The Belt Series in Montana

By CLYDE P. ROSS

With a geologic map compiled by
BETTY A. L. SKIPP

and a section on PALEONTOLOGIC CRITERIA
By RICHARD REZAK

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Data on the Belt series in Montana, and related rocks, with a stratigraphic classification

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ABSTRACT

Problems related to the Belt series have been discussed for nearly a hundred years, but the amount of detailed mapping and other quantitative data available, even now, is not commensurate with the importance of the problem. The present paper is an attempt to summarize existing data in the light of recent fieldwork by the writer and his associates. While the report deals mainly with the rocks in Montana, where the series was first recognized, it necessarily includes consideration of the series in Idaho and parts of Canada as well as of related rocks in more distant areas.

Throughout the Rocky Mountain region great thicknesses of dominantly sedimentary rocks intervene in many localities between the basal complex and strata of indisputably Cambrian age. In a number of places these pre-Paleozoic sedimentary rocks are divisible into at least two major subdivisions. Hinds has proposed calling the lower one Uncompahgran and the upper one Beltian, recognizing that the Belt series is one of its principal representatives. He and others, realizing that the boundary at the top of the Belt series and kindred rocks is indefinite in many places, suggest regarding all these rocks as Paleozoic although older than Cambrian. There are serious objections to this and similar juggling with long-established nomenclature, and they have not received general acceptance. In the Rocky Mountain region strata that may be allied to the Belt series include the Bannock volcanic formation of Anderson (1928) and Pocatello formation of Ludlum (1942) in southern Idaho, the Big Cottonwood formation, Mineral Fork tillite, Mutual formation and possibly others in Utah, perhaps some in Wyoming, the Grand Canyon series and Apache group in Arizona, presumably the upper part of the Precambrian sequence in southern Nevada and nearby parts of California, as well as a few units in other regions. Nomenclature and correlation in the part of the stratigraphic column that includes the Belt series are in a state of flux, pending further data.

The principal previous correlations of the Belt series are reviewed, including some involving related units in Canada. Walcott's original correlation is essentially similar to that offered in the present paper, a fact that is not obvious from the form in which it was tabulated. Correlations by Calkins, Daly, Schofield, Drysdale, Wilson and his coworkers, Sampson, and Clapp and Deiss differ among themselves in various details, but those of the American geologists agree in that they are based on the attempt to correlate carbonate-rich units throughout Montana and northern Idaho. Relatively recent work by Walker, Little, Okulitch, and other geologists of the Canadian Survey has raised queries as to their Windermere series, originally supposed to be of Precambrian age. Much of that series is now regarded as early Cambrian, whereas the lower part remains assigned to the Precambrian, with no recognized unconformity intervening. Part or all of the Windermere may be equivalent to the upper part of the Belt series in Montana and Idaho, but direct correlation of formational units is not practicable at present.

A general stratigraphic scheme is set up for the Belt series in Montana. The names employed in this scheme are in use, but the definitions have been modified. The oldest formations in the Belt series, not yet satisfactorily intercorrelated, are not included in named groups, but all others are. These groups are the Ravalli, Piegan, and Missoula, in ascending order, and the North Boulder group, which is distinguished because of its exceptionally coarse character. Much of the North Boulder group is probably equivalent to part of the Missoula group but some may be older. The formation names differ from place to place, in part merely because of local custom, in part because lateral variations are an outstanding characteristic of the series. This characteristic is among the reasons why attempts at long-distance stratigraphic correlations without adequate mapping have introduced errors.

Throughout the investigation attempts were made to find quantitative distinctions among the various units that make up the Belt series. Lithologic, petrographic, and chemical criteria are all potentially useful in this connection and some are so used at present. Stromatolites are useful as local horizon markers and may eventually aid in regional stratigraphic work. The present report points out the need for quantitative work and points out some of the sorts of data that would be useful. The data now at hand are so limited that attempts to use them can be only partially successful.

The Belt series, on the whole, is a monotonous assemblage of very thick units composed mainly of rather somber-hued, fine-grained clastic rocks, with subordinate carbonate rock. Nearly all contacts are gradational. Broad similarities throughout the Belt basin in Montana and adjacent regions are sufficient to imply essential unity. On the other hand, variations in details are numerous so that careful study permits subdivisions to be recognized and mapped. Many such subdivisions are of more limited extent than has been appreciated hitherto. The series was laid down in a wide basin whose floor and borders are nowhere well enough exposed so that its shape can be determined with accuracy. A small segment of the shore in southwestern Montana may be an old fault line. Most of the series has features that indicate deposition in shallow water, in part so shallow that exposures of parts of the sediment to the air were frequent. In such a basin local fluctuations can be assumed to have been frequent and lateral variations in stratigraphic details should be expected to have been commonplace. The character of these minor variations have only begun to be worked out.

The sediments composing the series are inferred to have been derived mainly from areas of gneiss and schist. They are thought to have been carried into the Belt basin by streams of low gradient in a fairly arid climate. Erosion in the basin at intervals during Belt time may have contributed material to later Belt deposits. Deposition within the basin was more nearly of marine than of lacustrine character but the extreme shallowness of the water made some of the deposits not typical

of either environment. The carbonate rocks, which are more plentiful than has been fully realized in the past, were derived mainly, directly or indirectly, from organisms, including algae. These rocks are not dominantly chemical precipitates. Thus the carbonate rocks are not fundamentally different in origin from similar rocks of post-Belt age.

Most of the series is so little metamorphosed that its sedimentary character is obvious. Some exposures of metamorphic rocks formerly regarded as of pre-Belt age, largely on this account, are now, with varying degrees of confidence, correlated with the Belt series.

Rocks similar in various ways to the Belt series in Montana are present in many places from Alaska to northern Mexico. Too little is known to warrant positive correlations but the Belt basin may have had connections northward to the Arctic Ocean or to an embayment west of the modern Hudson Bay. Southward there may have been connections with the ocean in the vicinity of southern California and northern Mexico. This support is afforded for the concept that a geosynclinal basin may have extended in Belt time from Alaska as far south as Mexico. Perhaps at times this basin was occupied by a continuous body of water but it is equally possible that at times there was a series of basins of deposition, with or without interconnections. Here as in similar situations throughout the world the relations between very thick accumulations of sedimentary rocks of Precambrian age and overlying Cambrian deposits are so far from being clear cut as to imply absence of any intervening major orogenic event. In Montana, in particular, unconformities at the top of rocks assigned to the Belt series are known at several localities but at other localities the contact with later rocks is conformable or nearly so. Perhaps some of the unfossiliferous rocks now assigned to the Belt series were laid down concomitantly with fossiliferous Lower Cambrian rocks elsewhere. Such a concept brings the Belt series in harmony with a worldwide pattern for similar rocks.

INTRODUCTION

PURPOSE AND SCOPE OF THE REPORT

The Belt series (Precambrian) is widespread in Montana, Idaho, Washington, and in adjacent areas in Canada. It has been studied and discussed for nearly a hundred years, but the amount of detailed mapping and quantitative study accomplished so far is not commensurate with the significance and intricacy of the problems involved in its study. The present paper attempts to summarize existing data, utilizing the writer's studies of the Belt series in Montana from 1946 through 1951, together with Rezak's studies of the fossils from 1951 through 1953 (Rezak, 1957) and work of Nelson and Dobell near Missoula (Nelson, 1959; Nelson and Dobell, 1959; Nelson and Dobell, 1961). A background is afforded by the writer's work in Idaho intermittently since 1923. The Belt series was first recognized in Montana, and the present report deals primarily with the rocks in that State, but necessarily involves consideration of the series in Idaho and parts of Canada as well as of related beds in more distant areas.

Although many papers have been published on the Belt series, much of the fieldwork has been rapid reconnaissance or incidental to study of mineral deposits in small districts. Some of the discussions have failed to consider these limitations. As a result, broad correlations and interpretations have been offered on the basis of few facts. We still lack enough facts and the series is so extensive that this will continue to be true for many years. Recently the writer, and his assistants, mapped six 30-minute quadrangles in and near Glacier National Park (Ross, 1959). This work was speeded and enhanced by the use of much hitherto unpublished data gathered in 1911 to 1914 by parties under M. R. Campbell, U.S. Geological Survey. Other less detailed studies were made by the writer and his associates near Helena, Missoula, Neihart, Whitehall, and southwest of Dillon. The mapping in these areas was needed in compiling a geologic map of the State (Ross, Andrews, and Witkind, 1955), and was one of the incentives that led to the present report.

In addition to the mapping just mentioned, key areas in Idaho were visited at various times since 1948. The value of visits to the Coeur d'Alene region, Idaho, was greatly enhanced by guidance in the field by P. J. Shenon, S. Warren Hobbs, Montis Klepper, and others who have worked there. In 1954 and 1955 Willis H. Nelson, assisted by J. P. Dobell, mapped an area close to Missoula, Mont., that is of vital significance in connection with the Belt problem (Nelson, 1959; Nelson and Dobell, 1959, 1961). Similarly recent studies from near Superior westward have been used in the present report (Wallace and Hosterman, 1956; Campbell, 1960).

After all available data relative to the Belt series are assembled and considered many aspects remain problematical. Suggestions and provisional conclusions in regard to various components of the problem of the series are presented below. However, the present paper is to be regarded as a progress report, not a final solution.

In the present report a general stratigraphic scheme is proposed for the Belt series in Montana. The stratigraphic names defined and used herein are not new but many have not been formally defined hitherto. An attempt is made to coordinate and modify the usage of the names of major subdivisions in such a way that they will not only be of value in presenting facts now available but will continue to be useful with further growth of knowledge. Units are defined primarily on the basis of their utility in mapping rather than of distinctions detectable in measured sections. The present investigation has shown lateral variations to be so prevalent that distinctions seen along any particular line of traverse can be expected to be difficult to reconcile with those elsewhere.

Plate 1 summarizes present knowledge as to the distribution of the components of the series in Montana. The large areas showing only group designations emphasize the many gaps in our knowledge. Plate 1 includes many data used for the geologic map of the State (Ross, Andrews, and Witkind, 1955) but is a new compilation by Betty A. L. Skipp showing all presently mapped subdivisions of the Belt series. It also incorporates much information not available at the time the State map was compiled.

Table 1 shows the revised nomenclature adopted in the present paper for units in northwestern Montana. Present information does not permit correlation among formations of the Missoula or Ravalli groups so that the spacing of these names in the table is correct only in the most general way. Studies now in progress by members of the U.S. Geological Survey will overcome this deficiency in part. The North Boulder group, prominent farther south, is not known in northwestern Montana and is, therefore, not shown on table 1. Much of this group is stratigraphically equivalent to parts of the Missoula group but the North Boulder group may include equivalents of parts of older groups as well, as is indicated in plate 1.

The main subdivisions, four of which are named groups, and many of the formations comprising the Belt series within Montana are now fairly well established. Detailed correlations and further subdivision of units now recognized in many parts of the State remain for the future and may lead to modifications of the established groups. Possibly the North Boulder group in southwestern Montana (not included in table 1) (Ross, 1949, p. 111, 113) will eventually be abandoned after the relations between it and other groups become better known. Correlation with neighboring regions has been attempted but must remain tentative, especially in respect to Canada.

Stratigraphic correlation between distant regions is difficult and uncertain in the absence of satisfactory maps for the intervening country, particularly for rocks like the Belt series. Past correlations between widely separated areas have necessarily been based on different geologists' personal field impressions of such features as color, texture, and composition. Identifications on such a basis are of doubtful validity unless checked by field mapping. When attempted through interpretation of printed descriptions of nonquantitative items, such as color, the result may have little value. These considerations led those concerned in the present investigation to seek quantitative criteria wherever possible, but with incomplete success.

TABLE 1.--*Correlation chart for the Belt series in northwestern Montana*
[Correlations between formations are implied only where horizontal lines are drawn]

Group	Formations			
	Vicinity of Libby	Glacier National Park	Vicinity of Missoula including Philipsburg area	Vicinity of Helena including Big and Little Belt Mountains
Missoula group of Clapp and Deiss (revised)	Top eroded Libby formation Striped Peak formation	Top eroded Mostly undivided but includes units that have been called Kintla, Roosville, Mount Rowe, Miller Peak, Shepard, and unnamed subdivisions and, near the base, the Purcell basalt	Pilcher quartzite Garnet Range quartzite Bonner quartzite McNamara argillite Miller Peak argillite with the Hellgate quartzite member locally	Greenhorn Mountain quartzite of Knopf Marsh shale Helena limestone Empire shale Spokane shale Greyson shale
Piegan group of Fenton and Fenton (revised)	Wallace formation	Siyeh limestone	Newland limestone	Newland limestone
Ravalli group of Calkins (revised)	St. Regis formation Revett quartzite Burke formation	Grinnell formation	Grinnell formation	Chamberlain shale
		Appekunny formation Altyn limestone Base not exposed	Appekunny formation	
Pre-Ravalli rocks	Prichard formation Base not exposed		Prichard formation Base not exposed	Neihart quartzite[1]
Rocks of pre-Belt age				Gneiss

[1] Relation to Prichard formation unknown.

Most components of the Belt series have broad similarities. Clearly definable and positively recognizable horizons marked by fossils or similarly useful features are rare. Thus essentially qualitative diagnostic features are now used and must be used in the future. Richard Rezak and, in an earlier stage of the investigation, Stephen Nordeng have shown during the present study that algal fossils (stromatolites) are widespread. In certain localities they are reliable horizon markers. They may prove similarly useful elsewhere when adequately studied and mapped. The stromatolites have been intensively studied only in and close to Glacier National Park. Clues to the environmental conditions during the growth of the algae obtained there are of interest in respect to the origin of the Belt series. The stromatolites are the only fossil remains known to be useful although other fossils probably exist. The incomplete petrographic and chemical data available are also summarized because of their potential value in regional correlations of formations and groups, and in ecologic studies. The areas underlain by the Belt series are so large and available information on many of them is so scanty that the application of quantitative criteria such as those just mentioned has barely begun.

PREVIOUS WORK

It is impractical to list all the numerous reports in which the Belt series is mentioned or discussed. A large proportion of the papers on mining geology in Idaho, western Montana, British Columbia, and Alberta refer to these rocks. In addition, many papers discuss in one way or another, the stratigraphy of the series. Nearly every discussion of the stratigraphic correlation contains ideas opposed to earlier ones. The present report is no exception. The comments and citations below are by no means exhaustive but are intended to outline the history of the Belt problem and to show the principal steps in investigations by many geologists that have led to current concepts.

The Belt series attracted attention as soon as geologists started to explore Montana and nearby parts of Canada. The rocks of the series were observed by Hayden as early as 1860 (Hayden, 1869), by Peale in 1872 (Peale, 1873), and were studied by Peale and others throughout the last quarter of the 19th century. Most of this work was exploratory and happened not to include those parts of Montana where the most complete exposures of the Belt series occur.

In 1884 rocks near the mouth of East Gallatin River now regarded as belonging to the Belt series were provisionally called "East Gallatin group" by Peale (Hayden, 1885, p. 50; 1888, p. 86). He later renamed these rocks the Belt formation (Peale, 1893, p. 12–20) because in the meantime Davis (1886, p. 691–703) had noted the presence of related rocks of greater aggregate thickness in the Big Belt Mountains.

At about the same time Dawson and his coworkers (Dawson, 1875) observed rocks now regarded as equivalents of the Belt series. Bauerman (1885, p. 5–32B) traversed the country near the 49th parallel in 1859–1861 and described, among others, rocks now known to belong to the Belt series (as that term is used in the United States). He included these rocks among the Paleozoic units on his structure sections, but his text indicates that he considered the idea that some of them might be as old as "Huronian." He seems to have been the first to liken structures in what is the Siyeh limestone to the markings on the molar tooth of an elephant (Bauerman, 1885, p. 26B). Daly (1912, p. 73–74) using Bauerman's simile, defined these features as molar-tooth structures and they are now known to be widespread. Bauerman seems to have confused red "shale" high in the Belt series with younger shale beds that Dawson regarded as Triassic (Bauerman, 1885, p. 30B, footnote by Dawson), a fact that emphasizes the paucity of conspicuous metamorphism in much of the Belt series.

Several of Dawson's papers (1875, 1877, 1887, p. 288–289; 1890, 1891, p. 167–173; 1895) give data on rocks of the Belt series in the general vicinity of the international boundary. In them he assigned most of the Belt series to the lower Paleozoic, but this assignment is in part a matter of definition rather than of confusion as to stratigraphic relations. In one paper (1891, p. 167–173) he recognized a great thickness of strata beneath those with recognized Cambrian fossils but was not then prepared to accept Walcott's ideas (Walcott, 1889; Powell, 1890, p. 19–20, 59–61, 66) as to the usage of the term "Algonkian" period (or system). This problem of terminology of the Precambrian system is still subject to discussion (King, 1949; American Commission on Stratigraphic Nomenclature, 1955; Raggatt, 1956, 1957; Kay, 1957; Harrison and Eade, 1957; Wilson, 1957).

Walcott, whose long-continued researches on Cambrian and Precambrian strata have done much to shape the ideas of geologists in the United States, began studies of the Belt series in 1898 in the Big and Little Belt Mountains (Walcott, 1899). At that time he set up the eight formations into which the series is still divided in those mountains and assigned a total thickness of 12,000 feet to the series there. After the initial paper, just cited, his principal contributions to knowledge of the Belt series in Montana related to their content of fossils and pseudofossils. From a more general point of view, his work did much to enhance our knowledge of the components of the Algonkian system and to define the usage of that term. Largely as a result of his efforts the U.S. Geological Survey in 1889 (Powell, 1890) defined the Algonkian as the period representing the time of deposition of clastic rocks older than the Cambrian. For Montana and adjacent regions the Algonkian period would correspond, on that basis, to the time during which the Belt series was deposited. The early history of the name Algonkian has been summarized by Wilmarth (1925, p. 103–108). Later Leith (1934) discussed the problem and concluded that most of the so-called correlations among rocks of Precambrian age do not carry the same connotation of time equivalence as those among later rock units. He proposed to redefine the Algonkian as "that part of the Pre-Cambrian sedimentary column in a particular region to which ordinary stratigraphic methods can be applied," and to speak of rocks of Algonkian type, without connotation as to strict equivalence in age between such rocks in different regions.

Calkins, working west of the area most familiar to Walcott, mapped components of the Belt series in Idaho and western Montana (Calkins, 1909; Calkins and Emmons, 1915; Calkins and Jones, 1913, 1914; Ransome and Calkins, 1908). In and near Idaho he used new formation names, most of which still stand in their original meanings, but in the Philipsburg quadrangle (pl. 2) in the central part of western Montana (Emmons and Calkins, 1913; Calkins and Emmons, 1915) he tried to use Walcott's 1899 classification set up farther east. The resulting difficulties are illustrated by the fact that in the second of the two reports just cited Calkins expressed doubts as to two of his earlier correlations. He entertains doubts as to still others among his original correlations (oral communication, 1948).

During the first 30 years of the present century numerous reports on mining districts and other restricted areas have contributed facts in regard to the Belt series. Some of them have already been cited and others will be mentioned in the descriptive part of the present report. Attention may be called to Willis' (1902) account of the rocks in part of what is now Glacier National Park. Several of the subdivisions of the Belt series now in general use were first proposed by Willis (1902). Barrell (1906, 1907) contributed ideas as to both correlation and origin. Daly (1912) mapped, described, and discussed the Belt series along the international boundary. He covered 2,500 square miles of geologically unknown country in six field seasons. Some of his mapping and, hence, his ideas as to local stratigraphic relations have been modified by later evidence gathered by Schofield (1914b, 1922, 1923) and later workers such as Drysdale (1917) but Daly's report remains an instructive and inspirational summary of many features of the Belt problem. For convenience in mapping along the 400 miles of the international boundary, Daly assigned local series names to certain geographic segments of the boundary area. Some of these series names are still in use in Canada but with somewhat modified meanings. Certain of the formation names used by Daly were applied to parts of several of his series, a practice that has contributed to the uncertainties that still exist in attempts to carry stratigraphic correlation across the international boundary. The most fundamental differences between Daly's work and that of his successors is that Daly regarded most of the Belt series as of Cambrian age and conformable under later Paleozoic rocks, whereas, largely on fossil evidence, later workers regard most of the series as Precambrian and in places apparently unconformable under rocks of Early Cambrian age. Differences in the correlations of Daly and his immediate successors are outlined on pages 13–17. Umpleby and Jones (1923) summarized data on the Belt series in Shoshone County, Idaho, and Kirkham and Ellis (1926) did the same for Boundary County at the northern tip of that state and suggested correlations with units in Canada. Sampson (1927) made a general study of the series but published only a brief abstract. However, a summary of his unpublished conclusions has been utilized in the present report. During the 1920's geologists of the Montana Bureau of Mines and Geology and the University of Montana had undertaken general studies of the Belt series, but their conclusions were published only in part (Wilson, Lambert, and Clapp, 1924; Bevan, 1929).

By 1930 the general distribution and character of the Belt series in Idaho, Montana, and adjacent parts of Canada were becoming known and details were accumulating for some areas. Those regional stratigraphic correlations that had been attempted were based so largely on data from such widely spaced regions that they necessarily involved speculation and uncertainty. Problems of origin and of correlation with units at a distance had only begun to be considered even though the Belt series had long been recognized as a major factor in any scheme of subdivision and nomenclature of Precambrian rocks (Van Hise, 1892, p. 272–347; Walcott, 1899).

During the next 20 years data were gathered in various areas. Most of the areas were small, but one of the largest was an area of about 3,000 square miles in the general vicinity of Helena, including part of the Big Belt Mountains described by Pardee and Schrader (1933). Hume (1933) and Brock (1934) clarified points in regard to the Belt series and allied rocks in Canada. Between them, they indicated that the series is, on the whole, of Precambrian age and that much of the highly metamorphosed Shuswap series or terrane is to be correlated with it. A recent summary of pertinent data (Reesor, 1957, p. 172–175) shows that the Shuswap terrane is definitely of pre-Permian age and may be of early Precambrian age or more probably may include components of Belt and Paleozoic ages. Solution of the problem of the age of the assemblage thus remains open. For present purposes it may be assumed that these rocks may include some of the most westward exposures of rocks of Belt age.

Among the Canadian publications during this period are papers by Rice (1937, 1941), Okulitch (1949), and Little (1950) that contributed much toward knowledge of the Belt series and associated rocks of Idaho and northern Washington. It should be noted, to avoid confusion, that Canadian geologists, following the practice employed by Daly, apply various local series names to rocks equivalent in whole or in part to the Belt series, as that term is used in the United States. So little relationship exists between the stratigraphic names in use in this country and those used in Canada that summaries of the kind included in the papers just cited are very valuable, although beyond the scope of the present report to discuss exhaustively.

