Geology of the Bonner Quadrangle Montana

GEOLOGICAL SURVEY BULLETIN 1111-F

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By WILLIS H. NELSON *and* JOSEPH P. DOBELL

CONTRIBUTIONS TO GENERAL GEOLOGY

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With special reference to the tratigraphy and structure of he rocks of the Belt series

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GEOLOGY OF THE BONNER QUADRANGLE, MONTANA

By WILLIS H. NELSON and JOSEPH P. DOBELL

ABSTRACT

The geology of the Bonner quadrangle in western Montana was mapped principally to study the Missoula group of the Belt series, and the structure of the area. The type localities of the formations comprising this group are all in or near the Bonner quadrangle. These formations are here redefined and in part renamed. Like the rest of the Belt series, the Missoula group has been regarded heretofore as of late Precambrian age but in our opinion the results of the present study suggest that its upper part may be of Cambrian age.

The oldest rocks exposed in the quadrangle belong to the Newland limestone, the local representative of the Piegan group. That group is characterized by the prevalence of calcite-rich rocks and is near the middle of the Belt series. Here clayey and sandy tan to gray limestone is the major component of the Newland limestone. Thin layers of very dark gray argillite are intercalated with the limestone; and noncarbonatic shale, sandy shale, and quartzite occur locally in units as much as 100 feet thick. The thickness is at least 4,000 feet.

The Piegan group is succeeded upward by the Missoula group, which occupies most of the Bonner quadrangle. The Missoula group has an aggregate thickness of about 14,300 feet, and as subdivided here includes, from the base upward, the Miller Peak quartzite (including the Hellgate quartzite member), Bonner quartzite, McNamara argillite, Garnet Range quartzite, and Pilcher quartzite. All components of the Belt series within the quadrangle appear to have gradational contacts. Shallow-water features, such as mud cracks and ripple marks, are plentiful in argillite beds in both the Piegan and Missoula groups.

The Miller Peak argillite is composed of red and green silty argillite and argillaceous quartzite, and locally a little pale-red and tan quartzite, and light-gray impure limestone. Some argillite units several hundred feet thick are of uniform color, but in others the colors alternate every few feet. Few argillite beds are free from clastic quartz which locally is sufficiently abundant to warrant naming the rocks clayey quartzite, or even quartzite. The quartzite forms isolated lenticular beds and lenses with some intercalated argillite. The Hellgate quartzite member, high in the formation, is one of the thickest of these lenses. It is about 1,200 feet thick, and has about 3,800 feet of argillite below it and 1,000 feet of argillite above.

The Bonner quartzite (here newly named) corresponds to the middle member of the McNamara formation as previously defined. The formation is charac-

terized by its tendency to break into blocks speckled with grains of white, translucent, weathered feldspar. Pink vitreous quartzite predominates; the quartzite is commonly crossbedded and intercalated with minor amounts of green and red argillite. The formation is about 1,500 feet thick.

The McNamara argillite, as here restricted, is composed largely of red and green metamorphosed clayey siltstone and minor amounts of argillite and quartzite. The colors are partly secondary. The formation is characterized by clay balls and chips; many of these clay balls are predominantly clay, but some are mainly very fine grained quartz, and others are mixtures. The formation is about 4,000 feet thick.

The Garnet Range quartzite consists of greenish-gray quartzite and dark-green and dark-gray argillite, shale, and sandy argillite; the formation weathers moderate to light brown. The color, abundant detrital mica, and intense internal deformation serve to distinguish the formation. The formation is about 1,800 feet thick.

The next succeeding formation is well exposed along Pilcher Creek, and is here named the Pilcher quartzite. Most of the formation consists of pale- to moderate-red crossbedded quartzite. The upper 10 to 20 feet of the formation contains red and green argillite and constitutes a transition zone between the Pilcher quartzite and the shale unit that overlies it. The Pilcher quartzite is from 700 to 1,800 feet thick.

The contact between the Missoula group and the overlying rocks may be either gradational or unconformable. The overlying rocks are probably of Middle Cambrian age; so that if the contact is gradational, as we believe is possible, the uppermost part of the Missoula group must be of Cambrian age. If the contact is an unconformity all the Belt series may be of Precambrian age, or there may have been ample time within the Cambrian period, before Middle Cambrian time, for an episode of erosion to have followed the deposition of some of the uppermost rocks of the Missoula group.

Above the Pilcher quartzite two lithologic units are recognized: a lower shale unit and an upper limestone unit. The shale unit is various shades of green, brown, red, and dark gray, and it includes some quartzite and limestone; the unit is estimated to be 200 feet thick. The limestone unit is dominantly gray and in part mottled with grayish-orange spots; some of the limestone is oolitic. One hydrozoan of probable early Paleozoic age has been found. The greatest thickness within the quadrangle is 300 to 400 feet; the top is not exposed. The shale and limestone units are thought to correspond lithologically to the Wolsey shale, Meagher limestone, and possibly Flathead quartzite, all of Middle Cambrian age of other areas in Montana.

Poorly indurated shale, sandstone, and conglomerate of continental origin are exposed along the west edge of the quadrangle, both north and south of Missoula. These beds contain plant fossils considered to be of Oligocene or early Miocene age.

Cirques, U-shaped valleys, and small moraines in the northern part of the quadrangle indicate glaciation during Pleistocene time. Larger glaciers were present a short distance north of the quadrangle. Shoreline features at many places below an altitude of 4,200 feet testify to the former glacial Lake Missoula.

Sills, dikes, and stocklike bodies of diabase of probable late Precambrian age constitute the only intrusive masses within the mapped area.

The quadrangle is near the intersection of two major tectonic lineaments: one, the Flathead-Bitterroot Valley line, trends northward just west of the quadrangle boundary; the other, represented locally by the Clark Fork fault, cuts the quadrangle diagonally. The Flathead-Bitterroot Valley line seems to separate areas of contrasting structural behavior. East of the line there is large-scale thrusting along northwestward-trending faults, but apparently no major strike-slip movement; to the west large thrusts have not been recognized but large-scale strike-slip displacements are known. The Clark Fork fault is interpreted to be a steep normal fault with the northeast side uplifted about 6,500 feet, and with little or no strike-slip displacement. South of the Clark Fork fault the structure is dominated by thrust faulting. The thrusting was toward the north, and the crustal shortening is believed to be at least 15 miles.

The Bonner Mountain anticline lies a short distance northeast of the Clark Fork fault. In the west half of the quadrangle this anticline is overturned to the northeast. Folding and overturning probably were in response to the same forces that formed the other thrust faulting and folding in the quadrangle.

Northeast of the Bonner Mountain anticline the rocks are sliced by several thrusts with northeasterly displacements. The faulting seems to have been preceded by folding, and there may have been several miles of crustal shortening in a zone now less than 2 miles wide.

Northeast of this zone of faulting a broad open syncline, the Wishard Creek syncline, involves rocks probably as young as Middle Cambrian in age. Most of the structures in the quadrangle are believed to have formed during the Laramide orogeny.

INTRODUCTION

PURPOSE AND SCOPE OF THE REPORT

This report records the results of a study of the Bonner quadrangle (fig. 19). The principal objectives were to study the stratigraphy and structure of the Missoula group of the Belt series. Sedimentary rocks younger than Cambrian occupy only a very small area and they were given only cursory attention.

The fieldwork was done during the summers of 1954 and 1955, when Dobell spent 24 weeks in the field and Nelson 31 weeks. Dobell is responsible for the section of the report on intrusive rocks; the remainder of the report was written by Nelson.

The geology was mapped on aerial photographs, scale about 1:20,000, and was later transferred to the final map by inspection. The topographic map on which the geology (pl. 35) is plotted was prepared by Nelson. Horizontal detail is from planimetric maps by the U.S. Forest Service at a scale of 2 inches to the mile from aerial photographs. Contour lines were partly sketched from aerial photograph stereoscopic models, and partly modified by the same means, from the contour lines on the Lolo National Forest Map, Montana (U.S. Forest Service, 1939), itself a modification of the Bonner quadrangle, Montana, map published by the U.S. Geological Survey in 1903. Relative errors of altitude of the resulting map probably never approach 400 feet and horizontal errors probably do not exceed 400 feet.

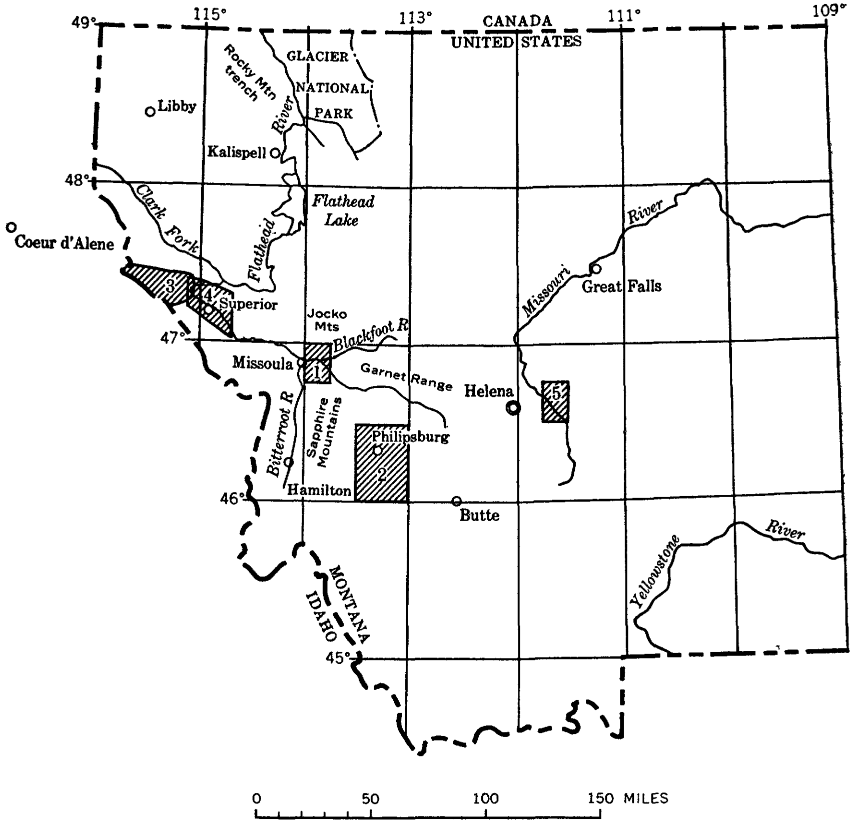

FIGURE 19.—Index map of western Montana showing relation of the Bonner quadrangle to regions discussed in previous reports. (1) Bonner quadrangle. (2) Philipsburg quadrangle; Calkins and Emmons, 1915. (3) St. Regis area; Wallace and Hosterman, 1956. (4) St. Regis-Superior area; Campbell, 1961. (5) Canyon Ferry quadrangle; Mertie, Fischer, and Hobbs, 1951.

TOPOGRAPHY

The Bonner area is in the Northern Rocky Mountains province of Fenneman (1931). It is almost completely mountainous, and, as is typical in the province, the mountains are not arranged in well-defined ranges. Three ranges that are more or less distinct outside the quadrangle meet here; within the quadrangle the ranges are arbitrarily separated by the larger stream valleys. The area south of the Clark Fork is the north end of the Sapphire Mountains. The mountains between the Clark Fork and the Blackfoot River are the west end of the Garnet Range. The mountains north and west of the Blackfoot River are part of the Jocko Mountains.

South of the Blackfoot River and Clark Fork the mountains have rounded crests, and are dissected by narrow steep-sided stream valleys. North of the Blackfoot River the lower crests are similarly rounded, but the higher mountains have been glaciated and some have sharp crests.

Little of the lower altitude part of the area is flat. The flat valley floor of the Clark Fork averages a little more than a mile wide, and the valley floors of the other large streams are somewhat narrower. A small area in the west edge of the quadrangle is part of an intermontane basin—the Missoula-Ninemile Valley of Pardee (1950, pl. 1). The city of Missoula is in the easternmost end of this valley.

Parts of the low flatlands are under cultivation; and lumbering and a little grazing are also carried on in the area.

BELT SERIES (PRECAMBRIAN)

PIEGAN GROUP

Newland limestone, Siyeh limestone, and Wallace limestone are three roughly synonymous terms used to designate a thick, characteristically calcite-bearing sequence of rocks about midway in the Belt series. C. P. Ross (1949) has redefined the Piegan group to include these three formations.

NEWLAND LIMESTONE

NAME

Current usage as reflected by the "Geologic Map of Montana" (Ross, Andrews, and Witkind, 1955) suggests that the term "Newland limestone" is the preferred designation for the rocks of the Piegan group in the Bonner quadrangle. C. M. Langton (1935) used "Newland limestone" in this area, and Calkins and Emmons (1915) used it in the Philipsburg quadrangle, which is about 22 miles southeast of the Bonner quadrangle. Clapp and Deiss (1931) used the term "Wallace limestone" within this area, but its present use is restricted to areas farther west. Wallace and Hosterman (1956) used the term "Wallace" in the vicinity of St. Regis, Mont., which is about 60 miles northwest of the Bonner quadrangle.

DISTRIBUTION

Small outcrop areas of the Newland limestone, bounded in part by faults, are found in the southwest corner of the quadrangle, along the south edge of the quadrangle near the middle, just north of Missoula, and just east of Bonner. A larger outcrop area extends northwestward from Rattlesnake Creek to the west edge of the quadrangle.

