middle or late Oligocene time it was no longer receiving significant volumes of deposits. Beginning in middle or late Oligocene time and continuing intermittently until late Miocene time, easterly exterior drainage resulted in the removal of large volumes of Eocene and Oligocene deposits.

In late Miocene and early Pliocene time, a closed basin like that of Dunbar Creek time again developed, but centering east of the quadrangle. In it the Madison Valley formation of Douglass (1908) was deposited, largely from fine pyroclastic materials and under conditions reminiscent of the Dunbar Creek formation. The quadrangle was apparently on the western margin of the Madison Valley basin and received coarse clastic channel, fan, and delta deposits during late Miocene and early Pliocene time. By late Pliocene time the Madison Valley basin, and probably the entire Three Forks basin as well, was filled, and sluggish exterior drainage to the east was reestablished. The Three Forks quadrangle was a graveled plain with a few low hills rising above it.

In latest Pliocene or earliest Quaternary time a northerly draining consequent stream system developed, probably initiated by uplift with northwesterly tilt in the ranges to the south, and the formerly eastward-draining or closed basin became part of the Missouri River system. The ancestral Madison and Jefferson rivers and their primary tributaries cut rapidly down through the soft Tertiary rocks and became superposed on the rugged pre-Tertiary surface. The cutting proceeded in a series of giant terrace steps controlled by discontinuities in the rate of tilting, or by varying resistance of rocks encountered in the downstream canyon outlet, or by major climatic changes, or varying combinations of these. Rounded gravels were deposited on the terraces by the master streams, and subrounded gravels in the tributary valleys. In dry intervals, in very late Pleistocene and Recent time, thin deposits of fan gravel and windblown silt formed on the benchlands. The present flood plains were built mostly in relatively humid Quaternary time. In the past century a semiarid climate developed; vegetation died out, and the flood plains of tributaries were deeply trenched, becoming climatic terraces. Little affected by the changing climate, because supplied by the persistently humid Yellowstone Plateau, the master streams did not become entrenched.

ECONOMIC GEOLOGY

CONSTRUCTION MATERIALS

The flood plains of the through rivers offer endless sources of gravel and sand, though much testing would be necessary to outline suitably large bodies of these materials that would meet the requirements for construction of large dams and heavy-duty roads. A few small gravel and sand pits have been opened but none have produced large tonnages to judge from the small size of the pits, and none was in continuous operation during the period 1952–57. Pits at flood-plain level are of course somewhat undesirable because of the danger of recurrent flooding by seepage or overflow, and the necessity of lifting the materials. More economic sources for sand and gravel would be older stream materials of Quaternary or even Tertiary age, exposed in the scarps between benches. Little sand, however, is likely to be preserved at bench fronts and sand is better sought on the flood plains. A few desultory attempts have been made to develop the higher gravels but were quickly abandoned, possibly due to lack of demand rather than to poor quality.

Two main ingredients in cement manufacture, limestone and quartz sand, are available in large quantities of high purity. Parts of the Paleozoic limestones, especially the Mission Canyon limestone, offer unlimited amounts of cement-grade limestone, and much of the white quartzitic sandstone of the Flathead and Quadrant formations could supply suitable sand.

The best local source of riprap is probably the brown sandstone of the Ellis formation because of its natural tendency to separate into blocks with dimensions of a few feet. The most favorable localities are east of Milligan Creek. With less predictable splitting properties, the sandstones of the Flathead and Quadrant are not so promising but should be suitable.

The bentonitic clays of the Bozeman group appear to have swelling ratios and other properties far below specifications for permanent canal linings, but might be useful temporary sources in emergencies.

The white tuffaceous siltstones of the Bozeman group contain too many impurities in the form of admixed sand and thin interbeds of other materials to be promising sources of commercial pumicite.

GLASS SAND

Very small fractions of the sandstone of the Quadrant and Flathead formations may be potential sources of glass sand. At the south end of the hill of Quadrant that occupies E½NE¼ sec. 5, T. 2 N., R. 1 E., is a small quarry reported to have produced glass sand. It was not in operation during any of several visits in the period 1953–56. However pure the sand from these formations may be, it is invariably cemented enough so that some grinding would be necessary to prepare it for use as glass sand.

MICA

A little book mica occurs in one area of Precambrian gneiss. The largest books of muscovite are 8 inches across and 3 inches thick. Some books of biotite are even broader and thicker. Most of the mica, however, is in much smaller aggregates. The mica is intimately associated with both granite pegmatite and vein quartz that cut the gneiss. Mica-bearing pegmatite is exposed in a single zone in SE¼SE¼ sec. 9, T. 1 S., R. 1 W., near the base of the Flathead sandstone. The zone, trending N. 86° W., makes a small angle with the trace of the Flathead so that at its west end it passes below the Flathead. Book mica occurs notably only at this end, where it is exposed sporadically by a shallow trench 150 feet long, trending N. 15° W., and by a few small pits nearby. No mica books crop out elsewhere but the associated pegmatite continues eastward for 600 feet. A vein of milky white quartz that lies next to the mica and the pegmatite in the trench is traceable by prominent outcrops and float for at least a mile farther eastward.

The trench exposes a small anticlinal fold, plunging steeply about N. 50°–60° W., in gneiss. The schistosity, striking N. 85° W. and dipping 80° S., trends unbroken across the crest of this fold. The crestal zone is intricately brecciated; among horses of gneiss, are masses several feet long of coarse biotite, muscovite, feldspar, granite pegmatite, and milky quartz. The distribution of mica seems to be controlled by the schistosity rather than by the fold.

There is no information that any mica has been shipped.

PHOSPHATE ROCK

The phosphatic sandstone and shale of the Phosphoria formation are mined in many parts of southwestern Montana. The Phosphoria in the Three Forks quadrangle, however, is dominated by chert and limestone and contains little phosphatic material. A few strata, especially in the Milligan Creek sector, are seemingly of commercial grade but are far too thin and impersistent to warrant mining.

IRON AND MANGANESE

Small bodies of hematite and related iron and manganese oxides crop out at many places in the quadrangle. All have been prospected without leading to production. They are of two main types: red sedimentary deposits of hematite-rich sandstones at the base of the Amsden and Big Snowy formations; and reddish-black hydrothermal deposits of mostly specularitic iron oxides and sooty black manganese oxides, with associated silica, in limestone and, rarely, quartzite at and near contacts with monzonitic intrusive rocks. In none is the body of ore minerals more than a few feet thick or continuous along strike for more than a few tens of feet. None has been followed for more than a few tens of feet downdip.

COPPER

No minable copper deposits are known in the quadrangle and no copper production has been recorded. Along the northern and eastern edges of the 10N pluton in both the monzonite and intruded carbonate rocks are many small showings of copper, largely in the form of surface stains of blue and green copper salts. All have been prospected. Copper sulfides appear in only one prospect, in north-central part sec. 26, T. 3 N., R. 2 E. (See pl. 1.) Here, a shear zone, 3–5 feet wide, striking N. 88° E. and dipping 83° N. at the contact between quartz monzonite and a xenolith or pendant of limestone, has been explored by a deep shaft. To judge by material on the dump beside the shaft, the zone is mineralized with a little chalcocite, covellite(?), and magnetite. The sulfides are thinly coated with malachite, azurite, chrysocolla, and a moderate yellowish-green mineral, in radiate clusters of curved brittle fibres, that is probably a copper salt. Monzonite of the footwall is for the most part remarkably fresh; locally, it is chloritized. The hanging wall is shot with veinlets of cross-fibre carbonate minerals.

At the Copper City workings southeast of the center of sec. 25, T. 3 N., R. 2 E., many hundreds of feet of shafts and drifts have been driven but there is no sign of copper ore beyond faint copper-salt stains on any of the dumps or outcrops, and no record of production. The prospect seems to have been last worked in 1949 and the claims patented by Mr. Herbert Dunbar, a rancher resident near Three Forks, as of July 5, 1950. The shaft, well timbered, had a hoist stuck about 50 feet down during several attempts at examination so I was unable to go underground. A few outcrops and material on the dumps indicate that the workings are in recrystallized white Lodgepole limestone, permeated with dikelets and with irregularly shaped masses of hornblende- and biotite-bearing monzonite. Locally, the limestone is altered to spotted calc-silicate rock, as described previously.