Fenton and Fenton in a series of papers (1931, 1933a, 1933b, 1936, 1937a, 1937b) have discussed the Belt series in all its aspects. Their data on the fossil content of the rocks of the series are especially interesting and furnished inspiration for the studies of algal fossils more recently undertaken under the general direction of the present writer (Rezak, 1957). Rezak has contributed a summary of his ideas on the subject of fossil remains in the Belt series to the present report. Fenton and Fenton made suggestions for detailed stratigraphic subdivisions but based them on individual sections rather than on geologic maps. Consequently their suggestions are not of regional usefulness. The same can be said of many of the published papers relative to stratigraphic nomenclature and correlation in the Belt series.

C. H. Clapp, while serving as president of the University of Montana, devoted much attention to problems of the Belt series. He published in 1932 a summary paper that includes a map of a large part of northwestern Montana at a scale of 1:500,000, a stratigraphic table, and a bibliography. That paper is too condensed to permit the citation of many of the facts on which his conclusions are based but serves to present his general ideas in orderly fashion. Another paper by Clapp and his associate, Deiss (Clapp and Deiss, 1931), gives additional details in regard to the Belt series but without including any geologic maps. Their paper is especially notable for calling attention to the widespread occurrence of calcareous beds near the middle of the series and for naming and describing a group of beds characterized by red and green argillaceous rocks in the upper part of the series (Missoula group). Clapp and Deiss emphasize the value of the calcareous rocks in correlations throughout the region in which the Belt series is exposed. Their conclusions reached on this basis are fundamental but need review because of the marked lateral variations in the rocks that are now believed to exist.

Langton (1935) later mapped some of the rocks of the series that Clapp and Deiss had named. Clapp died in 1935 while in the midst of his long study of the Belt series in Montana. Had he been able to summarize his ideas our knowledge of these rocks would have been further enriched. Deiss continued to work in Montana for some years after Clapp's death. He has published two papers, without geologic maps, describing the geology of specific areas in Montana (Deiss, 1943a, 1943b). These and the paper by Clapp and Deiss, just cited, show so many differences in the rocks of the Belt series in localities fairly close together as to constitute examples of the lateral variations that are characteristic of that series. Other general papers by Deiss (1935, 1936, 1938, 1940, 1941) and others deal primarily with Paleozoic (especially Cambrian) rocks, but they include information and concepts relative to the Belt problem. They are especially helpful in connection with the character of the contact at the top of the Belt series (p. 99).

Some of the more westerly of the exposures of strata of apparent Belt age within the United States are in the northeast corner of Washington (Park and Cannon, 1943, p. 5–13). The data assembled by Park and Cannon are supplemented by several illuminating papers on areas in Canada (Rice, 1937, 1941; Okulitch, 1949, 1956; Little, 1950; Reesor, 1957). The problem in this general region is to determine where the boundary between Cambrian and Precambrian strata is. Though the writers cited differ as to its exact position, they agree in postulating that the bulk of the strata in question are Cambrian. Daly (1950, p. 741–743) in a recent paper regarded all the strata as Precambrian and allied to the Belt series.

Gibson and his associates (Gibson, Jenks, and Campbell, 1941) have described the stratigraphy of the Belt series along the northern part of the border between Montana and Idaho. Gibson (1948) later published a report containing a geologic map of part of the region whose stratigraphy was discussed in the 1941 paper. Since then detailed mapping in and near the Coeur d'Alene region, Idaho, has been undertaken for the U.S. Geological Survey by S. Warren Hobbs and his associates. Their work has been carried as far east as Superior, Mont., but early in 1959 only a few preliminary maps and papers had been published. All information has been offered generously for use in the present study. It shows that broad similarities with the stratigraphy of the Coeur d'Alene region, as worked out long ago by Calkins, persist as far east as Superior, but it serves equally forcibly to show that there are numerous variations in stratigraphic characteristics throughout the broad region being studied.

Studies by the present writer and his associates, notably Richard Rezak, in and south of Glacier National Park (Ross, 1959; Rezak, 1957) have added to data on the Belt series in that part of Montana. In addition, reconnaissance work in 1946 through 1948 by Ross, Richard Harris, and H. W. Burke for use in compiling a new geologic map of Montana (Ross, Andrews, and Witkind, 1955) furnished other information on the series. A brief paper by Ross (1949) seeks to clarify the nomenclature of the series. Since the late 1940's many students from the Universities of Montana and Indiana, the Montana School of Mines, and other institutions have worked in the southwestern part of Montana in fulfilling the requirements for Master's and Doctor's degrees. Some of them worked in areas containing rocks of the Belt series, and data accumulated by them have been used in the present report.

In 1946 Mertie, Fischer, Hobbs, and their associates (1951) mapped the Canyon Ferry quadrangle (pl. 2) east of Helena that includes part of the Big Belt Mountains, the type locality of the Belt series. Among other things, their work, like most recent studies, emphasizes the gradational character of contacts between the components of the series. At the same time and in subsequent years Adolph Knopf (1950) and his wife, Eleanora B. Knopf, have worked near Helena. Montis Klepper and other members of the U.S. Geological Survey are working in the region near Boulder, and parts of their results have been published (Klepper, Weeks, and Ruppel, 1957). Field conferences with Klepper have been especially enlightening. Perry (1950) of the Montana School of Mines has furnished a convenient summary of information on the series.

Many geologists in addition to those mentioned here have dealt with components of the Belt series but the summary above will give an idea of the scope of available data.

THE SERIES NAME AND ITS HISTORY

Within the United States, usage of the term "Belt series" is now fairly well standardized. It corresponds in essence to the definition (Wilmarth, 1925, p. 112):

> The United States Geological Survey classifies the Belt as a provincial series belonging to the Algonkian system. The name is derived from the extensive development of the rocks in the Big Belt and Little Belt mountains of central Montana.

The principal recent official departure from this definition is in the fact that the U.S. Geological Survey does not now recognize formal subdivisions of Proterozoic time (Geologic Names Committee, 1953, p. 15). Hence, the Belt series is no longer referred to as belonging to the Algonkian system, although the series remains high in the Precambrian sequence. This holds even though lead in veins cutting part of the series has been determined to be older than would have been expected on other grounds (Kerr and Kulp, 1952; Eckelmann and Miller, 1956). The age based on these determinations is still far less than ages assigned to certain other Precambrian rocks on the basis of similar determinations. The significance of age determinations based on laboratory determinations of this character remains a subject of debate.

The first formal name applied to rocks now included in the Belt series was the East Gallatin group. This name was "provisionally" applied by Peale, then Hayden's assistant, (Hayden, 1885, p. 49-50) to rocks near the mouth of the East Gallatin River, east of Three Forks, Mont. The same name was later used by Hayden (1888, p. 86) for rocks in the Bridger Range, evidently including the exposures to which the name had been applied originally by Peale. Hayden then thought of these rocks as probably of Middle Cambrian age. Somewhat later Peale (1893, p. 12–20) spoke of them as the "Belt beds" or "Belt formation." He said that the "group could be only of Lower Cambrian or of Algonkian age," and preferred the latter designation. Clearly "East Gallatin" has priority over "Belt" as the formal designation of the series. Keyes (1927) has pointed this out and suggested that the assemblage should be called, in consequence, the East Gallatin group. However, that name was originally applied only to a small part of the series and it was soon abandoned by Peale. Belt series, or some modification or derivative of the term, has been applied for such a long time and so generally within the United States that a change at present would be obviously undesirable.

The first geologic study of the rocks under discussion in the Big and Little Belt Mountains was by Davis (1886), but no formal name was used. Instead, the rocks were referred to simply as "barren slates" and regarded as of Early Cambrian age. After Walcott began his studies in the same region he used the expression "Belt terrane" (1899, p. 201). When Van Hise (1892) prepared his first correlation papers on the Archean and Algonkian, he felt that so little was known in regard to the rocks now termed the Belt series that his descriptions of them were brief and without formal names. In a later paper (Van Hise and Leith, 1909) the term "Belt series" was used extensively. By the time that paper was written the term had evidently come into general use in Montana, although names such a formation, group, and terrane appear to have been used by some as more or less synonymous with series. Correlations between the Belt series in Montana and similar rocks in northern Idaho and in British Columbia and Alberta were being suggested. Since then the term "Belt series" has come to be generally accepted for the sedimentary rocks in Montana, northern Idaho, and adjacent parts of Washington that are older than recognized Cambrian strata and younger than certain more metamorphosed rocks, even though the metamorphosed rocks of that type can not be closely dated. Bailey Willis (1912, p. 14, 16, 62, 64, 67) in his general summary of stratigraphic data cited the name in this sense. At present the name is not definitely applied to rocks south of the 44th parallel, and some problems remain as to its application in certain places north of this line; for example, the Metaline quadrangle in Washington (Park and Cannon, 1943, p. 6–11), where the distinction between Precambrian and Cambrian rocks is not entirely clear.

Canadian geologists appreciate the fact that many rocks in British Columbia and Alberta, and perhaps also farther north, are equivalent to the Belt series in the United States, but that name is not now used in Canada for specific map units. Some Canadians use Belt, or its derivative Beltian, in a time sense (Schofield, 1914b, 1922). Among the major units formerly employed by Canadian geologists for rocks that appear to be in whole or in part correlative with components of the Belt series may be mentioned the Galton, Lewis, Purcell, Nisconlith, Selkirk, Shuswap, and Windermere series, and Priest River terrane. In recent Canadian papers the rocks equivalent to the Belt series of the United States are grouped as the Purcell and Windermere series. The age relations of the highly metamorphosed Shuswap rocks remain uncertain but rocks of Belt age may be included.

The Belt series is regarded by most students as of distinctly late Precambrian age. This assignment is in harmony with the data gathered during the present investigation except that, as will be shown below, there is a possibility that part of the rock now regarded as belonging to the series may have been deposited simultaneously with rocks whose fossil content shows them to be of Cambrian age. Many of the Belt strata contain fossils but not such as to be of value at present in regional correlations. Future geologic mapping over large areas now imperfectly known will aid in correlations within the series and in discovering relations between parts of it and rocks to the north and west that contain Lower Cambrian fossils. Recent determinations of absolute age by means of studies on radioactivity range within rather wide limits. One figure is 750 million years (Kerr and Kulp, 1952); another is at least 1,200 million years (Eckelmann and Miller, 1956). These suggest a greater age for parts of the Belt series than might otherwise be assumed, but neither figure is old in comparison with other determinations from Precambrian rocks (Farquhar and Russell, 1957). When techniques are perfected and many more samples studied, one can hope that radioactive determinations will go far to substitute for the lack of diagnostic fossils in stratigraphic work in Precambrian rocks.

In view of the fact that the stratigraphic column in the Lake Superior region has, in the past, been thought of as a standard for the Precambrian rocks of North America, it is of some interest to note that the Belt series and similar strata have been referred (Cooke, 1933, p. 471–474) "doubtfully" to the Keweenawan epoch. The reference is logical but obviously is not susceptible of proof on the basis of the data now available. So much of the area between Lake Superior and Montana is underlain by post-Proterozoic rocks that direct correlations by means of mapping are impossible. Perhaps some support for the concept is afforded by the fact that the boundary between Keweenawan and Cambrian rocks is locally a matter of controversy (Raasch, 1950). Rocks termed the "Bayfield group" by Raasch are regarded by him as of Middle and possibly also of Early Cambrian age, whereas it has been argued that the group may be of Keweenawan (late Precambrian) age.

The Ocoee series in the Appalachian region has been likened to the Belt series (King, 1949, p. 638). The differences are almost as numerous as the resemblances, but the two widely separated provincial series are alike in that interpretations as to the position and character of their upper limits and the relations to strata containing Lower Cambrian fossils continue to vary with different authors (King, 1949; Rodgers, 1956). This fact suggests possible approximate similarities in age between the two series. Uncertainties as to the Precambrian-Cambrian boundary are common throughout the world and have led to such suggestions as the one to set up a Beltian system (p. 9). In figure 7 the distinctions for the Appalachian region between Lower Cambrian rocks and Precambrian rocks just beneath them have been made by J. B. Hadley. Boundaries, mainly, are taken from the new geologic map of North America now in preparation.

Resser (1933, p. 743, 745, 746; 1938a, p. 2) advocated the use of the name Beltian for a system of rocks of Precambrian age but which he included in the rocks of the Paleozoic era, thus extending the time span of the Paleozoic beyond that ordinarily assigned to the era. His opinions are based largely on data from the Appalachian region, although he also had acquaintance with the Belt series in Montana. Use of the name Beltian system has been considered for broadly similar rocks in Asia (Grabau, 1922, p. 72–79; Endo and Resser, 1937, p. 43) but the older name Sinian has been elevated to systemic rank and applied in China and Manchoukuo (Manchuria) (Matsushita, 1934, 1935; Nakamuro and Matsushita, 1940, p. 311–319) to the rocks there that Resser evidently regarded as correlatives of his Beltian system.

Rocks that resemble the Belt series in their approximate position in the stratigraphic column, lithologic characteristics, and stromatolite content are known in many other parts of the world. Among these are Brazil (Almeida, 1944), Mexico (Arellana, 1956), South Africa (Macgregor, 1940; Schwellnus and le Roex, 1945; Young and Mendelssohn, 1949), Finland (Sederholm, 1932; Von Eckermann, 1937), Greenland (Katz, 1953; Koch, 1929a, 1929b, 1933; Troelsen, 1949, 1950), Australia (Fairbridge, 1950), India (Rao, 1949), and Siberia (Maslov, 1937, 1938). The rocks concerned are all so little metamorphosed that their sedimentary character is readily apparent. Like the Belt series, doubts have been expressed as to whether or not they are of marine origin and whether they should be regarded as early Paleozoic or older. In some instances gradation between them and beds of Cambrian age has been suggested. Nearly all contain stromatolites. Perhaps when these fossils are better understood their presence will aid in stratigraphic correlations between the widely separated assemblages of ancient sedimentary rocks, each of which is a thick and conspicuous component of the stratigraphic column in its own region.

In the western United States, Hinds (1936a, p. 6–7) advocates the use of Beltian in virtually the sense that Resser proposed. In 1935 he spoke of Beltian time as the upper part of the Algonkian period. Later he (1936b, p. 58) said that "pre-Beltian time probably should be established as a separate period of the Paleozoic." He proposed to call his new subdivision of the Paleozoic era the Uncompahgran period, a name taken from the Uncompahgre formation in southwestern Colorado. Thus, Hinds suggested that the Algonkian period as then defined by the U.S. Geological Survey be made into a part of the Paleozoic era and subdivided into a lower Uncompahgran period and an upper Beltian period. In a later paper (1940, p. 307) he spoke of his Uncompahgran deposits as unconformably underlying the Algonkian and unconformably overlying the Archean or earlier post-Archean complexes. In that paper he appears to regard Algonkian and Beltian as synonymous terms.

Okulitch (1956, p. 729) has suggested tentatively that the lower part of the Windermere series of Walker (1926) be raised to systemic rank, as the earliest system of the Paleozoic sequence. The rocks involved are below the fossiliferous Lower Cambrian strata and above the Purcell series. They belong in the upper part of the Belt series of the United States. Gussow (1957, p. 9) has suggested that the proposed system be called the Lipalian system. He further writes that

> the Purcell series and the Belt series are assigned to the Beltian system (as are the Grand Canyon series, the Llano series, et cetera—Walcott) and the Grenville series in eastern Canada, based on age dating and the presence of Cryptozoans or *Eozoon canadense*.

He does not regard either of his two proposed systems as Paleozoic. One of the difficulties that would be met in applying Gussow's ideas is that the major unconformity between his Lipalian and Beltian systems has been recognized only in parts of British Columbia near long. 117°. It is not known anywhere in Montana although beds believed to be equivalent to the Precambrian part of the Windermere series are plentiful there.

The suggestions made by Resser, Hinds, Okulitch, and Gussow have attractive features and serve to emphasize the fact that in many places throughout the world there are thick assemblages of sedimentary rocks that are now regarded as Precambrian but in character and relations have resemblances to overlying rocks assigned with reasonable certainty to the Cambrian. In some places evidence of unconformity at the base of the Cambrian rocks is obscure or lacking. Possible correlatives of the Belt series are exposed at intervals from Alaska to northern Mexico and others with some stratigraphic similarities are scattered from Greenland into the region around Lake Superior and stretch from Newfoundland and Nova Scotia southwestward through the Appalachian Mountains (Fig. 7). It would be convenient to have a single name designating these North American rocks and the apparently similar ones in other parts of the world. The name that has received most favor so far is Beltian system, but that name has the disadvantage that a root word in use for a provincial series would be applied to a system of worldwide scope. This reason alone seems sufficient to justify rejection of the term Beltian. Further, expansion of the term Paleozoic era so that rocks, such as the Belt series, could be regarded as of that age would seem to introduce more confusion than compensating advantages, especially as the rocks to be added are without useful fossils. The matter is involved in the general problem of correlation and subdivision of Precambrian rocks, whose solution is beset with so many difficulties that it is not being attempted by geologists at the present time. The resemblances between the various very thick assemblages cited previously do seem real and significant, but a major change in long-established nomenclature does not seem the best way to emphasize them.

STRATIGRAPHIC CORRELATIONS OF THE SERIES

Correlation of components of the Belt series is among the fundamental problems in the geology of western Montana, Idaho, and adjacent areas. Most papers that include data on the Belt series deal with this problem and several different correlation schemes have been published.

Although available information remains inadequate in some respects, the Belt series in Montana can now be divided with considerable confidence into several groups. Many named formations are on record. Some have been assumed to be of widespread distribution, but existing geologic maps are inadequate to put such assumptions on a firm basis, a fact that has given rise to many differences of opinions. Some of the formations are of only local significance. Most of the broad correlations that have been offered are based on relationships among carbonate-rich rocks, but such rocks throughout the series tend to have more resemblances than diagnostic differences so that it is easy to make errors in interpretation. The absence of diagnostic fossils, coupled with the abundance of gradational contacts and marked lateral variations, contribute to the difficulties. During the present study factors that might give a quantitative basis for recognition of stratigraphic units were sought but with only partial success. Lithologic, petrographic, chemical, and paleontologic data were assembled for this purpose.

The formally named groups and formations are defined here as precisely as circumstances permit, with the modifications of published data that now seem warranted. The descriptions are as quantitative as possible, although deficiencies in this respect are inherent in the character of the rocks and in the perennial incompleteness of knowledge. Perhaps one should say that present descriptions point the way to quantitative statement of essential criteria, rather than that this aim has been attained. Each of the stratigraphic subdivisions of the series tends to grade into the next unit. Many of the characteristics regarded as critical in making stratigraphic distinctions comprise minor, subtile details. Expression of criteria of this kind in exact terms is difficult but is even more desirable for rocks like the Belt series than the listing of diagnostic fossils or readily recognizable lithologic characteristics for units that possess such features. Subjective factors in identification of the stratigraphic components of the Belt series need to be minimized, although they must, obviously, continue to be taken into account.

PREVIOUS CORRELATIONS

WALCOTT'S CORRELATION

The first correlation of components of the Belt series was published in 1906 (Walcott, p. 1–28). By that time most of the region containing the series already had been visited by geologists. The results of Daly's work in 1901 to 1906 (Daly, 1912) were available to Walcott, although published later. Daly's report, although partly revised by later workers, remains the best available summary of the geology of the western part of the international boundary. Within Montana much of the early work on the Belt series was done either by Walcott personally or by his associates. Many of his conclusions still stand. His principal correlations are summarized in table 2, taken directly from his report of 1906.

Walcott in 1906 regarded the Newland limestone, the Altyn limestone, his Blackfoot series, and the Wallace formation as stratigraphic equivalents. In consequence he equated the Helena and Siyeh limestones and thought that the Grinnell and Appekunny argillites together were equivalent to the assemblage known as the Empire, Spokane, and Greyson shales. He thought of the carbonate rocks at the north end of the Swan Range (Badrock Canyon) as equivalent to his Blackfoot series. These carbonate rocks are now known to belong to the Siyeh limestone (Ross, 1959), which Walcott put much above his Blackfoot series. Apparently Walcott changed his mind on this point for Daly (1912, p. 183) cites a letter from Walcott stating that he had proved the equivalence of the Blackfoot with the Siyeh. This is in accord with modern ideas (Clapp, 1932; Ross, 1959). Clapp and Deiss (1931, p. 675) state that the limestone Walcott thought was overlain by his Camp Creek series on Camp Creek in the Mission Range is actually a downfaulted block of Paleozoic limestone. Although this statement is based on later mapping and can be accepted as presumably correct, there can be little question that Walcott saw extensive exposures of limestone of Precambrian age in and near the Mission Range that he correlated with his Blackfoot series. Perhaps the confusions as to correlation may be re-

TABLE 2.--*Precambrian sections of northwestern Montana and northern Idaho*

[Quoted from Walcott (1906, p. 18)]

Belt Mountains, Montana (Walcott)	Dearborn area, Montana (Walcott)	Lewis and Clark area, Montana (Walcott)	Lewis and Livingston Ranges, Montana (Willis)	Camp Creek, Mission Range section, Montana (Walcott)	North and northeast of Coeur d'Alene, Idaho (Calkins)	Boundary section east from Kootenay River (Daly)
Cambrian. Unconformity. MARSH, 800 ft.[1] HELENA. Calcareous. 2,400 ft. EMPIRE, 600+ ft. SPOKANE, 1,500+ ft. GREYSON, 3,000+ ft. Arenaceous strata. 5,100 ft. NEWLAND. Calcareous, 2,200+ ft. CHAMBERLAIN. Siliceous. 1,500 ft. NEIHART sandstone. 700 ft. Unconformity. Archean. Total section, 12,000 ft.	Cambrian. Unconformity. Siliceous and calcareous. 945 ft. Greenish and purple, arenaceous and siliceous strata. 5,772 ft. Base concealed. Total section, 6,718 ft.	Cambrian. Unconformity. Arenaceous. 1,015 ft. Limestone, 285 ft. Arenaceous. 1,210 ft. Base concealed. Total section, 2,510 ft.	No superjacent strata. KINTLA. SHEPPARD. Quartzites, 1,200 ft. SIYEH limestone. 4,000 ft. GRINNELL. APPEKUNNY. Siliceous. 3,800 ft. ALTYN, calcareous and siliceous. 700 ft. Base concealed. Total section, 9,700 ft.	Cambrian. Unconformity. CAMP CREEK SERIES. Arenaceous-gray. 1,762 ft. Calcareous and arenaceous. 1,560 ft. Arenaceous, mostly reddish color. 4,491 ft. Arenaceous red and gray colors. 198 ft of limestone near 700 ft from summit. BLACKFOOT. Calcareous and siliceous. 4,805 ft. RAVALLI. Siliceous and arenaceous. Purple, greenish, and gray beds. 8,255 ft. Base concealed. Total section, 24,770 ft.	No superjacent strata. STRIPED PEAK. 2,000 ft. WALLACE. Calcareous and siliceous. 6,000+ ft. BURKE-SAINT REGIS. Siliceous and arenaceous. Purple, greenish, and gray beds. 8,000 ft. PRICHARD. Bonded, dark blue gray, blue-black, and gray. Siliceous series. 10,000 ft. Base concealed. Total section, 25,000 ft.	No superjacent strata. YAHK, 500 ft. MOYIE. Argillite. 3,200 ft. KITCHENER quartzite. 7,400 ft. CRESTON quartzite. 9,500+ ft. Base concealed. Total section, 20,600 ft.