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CHARACTER

The dominant rock is banded or laminated argillaceous and sandy limestone. In addition, beds of shale, sandy shale, and impure quartzite that range from about an inch to a foot in thickness occur throughout the formation and make up perhaps 10 percent of it. Locally, sections from a few tens to perhaps 100 feet or more thick are made up of noncarbonatic argillaceous and quartzose rocks. At a few places there are beds of relatively pure, faintly bluish-gray limestone. Mud cracks are present, and some quartzite beds are faintly crossbedded. A few stromatolites were noted.

Most of the areas underlain by these rocks are mantled with soil and have few outcrops. Chips of gray argillite that contain very dark gray argillite laminae are the best criterion for recognizing areas underlain by the Newland. These chips have been largely leached of calcite and do not reflect the original composition of the bedrock. The absence of the red argillite so common in the other argillite formations is also useful in mapping this formation.

The laminated limestone is generally tan or gray argillaceous limestone interbedded with thinner layers of very dark gray argillite. (Argillite is used in this paper to mean a rock derived by low-grade metamorphism from a clay-rich sedimentary rock. Most of the clay-mineral fraction has been chemically rearranged and is now represented by newly formed chlorite and sericite. Slaty cleavage is weak or nonexistent, and bedding planes are the principal planar structures. The rock is mineralogically similar to slate, but lacks the cleavage of slate.) The very dark gray argillite forms layers as much as one-eighth inch thick. The tan or gray argillaceous limestone layers range from about 1/8 to 4 inches in thickness and average about 1 inch. The carbonate in these rocks is principally calcite, but locally the rocks contain a little dolomite. The weathered rind on the carbonate-bearing rocks and the froth produced by effervescence with acid is commonly brown which suggests a significant iron content, possibly originally in the carbonate.

THICKNESS

The base of the Newland limestone is not exposed, and the beds are so deformed that measurement is impossible, but the thickness can hardly be less than 4,000 feet.

CONTACT WITH THE MILLER PEAK ARGILLITE

The contact between the Newland limestone and the Miller Peak argillite is gradational. The lowest part of the Miller Peak is a transitional unit of green argillite that is not sharply separable from the rocks above or below. The green argillite grades downward into the limestone of the Newland and upward into the higher parts of the Miller Peak argillite.

CORRELATION

In the southwest corner of the quadrangle argillite and limestone crop out in several alternating bands. We believe this pattern is due to repetition by faulting of the Newland limestone–Miller Peak argillite sequence (see section *E–E'*, pl. 35). C. M. Langton (1935) on a small-scale map of this same area showed the same general areal distribution, but he interpreted it as representing four units: his Newland limestone, Spokane shale, Helena limestone, and Miller Peak argillite. Clapp and Deiss (1931) correlated the Spokane shale and Helena limestone of the Helena vicinity with the upper part of the Wallace limestone (Wallace limestone equals Newland limestone) of the Missoula area; but neither their correlation chart nor their stratigraphic descriptions show the Spokane shale and Helena limestone this far west.

MISSOULA GROUP

The Missoula group is the uppermost part of the Belt series. It is well exposed in the mountains east of Missoula, Mont., where Clapp and Deiss (1931) originally named and described it and its component formations. The type localities of all but one of their formations are within the Bonner quadrangle; the type locality of the McNamara argillite is about 4 miles east of the quadrangle, where it is well exposed. The original classification of Clapp and Deiss has been followed as far as possible, but some modifications seem necessary. The following table compares the nomenclature of this report with that of Clapp and Deiss (1931).

Subdivisions of the Missoula group

<table>
<tr><th colspan="2">Clapp and Deis (1931)</th><th>This report</th></tr>
<tr><td colspan="2">Sheep Mountain formation</td><td>Pilcher quartzite</td></tr>
<tr><td colspan="2">Garnet Range formation</td><td>Garnet Range quartzite</td></tr>
<tr><td rowspan="3">McNamara formation</td><td>Upper memher</td><td>McNamara argillite</td></tr>
<tr><td>Middle member</td><td>Bonner quartzite</td></tr>
<tr><td>Lower member</td><td rowspan="3">Miller Peak argillite
Hellgate quartzite member</td></tr>
<tr><td colspan="2">Hellgate formation</td></tr>
<tr><td colspan="2">Miller Peak formation</td></tr>
</table>

MILLER PEAK ARGILLITE

NAME

The Miller Peak argillite is here redefined to include the original Miller Peak formation of Clapp and Deiss (1931), their Hellgate formation, here redefined as a member, and the lower member of their McNamara formation. As thus amended, the Miller Peak argillite underlies more of the quadrangle than any other formation.

CHARACTER

The Miller Peak argillite is composed of red and green silty argillite and argillaceous quartzite. Silty argillite is somewhat more abundant than argillaceous quartzite, and reddish hues are more common than greens. Locally pale-red, tan, and light-gray quartzite, and light-gray impure limestone occur. Red and green may alternate every few feet, or sequences several hundred feet thick may be all one color.

The rocks of this unit weather so readily that most of the area it underlies is soil covered; outcrops are scattered and most are small.

Most of the argillite is in laminae from $\frac{1}{16}$ to 1 inch thick. The laminae differ from one another in relative abundance of clay and quartz and in the maximum size of grains. Small-scale scour-and-fill structures, mostly about half an inch deep, are common, as are mud cracks and ripple marks.

The purer argillite laminae are composed almost wholly of a fine-grained intergrowth of chlorite and sericite. Many bedding planes have a sheen due to the chlorite and sericite; at a few localities these new minerals are weakly oriented at an angle to bedding, giving a poor slaty cleavage. Rocks with this cleavage are common along the north side of the Clark Fork from Marshall Creek to Kendall Creek. Shreds of detrital muscovite, from about 0.1 to 0.4 mm long and 0.01 to 0.04 mm thick, make up a small percentage of the rock. The detrital muscovite gives some of the bedding planes a sheen. Very small chips of opaque material, from 0.04 to 0.4 mm long and from 0.01 to 0.04 mm thick, are scattered through some of these rocks. The chips are commonly curled and are perhaps fragments of thin layers of some iron-rich mineral or mixture deposited on top of the layers in quiet water.

Most of the argillite layers contain some detrital quartz and a few scattered detrital feldspar grains. These grains are angular and range from silt to very fine sand; they make up from a few to locally as much as 50 percent of the argillite.

The red argillite ranges from dusky red purple (5*RP* 3/2)[1] to

[1] Color designations in parentheses are from the "Rock-Color Chart," National Research Council, 1948.

grayish red purple (5*RP* 4/2). Under the microscope the red argillite is partly opaque owing to disseminated dust, probably mostly hematite, which makes the rock red. The more quartzose laminae in the red argillite contain the coarsest recrystallized sericite and chlorite noted in the formation; these commonly attain a diameter of 0.1 mm. The green of the large chlorite flakes in these layers combines with the red to become a medium gray (*N* 5) or greenish gray (5*GY* 5/1). These gray layers become grayish orange (10*YR* 6/4) when weathered. Grayish orange laminae in weathered outcrops are diagnostic of this formation.

The green argillite ranges from grayish yellow green (7*GY* 6/2) to dusky yellow green (7*GY* 4/2), and weathers to grayish orange (10*YR* 6/4). The darker layers weather more easily than the lighter ones; the lighter layers commonly develop no more than a weathered surface stain. Hematite is absent from these rocks, and chlorite is the principal coloring agent. Generally the darker layers are more quartzose than the lighter ones. Layers of pure argillite are much more commonly green than red.

The lower few hundred feet of the Miller Peak argillite is green argillite of slightly different hue and composition than the green argillite of the rest of the formation. These lower rocks are grayish yellow green (5*GY* 6/2) and commonly weather to a light yellowish brown (1*Y* 6/3). They are rather pure argillite that contains disseminated carbonate as well as rare carbonate beds. They weather easily and are poorly exposed. This basal part of the Miller Peak argillite grades upward into the main body of the formation.

Clay balls, like those common in the McNamara argillite, were found in probable Miller Peak argillite only on the trail near the east edge of the quadrangle in sec. 9, T. 14 N., R. 17 W., and in NE¼NE¼ sec. 8, T. 12 N., R. 18 W.

X-ray diffraction analyses of the fine fraction of four samples from the Miller Peak were made by A. J. Gude 3d, of the U.S. Geological Survey. A sample of red argillaceous quartzite from the top of Miller Peak contained a little montmorillonite. A little illite was recognized in hornfelsed argillite near the small diabase mass on the east side of Rattlesnake Creek, but samples of green argillite from just west of Kendall Creek, and sandy argillite from lower Pattee Creek, contained no recognizable clay minerals.

Locally some of the rocks in the Miller Peak argillite contain enough quartz to be called argillaceous quartzite, or even rarely, rather pure quartzite. The impure quartzite may form isolated beds, or sequences of beds in which quartzite predominates.

Most of the quartzite has light-grayish weakly saturated colors. In order of decreasing abundance they are, grayish orange pink (5*YR* 6/2), pale red (10*R* 6/2), and grayish yellow green (10 *GY* 6/2). Locally some of the quartzite beds are moderate red (5*R* 5/4) and light brown (5*YR* 6/4), and, if relatively pure, can easily be confused with the overlying Bonner quartzite. The quartzite beds range from 1 inch to 3 feet in thickness, and average about 8 or 10 inches. Many layers show crossbedding and some have scour-and-fill structures ½ to 1½ inches deep. Most of the quartzite beds are capped by argillite, usually red, which ranges from a thin veneer to a layer a few inches in thickness.

The quartzite contains subrounded to subangular sand grains in a matrix of fine-grained chlorite and a little sericite. The matrix is commonly 30 to 50 percent of the volume of the rock. Most of the detrital grains are quartz, a few are argillite and silty argillite, and a few are feldspar, both sodic plagioclase and microcline. Little or no secondary quartz has been added to the clastic quartz grains. Most of the quartz grains are free of strain shadows, probably because most are isolated, and thus were under hydrostatic, rather than directed pressure. A carbonate mineral, probably calcite, which occurs both as clastic grains and as matrix, makes up a small percentage of the rock. Detrital muscovite occurs as scattered flakes. Much of the quartzite has scattered dark seams, most of which are so thin that they look like penciled lines but some of which are three-quarters of an inch thick. The dominant mineral of the dark seams is ilmenite, probably with some hematite, accompanied by zircon, epidote, rutile, garnet, and tourmaline (pleochroic from bluish green to very pale pink); all are subrounded.

HELLGATE QUARTZITE MEMBER

The Hellgate quartzite is here considered a member of the Miller Peak argillite. The unit was considered a formation by Clapp and Deiss (1931, p. 679), who described it from exposures in Hellgate Canyon just east of Missoula. This study has shown that the unit is not sufficiently distinctive or widespread to warrant formational rank; it is but one of several similar quartzite sequences, all local and none distinctive.

THICKNESS

The thickness of the lower part of the Miller Peak argillite cannot be accurately measured in this area. The Hellgate quartzite member is about 1,200 feet thick on University Mountain. Clapp and Deiss (1931, p. 681) considered the beds here classed as the upper part of the Miller Peak argillite to be 410 feet thick but the rocks they measured are faulted against the overlying unit. As mapped on the upper part of University Mountain there is about 1,000 feet of argillite above the Hellgate quartzite member. The Miller Peak argillite as shown on section *D–D′* (pl. 35) is about 6,000 feet thick and the incomplete section below the Miller Peak thrust shown on section *B–B′* (pl. 35) is about 6,000 feet thick. Although both of these sections may contain unrecognized repetitions, these figures probably indicate roughly the thickness of the formation.

CONTACT WITH THE BONNER QUARTZITE

The contact of the Miller Peak argillite with the overlying Bonner quartzite is gradational through 50 to 100 feet, and is accompanied by a gradual upward decrease in argillite, mostly red, and a corresponding increase in quartzite. The transitional sequence is not known to be well exposed except in secs. 12 and 13, T. 14 N., R. 19 W.

BONNER QUARTZITE

NAME

The formation here named the Bonner quartzite is composed of a sequence of distinctive pink quartzites that Clapp and Deiss (1931, p. 680 and 681) included as a middle member in their McNamara formation. The formation is well exposed on both sides of the Blackfoot River just above the town of Bonner.

The Bonner quartzite is one of the most easily identified formations in the Missoula group. It is also widely distributed and thus is an ideal marker for subdividing the lower part of the Missoula group.

DISTRIBUTION

The Bonner quartzite is widely distributed throughout the Bonner quadrangle. This formation parallels many major faults and, here acting as a unit, is commonly in large-scale folds. The crests of many mountains are upheld by this resistant unit. Outcrops of the Bonner quartzite are rare except in areas of glacial scour in the northern part of the quadrangle. Elsewhere this formation is largely represented by blocky accumulations of pink feldspathic quartzite.

CHARACTER

The formation is predominantly vitreous arkosic quartzite, usually pink and it includes minor amounts of interbedded argillite. Most of the quartzite ranges between grayish pink (5*R* 8/2) and pale red (5*R* 6/2) with grayish pink of about 5*R* 7/2 being most common. Less common colors are reddish gray (5*R* 5/1), grayish red (5*R* 4/2), moderate red (5*R* 4/4), and greenish gray (5*GY* 7/1), the last being more characteristic of the fine-grained quartzite. Crossbedding is common. Quartzite beds range from 4 inches to 6 feet thick, and average about 2 feet. Through much of the formation the bedding planes are marked by indistinct pale-yellowish-green (10*GY* 7/2) layers that grade into the pink rock above and below; these green layers range from about ½ to 4 inches in thickness.