OUTLOOK FOR FUTURE DISCOVERIES OF ORE DEPOSITS

By current standards, the Three Forks quadrangle is an unpromising place to seek ore deposits. But an intriguing idea offers hope of discoveries in the distant future when outcropping ore bodies have been exhausted and new discoveries, if they are to come at all, must come from locating completely buried deposits. The idea follows from the now commonplace observation that individual ore deposits tend to be controlled by local structures in the country rocks, and that ore districts are controlled by larger scale structures. Pre-ore and intra-ore folds and, especially, faults of suitable orientation have commonly determined the loci of minable deposits. Especially favorable sites for ore deposition are the junctions of two or more sets of faults or persistent fractures of appropriate age.

The ore deposits on the east flank of the Boulder batholith, as shown by many workers (for example, Billingsley, 1915; Pardee and Schrader, 1933), are closely associated in time and space with the monzonitic intrusive rocks of the batholith. Appearing within intrusive rocks as well as in the surrounding rocks, the ore deposits are somewhat younger than the intrusions. The intrusive rocks were emplaced during and after Laramide folding. Their relation to major faulting is not clear, either locally or regionally, but is probably an overlapping one; in at least one place in southwestern Montana—the Willis quadrangle (W. B. Myers, oral communication, 1957)—a large Laramide monzonitic intrusion definitely cuts a major thrust. Most workers (for example, Klepper and others, 1957, p. 44, 60) regard the main episode of intrusion as later than the main period of folding and faulting. Thus the ore deposits of the region were probably deposited late in or after a major interval of faulting.

It is, then, conceivable that the juncture between the north-south Sixteenmile zone of Laramide faulting, and the east-west Willow Creek fault zone, was the locus of an ore district that is concealed by the Tertiary and Quaternary deposits of the Three Forks basin.

How might this hypothesis be tested? Random drilling through the Cenozoic cover, even though it probably averages little more than 1,000 feet thick in the target area, would be prohibitively expensive. Perhaps geochemical prospecting methods could be employed. From time to time, deep water wells and oil tests are drilled in the basin areas. At such times cuttings and fluids might be tested by available quick and cheap field tests for metallic anomalies, for it is well known (Hawkes, 1957, p. 267–289) that geochemical halos migrate upward into younger overburden. If striking anomalies appear they might form the basis for a drilling program. Surface or subsurface geophysical approaches might also be tried, but would seem less promising in view of the thickness of the cover.

PETROLEUM POSSIBILITIES

The geology of the Three Forks quadrangle, and its regional setting, has not encouraged petroleum exploration, and as a result its oil and gas possibilities are largely untested. No experienced exploration company has ever drilled a test well in or near the quadrangle, and locally sponsored tests, of which seven are recorded (see pl. 2), have yielded little information other than some terse drillers' logs with highly questionable stratigraphic interpretations or none. (I am indebted to C. E. Erdmann, of the Geological Survey, for copies of these logs.) None of these tests was as deep as 2,300 feet, three were less than 1,000 feet deep, and all were collared in Cenozoic deposits, so that in the absence of properly made logs it is impossible to determine what pre-Tertiary formations were eventually penetrated, if any.

About the only encouraging element from the standpoint of petroleum is the rapid thinning and disappearance, by wedgeout and erosion combined, of the Big Snowy formation within the quadrangle. The Big Snowy is a producing interval farther east in the State, and in the adjoining Toston quadrangle many of its outcropping strata are notably petroliferous. The associated solution breccia zone in the upper part of the Mission Canyon limestone has evident possibilities as a reservoir but there is no positive reason to think it acts as one in this region.

Parts of the Radersburg syncline offer the only places where the Big Snowy might be tested at reasonable depth and with reasonable hope of avoiding unfavorable structural complexities or large intrusions. Such places are upsection from the Big Snowy outcrops in the Milligan Creek sector, but well away from the andesite masses. There is little reason to suspect that this part of the Tertiary basin is fault controlled, so that tests might start in Cenozoic deposits.

Attempts to test the Big Snowy beneath the Lombard thrust within the quadrangle would seem inadvisable because the formation is presumably involved in unfavorably steep folding. Northeast of the quadrangle, where the remnants of Big Snowy rocks have moderate to low dips in certain thrust slices, the situation is perhaps a little more inviting.

Worth noting here is a thin zone of hydrocarbon enrichment in bleached Jefferson dolomite that might be called a dried oil seep. The occurrence is on the north wall of the west-draining gulch in NW¼NW¼ sec. 36, T. 3 N., R. 1 E., half a mile south of Copper City. The zone appears as a vivid irregular dark streak in a low-dipping fracture in white limestone of the Jefferson dolomite, bleached (and dedolomitized) due to intrusion of the 10N pluton. For a length of about 100 feet, the limestone for 8 inches to 2 feet on both sides of the fracture is darkened by hydrocarbons. In places hydrocarbon rich "dikes" extend as much as 4 feet above and below the central fracture. Perhaps this is an orthodox seep, but the setting suggests that it is a place where hydrocarbons driven away from the Jefferson during intrusion seeped backed down after the intrusive and the country rocks cooled.

MEASURED SECTIONS IN PRE-TERTIARY ROCKS

As complete a section as practicable waas measured in each area of pre-Tertiary rocks: Section A in the Milligan Creek sector; section B, Mud Spring Gulch sector; section C, Hossfeldt Hills sector; section D, Willow Creek sector. Each section is located on plate 2.

Total formation thicknesses were measured by planetable methods, except for the Mission Canyon and North Boulder formations, the thicknesses of which were estimated from a series of map measurements. Thicknesses of lithologic units within each formation were measured with steel tape and Brunton compass. Within formations, taped thicknesses are not corrected for curvature or bedding slip. Total formation thicknesses are adjusted to an arbitrary round number to allow for structural thickening and at the same time give a figure which is both reasonable and easily handled in discussion. For convenient reference, each measured interval has been given a number, prefixed by the symbol for its section (A, B, C, or D); each number also represents a specimen collection.

The Milligan Creek area offers much the best conditions in the quadrangle for measuring the pre-Tertiary section; elsewhere, in fact, confidence in measurements is weakened because of steep dips and other structural complications. Accordingly, section A is more detailed and, in general, more accurate than sections B, C, and D.

In section A, effort was made to record in the field the following features for the main rocks in each measured interval: common name; fresh color; weathered color if much different from fresh; type of stratification and of parting if prominent, based on McKee-Weir scale (1953); degree of roundness; grain size, based on Wentworth scale for clastic rocks and on resolution with 10-power hand lens for crystalline carbonate rocks; composition; and any special characteristics useful in identifying the unit. In sections B, C, and D, much of the petrographic detail is omitted or generalized.

SECTION A

Above Jefferson Canyon thrust, west-central part of quadrangle, in the drainage of Milligan Creek. From upper part of North Boulder formation to large andesite intrusive in lower part of the Kootenai formation. Formations from the North Boulder through the Lodgepole are represented by essentially continuous measurements which begin in the bed of the large creek that flows south along the east edge of sec. 10, T. 1 N., R. 1 W. Here, the beds strike N. 75°–80° W. and dip 30°–35° N. Upsection above the Wolsey strikes swing more northerly and dips decline slightly so that in sec. 2 the strike averages N. 85° W., and dips are 25°–30° N.

Above the Mission Canyon the measured section shifts 3 miles northeast, where the remaining measurements are in two segments. The first begins at the top of the Mission Canyon and goes to the base of the Phosphoria across the NE corner of sec. 25, T. 2 N., R. 1 W. and into the SE corner of sec. 24. Beds in the lower 200 feet of this segment (Big Snowy and Amsden formations) are somewhat crumpled and the reported thicknesses may be greatly in error; an average strike of N. 40° E. is assumed. Dips, except in the crumpled zone, are near 30° N. The second segment begins 3,200 feet west of the upsection end of the first, and extends from the base of the Phosphoria into the lower part of the Kootenai; in the Phosphoria and Ellis the strike is N. 70° E., swinging easterly in the Cretaceous formations to N. 85° E.; dips in the lower part average 32° N., flattening to about 29° N. in the upper part.