Total thickness of section of Precambrian rocks in northwestern Montana and northern Idaho, as now known, 37,000 ft.

[1] Walcott (1899) gives the thickness as 300 ft.

lated to the error in interpretation of structure to which Clapp and Deiss call attention. The type locality of the Blackfoot series (Walcott, 1906, p. 5) is in the canyon of the North Fork of the Blackfoot River, far to the east of the Mission Range. At that locality Clapp and Deiss (1931, p. 675) agree that the limestone belongs to the Belt series. They (Clapp and Deiss, 1931, p. 691) think of this limestone as equivalent to the Helena, whereas Walcott (1906, p. 20) equated it with the Newland. Barrell (1906) followed Walcott.

The heights of the blocks in Walcott's table (table 2 of this report) are approximately proportional to the thickness of the stratigraphic units. Consequently equivalent units in different areas are not in line with each other horizontally as they are in most correlation tables. Further, certain of the columns in the table fail to include formation names that are given in the text. These items, together with the fact that the geographic terms heading some of the columns are either obsolete or indefinite, render the table somewhat confusing to readers at present. Therefore, table 3 is appended which is intended as a clarification of Walcott's original. Where horizontal lines are omitted in the second table, the correlations intended by Walcott are not known.

Apparently the Camp Creek mentioned in Walcott's table (table 2 of this report) is the same one so designated on present-day maps. This stream is not in the Mission Range, as would be inferred from his table, but is farther east in what is now known as the Flathead Range.

Walcott's (1906, p. 19) original correlation of the Newland and Altyn limestones is based in part on their fossil content. Rezak's work (p. 33) shows that similarities in fossils exist but that they hardly warrant the confidence Walcott places in them. The forms now regarded as of organic origin are stromatolites that appear to have a wide stratigraphic range. The evidence of stromatolites could be used at least equally well in arguing for correlation of the Newland with the Siyeh, far above the Altyn in stratigraphic position. Walcott's Blackfoot series would, then, according to his

TABLE 3.--*Walcott's correlation table of 1906 as interpreted by C. P. Ross*

Big and Little Belt Mountains (Walcott)	South of the upper reaches of Dearborn River (Walcott)	Rogers Pass (Walcott)	Glacier National Park (Willis)	Camp Creek in Flathead Range (Walcott)	North and northeast from Coeur d'Alene area, Idaho (Calkins)	Area along international boundary east from Kootenay River (Daly)
Cambrian	Cambrian	Cambrian	No superjacent strata.	Cambrian		
Marsh shale, 800 ft.[1]	Sandstone, 510 ft.	Marsh shale, 1,015 ft.	Kintla argillite and Shepard formation, 1,200 ft together.	Camp Creek series: Sandstone, 1,762 ft.		
Helena limestone, 2,400 ft.	Helena limestone, 435 ft.	Helena limestone, 285 ft.	Siyeh limestone, 4,000 ft.	Sandstone, shale, and limestone, 1,560 ft.		
Empire shale, 600+ ft. Spokane shale, 1,500+ ft. Greyson shale, 3,000+ ft.	Sandstone, 5,772 ft.	Empire shale, 1,210 ft. Base concealed.	Grinnell argillite and Appekunny argillite, 3,800 ft together.	Sandstone, 4,491 ft.	Striped Peak formation, 2,000 ft.	
Newland limestone, 2,200+ ft.	Blue-gray limestone, 15 ft. Base concealed.		Altyn limestone, 700 ft. Base concealed.	Blackfoot series 4,805 ft.	Wallace formation, 5,000 ft.	
Chamberlain shale, 1,500 ft. Neihart sandstone, 700 ft.				Ravalli series 8,255 ft.	St. Regis formation, Revett quartzite, and Burke quartzite, 8,000 ft together.	Yahk quartzite, 500 ft. Moyie argillite, 2,200 ft.
					Prichard formation, 10,000 ft.	Kitchener quartzite, 7,400 ft. Creston quartzite, 9,500+ ft.

[1] Walcott (1899) gives the thickness as 300 ft.

original concepts, also be equivalent to the Siyeh limestone. Calkins (1909, p. 38, 40; Emmons, 1907, p. 34; Ransome and Calkins, 1908, p. 26–27) regarded the Blackfoot series as equivalent stratigraphically to the Wallace formation of the Coeur d'Alene area, the Newland limestone of "The original Belt section," and the unit in the Philipsburg quadrangle that was later (Emmons and Calkins, 1913, p. 41-44) designated as Newland. This is the correlation accepted in the present report. Calkins, however, at that time, retained the concept that the Altyn, rather than the Siyeh, was the stratigraphic equivalent of the Newland.

The Camp Creek series of Walcott was regarded by Walcott as overlying the Blackfoot series (of his nomenclature) in the Mission Range. This and its general lithologic character would make it correspond to the Missoula group of present usage, or to a large part of that group. In the light of present data, Walcott's correlation table agrees with such an assignment. In the first report on the Philipsburg quadrangle (Emmons, 1907, p. 34) Calkins is quoted as using the term Camp Creek formation for the rocks he later called the Spokane formation (Emmons and Calkins, 1913, p. 45–48). The latter constitute the representatives of the Missoula group in the Philipsburg quadrangle. Farther west, in 1903–05, Calkins (1909, p. 38–41) concluded that the Striped Peak formation corresponds in lithologic character and position to the lowest part of Walcott's Camp Creek series and, further that the

Algonkian beds above the Newland, including those referred to the Striped Peak and those supposed to be younger, are probably roughly equivalent to Walcott's Camp Creek "series", although a definite correlation is hardly justifiable as yet.

In a more recent published correlation table (Calkins and Emmons, 1915, p. 3), Calkins indicated that the Camp Creek series of Walcott is equivalent to the whole of the Missoula group as that term is used in the present report. As the term Missoula group had not then been proposed, it does not appear in Calkins' table, but the intent of his correlation is plain.

The above remarks show that a single adjustment in Walcott's correlation table would bring it into virtual agreement with the correlations proposed in the present report so far as localities within the United States are concerned. All that is necessary is to shift his entire column representing Willis' work in what is now Glacier National Park downward so that the Siyeh is shown as equivalent to the Blackfoot, Newland, and Wallace. All other units could remain in their same relative positions. The resemblance between Walcott's ideas and those of the present writer is brought out better in table 3 than in Walcott's original in which some formation names are omitted. The correlations suggested in Walcott's table between sections in the United States and the one in Canada that he lists differ radically from present concepts. This may be attributed to the scanty knowledge available to Walcott in 1906 regarding the geology north of the international boundary.

CALKINS' CORRELATIONS

Since 1907 Calkins has been a principal contributor to knowledge of the Belt series in western Montana and northern Idaho. Some of his results have been referred to in the discussion of Walcott's correlations. Table 4 was prepared by Calkins (1909) after a reconnaissance in western Montana and northern Idaho, in 1905. The principal difference between this table and an earlier one prepared in connection with work in the Coeur d'Alene area, Idaho (Ransome and Calkins, 1908, p. 27) is in the column relative to the Philipsburg quadrangle. In the earlier table the names Chamberlain and Neihart, taken from Walcott's work in the Big and Little Belt Mountains, are applied to the lowest two units in the Philipsburg area. In table 3 the formation in the Philipsburg area is called Prichard (a name that originated in the Coeur d'Alene area) and the unit beneath it (called Neihart in the earlier table) is unnamed. Table 3 indicates that the supposed fossil *Beltina danai* was used in correlating the Newland, Altyn, and Blackfoot, a procedure that few would accept today.

Table 5, a later correlation table by Calkins, is reproduced on page 14. The most essential difference from table 4, published 6 years earlier, is the substitution of the terms "Spokane" and "Greyson" for "Camp Creek" in the column for the Philipsburg area. The name Neihart is revived for rocks thought to be the lowest exposed in the Philipsburg area.

DALY'S ORIGINAL CORRELATION

In his report on the geology along the international boundary Daly (1912) discusses the stratigraphic relations of the Belt series in some detail. His own fieldwork was limited to the immediate vicinity of the boundary, but table 6, modified from one in his memoir, shows his ideas as to correlation in various localities that seemed to him to bear on the stratigraphy in the area of his investigation. Daly's correlations were hampered by his concept that most of the rocks of what he termed the Beltian system were of Cambrian age. Most of the formations listed in table 6 are now known to be of Precambrian age but some of these are correlated in the table with formations that are definitely of Cambrian age.

TABLE 4.--*Probable correlation of principal sections of Algonkian sediments in Montana and Idaho*
[Calkins, 1909, p. 40-41]

Belt Mountains (Walcott, 1899, p. 199-244)	Lewis and Livingston Ranges (Willis, 1902, p. 305-352)	Philipsburg district (Calkins and Emmons, 1915)	Mission Range (Walcott, 1906, p. 1-28)	Coeur d'Alene district (Ransome and Calkins, 1908)	Cabinet Range, western and central parts (Calkins)	Forty-ninth parallel, between crossings of Kootenai River (Daly, 1905, p. 91-100)
Cambrian. —Unconformity— *Marsh.* Shale, red. 800 ft.[1]	Top not seen. *Kintla.* Shale, maroon-red; ripple marks, etc; some quartzitic and calcareous beds. 800 ft. *Sheppard.* Quartzite, yellow, ferruginous. 700 ft.	Cambrian. —Unconformity—	Cambrian. —Unconformity— *Camp Creek.* Sandstones, gray, rather thin bedded. 1,762 ft.	Upper part of section eroded.	Shales and sandstones, medium- to thin-bedded; color prevailingly greenish-gray, but in part red and purple. Shales partly calcareous and weathering buff. A little white crystalline limestone, weathering yellow, at several horizons. Base not seen. 10,000± ft.	Upper part of section eroded.
Helena. Limestone, with some shale. 2,400 ft. *Empire.* Shales, greenish-gray. 600 ft. *Spokane.* Shales, with thin beds of sandstone; deep red. 1,500 ft. *Greyson.* Shales, mostly dark gray. 3,000 ft.	*Siyeh.* Limestone, dark-blue or gray, weathering buff, with shale interbedded. 4,000 ft. *Grinnell.* Shale, partly arenaceous; dark-red; ripple-marked and suncracked. 1,800 ft. *Appekunny.* Shale, gray, black, and greenish, interbedded with white quartzite. 2,000± ft.	*Camp Creek.* Shale and sandstone, the latter prevailing in upper portion. Color chiefly red. 5,000+ ft to 0 ft.	Shales, sandstones, and limestones. 1,560 ft. Sandstones, mostly reddish. 4,491 ft. Sandstones, largely shaly, colors red and gray, with 198 ft of limestone 700 ft below top. 3,887 ft.	*Striped Peak.* Shales and sandstone, red and green. 1,000+ ft.	*Striped Peak.* Shales and shaly sandstones, prevailingly dark-red; ripple marks, etc. 2,000+ ft.	*Yaak.* Quartzite. 500 ft.
Newland. Limestone, impure, weathering buff, with interbedded shale. 2,200 ft. *Beltina danai.*	*Altyn.* Limestone, upper part thin-bedded and ferruginous; lower part grayish-blue, massive, siliceous. 1,400 ft. *Beltina danai.* Base not exposed.	*Newland.* Limestone, thin-bedded, more or less siliceous and ferruginous, passing into shale; generally buff on weathered surface. 4,000 ft.	*Blackfoot.* Limestone, thin-bedded; more or less siliceous; siliceous layers, weathering buff, interbedded with calcareo-arenaceous shales. 4,805 ft. *Beltina danai.*	*Wallace.* Shales, more or less calcareous, interbedded with thin layers of siliceous and ferruginous limestone and calcareous sandstone. Limestones and calcareous shales weather buff. 4,000 ft.	*Newland.* Limestones, thin-bedded, siliceous and ferruginous, interbedded with more or less calcareous shales. 5,000± ft.	*Mooyie.* Argillite. 3,400 ft.
Chamberlain. Shale, mostly black, with some sandstone. 1,500 ft.		*Ravalli.* Quartzite, gray, with some dark-bluish and greenish shale. 2,000 ft.	*Ravalli.* Sandstones, quartzitic, fine-grained, grayish-purple and gray. 2,550 ft. Sandstones, compact, gray. 1,060 ft. Sandstones, greenish-gray, fine-grained in layers 4 in. to 2 ft thick. 4,645 ft. Base not seen. Total Ravalli, 8,255 ft.	*St. Regis.* Shales and sandstones, purple and green. 1,000 ft. *Revett.* White quartzite, partly sericitic. 1,200 ft. *Burke.* Indurated siliceous shales, with sandstones and quartzites, prevailingly gray-green. 2,000 ft.	*Ravalli.* Quartzites, siliceous shales, and shaly sandstones; upper part green and purple; lower part gray, mostly greenish, locally with faint purple tinge; middle part thickest bedded, and most quartzitic, consisting locally of fairly pure white quartzite. 8,000± ft.	*Kitchener.* Ferruginous quartzite. 7,400 ft.
Neihart. Quartzite, with some shale in upper part. 700 ft.		*Prichard.* Shales, dark-bluish, interbedded with sandstone, rusty-brown on weathered surface. 5,000± ft. Quartzite, light-colored. Base not exposed. 1,000± ft.		*Prichard.* Argillite, blue-gray to black, with distinct and regular banding, interbedded with a subordinate amount of gray sandstone. Uppermost part arenaceous and marked with shallow-water features. Base not exposed. 8,000+ ft.	*Prichard formation.* Argillite, dark-bluish, banded. 2,000 ft. Sandstones, gray, thick-bedded to shaly, interbedded with more or less sandy bluish shales. The rocks become more argillaceous toward the southeast. 10,000± ft. Base not exposed.	*Creston.* Quartzitic sandstone, thick-platy, gray, interbedded with a subordinate amount of bluish argillaceous material. Base not exposed. 9,500+ ft.
Archean.						

[1] Walcott (1899) gives the thickness as 300 ft.

TABLE 5.--*Correlation of principal sections of Algonkian sedimentary rocks (Belt series) in Montana and Idaho*

[From Calkins and Emmons, 1915]

Belt Mountains (Walcott, 1899, p. 199-244)	Philipsburg district (Calkins)	Mission Range (Walcott, 1906, p. 1-28)	Coeur d'Alene district (Ransome and Calkins, 1908)
Cambrian.	Cambrian.	Cambrian.	
Unconformity	Unconformity	Unconformity	
Marsh: Shale, red; 800 ft.[1]		Camp Creek: Sandstones, gray, rather thin-bedded; 1,762 ft. Shales, sandstones, and limestones; 1,560 ft.	
Helena: Limestone, with some shale; 2,400 ft.	Absent.		
Empire: Shales, greenish-gray; 600 ft.		Sandstones, mostly reddish; 4,491 ft.	
Spokane: Shales, with thin beds of sandstone; deep-red; 1,500 ft.	Spokane: Shale and sandstone, the latter prevailing in upper portion; color chiefly red; 5,000 ft.	Sandstones, largely shaly, colors red and gray; 198 ft of limestone 700 ft below top; 3,887 ft.	Striped Peak: Shales and sandstone, red and green; 1,000 ft.
Greyson: Shales, mostly dark-gray; 3,000 ft.	Greyson may be present and included in Newland.	Total thickness of Camp Creek, 11,700 ft.	
Newland: Limestone, impure, weathering buff, with interbedded shale; 2,200 ft.	Newland: Limestone, thin-bedded, more or less siliceous and ferruginous, passing into shale; generally buff on weathered surface; 4,000 ft.	Blackfoot: Limestone, thin-bedded; more or less siliceous layers weathering buff; interbedded with calcareo-arenaceous shales; 4,805 ft.	Wallace: Shales more or less calcareous; interbedded with thin layers of siliceous and ferruginous limestones and calcareous sandstone; limestones and calcareous shales weather buff; 4,000 ft.
Chamberlain: Shale, mostly black, with some sandstone; 1,500 ft.	Ravalli: Quartzite, gray with some dark-bluish and greenish shale; 2,000 ft.	Ravalli: Sandstones, quartzitic, fine-grained, purplish-gray and gray; 2,550 ft. Sandstone, compact, gray; 1,060 ft. Sandstones, greenish-gray; fine-grained; in layers 4 in. to 2 ft thick; 4,645 ft; base not seen. Total thickness of Ravalli, 8,255 ft.	St. Regis: Shales and sandstones, purple and green; 1,000 ft. Revett: White quartzite, partly sericitic; 1,200 ft. Burke: Indurated siliceous shales with sandstones and quartzites, prevailingly gray-green; 2,000 ft.
	Prichard: Shales, dark-bluish; interbedded with sandstone; rusty brown on weathered surface; 5,000± ft.		Prichard: Argillite, blue-gray to black; with distinct and regular banding; interbedded with a subordinate amount of gray sandstone; uppermost part arenaceous and marked with shallow-water features; base not exposed; 8,000 ft.
Neihart: Quartzite, with some shale in upper part; 700 ft.	Neihart: Quartzite, light-colored; base not exposed; 1,000± ft.		
Archean.			

[1] Walcott gives the thickness as 300 ft.

Daly's belief that most of the present Belt series was of Cambrian age led him to make minor changes in previous correlation tables that are at least in part incorrect. That is, in his columns for the Philipsburg and Mission Range areas he correlated the Flathead quartzite, Wolsey shale and Meagher limestone, all of Cambrian age, with beds generally recognized as belonging to the Belt series. He correlated the beds in the Philipsburg area—then called the Camp Creek series and here assigned to the Missoula group—with the Blackfoot, Newland and Wallace of other localities. His table shows the Siyeh limestone (plus part of the Gateway in one place and Sheppard in another) as equivalent to the Blackfoot and Wallace, in approximate accord with present ideas. However, Daly's table shows the Camp Creek series in essentially its correct position in the section for the Mission Range, where the name originated. He split the Greyson shale, Sheppard (now Shepard) formation, Ravalli argillite, and Prichard formation into two parts each, perhaps to facilitate correlations with units in Canada.

SCHOFIELD'S CORRELATION

Soon after Daly's work was done, Schofield (1914a, p. 79–91), in an area about 20 miles north of the boundary at Gateway, reported strata of Middle Cambrian age resting on the Roosville formation. The latter is locally the uppermost formation supposed to be correlatable with the Belt series. In later papers Schofield (1922, p. 5–10; 1923, p. 5–11) indicates that the fossil-

TABLE 6.--*Correlations in the Rocky Mountain Geosynclinal*

[From Daly, 1912, opposite p. 178]

1	2	3	4	5	6	7	8	9	10	12
Summit series, Selkirk Range, lat 49° N.	Purcell series, Purcell Range, lat 49° N.	Coeur d'Alene series. [1]	Series in Cabinet Range. [1]	Series in Philipsburg district. [2]	Series in Mission Range. [3]	Galton series, Galton Range, lat 49° N.	Lewis series, Clarke and Lewis Ranges, lat 49° N.	Series in Belt Mountains. [4]	Castle Mountain Bow River series. [5]	System
Conformity with upper Paleozoic(?).	Erosion surface.	Erosion surface.	(?)	Conformity with upper Paleozoic.	(?)	Erosion surface.	Erosion surface.	Conformity with upper Paleozoic.	Conformity with upper Paleozoic.	
									Sherbrooke, 1,375 ft. Paget, 360+ ft. Bosworth, 1,855± ft.	Upper Cambrian.
Lone Star, 2,000+ ft.	Moyie, 3,400+ ft.	Striped Peak, 1,000+ ft.	Striped Peak, 2,000+ ft.	Meagher, 400 ft. Wolsey, 100-300 ft. Flathead, 50-300 ft.	Flathead. Thickness(?). Camp Creek, 11,700(?) ft.	Roosville, 600+ ft. Phillips, 550 ft. Gateway, 1,850 ft.	Kintla, 820 ft. Sheppard, 500 ft.	Gallatin. Flathead.		Chiefly Middle Cambrian.
Beehive, 7,000 ft.	Kitchener, upper part, 6,000± ft.	Wallace, 4,000 ft. St. Regis, 1,000 ft.	Blackfoot (called Newland by Calkins), 5,000± ft.	Camp Creek, 0-5,000 ft. Blackfoot, 4,000 ft.	Blackfoot, 4,805 ft.	Gateway, lower part, 125 ft. Siyeh, 4,000 ft.	Sheppard, lower part, 100 ft. Siyeh, 4,100 ft.	Marsh, 800 ft. Helena, 2,400 ft. Empire, 600 ft.	Eldon, 2,728 ft. Stephen, 640 ft. Cathedral, 1,595 ft.	Middle Cambrian.
Ripple, 1,650 ft. Dewdney, 2,000 ft. Wolf, upper part, 1,000± ft.	Kitchener, lower part, 1,400± ft. Creston, upper part, 3,000+ ft.	Revett, 1,200 ft. Burke, 2,000 ft. Prichard, upper part, 1,500± ft.	Ravalli, upper part, 5,000± ft. Ravalli, lower part, 3,000± ft.	Ravalli, 2,000 ft. Prichard, upper part, 2,000± ft.	Ravalli, upper part, 4,550 ft.	Wigwam, 1,200 ft. MacDonald, 2,350 ft. Hefty, 775 ft.	Grinnell, 1,600 ft. Appekunny, 2,600 ft.	Spokane, 1,500 ft. Greyson, 2,000± ft.	Mount Whyte, 390 ft. St. Piran, 2,705 ft. Lake Louise, 105 ft. Fairview, 600+ ft.	Lower Cambrian.
Wolf, lower part, 1,900+ ft. Monk, 5,500 ft. Irene volcanic formation, 6,000+ ft. Irene conglomerate, 5,000+ ft. Total, 32,050+ ft.	Creston, lower part, 6,500± ft. Base concealed. Total, 20,300+ ft.	Prichard, lower part, 6,500+ ft. Base concealed. Total, 17,200+ ft.	Prichard, 12,000± ft. Base concealed. Total, 27,000+ ft.	Prichard, lower part, 3,000± ft. Neihart, 1,000± ft. Base concealed. Total 12,550-18,000 ft.	Ravalli, lower part, 4,000+ ft. Base concealed. Total, 25,055 ft.	Altyn, 650 ft. Base concealed. Total, 12,100 ft.	Altyn, 3,500± ft. Waterton, 200+ ft. Base concealed. Total, 13,420 ft.	Greyson, 1,000± ft. Newland, 2,200 ft. Chamberlain, 1,500 ft. Neihart, 700 ft. Total, 14,000± ft.	Continuation of Bow River argillites. Base concealed. Total, 12,353+ ft.	Beltian.
Unconformity.								Unconformity.		
Priest River terrane.								Cherry Creek beds.		
								Unconformity.		
								"Archaean".		