The quartzite is made up of moderately well rounded detrital grains that range in size from about ½ to about 12 mm and average about ½ mm. The quartzite is 70 to 80 percent quartz, 10 to 15 percent microcline, and 5 to 10 percent argillite; shreds of muscovite, and grains of hematite, zircon, rutile, leucoxene, garnet, and tourmaline (pleochroic from bluish green to very pale pink) are accessory. Locally a few red argillite chips, as much as 2½ inches long, are included at or near the base of a quartzite bed. About four-fifths of the quartz is in rounded detrital grains, the remainder is in secondary overgrowths. The original rounded margins of the detrital grains are outlined by a thin coating of dust, probably hematite, which gives the rock part of its reddish hue. Many of the quartz grains are strained in such a way as to show that the strain mostly resulted from pressure between grains as they are now distributed. The fragments of argillite and silty argillite are red and also contribute to the color of the rock. The purer quartzite contains very little matrix, but a few grains show thin coatings of a very fine grained sericite-chlorite mixture.

All gradations exist from argillite to quartzite, and red (about 5*R* 4/3) argillite layers are interbedded with the quartzite in much of the formation, especially in the lower part. Where present an argillite layer, usually an inch or so thick, tops most of the quartzite beds; and locally red argillite composes entire sections a few feet thick. In these finer grained rocks half or more of the rock is recrystallized matrix of very fine grained sericite and chlorite. The detrital grains are like those of the purer quartzite in mineralogy and proportions, but are finer and more angular. Flakes of detrital mica generally oriented parallel to bedding are common in the argillite. These rocks are red from disseminated hematite dust.

The green layers commonly separating pink quartzite beds in some parts of the formation are very similar to the red argillite in that they contain subangular fine sand-sized detrital grains in a chlorite-sericite matrix. Here, however, chlorite colors the rock, and hematite is subordinate or absent. The fine fraction of such a green rock was analyzed by X-ray diffraction by A. J. Gude 3d, but no clay minerals were noted.

Argillaceous and silty layers, both red and green like the rocks of the McNamara argillite, are common high in the formation and their abundance increases upward.

The only local rocks that might be confused with this formation are some of the pink quartzite beds in the Miller Peak argillite, such as the Hellgate quartzite member. The quartzite beds of the Miller Peak argillite, however, are less feldspathic, somewhat more orange, and are nowhere as thick. The quartzite of this formation generally contains much less disseminated argillaceous material than the quartzite of the Miller Peak argillite, including the Hellgate quartzite member.

THICKNESS

The thickness of the Bonner quartzite probably can be measured most accurately on the north side of Mount Sentinel, where, as shown on section *B–B′* (pl. 35), it is about 1,500 feet. A somewhat thicker section is shown near the north end of section *B–B′*. The thickness at the type locality about a mile northeast of Bonner, as determined graphically, is about 800 feet. This is about the thickness given by Clapp and Deiss (1931, p. 681) for the middle member of their McNamara formation and probably this is the section they measured; here the rock is rather strongly folded and may consequently be somewhat thinned.

CONTACT WITH THE McNAMARA ARGILLITE

The contact between the Bonner quartzite and the overlying McNamara argillite is gradational. The top of the Bonner is taken as the top of the highest typical pink quartzite, so that most of the transitional beds are assigned to the Bonner quartzite. The argillite and siltstone strata in the transitional zone are like those of the McNamara; they are interbedded with pink quartzite, and they become thicker and more numerous upward through the upper 100 feet or so of the Bonner.

McNAMARA ARGILLITE

NAME

The McNamara argillite is here restricted to include only the upper part of the formation as originally defined by Clapp and Deiss (1931). So restricted, the formation includes the rocks at McNamara's Landing, a small community, now abandoned, at the junction of Camas Creek, and the Blackfoot River, 3.8 miles east of the Bonner quadrangle. This formation is widespread in the Bonner quadrangle.

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CHARACTER

The McNamara argillite is composed of red and green slightly metamorphosed argillaceous siltstone and argillite and lesser amounts of more quartzose rock. Many sections are difficult and locally impossible to distinguish from the Miller Peak argillite. Clay balls and chips are present in most sections more than a few tens of feet thick, and offer a criterion for distinguishing the McNamara from the Miller Peak. Ripple marks and mud cracks are common.

Metamorphosed siltstone is the most abundant rock. Colors of all are shades of either red or green; the red ranges from grayish red (5*R* 4/2) to grayish pink (5*R* 7/2) and pinkish gray (5*R* 7/1), and the green from grayish green (10*GY* 5/2) to pale yellowish green (10*GY* 8/2) and light greenish gray (10*GY* 8/1). Generally, but not always, the smaller the maximum grain size the darker the color; so that most siltstone is somewhat lighter colored than most argillite. Red and green are about equally abundant. The siltstone forms layers that range from about 1/16 of an inch to about 8 inches in thickness. Most layers are fairly well sorted and with fairly uniform texture throughout, although a slight upward decrease in maximum grain size through a layer is common.

The siltstone is composed of a mixture of the same materials that make up the pure argillite and fine sandstone, and the siltstone beds are interbedded with and grade into both pure argillite and fine sandstone.

Some of the argillite is relatively free of sand and silt, and the color ranges from grayish red (5*R* 4/2) to pale red (5*R* 6/2) and from grayish green (10*GY* 6/2) to light greenish gray (10*GY* 7/1). The layers of argillite are from about 1/16 inch to perhaps as much as 4 inches thick, and average about 1/2 inch. The argillite is composed chiefly of fine-grained sericite and chlorite in various proportions. Commonly, scales of these minerals are fairly well oriented and range from 0.02 to 0.05 mm in diameter. Detrital mica flakes 0.2 to 0.3 mm in diameter oriented parallel to bedding are fairly common in both argillite and siltstone.

The red argillite contains disseminated red dust, probably hematite. The green argillite contains scattered opaque grains, 0.005 to 0.01 mm across, which are white in reflected light and which may be leucoxene.

A single specimen of green argillite was checked for clay minerals by X-ray diffraction, and none were found.

The quartzite of the McNamara argillite is fine grained, and the color ranges from grayish red (5*R* 5/2) to grayish pink (5*R* 7/1). The beds are from about one-tenth of an inch to several feet thick, and most commonly are 2 or 3 inches. Most are composed of fairly well sorted material showing slightly graded bedding in units that grade from grayish-pink (5*R* 7/2) fine-grained quartzite at the bottom to a thin layer of grayish-red (5*R* 5/2) argillite at the top. Much of the fine-grained quartzite has small-scale crossbedding, in which the depth of scour and fill is about 1 inch. These thin fill layers are faintly laminated with laminae from $1/20$ to about $1/10$ inch thick. Small grayish-red argillite chips are common in these layers. The quartzite differs from the pure argillite only in that it contains coarser clastic grains. It also contains abundant chlorite and sericite mostly recrystallized from argillaceous material. The coarser grained fraction is nearly all angular quartz, but sparse mica flakes, 0.2 to 0.3 mm across, and a very few grains of microcline, zircon, and tourmaline (pleochroic from bluish green to very pale pink) were noted. Some of the more quartzose layers contain 1 or 2 percent calcite, some of which is clastic and some matrix. Fine-grained red quartzite is especially abundant in the upper part of the formation and predominates in the upper few hundred feet.

The clay balls and chips characteristic of this formation range from about $1/16$ to 1 inch in thickness and from about $1/8$ to 2 inches in length. The clay balls are usually well rounded, somewhat flattish discs, or ovoids, and are much more abundant than the chips. Many of the clay balls and chips are composed of very fine grained sericite and chlorite. Others contain as much as 70 to 80 percent of very fine grained quartz in equidimensional grains from 1 to 2 microns in diameter; this very fine grained quartz is probably best classified as chert.

The color of some of these rocks has apparently been altered subsequent to deposition. In places where the rocks have been strongly folded, green colors predominate, locally red colors are absent. In the NW$1/4$ sec. 29, T. 14 N., R. 18 W. the rocks are overturned, and all are now green. The distribution of colors in the clay balls also suggests change after deposition. The boundary between red and green layers usually parallels bedding in the enclosing rocks, and some of these color boundaries bisect a clay ball. A few clay balls show the bedding of the source layers, some are so tilted that the bedding in them does not parallel the bedding of the enclosing rock. Some of the tilted clay balls are two-colored, and the color boundary parallels the bedding in the rock, thus crossing that in the ball.

Most of the rocks in the McNamara argillite resemble those in the Miller Peak argillite. The principal differences between the argillites of the two formations are: the rocks of the McNamara are generally lighter colored and less saturated in color than those of the Miller Peak. The presence of clay balls and chips through any considerable thickness of argillite labels those rocks as McNamara. Argillite free of course grains is much more common in the McNamara than in the Miller Peak and there is less and smaller sized detrital mica in the McNamara; new chlorite flakes are less well developed in the McNamara, few can be seen megascopically. The individual rock layers of the McNamara are more homogeneous than those of the Miller Peak, and no rocks in the McNamara are as course as much of the quartzite in the Miller Peak.

THICKNESS

The only uninterrupted and structurally simple section of the McNamara argillite in the quadrangle is in the northeast quarter northeast of Sheep Mountain. Here the thickness as measured from the geologic section *B–B′* (pl. 35) is about 4,000 feet.

CONTACT WITH THE GARNET RANGE QUARTZITE

The McNamara argillite is gradational through a few tens of feet to the Garnet Range quartzite and the contact is placed where the color changes from the dominant red of the upper part of the McNamara to the dominant green of the Garnet Range. This color change is accompanied by an upward increase in detrital muscovite; this increase in muscovite apparently takes place slightly lower in the section than the color change, so that some of the uppermost red quartzite of the McNamara contain as much muscovite as the Garnet Range quartzite. Beds of green argillite are found throughout both formations, but few are near the top of the McNamara. Red quartzite, with some red argillite, is especially abundant in the upper part of the McNamara, but is absent from all but a very few beds in the lower 100 feet or so of the Garnet Range quartzite.

GARNET RANGE QUARTZITE

NAME AND DISTRIBUTION

The Garnet Range formation was named by Clapp and Deiss (1931) for exposures along the Blackfoot River at the northwest margin of the Garnet Range. Within the Bonner quadrangle the Garnet Range is chiefly exposed in the Wishard Creek syncline (pl. 35), which crosses the northeastern part of the quadrangle; the only other exposures are along narrow fault slivers south of the Clark Fork.

CHARACTER

The Garnet Range quartzite is predominantly greenish-gray quartzite interbedded with slightly lesser amounts of dark green and gray argillite, shale, and sandy argillite. The rocks are distinctive and easily recognized by the ubiquitous detrital mica, and the characteristic colors, grayish green when fresh, and moderate to light brown (5*YR* 4/4 to 5*YR* 5/6) when weathered.

Most of the quartzite is relatively pure, fine grained, and greenish gray (5*G* 7/1). The quartzite is in beds that range from about ¼ inch to about 6 feet in thickness; most beds are between 1 inch and 2 feet thick, and the average is about 1 foot. The thinner beds are commonly laminated, and the laminae are from ⅛ to ¼ inch thick; they differ from one another in grain size and content of argillaceous material.

Ten to fifteen percent of the quartzite is a fine-grained matrix of chlorite and a little intergrown sericite. Clastic grains as much as 2 mm across make up about 70 to 80 percent of the rock. Most of the grains are subangular to subrounded quartz, with overgrowths that constitute about 10 percent of the total quartz in the rock. Rare feldspar grains and scattered shreds of detrital muscovite, 0.1 to 0.2 mm across, each make up less than 1 percent of a representative rock. Detrital mica is all pervasive. It generally is oriented about parallel to bedding and is common slightly concentrated in finer grained layers and at bedding planes. The muscovite is ubiquitous in the Garnet Range quartzite, whereas it is common only in some of the finer grained beds of the other formations. Subrounded aggregates of fine-grained chlorite similar in size and shape to the clastic quartz grains, make up 1 or 2 percent of the rock and probably represent clastic grains of argillite. A few grains are composed of muscovite and chlorite interlayered parallel to basal cleavage; these may be alteration products of some preexisting mineral, possibly biotite. Most of the more massive quartzite contains spots 1 to 4 mm across made up of clusters of a brown, limonite-colored mineral, probably ankerite, which forms from a few percent to perhaps as much as 10 percent of the rock. This brown mineral occupies interstitial spaces elsewhere occupied by fine-grained chlorite; it may have replaced the chlorite, but does not seem to have replaced quartz. Accessory minerals include scattered detrital grains of zircon, tourmaline (pleochroic from bluish green to very pale pink), and an opaque mineral, possibly leucoxene.

The uppermost rock in the Garnet Range is quartzite, somewhat lighter colored than the rest of the formation. It is yellowish gray (5*Y* 6/2) and locally contains grayish-red-purple (5*RP* 4/2) spots a few millimeters across. North of Rattlesnake Creek this rock makes the upper 300 or 400 feet of the formation.

A few feet of grayish-red (5*R* 5/2) quartzite is sporadic in the upper and in the lowermost parts of the formation.

Argillite, shale, sandy agrillite, and silty argillite are interbedded with the quartzite. These finer grained rocks predominate in some parts of the formation, and in others they make up only small parts of it. Quartzite is nowhere lacking, and even in dominantly fine-grained sequences most of the beds are some what silty or sandy. In general, those parts of the formation with the thickest quartzite beds contain the least finer grained rocks. The fine-grained argillaceous and shaly rocks range from dark greenish gray (5*GY* 3/1) to dark gray (*N* 3), and in general the finer rocks are the darkest. These rocks are composed of fine-grained chlorite and a little intergrown sericite, which encloses coarser clastic grains. Detrital mica is scattered throughout, and is commonly concentrated on bedding planes. The fine-grained fraction from one specimen was analyzed for clay minerals by X-ray diffraction, and found to contain a small amount of montmorillonite.