	Unit thickness (feet)
Andesitic intrusive.	
Kootenai formation:	
A–62. Poorly exposed. Sandstone and siltstone, thin-bedded, reddened and hornfelsed near plutonic contact. Main rock types, in scattered outcrops and float, are moderate- to pale-yellowish-brown calcareous subrounded fine sandstone; yellowish-gray salt-and-pepper subrounded medium sandstone, weathering grayish orange; dark-yellowish and reddish-brown siltstone	69
A–61. Sandstone, thick-bedded, faintly laminated. Basal 3 ft is grayish-orange calcareous limonitic subangular medium sandstone, like that in the Morrison formation (A–60), grading into yellowish-gray medium subrounded calcareous salt-and-pepper sandstone, with scattered grains of orange earthy limonite and chips of shale; a few thin lenses of pebble conglomerate. Lower 35 ft forms ledge	88
Total adjusted thickness	150
Morrison formation:	
A–60. Siltstone and fine subrounded sandstone, medium-bedded, yellowish-gray, calcareous, with a few lenses as much as 2 ft thick of mottled reddish-brown, olive-gray argillaceous microcrystalline limestone. In lower 30 ft are several thin beds of reddish-brown mudstone. Light-olive-gray limestone, like A–59, in beds 2–3 ft thick, at 15 ft and 40 ft from top	74
A–59. Limestone and siltstone. Three beds, each 2 ft thick, of light-olive-gray microcrystalline limestone, separated by 4-ft beds of yellow calcareous siltstone	14
A–58. Mudstone, like A–58 below but thin bedded	48
A–58. Mudstone. Mainly reddish brown, but in many places grades laterally and vertically into light olive gray, yellowish gray and medium light gray; irregularly thin to thick bedded; a few lenses, as much as 2 ft thick, of argillaceous microcrystalline grayish-olive limestone. In lower 40 ft, a few beds 1–2 ft thick of yellowish-gray calcareous siltstone	90
Total adjusted thickness	220
Ellis formation:	
A–57. Limestone, slabby, yellowish-gray, and light-olive-gray microcrystalline, weathering grayish-yellow, with a few thin beds of pale-reddish-brown calcareous micaceous medium sandstone, weathering moderate yellowish brown. Several short covered intervals are probably on siltstone	36
A–56. Sandstone, moderate-yellowish-brown to light-brown, hard, medium, calcareous. Lower 7 ft forms ledge; rest of unit yields dip slope covered with large slabs	22
A–55. Limestone. Grades laterally from pale-yellowish-brown to grayish-orange coarse reef calcite sandstone with many chert and quartz grains and fragments of pectens, cemented with calcite, to dense laminated finely crystalline medium-gray limestone. Basal 2 ft has many chert pebbles. Thin hard sandstone lenses in upper 4 ft; top channeled by overlying sandstone A–56	9
A–54. Chert pebble conglomerate. Subangular to rounded pebbles of dark chert and yellowish-orange siltstone in slightly calcareous sand matrix; weathers moderate yellowish brown, forms distinct ledge	8
Total adjusted thickness	75
Phosphoria formation:	
A–53. Chert and sandstone. Ledges a few feet thick of massive brownish-gray to yellowish-brown chert separated by poorly exposed intervals of similar thickness of thinly interbedded chert, platy yellowish-orange and yellowish-gray very fine sandstone, and dark-greenish-gray medium to coarse phosphatic sandstone. The phosphatic sandstone is made up of rounded to subangular chert grains, phosphatic oolites, shell fragments and iron-stained clay, cemented in places by calcite, in places by silica. A phosphatic bed 1½ ft thick crops out 3½ ft above the base of this unit; concentrations of float suggest 2 or 3 similar but much thinner beds in the upper part of the unit. Phosphatic grains weather bluish white, making distinctively speckled surfaces. Phosphatic parts have strong sour odor when wetted with HCl or struck sharply. Striking whitish outcrop	49

	Unit thickness (feet)
Phosphoria formation—Continued	
A–52. Dolomitic limestone, with chert nodules, like A–50	26
A–51. Largely covered. Sparse outcrops are of thin-bedded nodular grayish-orange fine quartzite, weathering medium yellowish brown; float suggests most of interval is thin-bedded brownish-gray chert	22
A–50. Dolomitic limestone, light-olive-gray to yellowish-gray, finely crystalline, with faint irregular bedding planes a few inches apart. Many pods and lenses of light-olive chert, locally rich in fusulinids	23
Total adjusted thickness	120
Quadrant formation:	
A–49. Quartzitic sandstone, thick-bedded blocky jointed, yellowish-gray, pinkish-gray and very pale orange weathering yellowish gray and yellowish orange; medium to coarse, with mixture of silica and carbonate cement; rounded. Forms ledge	85
Largely covered. Few outcrops are sandstone like A–49; float is of similar sandstone, with a little silty limestone	150
Largely covered. Sparse outcrops and float suggest interval made up of light-colored arenaceous limestone and dolomitic limestone (perhaps ⅓), and calcareous sandstone (perhaps ⅔) in beds a few feet thick	148
A–48. Sandstone and limestone. Beds a few feet thick of sandstone like that of A–49 above alternating with limestone like that of A–47 and A–46 below. Sandstone ledge 9 ft thick at top. This is gradation zone between Quadrant and Amsden formations	24
Total adjusted thickness	375–400
Amsden and Big Snowy formations:	
[NOTE.—The Amsden-Big Snowy contact is much in doubt here; perhaps it is best chosen in or at base of unit A–41]	
A–47. Limestone, olive-gray and light-olive-gray, fine to medium crystalline, in beds 5-10 ft thick, with interbeds 1–3 ft thick of yellowish-gray medium crystalline dolomitic limestone; forms low ledge	36
A–46. Sandstone and limestone. Mostly thick-bedded grayish-orange calcareous fine to medium subrounded sandstone, with thin interbeds of pale-olive medium crystalline arenaceous dolomitic limestone, pale-brown medium subangular calcareous chloritic sandstone, and medium-gray microcrystalline limestone. At base, 7 ft of fetid medium crystalline yellowish-gray limestone	41
A–45. Sandstone, pale-reddish-brown, thin-bedded, fine calcareous subangular. Near middle of interval, 2 ft bed of medium-light-gray, medium crystalline fetid limestone, and 7 ft dioritic sill (nearby, this sill thickens to as much as 20 ft); sandstone is hornfelsed near sill (excluding sill)	58
Covered. Float is largely yellowish-orange and grayish-yellow siltstone, with impersistent reddish-brown soil at top (red color here may be due to Quaternary weathering rather than to originally red rocks in this interval)	28
A–44. Limestone. Like A–43. Ledge former	17
Poorly exposed. Mostly thin-bedded silty and arenaceous limestone, in yellowish and greenish tones, interbedded with dark-gray limestone	24
A–43. Limestone, very thick bedded dark-gray fetid microcrystalline; weathers yellowish gray; many nodules and lentils of olive-black chert; locally rich in giant cup corals and biohermal debris; few interbeds of olive silty limestone. Ledge former	18
A–42. Limestone, dolomite, and mudstone. Thin-bedded, medium-gray nodular limestone and pale-red slabby finely crystalline dolomite interbedded with lenses of grayish-red-purple calcareous mudstone and fine sandstone. Dolomite layers weather to reddish-brown soil. Dioritic sill 3 ft thick at 15 ft below top (excluding sill)	70
A–41. Sandstone, mudstone, limestone, and dolomite. Alternations of thin beds of mottled grayish-yellow, yellowish-gray, grayish-orange-pink mudstone, yellowish-gray medium crystalline limestone, moderate orange-pink finely crystalline dolomite, and blocky mottled pale-red, grayish-orange-pink, very pale orange fine subangular calcareous sandstone	233
Covered in section. Good outcrops in this interval 300 ft to east show it to be much like A–42 above	147
A–40. Limestone, olive-gray, weathering light gray, with films of grayish-orange chert and a few nodules of olive-gray chert; one ledge-forming bed	4
Covered. Mostly thin-bedded light-gray finely crystalline dolomite and grayish-yellow arenaceous medium crystalline dolomitic limestone, with some grayish-red fine subangular sandstone, judging from float	79
A–39. Sandstone, thin-bedded, yellowish-gray, medium- to coarse subrounded, calcareous near base grading upward into fine subangular reddish-brown silty sandstone and greenish-gray medium subrounded quartzitic sandstone with a few dark chert grains; 3 ft of medium-grained medium-gray arenaceous limestone at top	19
A–38. Poorly exposed. Thinly interbedded light-colored calcareous mudstone, flaggy dolomite, and banded dolomitic limestone	7