[1] Calkins (1909, p. 40).
[2] Calkins (1909, p. 40; 1907, p. 33).
[3] Walcott (1906, p. 2).
[4] Walcott (1899, p. 201, and references therein to Peale and Weed).
[5] Walcott (1908).

iferous beds resting on the Roosville formation are regarded by different paleontologists as either lower Middle Cambrian or Lower Cambrian. This, plus Drysdale's idea (1917, p. 60–61, and table) that the upper part of the rocks, apparently equivalent to the Belt series in this part of Canada, is of Early Cambrian age, left the problem unsettled. The contact between the fossiliferous beds and the Roosville formation is reported to be conformable, but Schofield speaks of a basal conglomerate in the Cambrian strata and of differences in degree of metamorphism on the two sides of the contact.

On the basis of field evidence not available to Daly, Schofield made drastic changes in the Daly's Galton and Purcell series. Both these local series are equivalents of parts of the Belt series in the United States. Table 7 expresses Schofield's ideas and comparison with Daly's table will show the modifications Schofield made. Daly (1915, p. 94) appears to have accepted the revisions so far as the area north of Gateway is

TABLE 7.--*General correlation table of Schofield*
[From Schofield (1914, in front of p. 91)]

Clark and Lewis Range 49th Parallel	Rocky Mountains, British Columbia	Purcell Range, British Columbia	Coeur d'Alene district, Idaho	Cabinet Range, Montana	
Daly (1912).	The correlation of these two series by the author [Schofield] is based on sections described by Daly (1912) and subsequently modified by the author.		Ransome and Calkins (1908, p. 25).	Calkins (1909, p. 40).	
	Lowest Middle Cambrian.				Cambrian.
	Unconformity.				
Erosion surface.	Roosville, 1,000 ft.				
	Phillips, 500 ft.	Erosion surface.		Erosion surface.	
Kintla, 800 ft. Sheppard, 600 ft.	Gateway, 2,025 ft.	Gateway, 1,000 ft.			Pre-Cambrian (Beltian).
Purcell Lava.	Purcell Lava.	Purcell Lava.	Erosion surface.	Shale and sandstone 10,000 ft.	
Siyeh, 4,100 ft.	Siyeh, 4,000 ft.	Siyeh, 4,000 ft.	Striped Peak, 1,000 ft.	Striped Peak, 2,000 ft.	
Grinnell, 1,600 ft. Appekunny, 2,600 ft.	Wigwam, 1,200 ft. MacDonald, 2,350 ft. Hefty, 775 ft.	Kitchener, 4,500 ft.	Wallace, 4,000 ft.	Blackfoot, 5,000 ft.	
Altyn, 3,500 ft.	Altyn, 650 ft.				
		Creston, 5,000 ft.	St. Regis, 1,000 ft. Revett, 1,200 ft. Burke, 2,000 ft.	Ravalli, 8,000 ft.	
		Aldridge, 8,000± ft.	Prichard, 8,000 ft.	Prichard, 10,000 ft.	

concerned. Schofield transferred the whole Purcell series stratigraphically downward as compared with the Galton series and redefined the Kitchener formation of Daly, locally. Daly thought of that formation as a thick and variable unit whose main mass farther east is indistinguishable from the Siyeh limestone; Schofield's table shows the Siyeh above the Kitchener. Problems of this sort require solution before correlations across the international boundary can be made positively. However, a probable correlation for rocks in Canada near Gateway is indicated on plate 3.

DRYSDALE'S CORRELATIONS

Drysdale (1917, p. 58–63) further modified correlations of the Belt series in a study that grew out of a reconnaissance of 2 weeks near Marysville and Nelson, British Columbia. The towns mentioned are not shown on maps in the present report but can be readily found on a map of British Columbia. Drysdale's tentative correlations are shown in table 8, taken from his report. Drysdale's table omits Calkins' Spokane formation from the column on the Philipsburg area. It makes the Irene volcanics of Daly approximately equivalent to the Purcell lava although Daly (1915, p. 207–220) thought of the Purcell lava as above his Siyeh formation, much higher than his Irene volcanics.

Drysdale's correlations for areas in Canada are similar to those of the present report except that in table 7 he shows some units now considered Precambrian, as Lower Cambrian. The two columns in table 7 representative of areas in the United States should be shifted upward enough to make the Wallace and Newland equivalent stratigraphically to the Siyeh. The rocks Drysdale tabulated as pre-Beltian probably include metamorphosed representatives of units of Belt age.

CORRELATION BY WILSON, LAMBERT, AND CLAPP

As a result of extensive reconnaissance surveys in western Montana, several geologists of the Montana Bureau of Mines and Geology evolved a correlation of the various units of the Belt series there (Wilson and others, 1924, p. 91–92). They regarded the Altyn and Siyeh limestones as respectively correlative with the Newland and Helena and thought the upper part of the Belt series along the 49th parallel to be equivalent to the red rocks belonging to what they called the Spokane group. These red rocks are apparently the Spokane shale and, presumably, any other red rocks they correlated with the Spokane shale. In addition they thought there was a thinning and change in lithology to the east and west of a zone bounded by the 115° and 116° meridians. This last generalization does not agree closely with the facts given in the descriptive section of the present report and summarized graphically in plate 3.

TABLE 8.--*Drysdale's preliminary correlation table for Kootenay terranes, British Columbia*

[From Drysdale, 1917, opposite p. 62]

Epoch	Selkirk Range, southwest Kootenay	Selkirk Range, northwest Kootenay	Purcell Range, south-central Kootenay	Rocky Mountain Range, northeast Kootenay and Alberta	Rocky Mountain Range, southeast Kootenay and Alberta (Allan, 1914, p. 60)	Coeur d'Alene district, Idaho (Ransome and Calkins, 1908)	Philipsburg district, Montana (Emmons and Calkins, 1913, p. 32-34)
Carboniferous to Ordovician	Slocan, Pend-d'Oreille, and Niskonlith series, chiefly argillaceous and calcareous	Laurie argillite and limestone of Daly's Albert Canyon division ("Niskonlith series" of Dawson)	Jefferson limestone (Devonian)	Halysites beds (Silurian) Graptolite shales, Goodsir shales (Ordovician)			Quadrant quartzite Madison limestone Jefferson limestone Maywood limestone and shale
Upper Cambrian				Castle Mountain group: Ottertail limestone Chancellor argillite Sherbrooke limestone Paget limestone Bosworth limestone			
Middle Cambrian			Elko limestone, Burton shale, and sandstone	Castle Mountain group: Eldon limestone Stephen limestone Cathedral limestone Mount Whyte metargillite			
	Summit series (Daly): Ripple and Beehive quartzites and metargillites	Sir Donald quartzites and grit	Galton series: Roosville quartzite	Bow River series: St. Piran quartzite		Striped Peak shales and sandstones	Red Lion limestone Hasmark dolomite Silver Hill calcareous shales Flathead quartzite and conglomerate
Lower Cambrian	Summit series (Daly): Dewdney, Monk, and Wolf quartz grits and quartzites	Ross quartzite, grits, and metargillites	Galton series: Phillips metargillite Gateway sandstone	Bow River series: Lake Louise metargillite	Kintla metargillite Sheppard quartzite and dolomite		
	Summit series (Daly): Irene volcanics, limestone, and conglomerate	Basaltic lava, Nakimu limestone, and Cougar conglomerate	Galton series: Purcell lava Siyeh limestone and conglomerate	Bow River series: Fairview conglomerate and coarse sandstone	Siyeh limestone and Purcell lava		
	Priest River terrane (Daly): Kitchener		Purcell series: Kitchener		Grinnell and Wigwam	Wallace	Greyson shale Newland calcareous argillite
Beltian	Priest River terrane (Daly): Creston gray quartzite	Cougar quartzite	Purcell series: Creston gray quartzite	Hector metargillite Corral quartzite and sandstone	Appekunny and McDonald and Hefty	St. Regis, Revett, and Burke quartzites	Ravalli quartzite
	Priest River terrane (Daly): Aldridge: Upper, Lower	Illecillewaet Moose Basal quartzite	Purcell series: Aldridge rusty weathering quartzites		Altyn siliceous dolomite Waterton dolomite	Prichard metargillite	Prichard metargillite Neihart quartzite
Pre-Beltian or Archean	Shuswap quartz-mica schist series	Shuswap orthogneisses chiefly					

SAMPSON'S CORRELATION

Edward Sampson spent a season in the vicinity of Pend Oreille Lake, Idaho (Sampson, 1928) and later made a study of the literature concerning the Belt series. His published abstract relative to correlations in that series (1927, p. 111) is brief but table 9 taken from the files of the U.S. Geological Survey and Sampson's published comments (Wilson and others, 1924, p. 92) on the correlation by Wilson, Lambert, and Clapp show that his ideas may be summarized as follows: (1) the Striped Peak, Lone Star, Marsh and upper Siyeh are equivalents, (2) the Beehive, Wallace, Newland, Blackfoot, Kitchener and the Helena and Empire together are equivalents, (3) the Ripple, Dewdney, and Wolf together are equivalent to the Blacktail, St. Regis, and Revett together and these assemblages are in turn equivalent to the Creston, Wigwam, Grinnell, and Spokane, (4) the Monk, Burke, MacDonald, and Hefty together, and the Appekunny and the Greyson in turn are equivalents, (5) the Irene volcanics, Prichard, Aldridge, Altyn, and the Newland and Chamberlain together are equivalents, and (6) the Irene conglomerate is equivalent to the lower Prichard or perhaps the Waterton, which in turn is equivalent to the Neihart. Sampson's grouping so far as Idaho is concerned is essentially that still commonly accepted where his fieldwork was done. His ideas do not differ greatly from those accepted at the time for parts of Canada near Idaho. He essentially follows Willis for the succession in the Glacier National Park region. The Newland is listed in two widely different positions in the column; presumably an error. At one place it is equated with the Wallace, Blackfoot, Helena, and formations in Canada, whereas it is probable that the Helena, where identifiable, is stratigraphically higher than the others listed.

CORRELATION BY CLAPP AND DEISS

The paper on the correlation of the Belt series by Clapp and Deiss (1931) climaxed studies by its senior author that extended over many years. Its principal original feature is a set of measured sections of the uppermost part of the series from localities near and northeast of Missoula, mainly along tributaries of the Blackfoot River (table 10). The correlation for the series as a whole offered by Clapp and Deiss (1931, p. 677) is based on the belief that the assemblages of argillite and quartzite that separate the Helena and Newland limestones in the Big Belt Mountains "thin to the northwest and west to such an extent that the workers in the western ranges had mapped these limestones as a single formation." In view of the marked lateral variations in thickness and character in the various components of the Belt series, such a concept seems an oversimplification of the problem. In the present paper the alternate view is taken that the Helena limestone has not been traced far from its type locality and that a number of carbonate rock units roughly comparable to that formation in stratigraphic position are now known. These units tend to be discontinuous and erratic. In consequence, any classification that depends on long-range correlation of the Helena limestone is questionable.

Actually, the two viewpoints do not result in large differences in the classification of the series. The principal difference is that in the present paper the base of the Missoula group is placed somewhat lower stratigraphically than Clapp and Deiss advocated. For most areas, this facilitates mapping. It also tends to avoid long-range correlation of the kind involved in their use of the term "Spokane" for rocks more than a degree and a half in longitude west of the type locality of that formation. Langton, who worked under Clapp, (Langton, 1935, p. 36, fig. 3) applied both the names "Spokane formation" and "Helena limestone" to exposures even farther west. In some sections, as shown in table 10, Clapp and Deiss speak of Upper and Lower Siyeh limestone, separated either by an unnamed argillite body or by one that they label Spokane. Their "Lower Siyeh limestone" appears to be essentially the Siyeh limestone of Willis (1902, p. 316, 323). On this basis they appear to regard the lower part of the Wallace formation, the Newland limestone of the Big Belt Mountains, and the Siyeh limestone, as originally defined, as essential equivalents. They would correlate the upper part of the Wallace formation, the upper part of the unit Calkins termed Newland near Philipsburg, the Blackfoot series of Walcott, the Helena limestone, and their Upper Siyeh limestone. One difficulty with their concept is that the Wallace formation in and near its type locality, although in part calcareous, is not primarily a limestone as stated in their table. Most of it is argillite and quartzite with subordinate amounts of calcareous matter. Little study had been given to the corresponding beds from near the Idaho border eastward at the time Clapp and Deiss were doing their work. Hence they could not realize how scarce the limestone really was at appropriate horizons between Missoula and the border of Idaho. Langton's work cited above, and reconnaissances in connection with the present investigation leave doubt in the writer's mind as to how many limestone units are present in the region centered around Missoula and what their stratigraphic relationships may be.

A letter from Clapp (dated Feb. 10, 1932, to Francis A. Thomson, then president of the Montana School of

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TABLE 10.—*Correlation table of the Belt rocks of Montana (Clapp and Deiss, 1931)*

Coeur d'Alene (Calkins)	Striped Peak 1000 ft Wallace: limestone 4000 ft St.Regis 1000 ft Revett 1200 ft Burke: quartzite and argillite 2000 ft Pritchard: argillite and sandstone 8000 ft Base not exposed
Purcell Range (Schofield)	Gateway 1000 ft Purcell lava Siyeh: limestone and argillite 4000 ft Kitchener: dolomite, argillite and limy quartzite 4500 ft Creston: quartzite 5000 ft Aldridge: quartzite and argillite 8000 ft
Mission Range (Wilson)	Missoula group: argillite, sandstone, and quartzite; thin limestone 11,000+ ft Siyeh: limestone 5000 ft Ravalli: quartzite and argillite 10,700 ft
Glacier Park (Willis)	Kintla 800 ft Sheppard 700 ft Purcell lava Siyeh: limestone Shale Limestone 4000 ft Grinnell: shale 1800 ft Appekunny: quartzite and shale 2000 ft Altyn 1400 ft
Glacier Park (Clapp)	Missoula group: argillite 7600 ft Upper Siyeh: limestone 1500 ft Argillite 800 ft Lower Siyeh: limestone 3800 ft Grinnell: argillite 2800 ft Appekunny: argillite and siliceous quartzitic argillite 5300 ft
Missoula (Clapp and Deiss)	Sheep Mountain: quartzite 2300 ft Garnet Range: quartzite and sandy argillite 7600 ft McNamara: argillite and quartzite 3000 ft Hellgate: quartzite 2200 ft Miller Peak: argillite 2900 ft Wallace: limestone 7800 ft Base not exposed
Salmon Lake (Clapp and Deiss)	Miller Peak top not seen Upper Siyeh: limestone 4200 ft (Spokane) 1150 ft Lower Siyeh: limestone
McCabe Creek (Clapp and Deiss)	Hellgate: quartzite 2000 ft Miller Peak 1150 ft Upper Siyeh: limestone 4800 ft Spokane: argillite 2600 ft Lower Siyeh: Sill Lower Siyeh: limestone 5400 ft
Blackfoot Canyon (Clapp and Deiss)	Miller Peak 1400+ ft Upper Siyeh: limestone Blackfoot (Walcott)[1] 6100 ft (Spokane): argillite 3600 ft Lower Siyeh: limestone 4600 ft Grinnell: argillite 2600 ft Appekunny 2400 ft Base not exposed
Arrastre Creek (Clapp and Deiss)	Top not seen (Spokane): argillite 3700 ft Lower Siyeh: limestone 3800 ft Grinnell: argillite 2400 ft Appekunny 800 ft Base not exposed
Stonewall Creek (Clapp and Deiss)	Top not seen Spokane: argillite 2200 ft Lower Siyeh: limestone 3300 ft Grinnell: argillite 3100 ft Appekunny 300 ft Base not exposed
Prickly Pear Creek (Clapp and Deiss)	Top not seen Hellgate: quartzite 1700 ft Miller Peak Marsh (Walcott) [2]1800 ft [3]Helena: limestone 4000 ft Empire 600 ft Spokane: argillite 4500 ft Greyson: slaty argillite 2300 ft Base not exposed
Belt Mts. (Walcott)	Marsh 800 ft Helena: limestone 2400 ft Empire 600 ft Spokane 1500 ft Greyson: shale 3000 ft Newland: limestone 2200 ft Chamberlain 1500 ft Neihart 700 ft
Philipsburg (Calkins)	Spokane shale and sandstone 5000 ft Newland: limestone and shale 4000 ft Ravalli: quartzite and shale 2000 ft Neihart 1000 ft

[1] Walcott, 1906, Geol. Soc. America Bull., v. 17, p. 5–7.
[2] 1000 feet according to Barrell, 1907, U. S. Geol. Survey Prof. Paper 57, p. 9.
[3] Barrell, 1907, U. S. Geol. Survey Prof. Paper 57, p. 9.

Mines), written after the paper by Clapp and Deiss was published, records a modification of his ideas as to nomenclature. He notes that he mapped the Beaverhead Range in 1921 and 1922 and later concluded that the limestone that overlies the Ravalli should not be correlated with the Newland, which is only a part of this limestone, but "with the Wallace or Siyeh formation or group," and adds that the argillite and quartzite previously correlated by him with the Spokane, "have now been grouped as the Missoula group or formation." Dr. Clapp did not give his reasons for stating that the Newland limestone is only a part of the limestone above the Ravalli in the area of the Beaverhead Range, but the idea is not opposed to the concepts offered in the present report. His proposal to substitute the name Missoula for Spokane as far west as Missoula fits well with the ideas reached during the studies here summarized. It would require a rearrangement of the sections in table 10. The definition and description of the Missoula group as published, earlier than the letter just cited (Clapp and Deiss, 1931), constitute a notable advance in the study of the Belt series.

GIBSON'S CORRELATION

A correlation table for the Coeur d'Alene and Pend Oreille districts in northern Idaho and two quadrangles along the western border of Montana was presented by Gibson (1948, p. 9) in his report on the Libby quadrangle. Table 11 conveniently summarizes data at the date of publication relative to the parts of Idaho and Montana that are covered by it. Various modifications in detail have been made by later workers.

RECENT CANADIAN CORRELATIONS

Work in Canada prior to 1920 that bears on correlations of the Belt series has already been summarized. Most published correlations of the series in the United States deal with these early Canadian studies. The great mass of work done in Canada that concerns the "least deformed and metamorphosed" of the Precambrian rocks there has been recently digested and published in a single volume (Gill, 1957). One of the papers in this volume (Reesor, 1957) forms the principal basis for the summary below. Reesor states that the Precambrian rocks of southeastern British Columbia and southwestern Alberta "form a limited northward extension into Canada of the Belt Terrain of northwestern Montana and northern Idaho" and "consist of the dominantly clastic rocks of the Purcell and Windermere systems." The Purcell and Windermere are commonly referred to in the Canadian literature as series rather than systems. However, the Canadians also apply the term "series" to smaller divisions. There are several definitions of the Windermere series (Walker, 1926, 1929, p. 9, 1934; Evans, 1933; Okulitch, 1956, p. 706–721, 728–730). According to some of these definitions the Windermere includes some rocks with fossils of Cambrian age. Reesor (1957, p. 158) uses Walker's (1926) original definition that includes the Toby formation, the Horsethief Creek series redefined (Walker, 1934), and, locally, the Irene volcanic formation. As thus restricted, the Windermere series is of Precambrian age and corresponds to the upper part of the Belt series of the United States. Table 12 is figure 2 in Reesor's paper and shows the latest Canadian ideas as to correlation of Precambrian, Lower Cambrian, and Middle Cambrian units in the part of Canada north of western Montana, Idaho, and eastern Washington.

Several points are worth stressing in connection with table 12. Schofield's conclusions (1914b, 1922, 1923) have been confirmed and amplified by later work. In an area some 50 miles north of the international boundary and west of longitude 115°30′, Rice (1937, p. 7–13) studied a sequence similar to that listed by Schofield for the Rocky Mountains of British Columbia. He tentatively assigned an Early Cambrian age to strata above his Upper Purcell series (1937, p. 20–21), and fossils found later (Leech, 1954) agree with his assignment. Rice added a new formation, the Fort Steele, at the base and restricted the Siyeh formation to argillite and argillaceous quartzite, totaling 1,000 to 2,000 feet in thickness, leaving all of the carbonate rocks in the Kitchener formation. On this basis, the Siyeh of Rice may correspond to the basal part of the Missoula group of the present report. Strata lithologically resembling the Siyeh limestone in the United States have been placed by Rice in the Kitchener formation. The distance from the type locality of the Siyeh limestone is more than 150 miles. Whether the difference in nomenclature corresponds to lateral variations in the rocks is not clear. Reesor's table shows the name Siyeh everywhere just below the Purcell basalt, with the "Sheppard" overlying the lava. He (1957, p. 156) says the Siyeh formation is essentially argillite in the Purcell Range but becomes increasingly calcareous and dolomitic eastward. He cites a written communication from R. J. W. Douglas to the effect that in the Waterton area the lower Siyeh consists of grit and sandstone with salt crystal impressions, though the upper Siyeh is predominantly carbonate and shale. The Waterton area is immediately north of Glacier National Park, which contains the type locality of the Siyeh, there almost entirely composed of carbonate rocks. Douglas' map of the

TABLE 11.--*Gibson's correlation of Belt series in the Libby quadrangle with formations in nearby regions*

[Gibson (1948, p. 9). Approximate thickness given in feet]

	Coeur d'Alene district, Idaho[1]	Clark Fork district, Idaho[2]	Pend Oreille district, Idaho[3]	Trout Creek quadrangle, Montana-Idaho[4]	Libby quadrangle, Montana
Striped Peak formation	1,000+	4,000+	9,000+	3,000	Libby formation 6,000+; Striped Peak formation 2,000+
Wallace formation	4,000	6,000	6,000	10,500	12,000+
St. Regis formation	1,000	7,500	Blacktail formation 8,300	Revalli formation 5,300-9,800	Revalli formation 10,000
Revett quartzite	1,200	Burke formation 3,500	Burke formation 2,000+		
Burke formation	2,000				
Prichard formation	8,000+	20,000+	7,000+	7,800+	9,700+

[1] Ransome and Calkins (1908).
[2] Anderson (1930).
[3] Sampson (1928).
[4] W. F. Jenks, Harvard Univ. Ph. D. thesis, 1936.