The present mapping has shown that the "massive coarse-grained pink-white crossbedded pure siliceous quartzite" that Clapp and Deiss (1931, p. 682) included in the Garnet Range formation is a part of the Pilcher quartzite. Along the Blackfoot River, in the sequence they described in sec. 13, T. 13 N., R. 18 W., and sec. 18, T. 13 N., R. 17 W., the pink crossbedded quartzite overlies the Garnet Range and its top is faulted off against the Garnet Range to the northeast.

Clapp and Deiss (1931, p. 682) included the limestone and shale at the head of Lime Kiln Creek as a lens within the Garnet Range formation; the limestone and shale are now known to be of early Paleozoic age and are described later in this report.

THICKNESS

The thickness and ranges of thickness of the Garnet Range quartzite are difficult to determine because of tight folding and local faulting. The nature of this deformation is discussed on page 228. The thickness ranges from about 1,800 feet along the trail northeast of Sheep Mountain, to perhaps 3,800 feet, three-fourths of a mile to the northwest. We believe that most of this variation is the result of local differences in the amount of internal deformation. The deformation and thickness are minimum northeast of Sheep Mountain, which suggests that the thickness of about 1,800 feet there is probably about the original thickness. This thickness was determined graphically using a dip of 20° as indicated by trends of the contacts.

An alternate explanation for some of the range in thickness would be the existence of an unconformity at the top of the formation. If this is so, the thickness before erosion on the unconformity probably was greater than 1,800 feet, but because all the thicker sections are strongly deformed, no accurate estimate of its thickness under this interpretation can be made.

Clapp and Deiss (1931, p. 683) measured a thickness of 7,600 feet. They measured their section along the Blackfoot River from Johnson Gulch eastward to a point about 1½ miles east of the quadrangle, and thus measured across several faults and both limbs of the Wishard Creek syncline.

CONTACT WITH THE PILCHER QUARTZITE

Evidence concerning the nature of the contact between the Garnet Range quartzite and the Pilcher quartzite is inconclusive. Much of the evidence suggests that the contact is gradational, whereas other evidence suggests that it may be an unconformity. A discussion of these various interpretations follows.

The contact between the Garnet Range quartzite and the Pilcher quartzite is marked by a mixed zone which strongly suggests that the contact is gradational. This zone is 50 to 100 feet thick and is composed of interbedded greenish-gray mica-rich quartzite typical of the Garnet Range quartzite, red crossbedded quartzite typical of the Pilcher quartzite, and hybrid rocks, such as crossbedded greenish-gray mica-rich quartzite and red mica-rich quartzite. The nature of the bedding in the mixed zone also suggests that the contact may be gradational. Nearly all the red quartzite in the Pilcher quartzite immediately above the mixed zone is crossbedded, whereas many of the beds in the mixed zone are not. If the mixed zone is actually the lowest part of the Pilcher quartzite then it would seem that perhaps more of the beds in it should be crossbedded.

Alternately, the presence of greenish-gray rocks in the mixed zone may suggest that this contact is an unconformity. As thus interpreted these greenish-gray beds are considered to be composed of detritus from the Garnet Range quartzite, redeposited as part of the Pilcher quartzite. According to this interpretation, the local yellowish-gray color of some of the uppermost rocks of the Garnet Range quartzite can be explained as the result of weathering during erosion on the unconformity.

The most persuasive evidence in favor of the hypothesis of uncomformity at this horizon is the range in thickness of the Garnet Range quartzite, which may be the result of beveling beneath an uncomformity. As noted above, however, we believe that much of this variation is the result of deformation subsequent to deposition, and therefore we do not consider this sufficient evidence for an unconformity. It is noteworthy that the Garnet Range quartzite is relatively thick at the north edge of the quadrangle where the Pilcher quartzite is thinnest, and that the Pilcher quartzite is thick northeast of Sheep Mountain where the Garnet Range quartzite is thin. These observations suggest that the Pilcher quartzite may have been deposited upon an uneven surface cut into the Garnet Range quartzite.

Opposed to the interpretation that this contact is an uncomformity, is the distribution of internal deformation in these two formations. Most of the Garnet Range quartzite is strongly deformed, whereas most of the Pilcher quartzite is not. These relations do not, however, demonstrate an uncomformity at the top of the Gernet Range quartzite, because the deformation extends up into and is therefore younger than the lower few tens of feet of the Pilcher quartzite. The age and nature of this deformation is discussed further in the section on structure.

CORRELATION

To the west in the St. Regis-Superior area, Montana, A. B. Campbell (1961) has mapped similar-appearing rocks which are probably equivalent to the Garnet Range quartzite.

PILCHER QUARTZITE

NAME

Pilcher quartzite is a new name here applied to the uppermost formation of the Missoula group (Belt series) in the Bonner quadrangle. Clapp and Deiss (1931) called this the Sheep Mountain formation, but this name had earlier been applied to another formation elsewhere, and therefore is here abandoned for use in Montana. The Pilcher quartzite is named from the extensive exposures along the upper part of Pilcher Creek. In the lower reaches of the creek the bedrock is largely covered by glacial debris, but the upper end of the canyon has been glacially scoured, and here a complete, although thin, section of the formation is exposed.

DISTRIBUTION

The Pilcher quartzite crops out in the Wishard Creek syncline, which crosses the northeastern part of the quadrangle, and in a thin fault sliver south of the Clark Fork near the east edge of the quadrangle.

CHARACTER

Most of the formation, except the upper few hundred feet, is made up of crossbedded quartzite of uniform lithology (pl. 36*A*). The quartzite ranges from pale orange pink (about 5*YR* 8/2) to moderate red (5*R* 4/4); adjacent cross strata commonly contrast strongly in color. Beds range from about 1 inch to 10 feet in thickness, and are generally 2 or 3 feet. The sets of cross-strata range from 2 to 10 inches in thickness, and the individual cross-strata are from 1/8 to 1/2 inch or so thick (nomenclature of cross-stratification according to McKee and Weir, 1953). The crossbedding is more prominent in these rocks than in the Bonner quartzite, because the color contrasts between cross laminae are greater. Some of the rocks show bleached spots and streaks; somewhat less common are red spots in lighter colored rocks; some appear to have a dark mineral at their centers. These spots, both light and dark, commonly are 1/4 to 2 inches across, and from one-half inch to several inches long.

The quartzite is composed of about 90 percent fairly well sorted, subrounded sand-size clastic grains, mostly between about 0.2 and 0.5 mm in size—as much as 4 mm in a few beds. The matrix is fine-grained intergrown chlorite and sericite. Nearly all the grains are quartz; a few are quartzite, metasiltstone, argillite or silty argillite, and detrital muscovite, to 0.5 mm long; no feldspar was noted. There is no evidence of addition or recrystallization of the quartz, and only a few grains show strain effects, especially those with a small area in contact with an adjacent grain. Disseminated hematite dust gives the quartzite its red color. Barite makes up a fraction of a percent of one specimen from the top of Sheep Mountain; it replaces both clastic grains and matrix in areas as much as 0.2 mm across. Traces of zircon and tourmaline, pleochroic from bluish green to very pale pink, were noted.

The upper part of the formation is gradational with the lower part and resembles it; except that the upper part commonly lacks crossbedding, and much is as dark as dusky red (5*R* 3/4) or, in the uppermost part, even very dusky red (5*R* 2/4).

A few thin layers of argillite are interbedded in the uppermost 10 or 20 feet; these range from a fraction of an inch to a few inches in thickness. Some are grayish red (5*R* 4/2), are commonly ripple marked and mud cracked, and are very similar to argillite in the underlying formations. Some are dusky yellow green (about 5*GY* 5/2), and are indistinguishable from the overlying shale of Cambrian(?) age.

THICKNESS

Nearly complete sections of the Pilcher quartzite are probably present on the ridge in the southwest corner of sec. 30, T. 14 N., R. 17 W., and on the ridge in the S1/2 sec. 6, T. 13 N., R. 17 W. At both places the base is exposed, and the rock at the ridge tops is green argillite interbedded with dusky red quartzite, an assemblage normally confined to the uppermost Pilcher quartzite. The thickness at each place, as determined graphically from plate 35, is 1,000 feet. A complete section seems to be present in SE¼ sec. 6, T. 14 N., R. 18 W., and here the formation seems to be considerably thinner; 700 or 800 feet thick. Another complete section, on the ridge south of the Blackfoot River at the east edge of the quadrangle is about 1,800 feet thick although the structure is somewhat more complex.

CONTACT WITH SHALE OF CAMBRIAN(?) AGE

The contact between the Pilcher quartzite and the shale that overlies it is exposed at only a few places. In these exposures the change from relatively pure vitreous quartzite belonging to the Pilcher quartzite to somewhat thinner bedded, more argillaceous quartzite assigned the next unit above is fairly abrupt, rather more so than is common at contacts between formations of the Belt series. No discordance in attitudes is present, but shale like that of the unit above the Pilcher is interbedded with the uppermost quartzite assigned to the Pilcher in such a way as to suggest gradation. As the Pilcher quartzite is the highest formation locally assigned to the Belt series, the character of its contact with younger rocks is of particular significance. As is true of the base of the Pilcher quartzite, various interpretations are possible. In our opinion, the present evidence permits the concept that the Pilcher quartzite has a gradational contact with the next younger formation.

This concept is contrary to the commonly held belief that the top of the Belt series in and near northwestern Montana is marked everywhere by an unconformity. Evidence for this belief has been summarized by Deiss (1935), Knopf (1957, p. 86), and others. The positive evidence of angular discordance that has been recorded (Deiss, 1935, p. 119–122) is all in scattered localities farther east than the Bonner quadrangle, and therefore comparatively near to the edge of the basin in which the Belt series was deposited. More widespread evidence of discordance is believed to be furnished by the fact that the Flathead quartzite in the eastern part of northwestern Montana overlaps successively younger formations of the Belt series from east to west. Relations of this sort have been adduced near Helena (Knopf, 1957, p. 86), and Deiss (1935, pl. 8) offers a similar interpretation for a region that extends as far west as the Bonner quadrangle. His interpretation is vitiated, locally, by doubt as to the correctness of his observations. For example, he considered the Garnet Range quartzite in the Bonner quadrangle to be at least twice as thick as our mapping shows it to be (p. 207). C. P. Ross (1956, p. 692) suggests that lateral variations of lithology within the series are sufficient to explain the observed variations at the top of the series. Reesor (1957) reports from Canada that rocks equivalent to parts of the Belt series show

A. BOULDER OF CROSSBEDDED PILCHER QUARTZITE

Boulder is on the south side of the Blackfoot River about 3 miles northeast of Bonner

B. SHORELINES OF GLACIAL LAKE MISSOULA

View is eastward across Rattlesnake Creek valley to Jumbo Mountain

rapid variations from place to place. Our observations that the quartzite units, such as the Hellgate quartzite member, are lenticular are in accord with Reesor's conclusions.

CORRELATION

The Pilcher quartzite is considered to be the uppermost formation of the Belt series, and it is therefore assumed to be older than the Flathead quartzite of Middle Cambrian age. The Pilcher quartzite has so many characteristics in common with the Flathead quartzite, however, that correlation between the two is a possibility worthy of consideration. The composition, colors, and sedimentary structures and textures seem identical with those of the Flathead quartzite, which is widespread in other parts of Montana. Among its characteristics, the absence of feldspar from the Pilcher quartzite may be of diagnostic significance. The Flathead quartzite contains no feldspar, whereas this mineral is commonly present in quartzitic rocks of the Belt series. The outstanding difference between the Pilcher quartzite and Flathead quartzite is thickness. The Flathead quartzite in most areas that have been mapped is 300 feet or less in thickness (Hanson, 1952, p. 12); whereas, the Pilcher quartzite ranges from 700 feet to as much as 1,800 feet in thickness.

In an area 6 miles west of Superior, Mont., and about 55 miles west of Missoula, A. B. Campbell (oral communication, 1955) regards an assemblage of red, in part crossbedded, vitreous, feldspathic quartzite as equivalent to the Pilcher quartzite on the basis of its resemblance to the Pilcher quartzite. This assemblage is exposed in an area of less than one-tenth square mile entirely bounded by faults and deposits of Quaternary age. A thickness of about 700 feet of these rocks is exposed.

The uncertainties about the nature and position of the upper contact of the Belt series focus attention on the question of what features characterize the rocks of the series and differentiate them from younger rocks. The primary differences between the formations of the Belt series and those of early Paleozoic age seems to be the result of deposition in different tectonic environments. The formations of the Belt series are rarely less than 1,000 feet thick, which suggests deposition in an actively subsiding area, whereas the overlying early Paleozoic formations range from less than one hundred to a few hundred feet in thickness, which indicates that the crust had become more stable by the time they were deposited. In areas to the east of the Bonner quadrangle the formations of the Belt series are separated from younger formations by an unconformity, during the development of which subsidence not only became less active, but the formations of the Belt series were locally warped and eroded.

In the Bonner quadrangle the Garnet Range quartzite and older formations are thick and typical of the Belt series, and the shale and limestone of Cambrian (?) age are typical lower Paleozoic formations.

The thickness of the Pilcher quartzite is comparable to the thickness of some of the formations of the Belt series, and it is included in the series partly on that basis. If, however, it is eventually found to be equivalent to the Flathead quartzite of Middle Cambrian age, which it resembles in all respects except thickness, it will be reasonable to exclude it from the Belt series. Thus correlated, its thickness would indicate that the tectonic conditions that prevailed during the deposition of the Belt series persisted into Middle Cambrian time in the vicinity of the Bonner quadrangle, even though they had come to an end earlier farther to the east. Similar relations in Canada add credibility to this suggestion. In southeast British Columbia in the western part of the area occupied by rocks that are equivalent to Belt series, deposition was apparently continuous into early Cambrian time. V. J. Okulitch (1949, p. 17) reports that the Hamill series (a part of his Windermere series) and equivalent units in that area are of Precambrian age in one locality and of Cambrian age in another, and therefore seem to transgress the Precambrian-Cambrian boundary. Later (1956, p. 707–714) he reported conformable relations between rocks of early Cambrian age and older rocks at several places in the same general area of Canada. Reesor (1957, p. 158) would exclude the Hamill series from the Windermere system, but he agrees that the Windermere system, so restricted, is conformable below rocks of early Cambrian age. Farther to the east in Canada the equivalents of the Belt series are separated from rocks of Middle Cambrian age by an unconformity (Reesor, 1957).