Unit thickness (feet)

Amsden and Big Snowy formations—Continued

A–37. Sandstone, thick-bedded, pale-yellowish-gray, medium subrounded, calcareous, with a few dark chert grains 9

A–36. Dolomite and siltstone. Thin-bedded, light-olive-gray finely crystalline dolomite at base (19 ft) and thin-bedded mottled grayish-orange and yellowish-gray finely crystalline dolomite at top (4 ft) with largely covered interval between of dolomite interbedded with grayish-yellow siltstone (48 ft) 71

Total adjusted thickness 800

Mission Canyon limestone:

Largely covered. Base of formation forms massive ledge (A–34) 50–100 ft thick which caps cliff in Lodgepole formation. Bulk of formation underlies long grassy dip slope broken by a few low, much weathered ridges. Basal ledge, top ledge (A–35), and rare outcrops are of thick-bedded dark-gray to medium-gray microcrystalline limestone weathering yellowish gray to light olive gray, locally containing many dark chert nodules; fairly common are lenses of coarsely crystalline olive-gray limestone rich in fragments of crinoid stems, brachiopods, cup corals, and bryozoans. Thickness estimated .... ±900

Lodgepole limestone:

Largely covered. Ledges, 20–30 ft thick near base (A–32) and about 450 ft above base (A–33) are mostly of thin-bedded dark-gray to medium-gray microcrystalline limestone, weathering yellowish gray to light olive gray; a few beds are of coarsely crystalline olive-gray limestone containing many fragments of crinoid stems, brachiopods, and cup corals. Covered parts, as indicated by float and by exposures on strike within a few thousand feet of this section, are of similar dark thin-bedded limestone, with partings of yellowish-brown and grayish-orange calcareous shale. In the lower 200 ft, shaly partings are numerous and comparatively thin; they become fewer and thicker upward. Thick-bedded sequences appear in the upper 200 ft, and ultimately the formation grades into the thick-bedded Mission Canyon limestone. Fossiliferous beds are especially rich and abundant in the transition zone. Total thickness .... ±600

Three Forks shale:

[NOTE.—This thickness, though measured on well-exposed beds without visible structural complications, is great for the formation in this vicinity: 6 sections measured on the map at 1-mile intervals east and west of section A ranged from 200 to 350 ft. A good average thickness for this vicinity is 300 ft. The great measured thickness here may be due to local thickening of the lower shales (A–29 and A–30) or to near-bedding faulting; because the measured Jefferson here also seems too thick, duplication by unrecognized faults is the more reasonable explanation]

Three Forks shale—Continued

A–31. Sandstone and siltstone, thin-bedded, locally flaggy; pale-yellowish-orange, dark-yellowish-orange and grayish-orange; weathered rock darker and more brown; subrounded quartz grains and angular orange-stained calcite grains with a little clay in a calcite matrix. Near base, exposures poor, but sequence includes at least 3 beds 1–3 ft. thick of olive-gray clay shale, separated by a few feet of orange siltstone, as well as sporadic lenses of grayish-red, dark-reddish-brown, and grayish-yellow fossiliferous limestone; upsection, the beds are thicker, the clay content decreases, the quartz content increases, the average grain size increases 44

Covered. In basal 24 ft. a few exposures of olive-gray clay shale and masses of orange argillaceous limestone breccia like the basal breccia (A–28); rest of interval seems to be largely shale like A–29, with thin limestone beds like those near the base of A–31; top 10–20 ft. may include much orange siltstone like that of A–31 78

A–30. Limestone, irregularly bedded mottled argillaceous finely crystalline, yellowish-gray and olive-gray weathering pale yellowish brown and pinkish gray, fossiliferous 4

Covered (creek bed). Probably largely shale like A–29 81

A–29. Shale, thick-bedded, olive-gray, weathering grayish yellow to dusky yellow; many nodules of iron oxides and fossils replaced by iron oxides (fossil loc. A–29 from upper 30 ft.) 154

A–28½. Shale (4/5) thinly interbedded with siltstone (1/5) dark-yellowish-orange, calcareous; much float, but no outcrop, of light-gray microcrystalline limestone 28

A–28. Limestone breccia, yellowish and grayish-orange, weathering pale yellowish orange, thinly interbedded with grayish-orange medium crystalline silty limestone, weathering grayish orange pink 17

Total adjusted thickness 400

Jefferson dolomite:

A–27. Poorly exposed. Dolomite in alternating dark and light sequences 20–40 ft thick. Light sequences grayish yellow, fine to medium crystalline, with a few layers of limestone; dark sequences pale yellowish-brown medium crystalline, fetid. Grayish-yellow beds at top 180

A–26. Dolomite, pale-yellowish-brown, finely crystalline mostly thick-bedded; upper part has indistinct bedding planes a few inches apart; fetid; few thin lighter colored limestone beds, especially near top. Some dolomite beds finely streaked with lime-

	Unit thickness (feet)
Jefferson dolomite—Continued	
stone. Ledge former. Note: Some beds recorded in A–26, A–27 may be duplicated by near-strike fault	246
Covered	68
A–25. Dolomitic limestone and limestone, thick-bedded, light-brownish-gray to light-olive-gray, microcrystalline	27
A–24. Dolomite, with minor limestone, like A–22	52
A–23. Limestone and dolomite. Alternating beds, a few feet thick of several varieties of limestone (⅔ of interval) and dolomite (⅓). Limestone mostly light-gray microcrystalline and yellowish-gray medium crystalline. Dolomite medium crystalline, light-brownish-gray with streaks of yellowish-gray limestone	24
A–22. Dolomite, medium-bedded, light-brownish-gray to brownish-gray, medium crystalline, weathering yellowish brown. Few beds slightly lighter colored dolomitic limestone and limestone	23
A–21. Dolomite, thick-bedded, pale-yellowish-brown and dark-yellowish-brown, weathering pale-yellowish-brown; ledge former; medium crystalline, fetid. Few lenses, as much as 3 ft thick, of stromatoporoidal pale-yellowish-brown dolomitic limestone, near base	47
Total adjusted thickness	650
Maywood formation:	
Largely covered. Few outcrops of pale-yellowish-brown medium crystalline locally fetid limestone	13
A–20. Dolomitic limestone, thin-bedded, light-olive-gray and grayish-orange, interbedded with grayish-yellow flaggy calcareous siltstone. Sequence includes a few layers of pale-reddish-brown and moderate-reddish-orange calcareous siltstone	36
A–19. Largely covered. Probably much like A–20 above. At 13 ft above base, 2 ft outcrop of thin-bedded grayish-orange and pale-reddish-brown medium crystalline silty limestone	48
Total adjusted thickness	100
Pilgrim limestone:	
A–18. Dolomite and subordinate limestone, thick-bedded; ledge former. Medium-light-gray microcrystalline limestone at base, grading in less than 50 ft into yellowish-gray medium crystalline dolomite. Lower part of dolomite sequence weathers mottled yellowish gray, with the slightly darker patches coarser grained and less resistant than the lighter	226
Pilgrim limestone—Continued	
A–17. Largely covered. Small outcrops and float indicate interval is mostly banded gray limestone and orange dolomite, like A–16, but with some dark-yellowish-orange calcite sandstone	58
A–16. Limestone and dolomite, thin-bedded, with wavy bedding surfaces. Layers of medium-light-gray limestone, finely crystalline, ½–3 in. thick, alternating with silty medium-grained grayish-orange dolomite bands, less than ¼ in. thick; rarely, dolomite is pale reddish brown	70
A–16. Dolomite, thick-bedded mottled, medium-grained, yellowish-gray. Sporadic thin beds of light-olive-gray oolitic limestone, with fragments of trilobites and straight cephalopods	91
Total adjusted thickness	450
Park shale:	
A–15. Largely covered. Scattered outcrops are of dark-greenish-gray to pale-olive clay shale, with a few thin beds of fine sandstone near base	165
Total adjusted thickness	150
Meagher limestone:	
[NOTE.—Much of the formation has mild petroliferous odor]	
A–14. Limestone, medium-gray, finely crystalline, thick- to thin-bedded, with few lenses, less than 6 in. thick, of oolitic limestone and limestone-pebble conglomerate. Ledge former	40
A–14. Limestone, massive-looking but laminated where weathered, finely crystalline, mottled medium dark gray and grayish orange, with sporadic sequences a few feet thick of thin bands of modular medium-dark-gray fine-grained limestone alternating with bands of silty medium-grained grayish-orange limestone	40
A–13. Largely covered, but with poor outcrops of banded gray and orange limestone like upper 20 ft of A–12	46
A–12. Limestone, mottled; medium-dark-gray fine-grained limestone grading irregularly into coarser silty grayish-orange limestone. Ledge former	20
A–12. Limestone much like A–10, but with mottling slightly more evident, laminae thinner, slightly dolomitic in middle of sequence	137
A–11. Mostly covered. Sporadic outcrops of thin beds of calcite sandstone and siltstone, dark-yellowish-orange, alternating with medium-gray finely crystalline dolomite. (Concealed beds probably like A–10)	51