Waterton area (1952) describes the Siyeh formation as "limestone and dolomite; grey green and red argillite; algal limestone." The writer (Ross, 1959) placed the argillite above the Siyeh limestone in the Missoula group, excluding it from the Siyeh. However, at the international boundary there is not enough argillite between the limestone and the Purcell basalt to show on the map. Everywhere in Glacier National Park the argillaceous rocks between the limestone and the lava are thin. The differences in classification on the two sides of the boundary are of special interest because the Siyeh limestone is part of the Piegan group, rich in carbonate, that has been used by many as a starting point for stratigraphic correlations in the Belt series in general. Farther west the components of the Piegan group, there called the Newland and Wallace formations, seem safely correlatable with the Kitchener of Rice (Kirkham and Ellis, 1926, p. 27).

Largely on the basis of the relations of the Purcell basalt, it seems that the Missoula group of the present report corresponds to the small amount of clastic rocks of the lower part of the Purcell series below the basalt, the whole of the upper Purcell and that part of the Windermere series regarded as Precambrian. Toward the west uncertainties arise as to the relations between the Purcell basalt and the Irene volcanics. It has been supposed (Drysdale, 1917, p. 62; Kirkham and Ellis, 1926, p. 27) that these two are equivalent, but in table 12 the Irene volcanics are indicated as 2,000 feet or more above the Purcell basalt. Clarification of this point awaits further work. Table 6, taken from Daly (1912, p. 178), suggests that the Irene volcanics are low in the Belt series, perhaps as low as the Prichard, but this is negated by the fact that the Monk formation overlies the Irene volcanics. The Monk formation was considered Cambrian(?) by Park and Cannon. Apparently the Leola volcanics in Washington (Park and Cannon, 1943, p. 9–11) are equivalent to and coextensive with the Irene volcanics.

The Windermere series was first recognized near the intersection of long. 116° and lat. 51° (Walker, 1926, p. 13–20). There its outcrops extend a little east of 116°00′. Leech (1954) mapped it near long. 115°40′. Rocks correlated with the Windermere extend westward in Canada to beyond 117°00′ (Rice, 1941, p. 22; Walker, 1934, p. 2–10). Similar rocks extend short

TABLE 12.—*Correlation of Lower Cambrian and Cambrian formations, southeastern Canadian Cordillera*

[*Reesor, 1957*]

Age divisions (left margin, top to bottom): Middle Cambrian; Lower Cambrian; Windermere; Upper Purcell; Lower Purcell

Region between 49° and 49°30′	Rocky Mountains	Waterton (Douglas)	Kintla 1500 ft Top not exposed; Sheppard 600 ft; Basalt; Siyeh 3000 ft; Grinnell 750 ft; Appekunny 1100 ft; Altyn 500 ft; Waterton 617 ft Base not exposed
	Rocky Mountains	Beaver Mines (Hage)	Middle Cambrian; Uncon-formable; Kintla 2900 ft; Sheppard 470 ft; Basalt; Siyeh 3000 ft; Grinnell 760 ft; Appekunny 1600 ft Base not exposed
	Rocky Mountains	Galton Range (Schofield)	Burton; Discon-formable; Rooseville 600 ft; Phillips 550 ft; Gateway 1975 ft; Purcell lava; Siyeh 4000 ft; Wigwam 1200 ft; MacDonald 2350 ft; Hefty 775 ft; Altyn 650 ft Base not exposed
	Purcell Mountains	Cranbrook (Rice)	Eager 6000 ft ?; Cranbrook 600 ft; Discon-formable; Gateway; Purcell lava; Siyeh 2000 ft; Kitchener 6000 ft; Creston 6300 ft; Aldridge 16,000 ft; Fort Steele 7000 ft Base not exposed
	Purcell Mountains	Purcell Divide (Rice)	Eager 2000 ft; Cranbrook 600 ft ?; Uncon-formable; Kitchener-Siyeh 7500 ft; Creston 6300 ft; Aldridge 15,000 ft Base not exposed
	Selkirk Mountains	Western Purcell (Rice)	Milford group; Confor-mable?; Lardeau series 10,000 ft; Badshot 100 ft; Hamill series 10,000 ft; Horsethief Creek 5000 ft; Irene Volcanics 0–5000 ft; Toby 250–2000 ft; Uncon-formable; Mount Nelson 3200 ft; Dutch Creek 4300 ft Base not exposed
	Selkirk Mountains	Salmo (Little)	Nolway; Laib group 3000–5000 ft Reno 800 ft; Quartzite Range 4400–1800 ft; Three Sisters 4–6000 ft; Monk 4000 ft; Irene volcanics 6000 ft; Toby 5000? ft Base not exposed
Region between 50° and 53°	Rocky Mountains	Sunwapta (Hughes)	Mount Whyte; Peyto limestone 140 ft; Jonas Creek; Confor-mable; Hector 6000? ft
	Rocky Mountains	Bow Valley	Mount Whyte; Peyto limestone; St. Piran Lake Louise Fort Mountain 2000? ft; Uncon-formable; Hector 4600 ft; Corral Creek 1320 ft Base not exposed
	Rocky Mountains	Mount Assiniboine (Deiss)	Naiset; ?; Gog 1236 ft; Uncon-formable; Hector Base not exposed
	Rocky Mountains	Canal Flats (Leech)	Jubilee; ?; Confor-mable; Unnamed 770–1500 ft; Cranbrook 250–800 ft; Uncon-formable; Toby 45–1000 ft; Discon-formable; ?; Rooseville? Base not exposed
	Purcell Mountains	Windermere (Walker)	Jubilee; Uncon-formable; Horsethief Creek 4–5000 ft; Toby 50–2000 ft; Uncon-formable; Mount Nelson 3400 ft; Dutch Creek 3500 ft Base not exposed; ??
	Purcell Mountains	Dogtooth (Evans)	Canyon Creek; Donald 200–1500 ft; St. Piran Lake Louise Fort Mountain 1500–5500 ft; Slight uncon-formity; Horsethief Creek 5800 ft; Contact not exposed; Toby
	Selkirk Mountains	(Okulitch)	Milford?; Uncon-formable; Lardeau series 12,000 ft; Badshot 60 ft; Hamill series 10,000–4000 ft; Discon-formable; Horsethief Creek 2000 ft Base not exposed
	Selkirk Mountains	(Gunning)	?; Member 2; Member 1A and 1B; ?
	Monashee Mountains	(Jones)	?; Chapperon group; ?; Mount Ida group 60,000? ft; ?; Monashee group 50,000 ft; Archean??

distances into Idaho and Washington but appear to be absent in Montana. Judging by the descriptions given by Walker and by Rice, the series is extremely variable in thickness and in character. It includes conglomerate—in part very coarse, grit, lava flows and pyroclastic rocks—and also argillite, argillaceous schist, and subordinate amounts of limestone. The volcanic rocks prominent in its lower part are the Irene volcanics of Daly, which he thought to be much lower stratigraphically. The conglomerate in the Windermere series sets that series apart lithologically from most of the Belt series. The origin of this conglomerate is not clear. One hypothesis is that it is a fanglomerate (Walker, 1926, p. 16; Park and Cannon, 1943, p. 9); another is that it is a deposit along the shore of a rapidly spreading sea (Rice, 1941, p. 23). Okulitch (1949) regards it as deltaic.

There are such marked lithologic differences between the lower components of the Belt series and presumable counterparts in Canada that the papers already cited give a variety of possible correlations. In northern Idaho correlations have been indicated by the expedient of such hyphenated names as Aldrich-Prichard and Creston-Ravalli (Kirkham and Ellis, 1926, p. 15–16). The implied correlations are probable in a general way (table 10), but the hyphenated names have not received general acceptance.

Walker (1926, p. 16) correlated the Windermere series with the Summit series of Daly, although the latter's correlation table (table 5) gives no hint that the Summit series is as young as Walker believes the Windermere to be. Both he and Rice regarded the Windermere series in its eastern exposure as late Precambrian, separated from the Purcell series below by a marked angular unconformity and from beds of Early Cambrian age above by a much less marked unconformity, if any. Leech (1954, p. 5) likewise found no angular unconformity at the base of the Windermere.

The data presented by Walker and Rice seem to show rather definitely that the Windermere series, as mapped by them, is of late Precambrian age. It is stated to underlie disconformably, and locally unconformably, beds of Early Cambrian age and to rest with locally marked angular discordance on beds that correspond to the Belt series in the United States. In the western area of the Windermere series, however, part of the series as there mapped may be of Cambrian age (Park and Cannon, 1943, p. 5–17; Okulitch, 1949, p. 5–13; Little, 1950, p. 1–13). The level within the series at which the base of the Cambrian is postulated varies with the different authors and cannot be fixed definitely on paleontologic grounds. According to Park and Cannon, much the greater part of the Windermere series of the Canadians is probably of Paleozoic age. Their concepts as to the stratigraphic relations of the series are shown in table 13. The conflicting ideas of different workers as to the age of the Windermere series suggest that rocks assigned to that series in different localities are of different ages. Okulitch (1949) discussed this matter in detail and concluded that transgressive overlap of deltaic deposits has resulted in vast sheets of sandy detritus, continuous over large areas, but almost certainly of different ages in different localities. His concept is plausible but needs support from further fieldwork.

TABLE 13.—*Correlation of formations in Metaline quadrangle and Salmo map area*

[Quoted verbatim from Park and Cannon, 1943, p. 6]

Metaline quadrangle		Salmo map area: R. A. Daly (1912)			Salmo map area: J. F. Walker (1934)	
Limestone, 700 ft.	Devonian	Pend Oreille group	Upper Paleozoic		Pend Oreille group	Windermere, late pre-Cambrian
Ledbetter slate, 2,500 ft.	Ordovician					
Metaline limestone, 3,000 ft.	Middle Cambrian					
Maitlen phyllite, 5,000±ft.	Cambrian	Lone Star phyllite, 2,000+ ft.	Summit series	Cambrian		
Gypsy quartzite, 5,300-8,500 ft.		Beehive formation, 7,000 ft.			Reno formation, 3,500 ft.	
		Ripple quartzite, 1,650 ft.			Quartzite Range formation, 4,400 ft.	
		Dewdney formation, 2,000 ft.			Three Sisters series, 5,400 ft.	
Monk formation, 3,800+ ft.	Cambrian?	Wolf grits, 2,900 ft.		Beltian	Horsethief Creek series, 4,200 ft.	
		Monk formation, 5,500 ft.			Irene volcanic (?) formation (?)	
Unconformity	pre-Cambrian	Irene volcanics, 6,000 ft.				
Leola volcanics, 5,000+ ft.						
Shedroof conglomerate, 5,000+ ft.		Irene conglomerate, 5,000+ ft.				
Unconformity		Unconformity				
Priest River group (?)		Priest River terrane (?)				

More recently Okulitch (1956, p. 706-721, 728-730) has summarized his concepts in regard to the Windermere series and associated formations. He speaks of a marked unconformity between the Purcell series and the Windermere series and a less-distinct disconformity at the top of the lower part of the Windermere series as defined by Walker (1926). Okulitch suggests that the strata above the disconformity be regarded as Cambrian. According to usage in the United States the lower part of the Canadian Windermere series and any of the upper part that is older than Cambrian would be regarded as belonging to the Belt series. The part that Okulitch in his 1956 paper suggests may be Cambrian contains some fossils but not enough so that the base of the Cambrian part can be established on paleontological grounds. Lateral gradation, scanty fossils, and poor exposure leaves the problem of the position of the Cambrian-Precambrian boundary within the Windermere series of Walker still unsolved. It is clear, however, that, locally at least, this boundary is not marked by a definite unconformity.

The age assignment is not the only puzzling feature of the Windermere series, as judged from published accounts. Why is an assemblage of beds many thousands of feet thick near long. 116° in British Columbia not found anywhere east of there nor in Montana? The unit extends into the extreme northwestern corner of Idaho and adjacent parts of Washington but apparently is abruptly limited there also. It contains a diverse aggregate of rocks that ranges from conglomerate with boulders as much as 4 by 5 feet in size to fine-grained, argillaceous beds and includes much volcanic material. One can speculate that a single depression was rapidly filled with clastic deposits accompanied by the products of localized volcanism. In such a setting some fine-grained beds might well have been produced by ponding of streams, perhaps by lava, or through shifting of drainage channels. Any such situation would imply mountains bordering the depression being filled by the Windermere sediments, but direct evidence of the existence of mountains at the appropriate time and place has not been recorded. Okulitch (1949, p. 20) speculates that in late Precambrian time, land of fairly high relief existed near long. 117°20′, lat. 51°15′. If so, no evidence of the southward extension of this land mass has been definitely recognized within the United States. In Washington rocks related to the Windermere series have been referred to as possibly fanglomerate (Park and Cannon, 1943, p. 9), which would imply the presence of a rugged land mass. Nothing of the sort has been reported in western Montana or northern Idaho. Parts of these areas are little-known but, for the moment, the only possible conclusion is that the Windermere series as described and interpreted in British Columbia has no counterpart in Montana or in most of Idaho. The North Boulder group of southwestern Montana and some rocks with possible affinities to that group in southern Idaho and in Utah include material as coarse as much of the Windermere series. Like that series, each of these occurrences is known only in rather small areas. The North Boulder group has been inferred (p. 93) to have been deposited at the foot of a large fault-line scarp. One can speculate that each of these masses of coarse sediments, alien to most of the Belt series, results from special, local conditions that produced marked relief in a region that then had relatively featureless topography. Such conditions would necessarily be structural but the reason for widely scattered faults or similar features is unknown. The diverse ideas as to the origin of these coarse rocks listed on page 93, testify to present lack of detailed information. When further studies are made, data as to the governing tectonic conditions should be forthcoming.

Recent work in Washington (Weis, 1959) shows that Precambrian and Cambrian rocks that enter into the problems of correlation of the Belt series are exposed farther west in that state than was appreciated earlier. There are resemblances between these rocks and the Windermere series, but, in contrast to much of that series, they may include rocks deposited in deeper water than is characteristic of most of the Belt series.

REVISED CLASSIFICATION

BASIS OF CLASSIFICATION

The summaries already presented show the nomenclatures adopted by different students for the Belt series in Montana, with some data on names used in Idaho, Washington, and Canada. The revised classification presented below differs in details from any previously published. Because no single formation has been traced throughout western Montana, groups in some cases rather than formations are used to facilitate classification and discussion in this report. The definitions of groups offered apply primarily only to Montana but the scheme of classification could be extended to adjacent areas.

None of the formation names offered is original. Each is used in accord with established custom and, so far as possible, the intent of the original authors. It has been necessary, however, to present here definitions intended to clarify and make as precise as possible the meaning attached to each of the names of major units involved in the revision of classification of the series. Where previous definitions exist, they are cited. The descriptions below show past and present usage.

A basic concept in evolving the classification is that lateral variation is so prevalent that units of formational rank cannot be assumed to be recognizable very far from their type localities, in spite of the fact that most of them are very thick compared to formations in many other regions. The distances that particular formations have been traced vary widely, depending on their distinctiveness and on the present knowledge regarding each.

Plate 3 is a graphic representation of the concepts just outlined that should aid in coordinating the summaries on areas in northwestern Montana and northern Idaho (p. 57–85). It is impractical to include in this diagram data on southwestern Montana and central Idaho. Present data prevent adequate correlations with the units shown on plate 3, owing partly to the large distances involved. Posts are drawn on the fence diagram, plate 2, wherever stratigraphic data permit. Where districts lie close together information respecting them has had to be averaged into single posts in the diagram, and districts that include only fragments of the Belt series are omitted. Nevertheless, the wide spacing of the posts serves to emphasize the scattered character of available data. One post in Canada is shown to suggest correlations across the international boundary. The writer has no personal field acquaintance with rocks in British Columbia and feels less confidence in his Canadian correlations than the others. The names and correlations in plate 3 are those of the present report, rather than of the original author.

Most of the Belt series in Montana can be readily placed in four named groups: the Ravalli, Piegan, Missoula, and North Boulder; all except the last are named in ascending order. In addition, a few formations not included in named groups are at present inferred to be older than the Ravalli group. As there is only one formation regarded as pre-Ravalli in each locality and mutual relations between the old units are unknown, no formal group is proposed to include them. Ultimately a fifth group will be named to include the oldest units of the Belt series unless future investigations show that none is actually older than the present Ravalli group. If a fifth group becomes desirable, one possibility would be to raise the Prichard to group rank much as has been done for the Ravalli. The Prichard formation is the thickest and most widely exposed of the pre-Ravalli units in the United States and, at least in Idaho, has subdivisions that may, perhaps, justify formational rank in some localities when it is better known. In Montana exposures of pre-Ravalli strata are scattered, and mutual relations are obscure. Thus, a name for rocks of the series below the Ravalli group probably should come from Idaho or British Columbia rather than Montana. Plate 3, in which the Prichard formation is treated much as the groups above it are, demonstrates these points.

The various correlations of the Belt series have been based largely on the assumption, expressed or implied, that certain carbonate-rich units are mutually equivalent. The units thus correlated have varied with the concepts of the different authors. The review already given shows that Walcott and Calkins assumed that the Newland, Altyn, Blackfoot, and Wallace are equivalents. In their early work the Siyeh and Helena were correlated and thought to be younger than the other carbonate units just listed. Later Walcott realized that his Blackfoot series and the Siyeh limestone are closely related. Daly thought of his Kitchener and the Siyeh, Wallace, Blackfoot, and Helena, among others, as being equivalents (table 6). The part of the eastern phase of Daly's Kitchener formation in which carbonate-rich rock predominates must be, as he thought (Daly, 1912, p. 133–135), equivalent to the Siyeh limestone. Schofield took the Siyeh limestone out of the Kitchener formation and regarded it as equivalent to the Striped Peak formation, whereas he thought the Wallace and Blackfoot were a little older. These and other variations in the use of the name Kitchener are among the reasons why correlations between stratigraphic units in Canada and those in the United States are not readily made.

Clapp and Deiss correlated the Newland limestone with the lower parts of the Siyeh and Wallace, as defined by them, and regarded the Helena limestone as equivalent to their upper Siyeh. In the present paper their idea of an upper and a lower Siyeh limestone, separated by red beds, is not followed. Instead all red beds above the Lower Siyeh of Clapp and Deiss, plus all carbonate units interbedded therewith are placed in the Missoula group. This procedure is believed to restrict the name Siyeh essentially to the beds originally so designated by Willis. It also facilitates geological mapping by grouping together all the clastic beds that are similar lithologically. The carbonate beds above the Siyeh limestone, as here defined, are regarded as relatively discontinuous, local features. Attempts to correlate assemblages of these beds over long distances may lead to error. The general scheme of classification here proposed can be grasped by a glance at the correlation chart (table 1).

The present classification does not aid in correlating formations in and close to central Idaho with other parts of the series. Conceivably it will prove desirable to set up one or more groups to include these. More probably future work will permit correlation with existing groups. Suggestions as to correlations of formations have been offered (Umpleby, 1913a, p. 30–32; Ross, 1934, p. 19–20), but the evidence in their support is tenuous and in part contradictory.

It should be emphasized that in the definitions and descriptions of units in this report reliance has been placed, necessarily, on published reports by many geologists, who were by no means consistent among themselves in the use of lithologic terms, color designation, and so forth. Observations by the writer and his associates have aided in interpretation of divergent usages but, except where specifically noted, the terminology of the original author is followed here. For example, a rock may be spoken of here as a red shale whereas the present writer might have called it in the field a red-purple argillite or, perhaps, even an impure quartzite.

Most rocks of the Belt series in Montana have been partially recrystallized and few show distinctly formed shaly fissility. Thus terms such as shale and sandstone, although retained in accord with prior usage, are open to question. On the other hand, few of the rocks of the series in Montana are sufficiently metamorphosed so that slaty and schistose partings are well developed. In some places in Idaho and Canada the Belt series is more metamorphosed and rock names may appropriately show this difference. Unfortunately lithologic nomenclature is not sufficiently standardized to infer subtle differences safely from published descriptions, especially as some of the descriptions here drawn on were written many years ago.

DEFINITIONS

PRE-RAVALLI ROCKS

The units at present regarded as stratigraphically below the Ravalli group are the Prichard formation and the Neihart quartzite. The Prichard formation (or slate) is the argillaceous and arenaceous assemblage that constitutes the lowest exposed part of the Belt series in and near the Coeur d'Alene area in northern Idaho. It was named for Prichard Creek in that area (Ransome, 1905, p. 281). In part through actual mapping, in part through inference, the name has been carried almost as far eastward as long. 113°, south of Philipsburg, Mont. (pl. 1). In Idaho the Prichard formation has been tentatively identified as far south as lat. 46°40′ (Anderson, 1930b, p. 10). If the Orofino series of Anderson (1930b, p. 9–10) is, as he thinks, a part of the Prichard lower than any exposed elsewhere, the formation extends at least as far south as 46°30′. Recent work (Hietanen, 1956, p. 3–4) supports the idea that the Prichard extends southward, but it must be remembered that the rocks so assigned are so intensely metamorphosed that diagnostic features are obscured.

The Neihart quartzite is the dominantly arenaceous unit that underlies the Chamberlain shale and rests on granitic gneiss near Neihart (Weed, 1899; Wilmarth, 1938, p. 1473–1474) in the Little Belt Mountains, Mont. As broadly similar quartzitic rocks occur at various horizons in the Belt series and the Neihart cannot be traced far from its type locality, the name is here restricted to the area in which it originated. Future mapping may reveal the same quartzite in parts of the Little Belt Mountains where it is not now known, but use of the name in more distant areas would serve no useful purpose. Quartzite in small exposures south of Philipsburg (Calkins and Emmons, 1915) has been called Neihart quartzite, but, in a conversation in 1948, Calkins agreed that the correlation is questionable. As indicated on plate 1, the exposures south of Philipsburg may belong to the Missoula group.

The type locality of the Neihart quartzite is so far from the localities where the groups into which the Belt series is divided were established that relationships are uncertain. The unit is surely low in the Belt sequence near Neihart and has been regarded by Clapp and Deiss (table 10) as of pre-Ravalli age, an assignment that is here accepted as logical. However, the Fentons (1937b, p. 1880) speak of the Neihart as "basal Ravalli."

Among other units that might be regarded as of possible pre-Ravalli age, attention should be called to the Altyn limestone in Glacier National Park. At present (Clapp and Deiss, 1931; Fenton and Fenton, 1937b, p. 1880–1885) the Altyn is assumed to belong to the Ravalli group. It is the oldest formation in Glacier National Park, with its base not exposed. No limestone (or dolomite) in the Belt series stratigraphically as low as the Altyn is on record elsewhere.

RAVALLI GROUP

Calkins (1909, p. 37–38) applied the term "Ravalli" group to strata above the Prichard and below the formation he then called Newland, now commonly called Wallace, in northern Idaho and northwestern Montana. His usage is a modification of that of Walcott (1906, p. 7–9) who divided the Belt rocks in and near the Swan and Mission Ranges into three series, the Camp Creek, Blackfoot, and Ravalli, listed in descending order. These three series correspond approximately to the three principal groups of the present paper.