AGE OF THE BELT SERIES

Traditionally all the Belt series is considered to be Precambrian in age. This assumption is based largely on the belief, once widely held, that the rocks of the major subdivisions of geologic time were separated everywhere by unconformities. According to this concept the unconformity at the top of the Belt series, farther to the east, was identified as the break between the Precambrian and Paleozoic eras. Most geologists would now agree that unconformities are not reliable time markers. The earth movements that produced most features of this kind were restricted to limited areas, and the ages of the rocks both above and below unconformities vary from place to place. Gilluly (1949), Shepard (1923), Berry (1929), and Spieker (1956) have pointed out the undesirability of using unconformities as a basis for stratigraphic correlation, and Snyder (1947), Wheeler

(1947), and Wheeler and Beesley (1948) have dealt with the problem as it applies to the Precambrian-Cambrian boundary.

The data outlined above indicate that there is no demonstrable unconformity at the top of the Belt series in the Bonner quadrangle. Further, the rocks above the Belt series in the quadrangle are thought to be of Middle Cambrian age. On this basis the uppermost rocks of the Belt series in the quadrangle may include rocks of Early Cambrian and perhaps even of Middle Cambrian age.

Even if proof should eventually be found that the unconformity at the top of the Belt series is more widespread than can now be shown, the possibility would remain that the age of the unconformity is such that rocks of Early Cambrian age may lie beneath the unconformity. On the other hand, the possibility cannot be denied that an unconformity, if present, corresponds to a sufficiently long hiatus so that all the Belt series may be of Precambrian age.

One reason that the Belt series is commonly regarded as of Precambrian age is the lack of diagnostic fossils. To a degree the absence of fossils in many of the rocks of the series may have resulted from the unsuitable lithologic character of the material of which they are composed. Rather pure quartzite such as the Pilcher and, to a lesser extent, the Garnet Range quartzite, would scarcely be expected to be fossiliferous regardless of their age. For example, the lithologically similar Flathead quartzite has nowhere yielded useful fossils although it has been studied in many localities and is known to be of Cambrian age. The possibility of finding fossils in a rock like the McNamara argillite is better, but even this unit cannot be regarded as favorable for the preservation of fossil remains.

The conclusion best in accord with the facts gathered during the present study is that most of the Belt series is, as it has long been held to be, of Precambrian age. There is, however, a strong possibility that some of the uppermost beds of the series as mapped in the Bonner quadrangle may be of Cambrian age.

CONDITIONS OF DEPOSITION

Much has been written on the extent, character, and mode of origin of the Belt series; the reader is referred to Fenton and Fenton (1937 and 1957) and to Ross (1956) for more detailed discussion of these subjects than is given here.

The basin within which the Belt series was deposited was widespread and covered most of western Montana. It extended northward into Canada for an unknown distance, perhaps as far as the Arctic Ocean, and it or contemporaneous basins may have extended far to the south; the Fentons (1937, fig. 3; and 1957, fig. 1) show the inferred extent of the basin in which the Belt series was deposited.

In the Bonner quadrangle, as elsewhere, many rocks of the Belt series bear evidence of deposition in very shallow water, indicating that during most of this time sedimentation kept pace with subsidence. Mud cracks are common in the more argillaceous rocks, and they show that parts of the basin were just awash much of the time, so that the surface of a sediment layer dried and cracked before another layer was deposited. Rare raindrop impressions also testify that the sediments were exposed to the air.

Almost none of the detritus in the rocks of the Missoula group is larger than sand, which suggests that it may have been transported a great distance, that it may have come from a source area of low relief, or that it may be detritus that has been through more than one cycle of deposition. One possible source for multicycle sediments may have been lower parts of the Belt series.

It seems likely that much of the sorting of the sand from the finer grained detritus may have taken place during transportation and deposition as the result of differential transporting capacities of water currents of different velocity, and perhaps by winnowing action of waves along migrating shorelines. Some of the thinner, less widespread quartzite layers may represent channel or delta deposits of meandering streams.

The carbonate rocks in the Newland limestone and in the Missoula group probably were formed in shallow seas. Some of the carbonate was deposited by algae as stromatolites, some is detritus derived from stromatolites, some probably was ooze that micro-organisms helped precipitate, and some may have been deposited by direct chemical precipitation. At one time it was thought that most of the carbonate minerals in these rocks may have been the result of direct chemical precipitation (Daly, 1912, p. 643–675). The seas of Precambrian time were considered to have been different in composition from the seas of later times, so that the calcium that came into them was promptly precipitated. W. W. Rubey (1951, p. 1113–1114) has more recently concluded, from a variety of data, that "the composition of sea water and atmosphere has varied surprising little, at least since early in geologic time." Most of the carbonate rocks contain considerable argillaceous and quartzose detritus similar to, and probably from the same source as, the detritus in the noncarbonate parts of the Belt series.

DIABASE (PRECAMBRIAN?)

DISTRIBUTION AND NATURE OF OUTCROPS

All the larger diabase bodies are confined to, and seem to have been deformed with, the upper part of the Miller Peak argillite; they are chiefly sills although locally they are discordant. These larger bodies are located in a 3- to 5-mile-wide belt parallel to and north of the Clark Fork fault.

Most of the diabase sills and dikes are small; many were too small to map. These smaller bodies are commonly intruded into the McNamara argillite and older rocks.

Granophyric differentiates of the diabase were noted at two places, and aplitic dikes at another.

Outcrops are prominent where major stream valleys transect the intrusive bodies. A conspicuous brown granular soil and weathered fragments of diabase usually mark the surface extent of these intrusive bodies.

PETROGRAPHY

Fresh specimens were difficult to obtain and the sampling was, with a few exceptions, sporadic rather than systematic. Near the borders of the intrusive bodies the rocks are commonly somewhat finer grained and less altered than elsewhere and probably represent chilled contact phases. The rocks within these bodies all show some alteration, probably deuteric. The diabase is a dark, slightly greenish gray, fine- to coarse-grained aggregate of plagioclase and pyroxene. Pyroxene occurs in discrete grains between plagioclase laths in a typical diabasic texture.

Plagioclase (An_{50-65}), some unzoned and some with very faint normal zoning, constitutes 45 to 60 percent of the rock. In chilled-contact phases the plagioclase may be fresh or only slightly altered to sericite and chlorite, elsewhere the plagioclase is commonly partially altered to chlorite, epidote, zoisite, and albite. About equally abundant pale-brown pigeonite and augite together make up 30 to 45 percent of the diabase. These minerals occur as euhedral grains, as irregular intergrowths in optical continuity, and rarely as cores of pigeonite with overgrowths of augite. A low calcium content for the pigeonite is indicated by the optical data (optic plane $\perp$ 010, $2V \approx 0°$).

A little green to greenish-brown hornblende, and in most specimens somewhat less pale-green to greeninsh-blue amphibole, probably deuteric, are associated with, and commonly border the pyroxene, some of which is reduced to much-altered small cores within amphiboles. Olivine was noted in one specimen from the diabase body north of the Clark Fork, west of Kendall Creek. The olivine is associated with pyroxene and is altered to iddingsite and granular magnetite. A trace of biotite, pleochroic from yellowish to reddish brown, in discrete flakes, and associated with pyroxene, amphibole, and magnetite, occurs throughout the diabase. Micrographic intergrowths of quartz and orthoclase make up 1 to 3 percent of the diabase. Rare free quartz occurs in most specimens, and myrmekitic albite and quartz were noted in a few. Micrographic quartz and orthoclase are somewhat less abundant in chilled border phases than in the interiors of the diabase bodies. Accessory magnetite and apatite occur in all the specimens, and hematite, actinolitic hornblende, pyrite, and zircon were noted in some. Some of the magnetite is intergrown with a nearly opaque brown mineral which may be leucoxene altered from ilmenite. Granules of sphene enclosed in chlorite were noted in one specimen from east of Bonner.

Amphibole is particularly abundant in coarse-grained phases of the diabase body that crosses Donovan Creek at the east edge of the quadrangle. Here epidote in irregular veins as much as 2 inches thick was noted. The epidote crystals preserve the diabasic texture; later fractures within the epidotized bands are filled with randomly oriented later epidote.

A relatively light-greenish-gray and coarse-grained granophyric differentiate of the diabase crops out west of Kendall Creek in the E½ sec. 6, T. 12 N., R. 17 W. It is gradational to the diabase, but its geometric relation to the diabase body was not determined. Under the microscope the dominant granophyric texture seems to be modified by randomly distributed laths and radial aggregates of amphibole. Approximately 35 percent of the rock consists of coarse intergrowths of slightly turbid orthoclase and quartz. Part of this orthoclase is in the perthitic intergrowth with sodic oligoclase. Anhedral laths and radial clusters of hornblende (pleochroic from yellow green to greenish brown or blue) constitute 35 to 40 percent of the rock. The hornblende is commonly altered to chlorite and partly replaced by epidote. Sparse green biotite is intergrown with the hornblende. Perhaps 20 percent of the rock is plagioclase, completely altered to a fibrous aggregate of sericite, chlorite, albite, and epidote. Magnetite forms anhedral grains or somewhat skeletal intergrowths with hornblende, chlorite, sphene, or reddish-brown biotite. Other minor constituents include hematite, brown allanite (?), and apatite.

Minor light-tan aplitic dikes, from 1 to 6 inches in width, cut irregularly through the diabase near the mouth of Donovan Creek at an elevation of about 4,000 feet and near the south contact of the diabase. The outcrop covers about 20 square feet. The texture is xenomorphic granular with average maximum grain size of about 1 mm. Approximately 35 percent of the rock is quartz, 55 percent twinned and unaltered albite, and from 5 to 7 percent calcite. The calcite is interstitial to and locally replaces quartz and albite. Irregular granules of sphene are enclosed in the calcite. Minor chlorite, magnetite, hematite, and zircon also occur.

AGE

The age of the intrusive rocks is unknown. They were deformed along with the enclosing Belt series. As the major deformation in the region probably was of Laramide age, these rocks were emplaced before the Laramide orogeny. They have not been found in rocks younger than the McNamara argillite, and because of this we believe that they were probably emplaced during Belt time.

Daly (1912, p. 70, 161–163, 207–220), and Fenton and Fenton (1937, p. 1903–1904) report sills and dikes of similar composition cutting the Belt series in the vicinity of Glacier National Park; there the intrusive rocks are thought to be genetically related to, and therefore the same age as, lava flows that are interbedded with rocks of the Belt series. Similar rocks have been mapped in western Mineral County, Mont. Calkins and Jones (1914, p. 174–175) believe they were deposited before the Belt series was deformed, whereas Wallace and Hosterman (1956, p. 587–588, 595) believe they may have been intruded in Laramide or post-Laramide time. Mertie, Fischer, and Hobbs (1951, p. 47–48) in reference to similar rocks in the Canyon Ferry quadrangle say "It is possible that they are of Pre-Cambrian age, but they may be much younger. It is improbable that they are of Mesozoic or Tertiary age." Calkins and Emmons (1915, p. 12–13) believe that similar rocks in the Philipsburg quadrangle are post-Cambrian and probably younger than the late Cretaceous. Deiss (1943, p. 245–248) reports diorite and gabbro sills in the Saypo quadrangle, which differ from the other igneous rocks reported here in that they intrude rocks of Mesozoic age as well as the Belt series.

SHALE AND LIMESTONE OF CAMBRIAN(?) AGE

DISTRIBUTION

The shale and limestone of Cambrian(?) age form two mappable units. These units have not been given formal names because their areal extent is small, and because correlation with better known units to the east is uncertain. Outcrops of these rocks form isolated patches in the core of the Wishard Creek syncline.

Clapp and Deiss (1931, p. 682) noted the existence of these rocks, but they considered them to be a part of the Garnet Range quartzite.

CHARACTER OF THE SHALE

The lower unit of Cambrian(?) age is dominantly shale with some quartzite and a few limy layers. The shale is most commonly dusky yellow green (about 5*GY* 5/2), much is dusky red (about 5*R* 3/4) and some is greenish brown or dark gray. Much of the shale is without conspicuous banding, but beds, where visible, range from 1/16 to 2 inches in thickness. Commonly the shale is cut by fractures or weakly developed cleavage planes at ⅛- to ¼-inch intervals, so that much of the rock breaks up into elongated prisms bounded on two sides by the fractures and on two sides by bedding planes. The shale is composed of very fine grained sericite and chlorite mixed with various amounts of silty material. Small shreds of muscovite about 0.1 mm across make up a small percentage of the shale.

Beds of quartzite are scattered throughout the shale sequence and the lower 15 to 50 feet is made up dominantly of quartzite. The quartzite is most commonly pale brown (from 5*YR* 6/2 to 5*YR* 5/2), some beds are light olive gray (5*Y* 5/2) and some grayish yellow green (5*GY* 5/2). Locally some beds are dusky red (5*R* 3/2), like some of the quartzite in the upper part of the Pilcher quartzite. Beds range from 1 to 18 inches in thickness, and average about 1½ to 2 inches. Some of the beds are faintly laminated. In the lower part of this unit the quartzite beds are generally separated by thin shale interbeds, indistinguishable from those of the upper part of the unit. Many of these shale interbeds are irregularly marked; some have shallow round or oval depressions 5 to 20 mm across and 1 to 5 mm deep, which may be raindrop impressions; others have irregular, wandering markings 1 to 5 mm wide, which probably are worm casts or coprolites. Some of the quarzite beds are pierced by cylindrical holes, 1 to 5 mm in diameter. Each such hole is confined to one bed, and most are now partly filled with limonitic material. They cross the bedding at various angles; all can reasonably be called worm borings or scolithus tubes.