	Unit thickness (feet)
Meagher limestone—Continued	
A–10. Limestone, olive-gray to olive-black, microcrystalline in very thick beds; ledge former. Weathers light olive gray, showing thin crinkled laminae, and faint mottling in upper 100 ft. Scattered dark-yellowish-orange chert concretions and trilobite fragments. Ledge maker	136
A–9. Limestone, microcrystalline, dark-gray, faintly banded with darker and lighter layers 1–4 in. thick. In lower half, a few bands of silty moderate-olive-brown limestone, weathering yellowish gray	22
Total adjusted thickness	500
Wolsey shale:	
Covered (upper 25–75 ft may include beds gradational between Wolsey and Meagher lithologies. Rest of interval probably like A–8)	192
A–8. Clay shale, slightly micaceous; olive-gray weathering pale olive; few thin nodular limestone beds	44
A–7. Shale (2/3) and limestone (1/3). Beds as much as 1 ft thick of medium crystalline, hard light-olive-gray and greenish-gray limestone, weathering pale brown to dark yellowish brown, separated by poorly exposed beds, a few inches to 3 ft thick, of olive-gray to grayish-olive micaceous arenaceous shale	42
A–6. Shale (4/5) and sandstone (1/5). Alternating beds 1–4 in. thick of shale, grayish-olive with many worm casts on bedding surfaces, micaceous arenaceous calcareous, and pale-olive to grayish-olive sandstone, fine glauconitic micaceous calcareous subangular, weathering dark yellowish brown. Grades upward into A–7	28
Covered (probably much like A–5)	27
A–5. Argillaceous limestone, medium-gray, speckled with light olive, weathering medium gray with grayish-orange splotches; nodular thin-bedded, finely crystalline. A few beds are coarsely crystalline, pale yellow brown, with many trilobite fragments. A few shale partings, weathering grayish orange	13
Covered (probably 45 ft shale like A–3 overlain by 13 ft limestone like A–5)	58
A–4. Clay shale, fissile, pale-green to grayish-green, with lentils of moderate-yellowish-brown sandstone	7
A–3. Sandstone, thin-bedded, grayish-red to moderate-grayish-orange, coarse, quartzitic, subrounded	14

	Unit thickness (feet)
Wolsey shale—Continued	
A–2. Clay shale, fissile, grayish red, with streaks of grayish-green shale; interval divided into 3 equal parts by 2 beds, each 1 ft thick, of pale-reddish-brown glauconitic sandstone	14
Total adjusted thickness	400
Flathead sandstone:	
A–1. Quartzitic sandstone, very thinly crossbedded, in part flaggy, varicolored moderate-orange pink, grayish-pink, pale-reddish-brown, and grayish-red, coarse to medium; rounded; nearly pure quartz sand, cemented with quartz and tinted with iron oxides. Includes a few conglomerate lentils. Forms ledge	9
A–0. Sandstone, much like unit above but coarser grained and less well cemented, pale-reddish-brown and grayish-pink	26
Total adjusted thickness	35
North Boulder formation:	
Sandstone, grayish-red to dark-reddish-brown, coarse, subangular, micaceous arkose, with grayish-olive micaceous siltstone laminae; many lenses of gray pebble-to-boulder conglomerate in lower part. Measured only on map	>4, 000

SECTION B

Above Lombard thrust, northeast part of quadrangle, in drainage of Mud Spring Gulch. Maywood formation through most of Ellis formation. Formations from the Maywood through the Lodgepole are represented by a continuous section which begins in the NE¼NE¼ sec. 1, T. 2 N., R. 1 E., on the east side of 4,889-foot crest. The beds here strike N. 10° E. and dip 27° W. Due to folding and intrusion there is no full section of the Mission Canyon in this sector. Big Snowy and Amsden formations are measured in SE¼NE¼ sec. 32, T. 3 N., R. 1 E., beginning on the north side of the only prominent northwest-southeast depression in this quarter-section. The strike is N. 25° W., and the dip is vertical.

Because of folding and poor exposure, the Quadrant formation was not measured.

The Phosphoria and Ellis formations were measured in SE¼ sec. 20, T. 3 N., R. 1 E., beginning in the small saddle southwest of hill 4720. The strike averages N. 35° E.; dip 10° W.