Here the Ravalli group is regarded as including all formations of the Belt series, other than the Neihart quartzite and Prichard formation that are below the Newland limestone or its stratigraphic equivalents, such as the Wallace formation and Siyeh limestone. This definition agrees with most present usage but is phrased so as to be both broad and definite enough to provide a basis for regional mapping. The name is inferred (Wilmarth, 1938, p. 1776) to have been taken from the town of Ravalli. As just noted, uncertainties remain as to assignment of certain formations to the Ravalli, a fact that may require modification of the above definition on the basis of future work. For example, it may be found that the Altyn limestone is stratigraphically equivalent to some part of the Prichard formation, or, conversely, that the Neihart quartzite should be placed in the Ravalli group rather than below it.

In and near Idaho the Ravalli group includes the St. Regis formation, Revett quartzite, and Burke formation (Calkins, 1909, p. 37–38), all names of localities in or near the Coeur d'Alene area in Idaho (Ransome, 1905, p. 281–282) and not shown on maps in the present paper. Farther east, it includes the Appekunny and Grinnell formations and the Altyn limestone (Clapp, 1932, p. 22; Fenton and Fenton, 1937b, p. 1880–1890). The type localities of the Appekunny and Grinnell are mountains in Glacier National Park, and the Altyn was named for exposures on Appekunny Mountain north of the former town of Altyn (Willis, 1902, p. 316–321). In the Philipsburg area, Calkins (Calkins and Emmons, 1915, p. 3) used the term Ravalli formation in virtually the way Ravalli group is used here. From the Philipsburg area to near Neihart no rocks as old as the Ravalli group are known. In the latter locality the Chamberlain shale has been correlated with components of the Ravalli group (Clapp and Deiss, 1931, p. 693), an assignment that seemed probable. The name is derived from exposures near Chamberlain Creek in the Neihart area (Walcott, 1899, p. 206; Weed, 1899, 1900, p. 282).

PIEGAN GROUP

The assemblage above the Ravalli group has been variously designated by different geologists, most of whom use one or more formation names for it. This unit is distinguished by the fact that carbonate rocks predominate, and it is here termed the Piegan group. The base of the group is at the top of the Ravalli group and the top is at the place where largely carbonate-rich rocks give way to dominantly argillaceous ones, commonly reddish and greenish. Parts of the group, notably in the Wallace formation in and near Idaho, are carbonate-rich only by comparison with rocks above and below that are almost carbonate-free. The name Piegan group was first proposed by Fenton and Fenton (1937b, p. 1890-1900) from Piegan Mountain in Glacier National Park. In an earlier consideration of the Belt problem (Ross, 1947, p. 112–113) it was proposed, following Clapp and Deiss (1931, correlation table, p. 691) to restrict the Fentons' definition slightly. The presently proposed definition still further restricts the group but places its upper limit at a horizon that is readily recognizable in most places. As nearly all formational contacts in the Belt series are gradational, local uncertainties as to the precise position of boundaries are unavoidable.

The Fentons originally placed the top of their Piegan group at the top of the Sheppard (now Shepard) formation, named for the Shepard Glacier in Glacier National Park. Clapp and Deiss (table 10) did not use a group name but clearly intended to correlate the Shepard with beds elsewhere well above the main mass of carbonate-rich rocks. Thus, if the term "Piegan group" had been in use when their paper was written, Clapp and Deiss would have excluded the Shepard formation from it. They did intend to group together the carbonate rocks in the median portion of the Belt series, and their exclusion of the Shepard may have been because the unit was originally called the Sheppard quartzite (Willis, 1902, p. 316, 324). Actually much of it is dolomite, a fact impressed on the writer in 1950 during fieldwork in Glacier National Park, although already known to the Fentons (1937b, p. 1899). In the writer's first correlation (Ross, 1949, p. 111–113) the opinion of Clapp and Deiss was adopted. Further consideration, after additional fieldwork, strengthens this decision and leads to the opinion that the Shepard formation is only one of several carbonate-rich units at various horizons in the thick assemblage of reddish and greenish argillaceous and siliceous beds that constitute the upper part of the Belt series above the Piegan group.

Much further work should be done on the carbonate rocks in the upper part of the series, but it will be difficult, if not impossible, to determine their mutual interrelations throughout Montana. Much of the needed evidence is hidden by younger rocks. Besides the Shepard, the post-Piegan carbonate rocks include the Helena limestone and most or all of the beds called Upper Siyeh limestone by Clapp and Deiss (1931, p. 691). The contract between the thick mass of carbonate-rich rocks here included in the Piegan group and the reddish and greenish overlying beds, is as sharp and easily mapped as most contacts between units of the Belt series. If any of the carbonate-rich rocks above that contact are included in the Piegan group, as suggested by either Clapp and Deiss or the Fentons, uncertainties arise, mainly because so many of these carbonate units are discontinuous.

Although several different formation names have been applied to beds of the Piegan group in the United States, each is used only in a single region. Nomenclature has varied, but currently, in northern Idaho and the westernmost part of Montana, the term Wallace formation, derived from Wallace, Idaho (Ransome, 1905, p. 282), is used. From near Missoula to the Little Belt Mountains similar rocks (richer in carbonate) are called the Newland limestone (Walcott, 1899, p. 206) from exposures along Newland Creek north of White Sulphur Springs. In Glacier National Park and the surrounding region Siyeh limestone, derived from Mount Siyeh (Willis, 1902, p. 316–323), is the name usually used. When more detailed mapping is done the Wallace, Newland, and Siyeh will be broken up into many new formations, based largely on subdivisions already known to exist. Thus, in each region, the Piegan group will eventually comprise many formations instead of the individual units now recognized. Many of the new units will be of more restricted significance than the thick assemblages now recognized.

MISSOULA GROUP

The Missoula group, named for the city of Missoula (Clapp and Deiss, 1931, p. 677–683), includes all of the Belt series above the Piegan group, except parts of the North Boulder group in southwestern Montana. The Missoula group, as redefined here, embraces all the comparatively bright-colored reddish and greenish argillaceous and arenaceous strata high in the series. These strata commonly contain intercalated carbonate-rich masses and, locally, rather light-colored quartzite lenses. In some areas lateral variations in color and other features obscure the distinctive characteristics thus briefly summarized. As the thick assemblage of carbonate-bearing rocks of the Piegan group is commonly easily recognized, no major uncertainty as to the base of the Missoula group is likely to arise even where the distinctive features of the latter group are poorly shown. Confusion between the carbonate rocks intercalated in the Missoula group and closely similar strata belonging to the Piegan group is a possibility but only where relations to the clastic beds of the Missoula group are obscure.

As first defined (Clapp and Deiss, 1931, p. 677–683) the Missoula group contained all the Belt series near Missoula above the unit called Wallace limestone by Clapp and Deiss, more commonly called Newland in that area. This limestone, irrespective of its name, is the local representative of the Piegan group of this report. Clapp and Deiss (1931, p. 691) refer to an "Upper Wallace limestone" equivalent to the Helena limestone near Helena. The Newland and Helena limestones in the Helena region are separated by a great body of clastic rocks of the sort typical of the Missoula group (Mertie and others, 1951, p. 18–20). Clapp and Deiss suggest that these two limestones thin westward and become indistinguishable near Missoula. However, Clapp and Deiss (1931, p. 685) indicate that the two limestones remain separated by clastic beds near Salmon Lake, some 30 miles, air line, from Missoula, and that a similar separation is recognizable in Glacier National Park. Mapping is not complete, but the author's trips through country between and near Missoula and Helena and to Glacier National Park lead him to the concept that the Newland, Wallace, and Siyeh are broadly equivalent to each other and that the Helena, in its type locality, is merely one of the several separate limestone masses included in the Missoula group of the present report. This concept was adopted in Glacier National Park (Ross, 1959; Ross and Rezak, 1959, p. 436) and is the only one that furnishes a definite, practical basis for mapping the Piegan and Missoula groups there.

The Helena limestone in the Big Belt Mountains and in and north of Helena, which is the region in which the name is commonly applied, varies markedly in thickness and character from place to place. As the distance from Helena increases, variations within the formation and interruptions in exposures introduce enough uncertainty so that one would be rash to select any particular exposure of such rock and attempt to equate it with the Helena limestone of the type locality (p. 84–85).

The Missoula group contains many formations, but only a few have been traced far from the localities where they were first recognized. In much of northern Idaho the sole representative is the Striped Peak formation, named from a peak in the Coeur d'Alene area (Ransome, 1905, p. 282), but along the boundary between Idaho and Montana the Libby formation, named for a town in western Montana (Gibson, 1948, p. 17–19), overlies the Striped Peak. Near Superior, Mont., the group has been divided (Campbell, 1960, p. 560–568) into the Spruce formation, Lupine quartzite, Sloway formation, Bouchard formation, and an isolated unnamed quartzite unit, listed in ascending order. As suggested on plate 3, some or all of these may well be equivalent to formations near Missoula but knowledge of the intervening area is insufficient to establish this.

In the Bonner quadrangle near Missoula, the Missoula group includes the Miller Peak argillite, Bonner quartzite, McNamara argillite, Garnet Range quartzite and, at the top, the Pilcher quartzite. Most of these names were proposed originally by Clapp and Deiss (1931, p. 678–683) but have been redefined and redescribed as a result of later work (Nelson and Dobell, 1961).

Reconnaissances by Nelson, Dobell, and the writer lead to the opinion that some of the formations recognized in the Bonner quadrangle persist southeastward to and beyond Philipsburg. Both in this region and between Missoula and Superior lateral variations hinder correlations. The Spokane formation near Philipsburg, in the sense employed by Calkins (Calkins and Emmons, 1915, p. 3), is not the Spokane of the type locality, but instead is almost synonymous with the entire Missoula group as that term is used here.

From near Philipsburg southward almost to lat. 45° rocks with the characteristics of the Missoula group have been seen in reconnaissances by the writer (Ross, Andrews, and Witkind, 1955) studied in places by others (Shenon, 1931, p. 45–46; Myers, 1952, p. 4–6; P. A. Guttormsen, 1952, written communication, see pl. 1, index map). Some of the rocks in this area are so coarse as to imply that they are related to the North Boulder group, defined below, which grades into the Missoula group. On both sides of the border of Idaho near lat 45° there are wide expanses of the Belt series difficult to correlate with the scanty available data. On plate 1 these are tentatively indicated as belonging to the Ravalli and Piegan groups.

Northeast of Missoula further work is needed before the subdivisions of the Missoula group can be regarded as established. In published reports (Clapp and Deiss, 1931, p. 691, correlation chart; Deiss, 1943a, p. 211–218), the units above the Newland limestone are called the Spokane argillite, Upper Siyeh limestone, Miller Peak argillite, Cayuse limestone, Hoadley formation and, at the top, the Ahorn quartzite. To the writer it seems probable that the Spokane argillite of Clapp and Deiss north of Ovando is in large part equivalent to the Miller Peak argillite as redefined by Nelson near Missoula, and that the Upper Siyeh limestone of Clapp and Deiss is one or more of the limestone masses in the Missoula group, not necessarily equivalent to any formally named formation, although called Helena limestone by them along Prickley Pear Creek (1931, p. 691). This statement is based on concepts gained in the Flathead and Glacier National Park regions (Ross, 1959) and near Missoula, and a reconnaissance along two valleys north of Ovando with Nelson and Dobell (1961) in 1955. One of these valleys is that of McCabe Creek where Clapp and Deiss (1931, p. 685, correlation chart) measured a section.

The mountains east and north of Helena contain some of the longest known exposures of the Belt series, including those in the Big and Little Belt Mountains where the series got its name. The formations in that region that belong to the Missoula group, as here defined, are, in ascending order, the Greyson, Spokane, and Empire shales, Helena limestone, Marsh shale (Walcott, 1899, p. 206–207), and, at the top, a newly recognized formation, the Greenhorn Mountain quartzite of Knopf (1950, p. 839). These same units, with some variations in thickness and character, extend southward to near Three Forks (Klepper and others, 1957; Robinson, 1963). The type localities are in the Big and Little Belt Mountains and in neighboring areas to the west near Helena and Marysville. Most are stated approximately by Walcott (1899, p. 206–207). The Greenhorn Mountain quartzite has been recognized so far only on the mountain of that name northeast of Helena.

NORTH BOULDER GROUP

The name North Boulder group was originally proposed (Ross, 1949, p. 111, 113) for arkosic and conglomeratic components of the Belt series in isolated exposures near the Jefferson River southeast of Butte. The name was taken from the North Boulder River (Boulder River on some maps). The group is thought to rest directly on the pre-Belt complex of metamorphic rocks and where originally named lies without angular discordance beneath the Flathead quartzite (Middle Cambrian). It is known as far east as T. 2 N., R. 6 E., in Gallatin County (Ross, Andrews, and Witkind, 1955) and may be represented some distance southwest of its type locality. Apparently few coarse rocks like the North Boulder group extend north of lat 46°. The westernmost area reported in which the group may occur is near Wise River (pl. 1). In the Three Forks quadrangle these rocks are called the North Boulder formation (Robinson, 1963).

The original definition (Ross, 1949, p. 111) suggested that the group might be equivalent in age to parts of the Missoula and Piegan groups. The revised classification adopted in the present report places in the Missoula group all the formations discussed at the time the original definition was set up. However, recent work farther east near Logan and Three Forks suggests that part of the North Boulder group may be much older than any in the type locality. It now seems appropriate to apply the name North Boulder group to the entire assemblage of strata of the Belt series in southwestern Montana that includes significant quantities of material decidedly coarser than is characteristic of most of the Belt series. The coarse rocks are near-shore deposits and may range in age through the whole of Belt time. The group seems to interfinger with strata of the Spokane and Greyson shales and with older units that may be correlative with the Newland limestone and even the Chamberlain shale.

The North Boulder group has not yet been divided into named formations. Probably it contains in each of several localities three or four subdivisions of formational rank (Peale, 1896, p. 2; Tansley, Schafer, and Hart, 1933, p. 11–12; Berry, 1943, p. 6; Perry, E. S., 1950; written communication, 1947; Sahinen, 1950, p. 13–15; Alexander, 1955; McMannis, 1952, p. 11–14).

The interrelations between the subdivisions in the different localities are not known.

PURCELL BASALT AND SIMILAR ROCKS

The only effusive rock recognized so far in the Belt series in Montana is the Purcell basalt of Glacier National Park and vicinity. The name was given by Daly (1912, no. 2, p. 207) because of its occurrence in what he termed the McGilliviary division of the Purcell mountain system. Within Glacier National Park it is limited to the lower part of the Missoula group, mostly between the Shepard formation and the top of the Siyeh limestone (Ross, 1959). In Canada more lava of the Purcell type is known than has been found in Montana, and its value as a horizon marker has long been recognized (Daly, 1912, p. 162; Schofield, 1914b, p. 86, 89). The name Purcell is restricted to beds high in the sequence, that is, to those in the Missoula group of the present report. However, north of Glacier National Park similar lava has been reported at much lower horizons (Fenton and Fenton, 1937b, p. 1887–1888).

In northern Idaho and Washington and in neighboring parts of British Columbia volcanic rocks of late Precambrian age have been called the Irene volcanic formation and the Leola volcanics. Petrographically both have resemblances to the Purcell basalt except that they are more metamorphosed and much thicker. Daly and most later students (1912, p. 144–147; Park and Cannon, 1943, p. 9–11; Kirkham and Ellis, 1926, p. 39–40) regard the Irene and Leola volcanic rocks as coextensive with and therefore equivalent to each other. They place these rocks so high in the sequence that it seems clear that they are stratigraphic components of the Missoula group of the present report. The suggestion (Kirkham and Ellis, 1926, p. 40) that they are essential equivalents of the Purcell basalt farther east is here held to be the most probable one. They are so much thicker than the Purcell basalt that precise age equivalence may be questionable. Okulitch (1949, 1956) does not discuss the volcanic rocks, but from other publications (Walker, 1934, p. 3–6; Little, 1950) it is evident that the Irene volcanic formation, where present, is so high stratigraphically that it would be included in what Okulitch (1956, p. 728–730) proposes to include the part of the Windermere series that is older than Cambrian but younger than Proterozoic. He wishes to elevate this part of the Windermere series to systemic rank, making it the earliest system in the Paleozoic era. If the Irene and Purcell are stratigraphic equivalents, and Okulitch's idea should gain acceptance, part or all of the Missoula group would belong in the new Paleozoic system. So far his idea has not gained general acceptance (Lochman-Balk, 1956, p. 637–646).

The Huckleberry greenstone, as used by Campbell and Loofbourow (1946, 1957), may be stratigraphically equivalent to the Irene volcanic formation. This unit and associated sedimentary formations are thought to be of Belt age but are so widely separated from most of the Belt rocks that correlations with subdivisions of the latter may be questionable.

DIAGNOSTIC FEATURES

Most of the rocks that compose the Belt series lack features of color or texture that are striking enough to be recognized readily. Recognition of stratigraphic units, in consequence, has had to depend largely on the personal predilections of each observer, including his past experience with the series. To be of maximum value diagnostic features should be sufficiently quantitative to permit different observers to obtain comparable and consistent results. Various kinds of evidence that may help to eliminate the personal equation in this respect are outlined below. Inadequate data now hamper their effective application, but recent studies and others that are planned are tending to decrease this disadvantage.

LITHOLOGIC CRITERIA

The characteristics that have been relied on in the past and that, in large measure, must continue to be used by stratigraphers who study the Belt series are those that contribute to the overall appearance of the rock. These include color, texture, internal structural features, gross composition, and degree of metamorphism. Nearly all available descriptions are expressed in such nonquantitative terms as to be of limited use.

COLOR

Color is one of the criteria most widely used in Belt stratigraphy. Its use is essential, but it is difficult to standardize color terms. The lack of agreement as to the names of rock colors is well known. There have been many attempts to set up standard color names, at least two of which were intended for rock descriptions (Goldman and Merwin, 1928; Goddard and others, 1948). These postdate many published descriptions of Belt rocks and have not been widely used even in recent publications. Even where color names have been applied with care one cannot be sure of the exact color implied. In the present paper, color designations are based so far as possible on the Rock Color Chart of the National Research Council (Goddard and others, 1948), but in descriptions taken from published reports the original color names have had to be used. Though of great assistance, the chart has obvious handicaps, especially as by no means all possible rock color variations are represented by the colored squares reproduced in it. More elaborate books dealing with color are available (Ridgway, 1886, 1912; Maerz and Paul, 1930; Cooper, 1941; Nickerson, 1946) but have been used little in geologic work, in part because of the odd-sounding color names employed.

Certain units of the Belt series have sufficiently distinctive and uniform colors to be of great aid in correlation. Caution is needed, however, as particular beds in different units may be so nearly identical in color that this distinction, taken alone, is valueless. Also, metamorphic effects may change color drastically. No single characteristic is a safe guide in correlation of the subdivisions of the Belt series.

Any method of visual comparison requires care and some measure of artistic ability on the part of the user. Hence it cannot be as rigidly objective or quantitative as could be desired. Means of actual measurement are available and have been applied to the Munsell colors (Judd and Kelly, 1939; Reimann, Judd, and Keegan, 1946). In spite of the obvious advantages in the elimination of the personal equation these devices have not come into use by geologists.

TEXTURE

Most of the clastic beds of the Belt series consist of fine grains, with original textures obscured somewhat by recrystallization. In some rocks the original grain shapes of quartz and feldspar may remain discernible, but clays and the decomposition products of other minerals have been largely converted into a micaceous mat that tends to mask original textures. Partial recrystallization of the quartz and feldspar contribute to this effect. The result is a rock sufficiently characteristic of the Belt series so that the experienced observer operating in the general region where the series is exposed commonly can say that a particular rock mass does or does not have the general appearance of belonging to that series. That same observer would have difficulty in expressing in quantitative terms the reasons for his decision.

The carbonate-rich rocks are chiefly fine grained. Many are so impure as to grade into clastic rocks like those just commented on. The carbonate-rich rocks are distinctive because of color, internal structures, and fossil content rather than texture.

In general, with practice, the observer tends to associate certain textural peculiarities visible in the hand specimen with particular formations or groups. Such peculiarities should be used with caution. Diagnoses based on petrographic data are somewhat safer, but are available for a few areas only. In texture, as in color, variations within the formations that make up the Belt series are more plentiful and marked than might be inferred from many of the published descriptions.

STRUCTURE

The structural features that characterize the rocks that are not rich in carbonate include ripple marks, raindrop impressions, salt casts, mud flakes, mud cracks, and variations in the bedding. Wavy laminations are characteristic of some units. Crossbedding is conspicuous in some of the more quartzose strata. In a few localities the quartzitic rocks also have bulbous masses of unknown origin on bedding planes.

The carbonate-rich rocks are replete with features that have attracted the attention of all who have studied them but are still imperfectly understood. The most widely known of these features are the molar-tooth structures, seen as deeply etched patterns on weathered surfaces. Those that gave rise to the name simulate the irregularities on the grinding surfaces of the molar teeth of elephants. The etched surfaces, however, display a great variety of patterns, many that have no resemblance to markings on teeth. One of the problems connected with the features is whether the diverse patterns originate in one or several different ways. There are so many characteristics in common and such a tendency for the different patterns to merge into each other that a common origin seems probable, though not proved. They result from an unequal distribution within the limestone of silica, magnesian carbonates, and possibly other components, producing variations in solubility during weathering. Although the features are conspicuous where the surface is weathered, they are nearly or quite invisible on freshly fractured surfaces in the same rock.

The features most appropriately likened to molar-tooth markings consist of series of sigmoid ridges with outlines irregular in detail. Each edge rises a tenth of an inch or more above the background of the weathered joint face. Other structures are sufficiently unsystematic in shape or arrangement so that the resemblance to molar-tooth makings is lost, although other characteristics are much the same. Both of these varieties may cover the entire side of a wide joint face or may be limited to a small area within that surface without apparent system in form or arrangement. Some structures are fairly straight, subparallel ridges, either nearly normal to or at some moderately high angle to the bedding, reminiscent of cleavage fractures. Some are assemblages of subparallel closely spaced short ridges that look like the edges of flattish cleavage fragments. These give the rock the appearance of a breccia. Many of them are areally associated with stromatolites.

Some structures consist of numerous short, subrounded ridges, commonly with haphazard distribution. Others are identical except that they are depressions sunk in the weathered joint surface instead of ridges rising above it. In both these varieties individual ridges or depressions tend to have shapes that resemble cuneiform writing. In still another variety the raised areas are oval, giving somewhat the appearance of an oolitic limestone in which the oolites are rather sparsely and irregularly distributed and of more than ordinary size. These oval bodies are as large as half an inch in maximum diameter. They do not have, so far as is known, the internal characteristics of oolites. A number of the carbonate-rich rocks in the Belt series are oolitic but the oolites are individually too small to be readily visible without the aid of the microscope.