The quartzite is composed predominantly of clastic grains with only a small amount of matrix or intergranular material. Quartz makes up about 80 percent of the rock, and most of the rest is rounded pale-green aggregates of very fine grained (0.005 to 0.02 mm) material, which resembles glauconite. The green material was separated from a crushed sample of the quartzite by a Frantz magnetic separator, and analyzed by X-ray diffraction by A. J. Gude 3d. The mineral gave a muscovite pattern, and is called sericite because of its small size. Gude reported a high "fluorescent" background, probably from iron in sample, but no iron-mineral pattern. A few widely scattered grains of feldspar, both microcline and sodic plagioclase, were noted; and small chips and fragments of shale are common. Fine-grained green sericite with perhaps a little chlorite locally occurs as a matrix. Accessory minerals include clastic grains of zircon, pyrite, magnetite or ilmenite, tourmaline (pleochroic from bluish green to very pale pink), and hematite. Small shreds of muscovite are scattered through the quartzite.

THICKNESS OF THE SHALE

From the following section, and from the poorly exposed shale sequence in the NW¼ sec. 17, T. 14 N., R. 18 W., the shale is estimated to be about 200 feet thick.

Section of the shale unit of Cambrian(?) age in the center of sec. 36, T. 14 N., R. 18 W.

	Feet
Cambrian (?).	
Limestone:	
Top concealed.	
Limestone, impure, limonite-colored, and thin layers of interbedded shale. Probably near the base of the limestone sequence	2 or 3
Shale:	
Covered interval	about 100
Shale, dusky-red, and some interbedded grayish-green shale	30
Shale, sandy, grayish-green	1
Shale, grayish-green	10
Shale, dusky-red	20
Shale, somewhat sandy, olive-gray	10
Quartzite, dusky-red	15 to 20
Precambrian.	
Pilcher quartzite:	
Quartzite, moderate-red, in 1- to 2-inch beds with green shale partings	20 to 30

CHARACTER OF THE LIMESTONE

The limestone that makes up nearly all the upper unit is dark gray (*N* 3) on freshly broken surfaces, and weathers to medium gray (*N* 5). Bedding commonly ranges from 1 to 3 inches in thickness. Most of the limestone contains small calcite bodies of various shapes, some of which seem to be of organic origin. Many beds, especially in the upper part, are oolitic; the oolites are from 1 to 5 mm across, and in individual beds all of the oolites are of fairly uniform size. Other bodies included in the limestone range from small rodlike bodies to small irregular fragments and chips of limestone. Many bedding planes, especially in the lower part, are mottled with orange spots that average about 1 inch, and range from 1/16 to 4 inches across. The spots range from grayish orange (10*YR* 7/4) to pale reddish brown (10*R* 5/4), and are most commonly an intermediate color of about yellowish orange (10*YR* 7/6). The spots are the parts of discontinuous orange layers, as much as one-half inch thick, that locally lie parallel to bedding. Parts of one layer may show up as spots on several adjacent bedding planes, possibly as several spots on each. An individual orange layer crosses 1/16 to ½ inch of bedding between the places where it is exposed as a spot in each of two adjacent bedding planes. The uppermost rocks of the limestone sequence are a few tens of feet of fine-grained, slightly sandy limestone that is dark gray on both weathered and fresh surfaces. These rocks cap the conical hill in the middle of the S½ sec. 27, T. 14 N., R. 18 W. Bedding is from ½ to 2 inches thick. The dark limestone includes rods and chips of white limestone, about 1/16 inch wide and 1 or 2 inches long. Some of the uppermost limestone contains rounded to subrounded fragments an inch or so across in a matrix of white limestone.

Under the microscope most of the carbonate in the limestone that weathers light gray is seen to have been recrystallized to a mosaic that encloses rhombs of dolomite. The orange layers are composed of clayey limestone; the orange color is mostly due to limonitic dust largely confined to calcite grain boundaries.

THICKNESS OF THE LIMESTONE

The limestone is the uppermost bedrock unit so that its maximum original thickness cannot be determined. It is 300 to 400 feet thick in the hill in the SE¼ sec. 27, T. 14 N., R. 18 W.

CORRELATION

The shale sequence resembles the Wolsey shale, and the limestone resembles the Meagher limestone which overlies the Wolsey shale. Both the Wolsey and Meagher are Middle Cambrian age. Fossil remains from the limestone on the boundary between secs. 8 and 17, T. 14 N., R. 18 W. about 700 feet east of the southwest corner of sec. 8 were referred to Richard Rezak of the U.S. Geological Survey. He reported that "The only fossil recognized is a hydrozoan closely related to the genus *Clathrodictyon*. This genus is quite abundant in rocks ranging in age from Cambrian through Devonian." The range of these fossiles permits the Middle Cambrian age suggested by the lithologic character of the rocks.

The lower 15 to 20 feet of the shale sequence is quartzite that is tentatively correlated with the Flathead quartzite. This quartzite is about the right thickness to be the Flathead quartzite, but it is not dominantly red, it is more argillaceous, and it lacks the crossbedding that is typical of the Flathead quartzite. Alternately, as noted in the section on the correlation of the Pilcher quartzite, we believe that the Pilcher quartzite may eventually prove to be the equivalent of the Flathead quartzite.

Stratigraphic units that do not correspond exactly with the Wolsey shale and Meagher limestone of the well-known sequence of Middle Cambrian age in the Little Belt Mountains have been found by Calkins and Emmons (1915, p. 5) and Hanson (1952, p. 12) in the Philipsburg quadrangle (about 22 miles southeast of the Bonner quadrangle) and Deiss (1939, p. 54) some 50 miles to the northeast. Direct correlation between the local rocks of probable Middle Cambrian age and the better known formations farther east is thus at present impossible.

DEPOSITS OF CENOZOIC AGE

BASIN DEPOSITS OF TERTIARY AGE

The basin deposits of Tertiary age crop out both north and south of Missoula. They underlie only about 3 square miles at the west edge of the quadrangle, are poorly exposed, and were not studied in detail. They are composed of sandstone, conglomerate, shale, volcanic ash, and lignite. Many of these rocks, especially the sandstone and conglomerate, are orange, probably due to limonite cement. The shale is usually light gray, tan, or light orange; the volcanic ash is white, light gray, or light tan. Beds range from 1 inch to about 4 feet in thickness. These rocks are generally rather poorly sorted. The matrix of the conglomerate is shaly sandstone, and the matrix of the sandstone is shale or silty shale. The pebbles and cobbles are angular to subrounded, as much as 5 inches or more across, and average perhaps 1 or 2 inches. These fragments are derived almost entirely from the Belt series, and all the local rock varieties are represented. Most of these rocks are only weakly indurated; a little of the conglomerate is resistant. Lignite (Pardee, 1913, p. 229–244), has been mined from Tertiary deposits just west of the quadrangle, in sec. 4, T. 13 N., R. 19 W. Rowe (1906, p. 60) reports a 4-foot lignite bed.

Plant fossils abound in water-laid ash beds. Jennings (1920) considered them to be of Oligocene age. Chaney and Sanborn (1933, p. 61) suggest an early Miocene age.

The northeastward dip shown on the geologic map in the Tertiary rocks north of Missoula (pl. 35) was photogrammetrically determined. At least 1,000 feet of these rocks occur in this incomplete, faulted section.

DEPOSITS OF QUATERNARY AGE

GLACIAL DEPOSITS

Most of the cirques in the quadrangle contain small moraines, and the lower reaches of Pilcher Creek and Lime Kiln Creek are choked with debris, largely morainal. A terminal moraine across Rattlesnake Creek in sec. 19, T. 14 N., R. 18 W., is described by W. C. Alden (1953, p. 108–109). A marginal moraine lies along the east side of Rattlesnake Creek canyon in sec. 2, T. 14 N., R. 18 W., at an elevation of about 5,400 feet, and another is along the northeast side of Pilcher Creek in SW¼ sec. 5, and the N½ sec. 8, T. 14 N., R. 18 W.

DEPOSITS OF GLACIAL LAKE MISSOULA

Within the quadrangle the deposits of glacial Lake Missoula consist of beach deposits, bottom deposits, and ice-rafted rock fragments; none are extensive enough to be shown on the geologic map (pl. 35).

Distinct but delicately carved shorelines are confined to the western part of the quadrangle below 4,200 feet (pl. 36*B*). They can be seen on the slopes of Jumbo Mountain on the north side of the canyon of the Clark Fork as far east as Marshall Creek, and north of Jumbo Mountain on both sides of Rattlesnake Creek for about 5 miles, and on the west face of the mountains that extend from University Mountain south to and beyond Pattee Creek canyon.

The beach markings consist of thin horizontal beach deposits and shallow excavated nicks which, although plainly visible from a distance, especially when favorably illuminated, are scarcely perceptible to a person standing on them. These beach deposits consist of thin accumulations of poorly rounded gravel and sand. The shore markings show on open and grassy slopes that commonly face, or are near, broad open valleys that provided sizable fetch for the waves. Elsewhere wave action was too weak to produce marked beaches. The slopes that show the beach markings are comparatively free of coarse debris and rock ledges, and were more vulnerable to wave attack than many of the other slopes. Several small bar and spit deposits are in the saddle north of Jumbo Mountain.

The only other deposit attributed to the lake is the finely laminated silt in the northern part of sec. 24, and the southern part of sec. 13, T. 13 N., R. 19 W. The ridge shown on the map (pl. 35) is part of this deposit. The silt overlies, and thus is younger than, the alluvium, from which it is not differentiated on the geologic map. W. C. Alden (1953, p. 108) states that along the Blackfoot River, east of the Bonner quadrangle, a glacier terminated in glacial Lake Missoula, and that lacustrine silts were deposited on top of outwash laid down in front of the glacier as it retreated. Similar deposits of silt are abundant in the valley west of Missoula outside the Bonner quadrangle (Alden, 1953, p. 156). Erratic pebbles and boulders believed to have been ice rafted are common on many of the slopes below about 4,200 feet.

ALLUVIAL DEPOSITS

Most of the alluvium in the quadrangle is in the valleys of the larger streams. It consists of loose, poorly sorted material that ranges from fine silt and clay to coarse rounded gravel and boulders. Much of the alluvial fill is glacial outwash, and locally the alluvial deposits are associated with morainal and lacustrine deposits.

At least twice during Pleistocene time there were extensive glaciers in and adjacent to the upper reaches of the Blackfoot River (Alden, 1953, p. 88 and 106), and according to Alden (p. 109) the Blackfoot was handling a large amount of coarse bouldery material, some of which was deposited along with finer material in some of the stream valleys in the Bonner quadrangle.

Most of the alluvium along Rattlesnake Creek is apparently Wisconsin in age. The moraine in sec. 19, T. 14 N., R. 18 W., probably marks the maximum extent of a glacier of the last major stage of glaciation. Upstream from the moraine the floor of Rattlesnake Creek canyon was occupied by the glacier, so that most of the alluvium there must have been deposited after the glacier had receded. Downstream from the moraine most of the valley fill is glacial outwash that merges with the upper level of alluvial fill along the Clark Fork at Missoula; this relation suggests that the outwash from the Rattlesnake Creek glacier and the alluvium along the Clark Fork are about the same age.

COLLUVIAL DEPOSITS

Three areas are mapped as colluvium. One of these is along and mostly south of Pattee Creek. Here a layer of angular rock fragments mixed with soil fills part of the valley of Pattee Creek, and laps up onto the north slope of the ridge that includes Mitten Mountain, from which much of the material must have come. The shape and continuity of this deposit suggest that it partly filled a broad valley. Another deposit of rock debris lies in the sadddle between Wishard Creek and Johnson Gulch Creek about 2 miles south of Sheep Mountain. Most of the material in this deposit must have come from the nearby mountains; much of it probably was transported by creep. A third deposit of debris lies in the northeast corner of the quadrangle. The top of this deposit is mostly swampy. It apparently is on a remnant of an older somewhat subdued surface which has not yet been reached by the rejuvenated streams.

LANDSLIDE DEPOSITS

The two landslide deposits shown on the map (pl. 1) in the NE¼ sec. 13, T. 13 N., R. 19 W., and in secs. 2, 3, 4, 10, and 11, T. 13 N., R. 19 W., are composed of jumbled unsorted mixtures of soil and rock fragments, and were outlined by their characteristic topography as seen on aerial photographs.

A small deposit of angular frangments occupies the lower end of a gulch in the NE¼ sec. 26, T. 13 N., R. 19 W., and is probably either a landslide deposit, or a gravel bar built in glacial Lake Missoula. The front of this deposit forms part of the south wall of Hellgate Canyon, east of Missoula along the Clark Fork, and appears to have been steepened and truncated by moving water, possibly during a time when the lake was rapidly draining.

STRUCTURE

REGIONAL SETTING

The Bonner quadrangle is near the intersection of two regional lineaments, and only about 8 miles northeast of a northeast salient of the Idaho batholith (fig. 20). The development of the features must have affected the structure in the quadrangle, but the regional structure is not well enough known to determine the relation between the local and regional structures.