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	Unit thickness (feet)
Quaternary gravel.	
Ellis formation:	
B-33. Poorly exposed limestone and siltstone, thinly interbedded; varicolored fine- and microcrystalline clayey limestone, with subordinate calcareous siltstone that rarely crops out. At top, 5 ft of dark-greenish-gray mudstone	224
B-32. Mostly covered. Scanty outcrops are of hard iron-stained sandstone, rich in chert	32
B-31. Limestone, yellowish-gray, sandy, clastic; many pelecypod fragments	21
Covered, except for small exposure of chert pebble conglomerate 8 ft below top	31
B-30. Limestone, thick-bedded, olive-gray, finely crystalline, with scattered chert pebbles	30
Total adjusted thickness	>350
Phosphoria formation:	
Covered	8
B-29. Quartzite and quartzitic sandstone, thick-bedded, yellowish-gray; a few thin interbeds of olive chert	52
B-28. Chert, thin-bedded, olive-gray	51
B-27. Quartzite, yellowish-gray, one prominent bed	4
Covered saddle. Slope below known Phosphoria formation covered with chert and calcareous siltstone float, probably from overlying beds, but also has phosphatic sandstone float of Phosphoria type not present in outcrop above. (Opposite slope of saddle has only quartzitic sandstone float and is assigned to Quadrant formation)	60
Total adjusted thickness	150
Quadrant formation (individual units within formation not measured):	
Quartzitic sandstone, thick-bedded, yellowish-gray and dark-yellowish-orange, fine grained	>80
Amsden formation:	
B-26. Limestone and dolomite, thick-bedded, pale-olive and medium-gray, microcrystalline; forms ledge	108
B-25. Limestone, thick-bedded, medium-dark-gray, microcrystalline; fetid, with many chert nodules	82
B-24. Partly covered. Much reddish-brown soil at top; scattered exposures of reddish-brown and grayish-yellow clayey limestone and calcareous siltstone	98
Total adjusted thickness	300
Big Snowy formation:	
B-23. Largely covered. Two 14-ft covered intervals separated and bounded by 1-2 ft limestone ledges. Limestone is in single beds, light olive, microcrystalline, fetid; many dark chert nodules and fossil fragments	31
Big Snowy formation—Continued	
B-22. Shale, fissile, grayish-olive	5
Covered. Probably shale like B-22	18
B-21. Limestone, thick-bedded, dark-gray, microcrystalline, fetid, fossiliferous	13
Covered. Probably shale like B-22	52
Total adjusted thickness	100
Mission Canyon limestone, not measured.	
Lodgepole limestone:	
Largely covered. Ledge at base of section and scattered outcrops above are of thick-bedded dark-gray microcrystalline limestone with a few silty partings	709
Total adjusted thickness	700
Three Forks shale:	
[NOTE.—Details of Three Forks shale from NE¼SW¼ sec. 1, about 3,200 ft southwest of main measured section]	
B-20. Siltstone. Like B-19 but thick-bedded and flaggy. (Fossil loc. 252i, 35-40 ft below top)	69
B-19. Siltstone, thin-bedded, pebbly, ferruginous and calcareous, dark-yellowish-brown to dusky-brown, locally petroliferous. (Fossil loc. 252h.) Pale-yellowish-brown clayey limestone bed at base	6
Covered. Float is mostly black shale	23
B-18. Shale, clay, olive-gray, with scattered lentils of dark limestone. (Fossil loc. 252g from top of interval; 252f from 10 ft below top)	45
B-17. Poorly exposed. Scattered outcrops are grayish-black to greenish-black fissile clay shale, with lenses of dark-reddish-brown limestone coquina. (Fossil loc. 252e from 2-5 ft above base; 252d from basal 2 feet)	20
B-16. Shale, clay, olive-gray to brownish-gray, with a few lenses of dark-reddish-brown limestone coquina. (Fossil loc. 252c from 10-20 ft below top; 252b from basal 5 ft)	32
B-15. Largely covered. Scattered exposures are mostly of very thin-bedded yellowish-gray calcareous siltstone and olive-gray clay shale, with lenses 1-4 in. thick of fossiliferous dark-reddish-brown to pale-yellowish-brown calcareous siltstone. At base, a few thin crumpled layers of yellowish-gray to very pale orange medium crystalline clayey dolomitic limestone. (Fossil loc. 252a from 90 ft above base.)	127
Total adjusted thickness	300
Jefferson dolomite:	
B-14. Poorly exposed. Long dip slope with a few steplike ledges of thick-bedded yellowish-gray medium crystalline dolomite and dolomitic limestone	30

	Unit thickness (feet)
Jefferson dolomite—Continued	
B–13. Dolomite and dolomitic limestone. About ⅔ yellowish-gray and yellowish-brown dolomite, and ⅓ yellowish-gray limestone or dolomitic limestone	260
B–12. Dolomite, thick-bedded, dark-yellowish-brown, coarsely crystalline	66
Covered. Probably much like B–12	21
Total adjusted thickness	400
Maywood formation:	
B–11a. Poorly exposed. Scattered outcrops are of light-olive-gray and yellowish-gray thick-bedded medium crystalline limestone. Covered intervals probably underlain by yellowish-gray calcareous siltstone	32
B–11b. Largely covered. 2-ft beds at top and bottom of yellowish-gray medium crystalline limestone. Covered remainder probably underlain by yellow siltstone	26
B–11c. Grayish-yellow calcareous siltstone	11
Total adjusted thickness	70

Pilgrim limestone: Dolomite, thick-bedded, yellowish-gray.

SECTION C

Below Lombard thrust, northeast corner of quadrangle. From Lodgepole formation through Kootenai formation. Measurement is essentially along the east wall of a canyon which begins at the west edge of the SW¼SW¼ sec. 29, T. 3 N., R. 2 E., and trends SE into the NE¼NW¼ sec. 32, T. 3 N., R. 2 E. Strike averages N. 30° E. All the strata are overturned, with dips ranging from 40° N. to 80° N. Because the rocks are overturned and with variable dip, it is obvious that thickness measurements are not to be taken very seriously. Formations between the Lodgepole and the Pilgrim crop out in this sector but not well enough for satisfactory measurement.

	Unit thickness (feet)
Covered saddle (near top of Kootenai formation)	>100
Kootenai formation:	
C–27. Sandstone and siltstone. Many ledges 1–2 ft thick of several sandstone varieties, separated by thick beds of siltstone. Sandstone types include hard light-gray quartzitic, soft-grayish-yellow calcareous, and dark-greenish-gray arkosic. Many worm casts on siltstone bedding surfaces	>140
C–26. Limestone, thick-bedded, mottled yellowish-orange and yellowish-gray, coarsely crystalline, with abundant gastropods; forms ledge	7
Covered. Distinctive reddish-brown soil in upper 14 ft; interval probably underlain by siltstone	72
Kootenai formation—Continued	
C–25. Quartzitic sandstone, thick-bedded, medium; yellowish-gray, forms ledge	12
C–24. Partly covered. Mostly reddish-brown and dark-yellowish-orange mudstone with scattered outcrops of dark sandstone and limestone in beds 1–3 ft thick	198
C–23. Quartzitic sandstone, thick-bedded, yellowish-brown, with many chert fragments and lenses of pebble conglomerate	43
Covered. Interval probably underlain by reddish-brown mudstone	53
C–22. Sandstone, thick-bedded, yellowish-brown, with much chert	47
C–21. Largely covered. Scattered outcrops of reddish-brown, olive-gray, and light-gray mudstone and fine sandstone	180
C–20. Sandstone, thick-bedded, locally cross-stratified, yellowish-gray; medium, much chert; calcareous cement; ledge former	32
Total adjusted thickness	>600
Morrison formation:	
Covered. Reddish-brown and yellowish-brown soil and siltstone float	132
Total adjusted thickness	130
Ellis formation:	
Largely covered. Lacks red detritus and soil, typical of adjoining covered Morrison interval; probably underlain by yellowish-gray siltstone and limestone	74
C–19. Conglomerate. Rounded chert pebbles, dark-yellowish-brown	26
C–18. Limestone, thick-bedded, yellowish-gray, dense microcrystalline to medium crystalline with scattered pebbles; ledge former	20
C–17. Breccia zone. Uncemented angular masses of chert and quartzite. (Probably represents a weathered pre-Ellis surface.)	26
Total adjusted thickness	150
Phosphoria formation:	
C–16. Quartzitic sandstone, thick-bedded, yellowish-gray, with a few thin beds of dark chert	54
C–15. Partly covered. Scattered outcrops are of yellowish-gray medium crystalline limestone. Covered interval probably siltstone and chert	37
C–14. Chert, light-olive, thick-bedded, forms ledge	49
C–13. Quartzite, thick-bedded, fine-grained, pale-yellowish-brown	10
Largely covered. Scattered outcrops are of yellowish-gray medium crystalline limestone and dark chert	46
C–12. Limestone, thick-bedded, yellowish-gray medium crystalline, with large irregular lentils of dark chert and quartzite	15
Total adjusted thickness	200