Features resulting from irregular mingling of aggregates of different carbonates are locally conspicuous, especially in the Helena limestone near Helena. In these, dolomitic rock weathers to various shades of yellowish and grayish orange, and rock in which calcite predominates is gray. Locally, a mottled rock of distinctive appearance contains dark, nearly black, nodules and layers of chert. Limestone in the Belt series in other areas and in a wide range of stratigraphic positions includes material somewhat similarly mottled but in no locality familiar to the writer is such material as distinctively visible or as widespread as in the Helena limestone close to its type locality.

The lava is easily distinguishable from the rest of the Belt series by its igneous texture and, commonly, the conspicuous pillow structures. Therefore such rock is valuable as a local horizon marker. In and close to Glacier National Park it is called Purcell basalt and occurs in the Missoula group; mostly low in that group. However, as noted on page 31, lava may not be limited to this position throughout the Belt series. In addition to the possibility thus afforded for error in stratigraphic correlation, confusion might arise between lava flows and intrusive sills. The sills commonly show transgressive relations and may be associated with obviously crosscutting dikes. Further, sills that cut carbonate rocks commonly have conspicuous borders of light-colored marmarized limestone both above and below.

PALEONTOLOGIC CRITERIA

By Richard Rezak

Stromatolites are of outstanding interest among the structures of the rocks of the Belt series. Like the "molar-tooth" structures described, they are particularly abundant in the carbonate-rich rocks although they are by no means limited to these facies. The term "stromatolite" (Kalkowski, 1908, p. 68) has been generally accepted to designate certain laminated structures such as those that occur in the Belt series. These structures are formed through the growth of mats consisting of unicellular and filamentous green and blue-green algae. Stromatolites are distinguished from fossil algae because they do not have microstructures of a kind that would enable determination of the true biologic relationship of the organisms that contributed to the growth of the fossil. Consequently, classification of stromatolites necessarily involves the use of a purely artificial classification based upon "form genera" and "species," not true biologic entities.

The stromatolites are somewhat more visible than other structures on freshly broken surfaces because of the greater difference in composition from one layer to the next. On weathered surfaces the laminae are often etched into strong relief. (See fig. 1.) The stromatolites of the Belt series, especially those in and near Glacier National Park, have recently been studied and described by the writer (Rezak, 1957). That paper contains a classification with descriptions of the distinctive forms. The work demonstrates that stromatolites with distinctive characteristics are widespread in the Belt series. Where adequately mapped, they will prove of value for correlation within a given area and as a means of subdividing parts of the Belt series not divisible on other bases. At present, stromatolite zones have been used for such purposes only in the Glacier National Park region. Even there, limitations of time and inadequate base maps have prevented full use of these aids to stratigraphic study. Future work is expected to increase the number of subdivisions in that region that can be mapped on the basis of stromatolite zones and to disclose zones that can be used similarly in other areas. Comparisons between stromatolite zones in widely separated areas have little or no stratigraphic significance, because the same zone would not be continuous over the entire basin of sedimentation. However, some zones in one area may interfinger with zones of an adjacent area, and by gradually working out their relationships over the entire basin, it may be possible to make long-range correlations.

Eight stromatolite zones useful for local correlation have been described by the writer (Rezak, 1957, p. 135–144) in detail in the Glacier National Park region. The zones range throughout the formations of the series that are represented there, being absent only from the Appekunny argillite. They vary in thickness, composition, and areal extent. Some are widespread and extend into neighboring regions, while others occur only in restricted areas. Only two of the zones have been mapped. Stromatolites found at several localities in

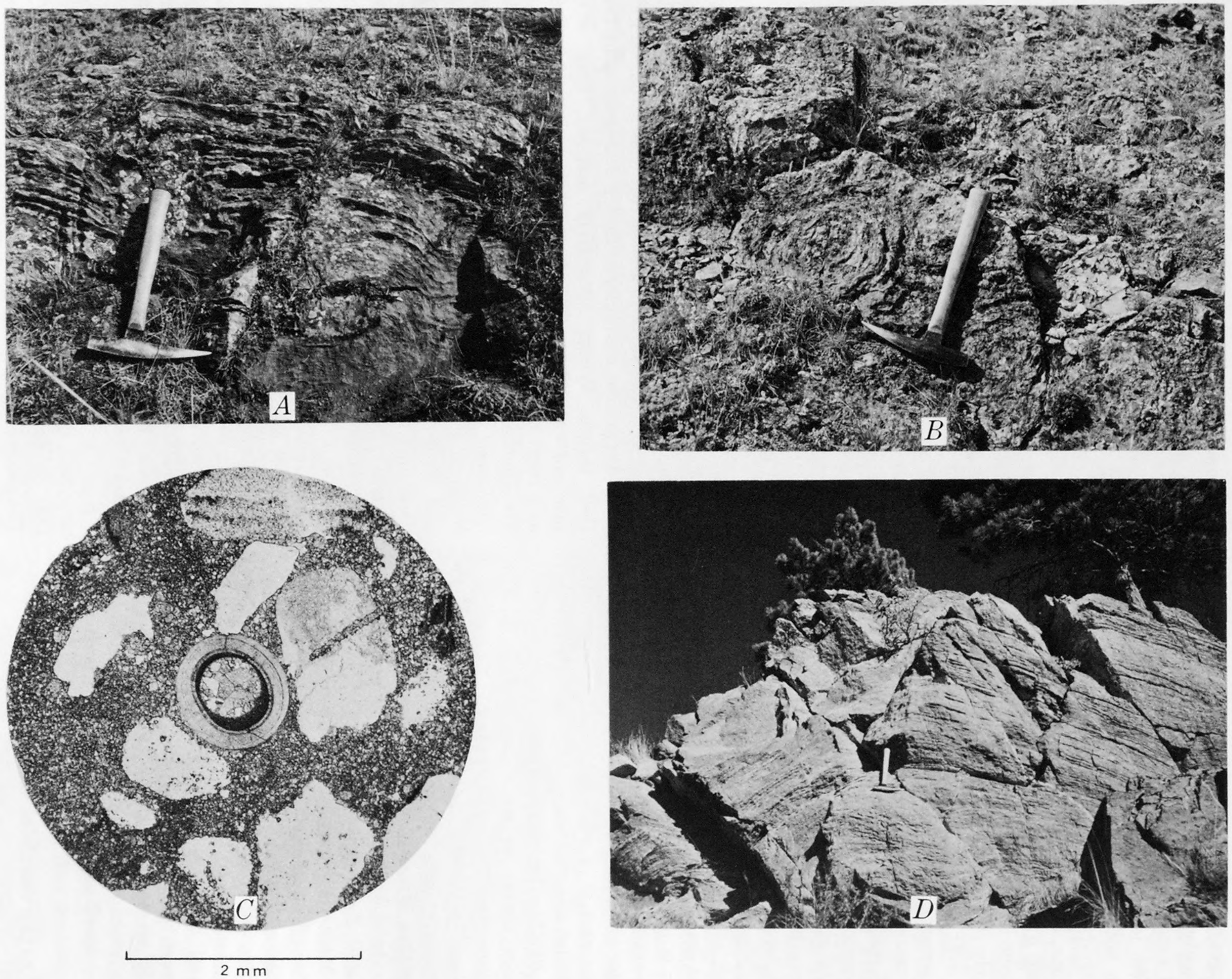

FIGURE 1.—Stromatolites in west-central Montana. *A, Cryptozoon occidentale* Dawson. Spokane shale, SE¼SE¼ sec. 23, T. 4 N., R. 2 W., Devils Fence quadrangle, Montana. *B, Collenia multiflabella* Rezak. Greyson shale, NW corner NW¼NE¼ sec. 22, T. 4 N., R. 2 W., Devils Fence quadrangle, Montana. *C,* Organic(?) structure in thin section of Altyn limestone. Specimen from near Swiftcurrent Falls, Glacier National Park, Mont. × 21.75. *D, Collenia symmetrica* Fenton and Fenton. Helena limestone, at top of hill 0.7 mile east of the west end of Helena Scenic Drive, Helena, Mont. SE¼ sec. 26, T. 10 N., R. 4 W.

northwestern Montana and northern Idaho give basis for the expectation that the Belt series in these areas is divisible on the basis of stromatolite zones.

Stromatolites are significant from standpoints other than that of stratigraphic correlation. They give clues as to the mode of origin of the series and of Precambrian rocks in other regions. The author (Rezak, 1957, p. 141–149) has discussed the environmental implications of stromatolites in some detail. However, the stromatolites serve to show that the beds containing them were laid down in shallow water, possibly on marginal mud flats that were periodically wet and dry. They also tend to support the concept of marine rather than lacustrine origin of the Belt series, as the stromatolites more nearly resemble present marine forms than those of lacustrine habitat (Rezak, 1957, p. 149; Ross and Rezak, 1958). The Belt algae may have formed under conditions approximating those of the Bahama Banks or some of the coral reefs in the southern Pacific Ocean.

In the Coeur d'Alene mining region, Idaho, stromatolites occur in the Prichard, Burke, and Wallace formations. To date the only recognized genus in this area is *Newlandia* Walcott. Several good exposures of these structures in the Prichard formation occur along the east side of Little Pine Creek valley about 2 miles southeast of Pinehurst, Idaho. (See fig. 2*D*.) They are scattered throughout a 15-foot zone of greenish-gray argillite beds, but are especially abundant in a 2-foot argillite bed that can be traced for 250 feet along the strike. This horizon is about 600 feet stratigraphically below the contact of the Prichard formation with the overlying Burke formation.

The *Newlandia* Walcott has also been observed near the middle of the Burke formation in an outcrop on a rocky spur east of the elementary schoolhouse in Kingston, Idaho. The bed containing the structures is a buff fine-grained slightly sericitic quartzite. The exact stratigraphic position is not known because of faulting in the vicinity and the lack of an index horizon. This occurrence is probably 1,500 to 2,000 feet stratigraphically above that in the Prichard formation.

Near Troy and Libby, Mont., stromatolites occur in the Wallace and Striped Peak formations. The Wallace formation contains an abundance of *Collenia symmetrica* Fenton and Fenton (fig. 2*B* and *C*). The following stratigraphic section shows the typical relationships of the stromatolites in this formation. (See fig. 2*B*.)

Section measured half a mile north of Troy Ranger Station, on northeast side of road along northeast side of Kootenai River, SW¼ sec. 36, T. 32 N., R. 34 W.

	Ft	In.
Sandstone, pale-orange and gray, fine-grained, calcareous	6	0
Limestone, pale-red, stromatolitic (*Collenia symmetrica*)	3	0
Sandstone, light-gray, fine-grained, calcareous, weathers to dark yellowish orange	2	8
Limestone, pale-red, stromatolitic (*Collenia symmetrica*)	3	0
Argillite, pale-brown, finely laminated, calcareous; massive but breaks into thin slabs	2	0
Sandstone, yellowish-gray, calcareous; contains thin black bands	2	6
Dolomite, dusky-red-purple, very fine grained, thinly bedded; weathers rusty brown	2	0
Limestone, stromatolitic; matrix moderate red and laminae black; species indeterminate	0	6
Oolite, pinkish-gray, calcareous; bed contains pebbles of oolitic rock as much as 2 inches long and three-fourths inch thick	1	6
Sandstone, red, very fine grained, calcareous, with laminations at base of unit	15	0
Total	38	2

Another exposure of *Collenia* beds worthy of mention occurs beneath the south abutment of the suspension bridge across Kootenai River just west of Kootenai Falls picnic ground, sec. 36, T. 31 N., R. 33 W. Here from 10 to 20 feet of greenish-gray stromatolitic limestone occurs in the Wallace formation. These stromatolites are similar to *Collenia symmetrica* and attain maximum diameters of 10 feet. (See fig. 2*A*.)

The writer has not observed stromatolites in the Striped Peak formation. However, C. E. Erdmann (written communication) mentions stromatolites in association with oolitic limestone and gray sandy limestone about 200 feet above the base of the Striped Peak formation on the south slope of Mount Berray above Windquist's prospect, at an elevation of about 4,500 feet, T. 27 N., R. 33 W., unsurveyed. Another of Erdmann's localities is in the Trout Creek 30-minute quadrangle. Here stromatolites occur associated with oolitic limestone and pink shaly limestone about 1,000 feet above the base of the Striped Peak on the ridge east of East Fork Elk Creek, 2 miles south of Delravik's ranch, T. 26 N., R. 33 W., unsurveyed.

The Helena limestone near Helena, Mont., contains stromatolites. Gigantic specimens of *Collenia symmetrica* may be observed on a hill 0.7 mile from the west end of the Mount Helena Scenic Drive, SE¼

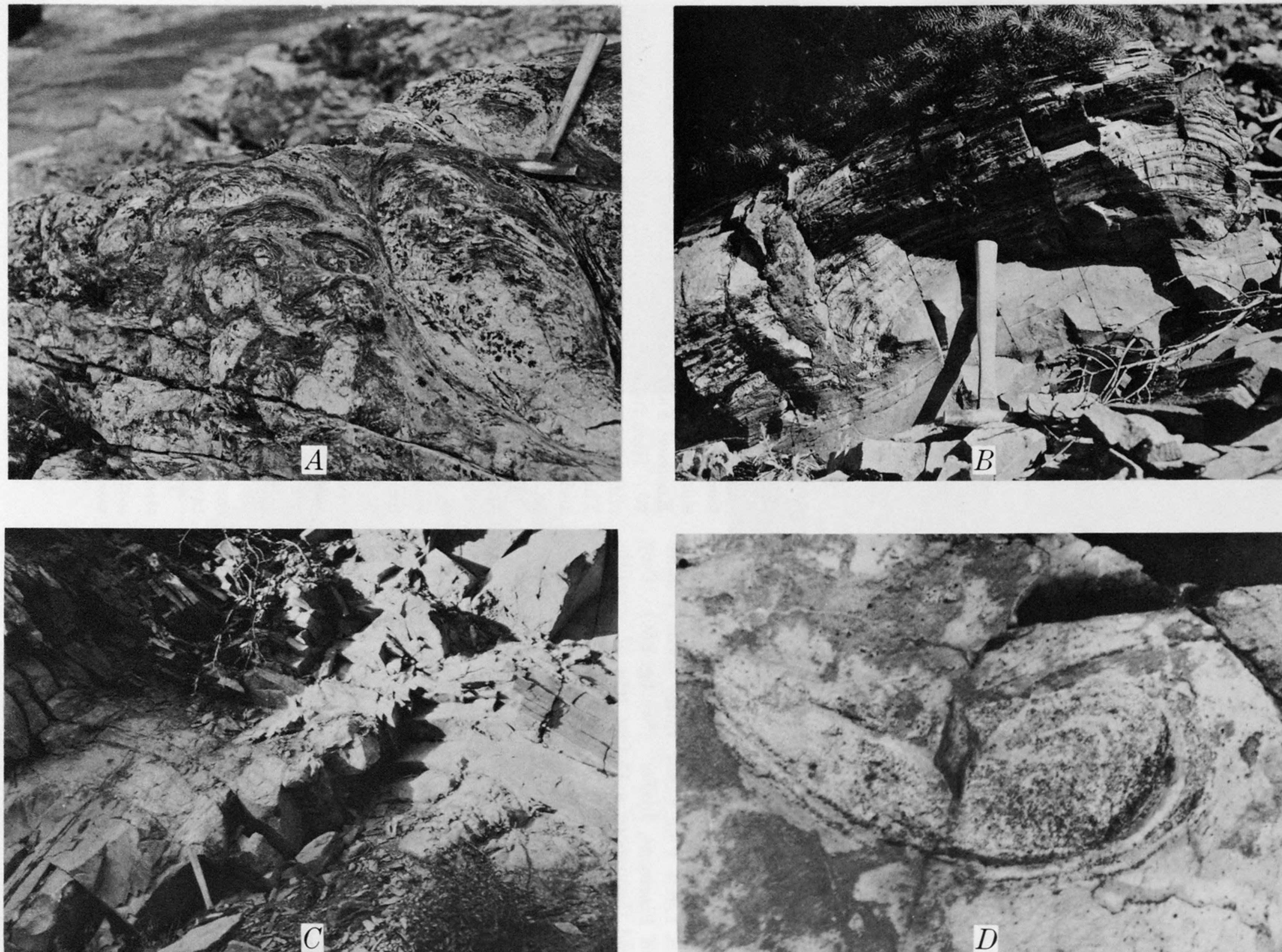

FIGURE 2.—Stromatolites in and near Idaho. *A, Collenia symmetrica* Fenton and Fenton. Wallace formation, on south side of Kootenai River, under suspension-bridge abutment, just downstream from picnic ground. Sec. 36, T. 31 N., R. 33W. *B, Collenia symmetrica* Fenton and Fenton. Wallace formation, ½ mile north of Troy Ranger Station on the northeast side of road along northeast side of Kootenai River. This outcrop illustrates the fact that in many instances stromatolites are not readily recognized as such. *C, Collenia symmetrica* Fenton and Fenton. Wallace formation, along lower road on northeast side of Kootenai River, about 1¼ miles southeast of Troy, Mont. *D, Newlandia* sp. Prichard formation, about 2 miles southeast of Pinehurst, Idaho. U.S.G.S. Algae No. a 1. × 1.

sec. 26, T. 10 N., R. 4 W. (See fig. 1*D*.) Stromatolites are found on the top and both flanks of the hill, but the best exposures are found high on the west flank.

South of Helena, near Radersburg, Mont., the Greyson shale and Spokane formation both contain abundant stromatolites. In the Devils Fence 15-minute quadrangle, northwest corner NW¼NE¼ sec. 22, T. 4 N., R. 2 W., the Greyson shale contains forms similar to *Collenia multiflabella* of the Glacier National Park area. These beds occur in limestone beds about 600 feet below the contact of the Greyson with the overlying Spokane shale (fig. 1*B*).

The Spokane shale contains stromatolite beds about 200 feet above its base. In the Devils Fence quadrangle, SE¼SE¼ sec. 23, T. 4 N., R. 2 W., several limestone beds contain *Collenia symmetrica* and *Cryptozoon occidentale*. (See fig. 1*A*.)

Farther south, in the Horseshoe Hills area near Logan, Mont., Peter Verrall reports (1955, p. 24–25) the existence of a basal stromatolite unit in what he regards as the Newland limestone. He identifies the stromatolites as *Collenia symmetrica* and, using these fossils, correlates the basal Newland with the basal Siyeh of Glacier National Park, some 200 miles to the north. Earlier in this section it has been indicated that long-range correlations on the basis of stromatolites is hazardous. Considering the abruptness of changes in facies in the Belt series and the stratigraphic range of the fossils, Verrall's correlation is vulnerable.

C. D. Walcott (1914b) and C. L. Fenton and M. A. Fenton (1937b) have described the occurrence of stromatolites in the Belt series of northwestern Montana. These papers are readily available and the occurrences reported in them will not be reviewed at this time.

Fossils other than stromatolites have been reported in the Belt series, mainly by Walcott (1899, 1914, 1915) and Fenton and Fenton (1931, 1933a, 1936, 1937b). These fossils include bacteria, algae, crustaceans, brachiopods, and indirect evidences of life such as burrows and trails.

The earliest report of the presence of fossil bacteria in the Belt series was by Walcott (1915, p. 256). He described and illustrated specimens of *Micrococcus* sp. from the Gallatin formation (of Keyes) on the north side of East Gallatin River, 5 miles east of Logan, Gallatin County, Mont. The discovery of algae and fungi in the Precambrian Gunflint iron-formation of the Canadian Shield (Tyler and Barghoorn, 1954, p. 606) proves the presence of these primitive organisms in ancient sedimentary rocks. Preservation of such fossils for any great length of time depends largely upon the type of host rock. Black chert appears most favorable for preservation, whereas limestone is least favorable. One of the places where chert beds have been observed in the Belt series is on the ridge between Dry Gulch and Deep Creek, SE¼ sec. 35, T. 14 N., R. 21 W., Missoula, Montana-Idaho 30-minute quadrangle. There several stromatolite-bearing beds are composed of alternations of light-gray and black chert. Specimens of the black chert were macerated in hydrofluoric acid and the residues examined under magnifications of as much as 1,200 times. The residues appear to be an organic hash too poorly preserved to permit positive identification of its components. Further searching for black chert may turn up well-preserved plant fossils.

The Fentons (1933b, p. 190) reported the discovery of oboloid brachiopods in the Newland limestone. In a later paper (1936, p. 616) they described a new species: *Lingulella montana* from the middle and upper Newland in cliffs along Little Birch Creek, NW¼ sec. 15, T. 9 N., R. 4 E., Meagher County, Mont. The area in which the fossils are found is cut by many faults which the Fentons (1936, p. 610) state are of small throw or follow bedding planes, and do not introduce Cambrian rocks into the sequence. The measured sections included in the paper show that the brachiopods are found 27 to 30 feet above beds containing Walcott's Precambrian flora. The sections record brecciation and fracturing in the beds below those containing the brachiopods. Rezak's observations raise the possibility that unit 2 of Fentons' section *A* (1936, p. 611) and unit 8 of their section *B* may well be fault slices that had displacement such that Cambrian strata were caused to rest directly on the upper Newland.

Fossil crustaceans have been described by Walcott (1899, p. 238) from the Greyson shale in Deep Creek canyon near the former town of Glenwood, and in Sawmill canyon 4 miles above Neihart, Mont. He proposed the name *Beltina* for carbonaceous films that he supposed to be fragments of Merostomata. David White (1929, p. 393) and the Fentons (1936, p. 616) conclude that these are partly inorganic and partly the remains of the soft fronds of brown algae.

Clyde P. Ross, while examining thin sections cut from specimens of the Altyn limestone collected near Swiftcurrent Falls in Glacier National Park, discovered a structure that has been tentatively identified by J. Harlan Johnson as organic, although he is not sure what it represents. (See fig. 1*C*.) Johnson adds (oral communication) that if found in lower Paleozoic limestone, he would not hesitate to say that it belonged to a trilobite or chitinous brachiopod.

Occasionally traces resembling fossil burrows and trails are found in the Belt series. These have been described in some detail by Walcott (1899, p. 236) and the Fentons (1937b, p. 1950). The structures are attributed to the work of annelids, mollusks, and crustaceans.

PETROGRAPHIC CRITERIA

The rocks that make up the Belt series were originally mainly sandstone, siliceous mudstone and siltstone, and limestone, mostly impure, and magnesian. Outstanding petrographic peculiarities in these sedimentary rocks useful in stratigraphic correlation are not abundant, but a few are worthy of comment. The igneous rocks locally are useful in correlations but are so distinctive as to present few problems. Petrographic data are so scattered and incomplete that the remarks below are far from conclusive.