One of the regional lineaments trends N. 50° W., and is represented in the quadrangle by the surface trace of the Clark Fork fault. West of the quadrangle, this lineament is the straight southwestern flank of the Squaw Peak Range, which Pardee (1950, p. 390–392, and pl. 1) considered to be controlled by a fault. Farther northwest, near and northwest of Superior, Mont., this lineament is the expression of several parallel faults, of which the Osburn fault (Ransome and Calkins, 1908, p. 62) is the most widely known.

The other regional lineament, the Flathead-Bitterroot Valley line, trends north-south just west of the quadrangle. This line joins the Rocky Mountain trench to the north near Flathead Lake. The Bitterroot Valley segment of the line is believed to be related to normal faulting (Lindgren, 1904, pl. 4, p. 47–51; Pardee, 1950, p. 389–390; and Ross, 1952). The Rocky Mountain trench and similar features have been considered to be related to normal faults (Daly, 1912, p. 90; Schofield, 1920, p. 73–81; Pardee, 1950, p. 393–395); to high-angle reverse faults that dip eastward (Chamberlain, 1925, p. 759–760; Clapp, 1932, p. 24–27; Flint, 1924, p. 413); to high-angle reverse faults that dip westward (Evans, 1933); and to a zone of intense structural deformation which includes tight folds as well as faults (Shepard, 1922 and 1926).

In the latitude of the Bonner quadrangle, the Flathead-Bitterroot Valley line seems to separate areas of contrasting structural behavior. East of the line, within the quadrangle, there is large-scale thrusting but no evidence of strike-slip displacement along the northwesterly faults. To the west of the line there is no obvious large-scale thrusting, but there is evidence of large-scale strike-slip displacement along northwesterly trending faults. Wallace and Hosterman (1956, p. 590 and 591) suggest a possible strike-slip displacement of about 16 miles along the Osburn fault near the Montana-Idaho State line west of Superior.

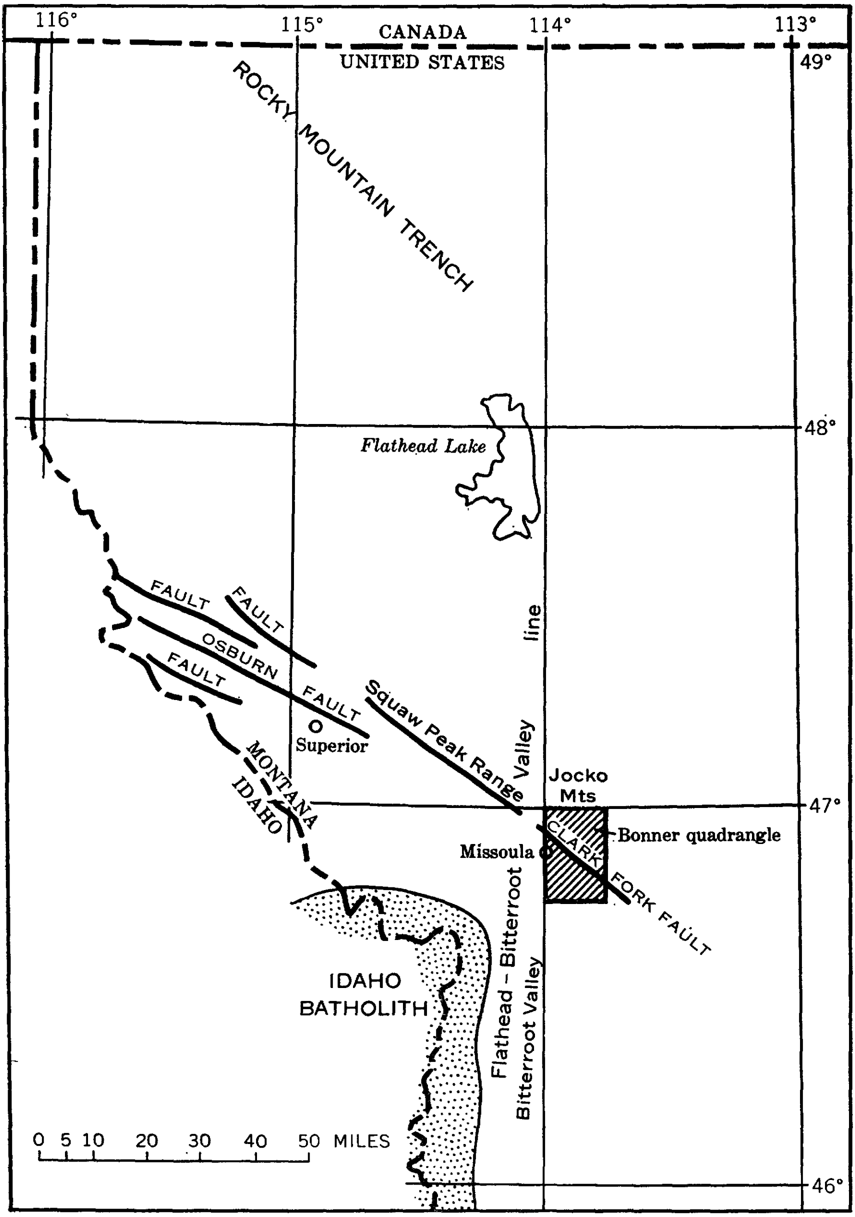

FIGURE 20.—Regional structure setting of the Bonner quadrangle.

It seems unlikely that two extensive structural features such as these lineaments could cross without one offsetting the other. The differences in the nature of the displacement on either side of the Flathead-Bitterroot Valley line suggest that the Clark Fork fault, east of the line, and the faults west of the line may be unrelated even though they are alined. On the other hand, the lineament that includes the surface expression of the Clark Fork fault may interrupt the Flathead-Bitterroot Valley line. This is suggested by a break in the continuity of the Flathead-Bitterroot Valley line by the Jocko Mountains just northwest of the quadrangle.

CLARK FORK FAULT

The surface trace of the Clark Fork fault is marked by the straight northwestward-trending segment of the Clark Fork in the east half of the quadrangle. The only probable outcrop of the fault is in the roadcut on the north side of U.S. Highway 10 just west of Marshall Creek. The amount and direction of displacement along this fault cannot be proved by the evidence in the quadrangle. Its straightness suggests that it is a high-angle fault. The structure along it is most simply interpreted as the result of relative downdropping of the south side, implying that the south limb of the Bonner Mountain anticline is the upfaulted northeast limb of the great synclinal structure that trends northward through the Miller Peak region.

The displacement on the Clark Fork fault shown by the offset in the contact between the Newland limestone and the Miller Peak argillite on section *B–B′* (pl. 35) is about 6,500 feet. Here the contact north of the fault is projected from Rattlesnake Creek, and the depth to the contact south of the fault is in part dependent on the inferred thickness of the Miller Peak argillite.

FAULTS SOUTH OF THE CLARK FORK FAULT

The block south of the Clark Fork fault is dominated by a series of southward-dipping thrust slices in which rocks of the Belt series generally dip southward. The whole faulted sequence has the form of a broad southward-plunging syncline. The trace of the synclinal axis lies to the east of the locus of maximum curvature of the thrusts, and may be shifted as much as 1½ miles transversely in adjacent slices. The thrusts are essentially parallel to bedding planes in the east limb, but depart from them and cut obliquely down section in the west limb.

The movement on these faults seems to be northward as indicated by an overturned fold in sec. 17, T. 12 N., R. 18 W. (section *B–B′*, pl. 35). As thus interpreted at least 6 miles of thrusting is required to explain the repetition of the Bonner quartzite along section *B–B′* (pl. 35). The offset of the axis of the southward-plunging syncline indicates a west-east component of movement in addition to the northward displacement.

The northeastward-trending outcrop bands of limestone and argillite in the southwest corner of the quadrangle, south of the Miller Peak fault, are interpreted as thrust slices of Newland limestone and Miller Peak argillite. (This interpretation is discussed on page 195.) As thus interpreted a minimum of about 9 miles of thrusting is required. (See section *E–E′*, pl. 35.)

The fault that cuts off the Bonner quartzite in the NW¼ sec. 7, T. 12 N., R. 18 W., cannot be located with certainty east of the Bonner quartzite because it brings McNamara argillite against McNamara argillite. It has been tentatively extended along a band of scattered diabase outcrops and float. As noted in the discussion of its age, the diabase is believed to have been intruded before the deformation of the Belt series. If so, and if the diabase actually marks the trace of the fault, then the diabase must have been dragged in along the fault. An alternative interpretation is that at least some of the diabase was intruded after some of the faulting.

FOLDS IN JUMBO MOUNTAIN

The geologic map (pl. 35), suggests an eastward-plunging anticline through Jumbo Mountain, and a syncline probably lies between it and the Bonner Mountain anticline. Section *B–B′* (pl. 35), illustrates this interpretation. The anticline that passes through Jumbo Mountain and the syncline north of it are apparently absent east of Bonner (see section *D–D′*, pl. 35).

BONNER MOUNTAIN ANTICLINE

The Bonner Mountain anticline extends from the east boundary of the quadrangle westward through Bonner Mountain. Near Bonner the axis swings northwestward and at Rattlesnake Creek it appears to turn northward. West of Bonner the anticline is overturned to the northeast apparently in response to the same compressive forces that caused the southwestward-dipping thrust faults just to the northeast. The direction of overturning on this fold indicates movement to the north and east.

WISHARD CREEK SYNCLINE

The rocks in the northeastern part of the quadrangle are folded into a broad syncline, here called the Wishard Creek syncline. The axis of this syncline is parallel to the other major structures north of the Clark Fork fault. Several small normal faults parallel the synclinal axis.

THRUST FAULTS NORTH OF THE CLARK FORK FAULT

The thrust faults north of the Clark Fork fault are between, and probably related to, the Bonner Mountain anticline and the Wishard Creek syncline. Their trends show that they dip southwestward, and drag folds such as the one in secs. 19, 20, 29, and 30, T. 14 N., R. 18 W., indicate that, at least locally, movement was toward the northeast. At several places rocks south of the faults are adjacent to older rocks north of the faults. (See sections *C–C′* and *D–D′*, pl. 35.) These relations require that folding preceded or accompanied the faulting. According to this interpretation, which is shown on the sections (pl. 35), crustal shortening of several miles has taken place along these faults. Conversely normal displacement (northeast sides up) could have produced these structural relations; movements in this direction, however, are opposite those indicated by drag folding. This interpretation also requires that the folding occur prior to or at the same time as the faulting. The map relations could be the result of strike-slip movements, and at a few places, such as sec. 3, T. 13 N., R. 18 W., no more displacement would have been required than would have been necessary to form the structures by dip-slip movements. Most of the map relations, however, would require much larger strike-slip movements, and all of the drag features indicate dip-slip rather than strike-slip movements.

Another fault, the Blackfoot fault, also apparently a thrust, is probably related to the above faults, but it diverges from them. It brings Miller Peak argillite into contact with the Pilcher quartzite; a stratigraphic displacement of at least 10,000 feet.

MINOR STRUCTURAL FEATURES

Small-scale folding and associated faulting seems to be more intensely developed in the Garnet Range quartzite than in the other formations of the Belt series. The folds are asymmetrical, fairly tight, and often isoclinal; they have amplitudes of 2 or 3 feet to several tens of feet, and wavelengths of 15 to 100 or more feet. The faults parallel the axial planes of the folds. This deformation is too small to affect the contracts on the geologic map (pl. 35), scale 1: 62,500; it is, however, indicated on the map by reversals of dip, and by the apparent disparity between the steepness of contacts and nearby bedding attitudes. This small-scale deformation extends up into and dies out in the lower few tens of feet of the Pilcher quartzite. As noted in the discussion of the age of the structures, we believe that this deformation occurred after both the Garnet Range and Pilcher quartzites had been deposited, and that the difference in the amount of internal deformation between the Garnet Range quartzite and Pilcher quartzite is due to differences in their competences.

Variations in the attitudes within the Newland limestone, the Miller Peak argillite, and the McNamara argillite, as well as small folds seen in rare good outcrops, show that these rocks are, locally at least, rather intensely deformed. These small structures parallel, and undoubtedly resulted from, the same stresses that formed the large structures of the area.

AGE OF THE STRUCTURAL FEATURES

Most of the deformation in the quadrangle appears to have occurred between Cambrian and middle Tertiary time; and most of the structural features seem to be related, and hence of similar age. The maximum age for the deformation is suggested by the deformation of the shale and limestone of Cambrian (?) age. The position of these rocks as part of the Wishard Creek syncline, one of the elements of the dominant structure of the area, suggests that the major deformation took place after these rocks were deposited. The minimum age for the deformation is provided by the basin deposits of Tertiary age which are less deformed than the older rocks. This observation indicates that the major deformation occurred before middle Tertiary time. The major regional orogeny within this interval is Laramide and it is assumed that most of the deformation in the quadrangle occurred then.

The fact that the internal deformation of the Garnet Range quartzite dies out in the lower few tens of feet of the Pilcher quartzite could indicate an episode of deformation early during the deposition of the Pilcher quartzite. This seems unlikely because the lithology of the Pilcher quartzite shows no change at this time. Another explanation, which we favor, is that the deformation occurred later. As thus interpreted, the difference in degree of internal deformation between the two formations is due to differences in the ways they responded to stress. The Garnet Range quartzite, with abundant included shale and argillite, may have been particularly susceptible to internal deformation; whereas the more homogenous Pilcher quartzite probably responded as a unit. Some of the bedding slip that would be distributed throughout a more homogenous sequence of rocks may here be largely absent from the Pilcher quartzite and concentrated in the less competent Garnet Range quartzite.

The Clark Fork fault and the other normal faults are not obviously due to compressive stresses as are the other structures of the area. This observation suggests that the normal faults may have formed at a different time than the other structures.