		Unit thickness (feet)
Quadrant formation:		
	Brecciated and sheared sandstone and quartzite, like C–16 above	30
C–11.	Quartzitic sandstone, thick-bedded, white to yellowish-gray, medium-grained; includes a few grayish-yellow limestone beds 1–3 ft thick	100
	Total adjusted thickness	150
Amsden formation:		
C–10.	Siltstone and limestone. Thin-bedded, yellowish-gray siltstone and some fine sandstone with calcareous cement, interbedded with subordinate olive-gray microcrystalline limestone	114
C–9.	Limestone, thick-bedded, light-olive-gray, with a few thin beds of moderate yellowish orange siltstone	27
C–9a.	Siltstone and limestone like C–10, but proportions of two types about even. At base 8 ft blackish-red microcrystalline coralline limestone	51
C–8b.	Mudstone, thick-bedded varicolored, mostly brownish-red and dark-bluish-gray	18
C–8a.	Sandstone and pebble conglomerate, thick-bedded, blackish red	17
	Total adjusted thickness	250
Big Snowy formation:		
C–7c.	Quartzitic sandstone, yellowish-gray	3
C–7b.	Shale, dark-greenish-gray	3
	Largely covered. Float suggests main rock type is yellowish-gray calcareous siltstone; some dark-gray shale exposed by digging	23
C–7a.	Limestone, thick-bedded, light-olive-gray, microcrystalline, with a few thin silty beds; medium crystalline near base; forms ledge	42
C–6.	Sandstone and limestone. Alternating thick beds of yellowish-brown calcareous sandstone and yellowish-brown fragmental sandy limestone	31
	Largely covered. Probably dark shale, except for 12-ft bed near middle of yellowish-brown fragmental limestone like that in C–6	116
C–5.	Limestone, thin-bedded, olive gray, with mottled weathering surfaces in yellowish-gray and yellowish-orange. Fossiliferous, especially in basal 3 ft	17
C–4.	Poorly exposed. Scattered outcrops of black shale	8
	Covered. Probably same as C–3	82
C–3.	Siltstone and mudstone, thin-bedded, yellowish-brown, calcareous	91
	Total adjusted thickness	400
Mission Canyon limestone:		
C–2.	Conglomerate and breccia, subrounded to angular fragments of limestone	4
Mission Canyon limestone—Continued		
C–1.	Limestone, very thick-bedded, dark-gray and grayish-olive microcrystalline to fine crystalline; forms monolithic outcrop cut by many small faults subparallel to bedding. Many thin zones of solution breccia	950
	Total adjusted thickness	950
Lodgepole formation:		
C–0.	Limestone, thin- to thick-bedded; dark-gray microcrystalline, with a few yellowish-brown silty partings near base	295
	Total adjusted thickness	300

SECTION D

Below the thrust zone in southwest corner of the quadrangle. From Precambrian hematitic phyllonite and gneiss to base of Ellis formation. From the crystalline rocks to the base of the Meagher formation the measured section is in sec. 8, T. 1 S., R. 1 W., partly in the Jefferson Island quadrangle. The section begins 15 ft northeast of the large north-flowing creek in E½ of the section and roughly follows the creek. The beds here strike N. 65° W. and dip 36° N.

To avoid structural complications, the section beginning with the basal Meagher shifts about 2,000 ft east to the NW¼NW¼ sec. 9, T. 1 S., R. 1 W., Three Forks quadrangle, beginning in the bed of the only west-flowing creek in this part of sec. 9. From this point to the top of the Jefferson dolomite the section is continuous and the strike and dip throughout are close to N. 85° E., 25° N. At the top of the Jefferson, there is an offset of 300 feet east to take advantage of better exposures in the Three Forks shale. The overlying Lodgepole limestone was also measured on this offset.

This part of the section ends in the SW¼NW¼ sec. 4, T. 1 S., R. 1 W., at the top of the Lodgepole. The Mission Canyon limestone was not measured with the planetable, but estimated from map measurements.

The remainder of the section from the top of the Mission Canyon to the base of the Ellis was measured in NW¼SE¼ sec. 34, T. 1 N., R. 1 W., about 1¼ miles northeast of the end of the preceding part of the section. This segment begins on the southwest side of the only prominent saddle in center SE¼ sec. 34 and continues almost to the Jefferson River flood plain. The strike is N. 55° W., dips are vertical. Exposures of the Phosphoria formation in this interval are poor and have been supplemented by lithologic details from measurements 4,500 ft to the east-southeast in the SW¼SE¼ sec. 35.

	Unit thickness (feet)
Quaternary alluvium and colluvium.	
Ellis formation:	
D-61. Chert-pebble conglomerate	3
Total adjusted thickness	>3
Phosphoria formation:	
D-60. Quartzitic sandstone, thick-bedded, slightly calcareous, grayish-yellow at base grading to pale-yellowish-brown at top	19
D-59. Siltstone, pale-yellowish-orange, chippy; with a little pale-yellowish-brown chert, olive-gray chert, and phosphatic sandstone like that in D-58. (Firm outcrops of siltstone are not present, but the float is so abundant and the lithology so unique that the siltstone must be in place.)	59
D-58. Phosphatic sandstone and chert. Three beds of medium to fine phosphatic sandstone, medium-gray to olive-gray, averaging 3 ft thick, separated by pale-yellowish-brown chert beds, 2 ft thick. Phosphate is in scattered ovoid and irregularly shaped grains	14
D-57. Chert. Like D-54	15
D-56. Sandstone, grayish-yellow, calcareous cement; with many rounded shale pebbles	2
Covered	22
D-55. Calcareous siltstone and dolomite, thinly interbedded, pale-yellowish-orange to very pale orange, flaggy	3
D-54. Chert, light-olive-gray to yellowish-gray, in one bed	3
Total adjusted thickness	150
Quadrant formation:	
D-53. Quartzitic sandstone, thin- to thick-bedded, yellowish-gray, medium; faintly cross-stratified locally. Main cement is quartz, but most beds also have a little calcareous cement. Includes a few thin beds of white to yellowish-gray limestone	157
D-52. Dolomitic limestone, light-olive-gray and yellowish-gray, in 3 bands, each 5–10 ft thick, separated by two covered intervals 6 ft. thick probably underlain by quartzitic sandstone	38
Total adjusted thickness	200
Amsden formation:	
Covered (except for 1-ft bed of olive dolomite in middle). Probably underlain mostly by thin-bedded siltstone and quartzite	37
D-51. Dolomite and dolomitic limestone, thick-bedded, yellowish-gray. A few thin beds of very pale orange calcareous siltstone	17
D-50. Limestone and dolomitic limestone, thin- to thick-bedded, grayish-yellow; includes a few thin beds of pale-red calcareous siltstone and of brownish-gray very fine quartzite; forms low ledge	64
Amsden formation—Continued	
D-49. Calcareous siltstone, thin-bedded, grayish-orange with a few pale-red beds. Upper part contains a few brachiopods	31
D-48. Limestone, laminated, light-olive-gray, medium crystalline; one bed forming low ledge	3
Covered. Float is yellowish-brown, pale-yellowish-orange and grayish-orange calcareous siltstone, and probably represents composition of interval	21
D-47. Limestone, light-olive-gray to light-gray, weathering slightly darker, locally pinkish; one ledge-forming bed	4
Covered. North side of saddle covered by moderate orangish-pink soil to base of next unit	55
Covered. South side of saddle covered by float of pinkish-gray quartzite and yellowish-gray and grayish-orange siltstone	170
D-46. Quartzitic and calcareous sandstone, thick-bedded, pinkish-gray, dark-yellowish-orange and pale-yellowish-brown, medium; cement alternately quartz and calcite. The quartz-cemented beds contain ovoid mottles	79
D-45. Quartzitic sandstone, thick-bedded yellowish-orange, medium, with a few pebbles of yellowish-gray siltstone	8
D-44. Calcareous siltstone, thick-bedded, pale-reddish-brown and light-brown, locally mottled to yellowish orange. A few thin discontinuous bands of nearly black hematite and silica	6
Total adjusted thickness	500
Big Snowy formation:	
Covered	9
D-43. Largely covered. Scattered outcrops of dolomite, light-olive gray to pale-yellowish-brown, flaggy. Unexposed beds probably shale	58
Total adjusted thickness	60
Mission Canyon limestone:	
D-42. Limestone, light-olive-gray, thick-bedded, fine crystalline	20
D-41. Limestone, thick-bedded, dark-gray, microcrystalline with a few yellowish-gray silty partings	134
(Remainder of Mission Canyon limestone not measured.)	
Total adjusted thickness	>150
Lodgepole limestone:	
D-40. Limestone, thin-bedded like D-39; abundant brachiopods and crinoid fragments	190
Covered. Probably thin-bedded limestone with many silty partings	218