Some of the movement on the Clark Fork fault probably took place later than the major deformation in the area. This fault seems to displace the basin deposits at the west edge of the quadrangle, indicating movement at least as late as middle Tertiary time. This late movement, however, may be only the latest episode of a series of movements, some of which may have occurred much earlier.

PHYSIOGRAPHY

The earliest event, in the Bonner quadrangle, that can be linked with the physiographic development of the region was the deposition of the basin deposits of Tertiary age. These deposits were laid down in basins that occupied approximately the same sites as the present intermontane valleys and that were formed by blocking of drainage during Oligocene or Miocene time. The basins were slowly and intermittently depressed. The basin deposits in the Bonner quadrangle are in the east end of the Missoula-Ninemile Valley (Pardee, 1950, pl. 1), which extends northwestward from Missoula for about 40 miles.

According to Pardee (1950, p. 366), a peneplain had been formed by the time the deposition of the basin deposits had ceased. Some of the present accordant summits and uplands of low relief in the guadrangle may be remnants of this peneplain. The mountains south of Clark Fork commonly have rounded summits, and many attain fairly uniform altitudes of between 6,000 and 7,000 feet. The summit of the Garnet Range within the quadrangle also lies at about this altitude, and is generally somewhat rounded; just outside the quadrangle to the east the summit is locally flat and even boggy.

Following the development of the peneplain, the region was uplifted so that stream dissection became more vigorous. The resulting downcutting was apparently temporarily slowed to produce broad valleys. Remnants of these old valley surfaces are preserved as local benches or places of gentler slope on some of the mountain spurs between 4,000 and 5,000 feet above sea level. On these gentler slopes there are accumulations of weathered subangular boulders of locally derived rocks.

Continued uplift brought the rocks of the region to their present altitudes, and continued downcutting by the streams produced the present deep canyons through the mountains.

Local wide valleys appear to be much younger than the benches at between 4,000 and 5,000 feet above sea level. Deer Creek and Pattee Creek occupy shallow canyons incised in the bottoms of broad open valleys which are connected by a low broad divide. The streams have not cut far below the broad valley bottoms, and on this basis the broad valleys are thought to be comparatively young.

The later part of the physiographic development of the area has been markedly influenced by glaciation. Glaciers extended into the northern part of the Bonner quadrangle and sculptured part of the present topography there.

A glacier extended down Rattlesnake Creek to the southwest corner of sec. 19, T. 14 N., R. 18 W.; and the canyon of Rattlesnake Creek is U-shaped and most of the tributary valleys are hanging upstream from there. Pilcher Creek is U-shaped throughout its length. Glacially carved cirques with floors between 5,800 and 6,800 feet altitude are present in the high mountains between the Blackfoot River and Rattlesnake Creek, and in the mountains north of Rattlesnake Creek. The mountains have been rather intensely glaciated for about 10 miles north of the area. Numerous cirques are along the ridges northwest and southeast of Sheep Mountain, an arête. Most of these cirques are on the northeast side of the ridges. Two cirques occur just south of the high point in SW¼ sec. 27, T. 14 N., R. 18 W. Beeskove Creek, Pilcher Creek, and Fraser Creek have cirques at their heads. All the lakes north of Rattlesnake Creek occupy cirques; and the mountain peak in sec. 4, T. 14 N., R. 18 W., is a glacial horn.

The terminal moraine of a glacier in the valley of Rattlesnake Creek lies in the SW¼ sec. 19, T. 14 N., R. 18 W.

Two small bedrock benches along Rattlesnake Creek appear to have been cut by ice marginal streams. One, less than 100 feet above Rattlesnake Creek, is just east of the mouth of Fraser Creek, and the other is south of Rattlesnake Creek between the 4,400- and 4,600-foot contour lines in sec. 20, T. 14 N., R. 18 W.

Glacial Lake Missoula (Pardee, 1910 and 1942; and Alden, 1953, p. 154–165) occupied most of the quadrangle below an altitude of about 4,200 feet; the lake may have been excluded from part of the canyon of Rattlesnake Creek by glacial ice. Several shallow closed depressions in the saddle immediately north of Jumbo Mountain probably are behind bars and spits built in glacial Lake Missoula while Jumbo Mountain was an island. The lake was contained by an ice dam in the gorge of the Clark Fork near the Idaho-Montana State line, and occupied several intermontaine basins and interconnecting valleys of northwestern Montana.

Most of the alluvium in the valleys of the larger streams in the Bonner quadrangle was deposited during an episode, or episodes, of stream aggradation during the Pleistocene. The alluvium along Rattlesnake Creek, which is mostly glacial outwash, merges with the alluvial fill in the Clark Fork valley. At least two major stages of glaciation are recorded in the region (Alden, 1953, p. 86 and 106); and many streams were then transporting large quantities of glacial outwash material.

The Clark Fork, the Blackfoot River, Miller Creek, and Rattlesnake Creek are at present meandering and widening their flat valley floors. Above the present valley floors there are one, and locally two, terraces cut in the alluvium when the streams were graded at levels slightly above the present stream levels.

The inner canyon of Deer Creek has been cut deepest just east of the divide between it and Pattee Creek in the SW¼ sec. 6, T. 12 N., R. 18 W. This, along with the low, broad nature of the divide suggests that Deer Creek may have beheaded Pattee Creek. The distribution of colluvium across the divide between Pattee Creek and Deer Creek suggests that any possible beheading probably took place prior to the deposition of the colluvium. An alternate explanation is that the debris mapped as colluvium in Deer Creek canyon is younger than the main mass of colluvium along Pattee Creek. The incised canyon occupied by the lower part of Deer Creek appears to be graded to the level of the highest terrace cut in alluvial fill by the Clark Fork. This suggests a late Pleistocene or early Recent age for the icising.

LITERATURE CITED

Alden W. C., 1953, Physiography and glacial geology of western Montana and adjacent areas: U.S. Geol. Survey Prof. Paper 231.

Berry, E. W., 1929, Shall we return to cataclysmal geology?: Am. Jour. Sci., 5th ser., v. 17, p. 1–12.

Calkins, F. C., and Emmons, W. H., 1915, Description of the Philipsburg quadrangle, Montana: U.S. Geol. Survey Atlas, Folio 196.

Calkins, F. C., and Jones, E. L., Jr., 1914, Economic geology and ore deposits of the Coeur d'Alene district, Idaho: U.S. Geol. Survey Bull. 540–E.

Campbell, A. B., 1961, Geology and mineral deposits of the St. Regis-Superior area, Mineral County, Montana: U.S. Geol. Survey Bull. 1082–I.

Chamberlain, R. T., 1925, The wedge theory of diastrophism: Jour. Geology, v. 33, no. 8, p. 755–792.

Chaney, R. W., and Sanborn, E. I., 1933, The Goshen flora of west central Oregon: Carnegie Inst. Wash. Pub. 439.

Clapp, C. H., 1932, Geology of a portion of the Rocky Mountains of northwestern Montana: Montana Bur. Mines and Geology, Mem. 4, 30 p.

Clapp, C. H., and Deiss, C. F., 1931, Correlation of Montana Algonkian formations: Geol. Soc. America Bull., v. 42, no. 3, p. 673–697.

Daly, R. A., 1912, Geology of the North American Cordillera at the forty-ninth parallel: Canadian Geol. Survey Mem. 38.

Deiss, C. F., 1935, Cambrian-Algonkian unconformities in western Montana: Geol. Soc. America Bull., v. 46, p. 95–124.

——— 1939, Cambrian stratigraphy and trilobites of northwestern Montana: Geol. Soc. America Spec. Paper 18, 135 p.

——— 1943, Stratigraphy and structure of the southwest Saypo quadrangle, Montana: Geol. Soc. America Bull., v. 54, p. 205–262.

Evans, C. S., 1933, Brisco-Dogtooth map area, British Columbia: Canada Geol. Survey, Summary Rept. 1932A, p. 106–178.

Fenneman, N. M., 1931, Physiography of Western United States: New York, McGraw-Hill Book Co., 534 p.

Fenton, C. L., and Fenton, M. A., 1937, Belt series of the north; Stratigraphy, sedimentation, paleontology: Geol. Soc. America Bull., v. 48, p. 1873–1969.

——— 1957, Paleoecology of the Precambrian of northwestern North America, *in* Treatise on marine ecology and paleoecology: Geol. Soc. America Mem. 67, pt. 2, p. 103–116.

Flint, R. F., 1924, A brief review of Rocky Mountain structure: Jour. Geology, v. 32, p. 410–431.

Gilluly, James, 1949, Distribution of mountain building in geologic time: Geol. Soc. America Bull., v. 60, no. 4, p. 561–590.

Hanson, A. M., 1952, Cambrian stratigraphy in southwestern Montana: Montana Bur. Mines and Geology Mem. 33.

Jennings, O. E., 1920, Fossil plants from the beds of volcanic ash near Missoula, western Montana: Pittsburg Carnegie Mus. Mem., v. 8, no. 2, p. 385–450.

Knopf, Adolph, 1957, The Boulder bathylith of Montana: Am. Jour. Sci., v. 255, no. 2, p. 81–103.

Langton, C. M., 1935, Geology of the northeastern part of the Idaho batholith and adjacent region in Montana: Jour. Geology, v. 43, no. 1, p. 27–60.

Lindgren, Waldemar, 1904, A geological reconnaissance across the Bitterroot Range and Clearwater Mountains in Montana and Idaho: U.S. Geol. Survey Prof. Paper 27.

McKee, E. D., and Weir, G. W., 1953, Terminology for stratification and cross-stratification in sedimentary rocks: Geol. Soc. America Bull., v. 64, no. 4, p. 381–390.

Mertie, J. B., Jr., Fischer, R. P., and Hobbs, S. W., 1951, Geology of the Canyon Ferry quadrangle, Montana: U.S. Geol. Survey Bull. 972.

Okulitch, V. J., 1949, Geology of part of the Selkirk Mountains in the vicinity of the main line of the Canadian Pacific Railway, British Columbia: Canada Geol. Survey Bull. 14, 26 p.

——— 1956, The Lower Cambrian of western Canada and Alaska, *in* Rodgers, J., ed., El sistema Cambrico, su paleogeografia y el problema de su base—symposium: Internat. Geol. Cong., 20th, Mexico, D.F., 1956, pt. 2, p. 701–734.

Pardee, J. T., 1910, The glacial Lake Missoula: Jour. Geology, v. 18, p. 376–386.

——— 1913, Coal in the Tertiary lake beds of southwestern Montana: U.S. Geol. Survey Bull. 531, p. 229–244.

——— 1942, Unusual currents in glacial Lake Missoula, Montana: Geol. Soc. America Bull., v. 53, no. 11, p. 1569–1599.

——— 1950, Late Cenozoic block faulting in western Montana: Geol. Soc. America Bull., v. 61, no. 4, p. 359–406.

Ransome, F. L., and Calkins, F. C., 1908, The geology and ore deposits of the Coeur d'Alene district, Idaho: U.S. Geol. Survey Prof. Paper 62.

Reesor, J. E., 1957, The Proterozoic of the Cordillera in southeastern British Columbia and southwestern Alberta, *in* The Proterozoic of Canada: Royal Soc. Canada Spec. Pub. no. 2, p. 150–177.

Ross, C. P., 1949, The Belt problem [abs.]: Wash. Acad. Sci. Jour., v. 39, no. 3, p. 111–113.

——— 1952, The eastern front of the Bitterroot Range, Montana: U.S. Geol. Survey Bull. 974–E.

Ross, C. P., 1956, The Belt series in relation to the problems of the base of the Cambrian system, *in* Rodgers, J., ed., El sistema Cambrico, su paleogeografia y el problema de su base—symposium: Internat. Geol. Cong., 20th, Mexico, D.F., 1956, pt. 2, p. 683–699.

Ross, C. P., Andrews, D. A., and Witkind, I. J., 1955, Geologic map of Montana: U.S. Geol. Survey.

Rowe, J. P., 1906, Montana coal and lignite deposits: Montana Univ. Bull. no. 37, 82 p.

Rubey, W. W., 1951, Geologic history of sea water: Geol. Soc. America Bull., v. 62, no. 9, p. 1111–1147.

Schofield, S. J., 1920, The origin of the Rocky Mountain Trench: Royal Soc. Canada Trans., sec. 4, p. 73–81.

Shepard, F. P., 1922, The structural relation of the Purcell Range and the Rocky Mountains of Canada: Jour. Geology, v. 30, no. 2, p. 130–139.

——— 1923, To question the theory of periodic diastrophism: Jour. Geology, v. 31, no. 7, p. 599–613.

——— 1926, Further investigations of the Rocky Mountain Trench: Jour. Geology, v. 34, no. 7, pt. 1, p. 623–641.

Snyder, F. G., 1947, The problem of the Lipalian interval; Jour. Geology, v. 55, no. 3, pt. 1, p. 146–152.

Spieker, E. M., 1956, Mountain-building chronology and nature of geologic time scale: Am. Assoc. Petroleum Geologists Bull., v. 40, no. 8, p. 1769–1815.

U.S. Forest Service, 1939, Lolo National Forest map (east half) Montana (scale 1 inch equals 2 miles).

Wallace, R. E., and Hosterman, J. W., 1956, Reconnaissance geology of western Mineral County, Montana: U.S. Geol. Survey Bull. 1027–M, p. 575–612.

Wheeler, H. E., 1947, Base of the Cambrian system: Jour. Geology, v. 55, no. 3, pt. 1, p. 153–159.

Wheeler, H. E., and Beesley, E. M., 1948, A critique of the time-stratigraphic concept: Geol. Soc. America Bull., v. 59, p. 75–86.

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