	Unit thickness (feet)
Lodgepole limestone—Continued	
D–39. Limestone, thin- to thick-bedded; olive-gray, many layers contain coral and crinoid fragments; forms ledge	64
Total adjusted thickness	450
Three Forks shale:	
D–38. Calcareous siltstone and nodular limestone, thin-bedded. Beds progressively thicker, more sandy, and less clayey toward top; most beds pale orange to grayish orange; some limestone layers yellowish-gray, grayish-red and dark-reddish-brown; entire sequence fossiliferous, especially the red and brown beds	92
D–37. Clay shale, fissile, yellowish-gray to pale-olive-gray	19
D–36. Siltstone and limestone. Thinly interbedded grayish-orange calcareous siltstone and olive-gray limestone, with scattered zones of limestone breccia	8
D–35. Limestone. Lower half thick-bedded, grayish-orange, medium crystalline, porous, with zones of solution breccia containing angular fragments of dark microcrystalline limestone, grading into upper half, thick-bedded light-olive-gray microcrystalline limestone which forms ledge	19
D–34. Limestone and mudstone, thick-bedded, light-brown, grayish-orange, and dark-yellowish-orange, clayey limestone and calcareous mudstone with lentils of light-gray shale, intricately contorted in places; locally, is a breccia with fragments several inches long	54
Total adjusted thickness	200
Jefferson dolomite:	
[NOTE.—Practically every bed has strong fetid odor]	
D–33. Dolomite. Like D–32 but beds are thinner	26
D–32. Dolomite, thin-bedded, grayish-orange, coarsely crystalline	27
D–31. Dolomite, thin-bedded, yellowish-gray, medium crystalline	24
D–30. Dolomite, medium-light-gray 6-ft ledge at base, grading upward into beds 1–3 ft thick medium dark gray; medium crystalline	33
D–29. Dolomite. Alternating thick beds of pale-yellowish-brown and yellowish-gray, grades into D–30	41
D–28. Dolomite, like D–25	55
D–27. Poorly exposed. Outcrops are very pale orange dolomitic limestone and lenses of dolomite-pebble conglomerate as much as 4 ft long and 1 ft thick	49
D–26. Dolomite, thick-bedded, olive-gray, fine crystalline; forms ledge	22

	Unit thickness (feet)
Jefferson dolomite—Continued	
D–25. Dolomite, thick-bedded medium-gray to brownish-gray, medium crystalline. Scattered "spaghetti beds" consisting of dark crystalline dolomite surrounding closely packed irregular tubes, ½ in. to 2 in. long and averaging ⅛ in. thick, of earthy yellowish-gray limestone, presumably stromatoporoid zones. Includes a few thin beds of light-olive-gray dolomite and dolomite-pebble conglomerate	28
Covered	40
D–24. Dolomitic limestone, thick-bedded, pale-yellowish-brown at base grading rapidly upward to dark yellowish brown, medium crystalline. Contains many irregular pods and masses of black and yellow chert, mostly less than an inch thick, but nearby are single masses of chert as much as 10 ft thick and 40 ft long	18
Total adjusted thickness	350
Maywood formation:	
D–23. Calcareous siltstone, yellowish-gray, pale-yellowish-orange and dusky-yellow	3
Covered. Float is reddish and yellowish siltstone	9
Total adjusted thickness	10
Pilgrim limestone:	
D–22. Dolomite and dolomitic limestone, thick-bedded, yellowish-gray, dusky-yellow, and light-olive-gray with scattered orange-pink splotches; slightly porous	28
D–21. Dolomite, thick-bedded, mottled light-olive-gray to olive-gray, medium to coarse crystalline. Rock markedly porous and resembles sandstone. Slight fetid odor. Forms top of cliff	29
D–20. Dolomitic limestone and dolomite, thin-bedded; dolomite content varies rapidly both laterally and vertically. Least dolomitic material is yellowish-gray fine crystalline; most dolomitic is brownish gray and medium crystalline	32
D–19. Dolomitic limestone and dolomite, thin-bedded, light-brownish-gray, locally mottled with pale yellowish brown. Includes thin zones of conglomerate with flat subangular dolomite pebbles, and slabs as much as 4 ft long	21
D–18. Limestone and dolomite. Alternating laminae of gray limestone and orange dolomite with thin layers of dolomite pebble conglomerate like that in D–19	6
Covered. (Float is mostly dark-gray, coarsely crystalline, fetid dolomite, but interval probably largely underlain by banded dolomite and limestone like D–17 and D–18.)	60

	Unit thickness (feet)
Pilgrim limestone—Continued	
D–17. Limestone and dolomite. Thin layers of medium-gray, finely crystalline limestone alternating with generally thinner ones of pale-yellowish-brown clayey dolomite; some beds nodular. Many thin dolomite streaks within limestone beds. Many chert stringers and lentils. Includes a few layers of dolomite-pebble conglomerate, many with reddish dolomitic cement. Forms ledge	92
Covered. Probably thick-bedded dolomite like D–16	77
D–16. Dolomite, thick-bedded, light-gray to medium-gray, microcrystalline, with thin zones of oolites; lower 34 ft poorly exposed, upper 12 ft forms ledge with distinctly mottled surfaces due to slightly darker oolitic zones which grade into lighter crystalline material; rare laminae of yellowish-gray limestone; many films of yellowish-orange chert on bedding planes and in joints	46
Total adjusted thickness	400
Park shale:	
D–15. Shale, dark-greenish-gray to dusky-yellowish-green, weathering light greenish gray to pale olive, locally calcareous	73
Covered	30
Total adjusted thickness	100
Meagher limestone:	
D–14. Limestone like D–13, but beds somewhat thinner and with more orange clayey layers	93
Covered	22
D–13. Limestone, alternating thin beds of medium-gray crystalline limestone and grayish-orange clayey limestone	93
D–12. Limestone, thick-bedded, medium-gray with many fine grayish-orange silty streaks	130
Covered. Rare outcrops nearby and float indicate this interval largely medium-gray, thin-bedded limestone, mottled pale orange to grayish orange, with many films and lentils of grayish-orange chert	107
D–11. Limestone, thin-bedded, medium-dark-gray, nodular, grading into calcareous mudstone. Locally, dusky yellow to yellowish gray silty streaks. (Sequence is broken into units a few feet thick by 11 latite sills ranging in thickness from 4 in. to 3 ft; sills not included)	22
Total adjusted thickness	500

	Unit thickness (feet)
Wolsey shale:	
D–10. Covered. Probably mostly gray shale	16
D–9. Limestone, thin-bedded, light-olive-gray and grayish-orange, clayey, nodular. Orange strata fossiliferous	86
D–8. Shale, thick-bedded pale-brown to pale-yellowish-brown and light-olive-gray, micaceous; a few yellowish-brown limestone laminae. (Latite sill, 5 ft thick, near base not included)	91
D–7. Shale and mudstone. Like D–5 but lacking limestone and quartzite	64
D–6. Shale and mudstone. 7 ft grayish-red shale impregnated with iron oxide (D–6a); overlain by 11 ft pale-reddish-purple mudstone flecked with white calcite crystals (D–6b)	18
D–5. Mudstone and shale. Pale-yellowish-brown mudstone and medium-gray micaceous shale, slightly hornfelsed, with a few thin beds of ferruginous clayey limestone and pale-yellowish-brown glauconitic quartzite	81
(Monzonite sill about 200 ft thick.)	
D–4. Largely covered. Small exposures of thick-bedded, medium- to dark-gray hornfelsed shale and thin-bedded quartzitic sandstone like D–3	50
Total adjusted thickness	400
Flathead sandstone:	
D–3. Quartzitic sandstone, thick-bedded, varicolored grayish-orange, pink to pinkish-gray and yellowish-gray; largely medium, but coarse near top. Near top, a few thin laminae of yellowish-gray hornfelsed shale. Forms low ledge	38
Covered	6
D–2. Pebble conglomerate, subrounded	3
Total adjusted thickness	50
Precambrian crystalline rocks:	
D–1. Alternating hematitic phyllonite and gneiss.	

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