CRITERIA IN IGNEOUS ROCKS

The Purcell basalt in the Missoula group in a few places in Montana is commonly so easily recognized that petrographic study is not needed in order to recognize the stratigraphic unit. It would be of interest to know whether petrographic data would aid in determining how far the various names applied to volcanic rocks intercalated in the series in British Columbia, Washington, and Idaho reflect stratigraphic differences, but available data are not decisive. The Purcell basalt, which was studied only in Glacier National Park, is believed to have had the composition of basalt originally. It is much altered, perhaps in part by hydrothermal processes related to eruption. Daly (1912, p. 144–147) regarded his Irene volcanic formation as andesitic lava, altered to greenstone. The mineral composition and structural features of the two volcanic assemblages are so similar that the fact that one is termed "basalt" and the other "andesite" is of doubtful significance.

The Leola volcanics in northeastern Washington (Park and Cannon, 1943, p. 9–11) is credited with being equivalent to the Irene volcanic formation of Daly. The major distinguishing features include the pronounced schistosity and the tuffaceous material in the Leola volcanics. Park and Cannon speak of the Leola volcanics as altered basalt. Petrographically, then, the Purcell, Irene, and Leola volcanics are all calcic extrusive rocks. Whether the differences in metamorphism noted are sufficient to offset other evidence tending to show that the three are approximately correlatable stratigraphically will remain a question until further information is available.

Intrusive dikes and sills invade horizons below the Purcell basalt in Glacier National Park. The sills, in particular, might add to difficulties in stratigraphic correlation. One readily made petrographic distinction is that the minerals of the flows are more thoroughly altered than those of the dikes and sills. Further, the texture of those parts of the effusive rocks that approach diabase in appearance is less typically diabase and less uniform than that of the diabasic parts of the intrusive rocks. Another distinction that is probably of greater diagnostic value is that the intrusive rocks commonly contain micropegmatite whereas the effusive ones appear to be free of this constituent.

CRITERIA IN CARBONATE ROCKS

The carbonate rocks at different horizons and localities vary in the proportions of calcium and magnesium carbonates, and some also contain significant quantities of iron carbonate. Most of the units customarily called limestones are magnesian and parts of them are dolomite. Some are detrital (fig. 3*E*), rather than chemical, aggregates of calcite and dolomite plus quartz, feldspar, and micaceous minerals (fig. 3*A*). These include edgewise conglomerate (fig. 3*A*) and oolitic aggregates, many of both being clearly visible only under the microscope (fig. 3 *B*, *C*, *D*, *F*). Much, possibly all, of the carbonate in rocks of the Belt series is genetically related to organic processes, mainly those that produced the stromatolites (fig. 3*B* and *C*). Nevertheless, most of the carbonate-bearing rocks are far from being pure. The microscope shows that the impurities consist dominantly of clastic grains identical with those of the argillite and quartzite beds.

Some of the carbonate rocks throughout the Belt series can be seen to be oolitic, even in the hand specimen. Under the microscope this is a conspicuous feature, and the oolites show differences in detail. Many have little or no internal structure preserved (fig. 3*F*), but some of the Siyeh limestone contains pebbles consisting of groups of oolites with intricate internal structure (fig. 3*B*). Perhaps when more data are accumulated, deductions of value in stratigraphic correlation can be drawn from items of this kind.

Structures of the kinds already commented on as prominent in outcrop and hand specimen are also conspicuous under the microscope (fig. 4). It is difficult to distinguish in thin sections between structures like that shown in figure 4*B*, a variant of molar-tooth structure, and those exemplified by figure 4*C*, supposed to be stromatolitic. Figure 4*A* shows intraformational conglomerate in limestone, a feature that records disturbance of strata prior to consolidation. Perhaps most of the widely variable forms in molar-tooth and similar structures result from slumping and shifting of the soft calcareous ooze before and during cementation. Where that ooze already contained organic or other structures, the results would be more complex than in their absence.

CRITERIA IN CLASTIC ROCKS

Petrographic study supplements field observation in showing that most of the Belt series consists of fine-grained clastic rocks with few notable characteristics, none so far valuable in correlation. Mud flakes are seen to be even more abundant than can be appreciated from inspection of outcrops and hand specimens (fig. 5*B* and *D*). Many have laminae of microscopic dimensions, commonly from 0.03 to 0.5 mm in width (fig. 5*B*, *C*, *E*, *F*). Typical material consists of grains a few thousandths to a few hundredths of a millimeter in diameter, that is of silt and clay sizes, embedded in a matrix in which flaky minerals are conspicuous. Quartzite beds that in the field seem conspicuously coarse mostly consist of grains a millimeter or less in diameter (fig. 5*A*). Even in conglomerate, with pebbles 100 mm and more in maximum dimension, much of the matrix consists of grains 0.01–0.10 mm in diameter. The nonflaky grains in nearly all the rocks are subangular to poorly rounded, where recrystallization has not obliterated original form.

It is difficult to discern the quantitative composition of these fine-grained rocks under the microscope. Broadly the rocks consisted originally of quartz, clays and other flaky minerals, feldspars, probably some ferromagnesian minerals, and, rarely, fragments of rocks, which are mostly argillaceous. More contain carbonate than might be suspected from the hand specimen, but, on the other hand, some rocks that superficially resemble limestone are high in noncarbonate clastic constituents. Some are now colored by fine grains of hematite.

In mineral composition these rocks resemble graywacke and subgraywacke (Pettijohn, 1949, 1957, p. 303). In the textures and structures that reflect their origin they differ from typical graywacke. They are finer grained and the component grains are better rounded than in graywacke. Also most show both crossbedding and ripple marks, which are reported not to be characteristic of graywackes. Their colors are, on the whole, far from brilliant but scarcely as dark as those of graywacke, in part because chlorite is less prevalent. Although close examination, in hand specimen and thin section, shows the fine-grained rock of the Belt series to be well bedded, many outcrops have a massive appearance, with joints of various kinds but rather commonly without shaly or slaty partings. This tendency to an absence of fissility is reminiscent of many graywackes.

Some of the coarser clastic material in the Belt series resembles arkose (figs. 3*A* and 5*E*). Examples can be found in parts of the Altyn limestone and the North Boulder group. However, near Neihart, one of the few localities where Belt rocks are well exposed resting on granitic gneiss, arkosic beds are absent. The sample of Neihart quartzite listed in table 16 contains more than 96 percent silica. This is in harmony with the abundance of quartz in the thin sections examined. (See p. 59.)

A larger proportion of the clastic rocks contain feldspar than would be suspected without close observation. Even though in many the amount present is small it may be diagnostic because the overlying Flathead quartzite of Cambrian age appears to be free of feldspar. In many places Flathead quartzite is difficult to distinguish from other quartzites intercalated in the Belt rocks, a fact that complicates correlation and interpretations.

CRITERIA RELATED TO DEGREE OF METAMORPHISM

The Belt series in Montana shows low-grade metamorphism except along the borders of large granitic masses such as the Idaho batholith. Locally the metamorphic effects obscure or obliterate the characteristics on which stratigraphic correlations are normally based. These features are beyond the scope of this report. The extent that regional metamorphism has affected large expanses does require consideration here. It is reflected in the recrystallization of sand grains, the conversion of clay into micaceous minerals, and the production of feldspars either by union of radicals derived from clastic grains of minerals of suitable composition or by recrystallization, presumably with modification in the composition of the feldspar. These and similar effects of low-grade metamorphism are observable in various degrees in the rocks of the Belt series whereever they have been studied. In many rocks small amounts of clay minerals remain unaltered and probably much of the feldspar is in original clastic grains with little or no change in composition since deposition in the sediments.

On the basis of the scattered observations so far made, the clastic rocks in the Missoula group appear less metamorphosed than those lower in the Belt series and there may even be a progressive change within the group.

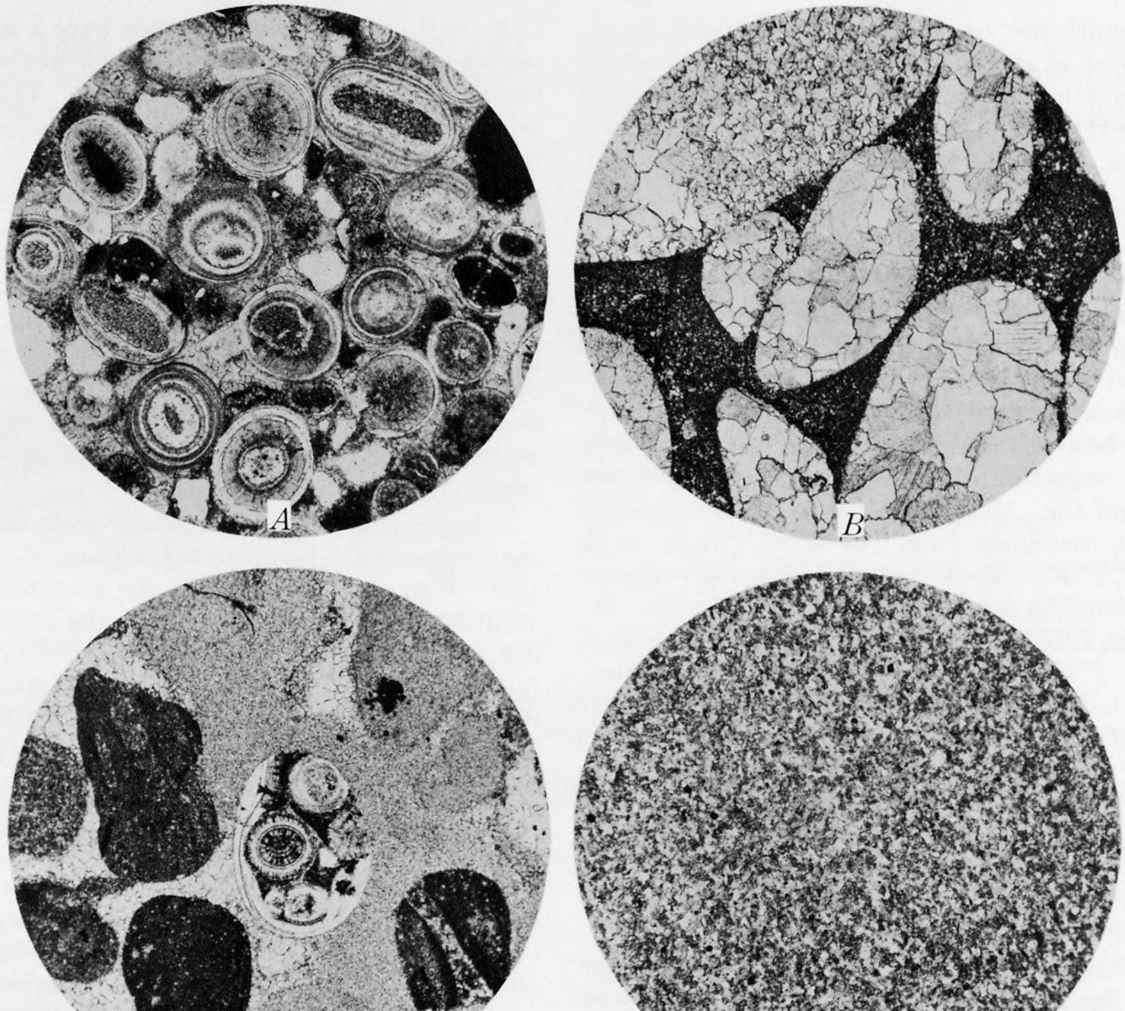

FIGURE 3.—Photomicrographs of carbonate rocks (photographs by Wendell Walker).

This statement is based mainly on the fact that the quartz in the Missoula group retains the outlines of the original detrital grains better than that in many pre-Missoula rocks. The distinction is none too positive and many quartz grains in the Ravalli group, particularly the coarser ones, retain faint traces of their original shapes. In thin section these traces consist mainly of thin and in part discontinuous lines of dust particles. The quartz beyond these lines, within the area of the cementing matrix of the original sediment, is at least in approximate optical continuity with that in the core of the grains.

In the parts of Montana familiar to the writer the Prichard formation (or slate) commonly appears to be more highly metamorphosed than most of the Belt series in Montana. In part this is certainly due to igneous metamorphism along the borders of the Idaho batholith, but whether any of the metamorphic effects are regional rather than local is unknown. If regional, their presence would be in harmony with the fact that the Prichard is one of the oldest units in the Belt series.

As table 14 below indicates, most of the rocks of the Belt series in Montana contain small quantities of clay minerals. The table records the results of X-ray examination of material of clay size separated from small samples. In all rocks the quantity of clay present is so small that it was not noted in ordinary petrogarphic study. Ogden Tweto [1] reported clay minerals in many Belt rocks from petrographic examination alone.

The specimens analyzed were unweathered, and the clay minerals in them are thought to be original, perhaps recrystallized to some extent during the long interval between the time of sedimentation and the present day. One conspicuous feature of the table is its uniformity. Montmorillonite, although not abundant, is present in most samples. Illite is almost as widespread. Mica is widespread, although not reported for some of the samples analyzed. However, the uniformity is sufficient to show that the clay components of the rocks of the Belt series are without stratigraphic value.

[1] Ogden Tweto, 1937, Petrography of pre-Cambrian and Paleozoic rocks of Montana and Yellowstone National Park: Unpub. Dissertation, Montana State Univ.

CHEMICAL CRITERIA

MAJOR COMPONENTS

Little information has been published in regard to the chemical composition of the rocks of the Belt series. During the present investigation 10 argillite samples of the Grinnell and Appekunny formations were submitted for analysis with the results shown in table 15. Three analyses from the lower part of the Belt series in scattered localities are given in table 16. The dominantly clastic rocks of the Missoula and North Boulder groups are represented in tables 17, 18, and 19, which together give 22 analyses from widely scattered areas in Montana. Table 20 gives 10 analyses of carbonate rocks from various places in the State. Table 21 gives 12 miscellaneous analyses from eastern Idaho and western Montana. Table 22, which shows the average composition of various kinds of sedimentary rocks, is added for comparison with the analyses from the Belt series. Both for these chemical analyses and for the X-ray data in table 14, it is impracticable to include here sufficiently detailed maps to show precise locations. Those interested can identify the locations given in the tables by consulting pertinent maps.

EXPLANATION OF FIGURE 3

A. Specimen A 13. Altyn limestone. This rock contains more than the average amount of clastic grains of quartz and feldspar, or the two together (0.4–2.0 mm in diameter) in a fine-grained, carbonate matrix (diameter 0.04–0.08 mm). Most of the feldspar is plainly twinned. The rock contains oolites, some of which are silicified, but no oolites are visible in this photograph.

B. Specimen FM 333. Siyeh limestone. Rock contains pebblelike forms as much as 1.5 mm long regarded as oolites without internal structure, but one in the view is a compound oolite with structure. The matrix is mainly fine-grained carbonate and quartz but there are also aggregates of coarser, later quartz.

C. Specimen G N P.–S–16a. Siyeh limestone. This limestone contains exceptionally abundant and well-preserved oolites, of which many are 0.6 mm in diameter, in a fine-grained matrix which includes some quartz, feldspar, and mica.

D. Specimen Lbby–w–1. Wallace formation. This rock contains plentiful oolites, with diameters as much as 0.6 mm, in a comparatively coarse, clastic-textured matrix. The internal structure of the oolites is not as well-preserved as in specimen G N P.–S–16a. Most have quartz nuclei. The matrix includes quartz, orthoclase, plagioclase, and a little mica imbedded in carbonate. The clastic grains are as much as 0.1 mm in diameter.

E. Specimen MM 4. Newland limestone. This is an impure limestone with nearly 50 percent carbonate in grains as much as 0.03 mm in diameter. The rest of the rock is quartz, 2 kinds of feldspar, mica, and a little epidote. Some of the quartz and feldspar grains are as much as 0.1 mm long.

F. Specimen FM 67. Shepard formation. This rock contains recrystallized oolites as much as 4.0 mm long in which the carbonate grains are about 0.2 mm in diameter. The matrix is mostly carbonate in grains about 0.01 mm in diameter.

FIGURE 4.—Photomicrographs of structures in carbonate rocks, Siyeh limestone. (Photographs by Wendell Walker.) *A*, Specimen FM319. This rock is especially irregularly laminated, perhaps because of movement prior to consolidation. Some laminae contain quartz and plagioclase grains as much as 0.03 mm in diameter but much of the carbonate is in grains 0.006 mm in diameter. *B*, Specimen FM341. Most of this rock consists of carbonate in grains about 0.005 mm in diameter, but some layers contain clastic quartz grains as much as 0.03 mm in diameter. The contorted layers are mainly rather coarse carbonate (0.1–0.5 mm in diameter). Their origin is not known. *C*, Specimen Belt 3. The groundmass is carbonate with some quartz plagioclase and mica, mostly 0.005 to 0.01 mm in diameter. Rare detrital grains of zircon, epidote and rutile are present. The conspicuous irregular structures throughout the rock may be of stromatolitic origin.

TABLE 14.—*Results of X-ray analyses of rocks of the Belt series in Montana for clay minerals*

[Analyses by A. J. Gude, 3d and sample N 3 by E. W. Tooker. 1, major constituent; 2, minor constituent; 3, trace]

Serial	Field	Formation or group	Location	Chlorite	Mica	Illite	Mont-morillo-nite	Mixed layer clays	Remarks
239381	SM 13	Missoula group, probably Miller Peak argillite	U.S. Highway 93, near mouth of Miller Creek near Missoula.		1				
239383	19-47	Missoula group	Near Lolo Hot Springs west of Missoula		1				
242154	32-GP	Shepard formation	Flattop Mountain, Glacier National Park	1		2	3		Argillite.
242148	FM 67	Shepard formation	Flattop Mountain, Glacier National Park, NW¼ sec. 22, T. 35 N., R. 17 W.			2	2	2	Dolomite.
242155	Belt No. 10	Missoula group	Middle Fork of Flathead River, Glacier National Park near Walton.	2		2	3?		Limestone.
242137	FM 7	Missoula group	Argosy Mountain, Nyack quadrangle, Flathead region.	2	1		2		Red argillite.
242138	FM 9	Missoula group	Kah Mountain, Nyack quadrangle, Flathead region.			2	3		Pink quartzite.
242161	FM 22	Missoula group	Miner Creek, Marias Pass quadrangle, Flathead region.	3	2		3		Quartzitic argillite.
242133	FM 33	Missoula group	Pioneer Ridge, Nyack quadrangle, Flathead region.	2	2		3		Green argillite.
239390	10-4-15	Garnet Range quartzite	Clark Fork between Allan and Martinez Creeks, Bonner quadrangle, NE¼ sec. 19, T. 12 N., R. 17 W.	2	2		3		Black shale.
239385	4-57-13	McNamara argillite	Blackfoot Highway, Bonner quadrangle, SW¼ sec. 14, T. 13 N., R. 18 W.	2		3			
239384	4-57-10a	Bonner quartzite	Blackfoot Highway, Bonner quadrangle, sec. 14, T. 13 N., R. 18 W.		2				
239388	7-17-22	Bonner quartzite	Pattee Creek, Bonner quadrangle, NE¼ sec. 3, T. 12 N., R. 19 W.		1		3		
239391	11-42-2	Miller Peak argillite	Miller Peak, south of Missoula. Type locality, Bonner quadrangle, NW¼ sec. 32, T. 12 N., R. 19 W.	3	2		3		Red argillite.
239387	7-17-17	Miller Peak argillite	Pattee Creek, Bonner quadrangle, NW¼ sec. 3, T. 12 N., R. 19 W.	3	1				Laminated quartzite and argillite.
239386	6-45-7	Miller Peak argillite	South side of Rattlesnake Creek, Bonner quadrangle, NW¼ sec. 30, T. 14 N., R. 18 W.	2		3			Hornfels.
239389	7-106-5a	Miller Peak argillite	U.S. Highway 10 at Kendall Creek, Bonner quadrangle, SW¼ sec. 7, T. 12 N., R 17 W.	3	1				Green argillite.
242160	MM 1	Marsh shale	North of Helena, Helena quadrangle	3	2		3?		
242156	MM 5	Helena limestone	Avalanche Creek, Canyon Ferry quadrangle, Big Belt Mountains.	2	2		3		
242159	MM 3	Empire shale	Magpie Creek, Canyon Ferry quadrangle, Big Belt Mountains.			3?			
242158	MM 6	Spokane shale	Spokane Hills, east of Helena and west of Canyon Ferry quadrangle.	3	1		3?		
242157	MM 4	Newland limestone	Magpie Creek, Big Belt Mountains, Canyon Ferry quadrangle.			2	3?		
242150	FM 26	Siyeh limestone	Graves Creek, Nyack quadrangle, Flathead region.	2		2	2	2	
242149	FM 22	Siyeh limestone	Twin Lakes, Nyack quadrangle, Flathead region.	2	2		3?		
242147	FM 21	Siyeh limestone	Noisy Creek, Nyack quadrangle	2		2	3		
242146	FM 28	Siyeh limestone	Tongue Mountain, Nyack quadrangle			2	2	2	
242145	FM 29	Siyeh limestone	Head of Clayton Creek, Nyack quadrangle	2	2		3		Argillaceous.
242141	FM 18	Gradation zone between Siyeh and Grinnell.	Near head of Wheeler Creek, Nyack quadrangle.	2	2		3		
242139	FM 19	Grinnell formation	Near head of Wheeler Creek, Nyack quadrangle.	2		2	3		Coarse quartzite near top.
242131	FM 23	Grinnell formation	West slope of Swan Range, Nyack quadrangle	2	1		3		Purple argillite.
242132	FM 24	Grinnell formation	West slope of Swan Range, Nyack quadrangle	2	1		3		Green argillite.
242134	FM 16	Grinnell formation	Tom Tom Lookout, Nyack quadrangle	2	2		3		Intraformational conglomerate.
242135	FM 53	Grinnell formation	Mount Baptiste, Nyack quadrangle			2	3		Siliceous argillite.
242136	FM 45	Grinnell formation	Felix Basin, Nyack quadrangle	2		2	3		Quartzite in metamorphic zone near intrusive mass.
242143	FM 62	do	Two Medicine Pass, Marias Pass quadrangle	2	1		3		Purple-red argillite.
242152	41C-GP	do	McDonald Creek, Going-to-the-Sun Highway, Glacier National Park.	2	2		2		Green argillite.
239379	R-3	do	Rock Creek, southeast part of Bonner quadrangle.	2		3	2		Quartzite near intrusive plug.
239380	R-4	do	Rock Creek, southeast part of Bonner quadrangle.	3	2		2		Same as 239379, farther from plug.
242140	FM 60	do	Head of Park Creek, Marias Pass quadrangle	2		2	3		White quartzite with clay spalls.
239382	SM 9	Appekunny formation(?)	Hamilton quadrangle		1				Border zone of Idaho batholith.
242142	FM 20	Appekunny formation	Wolf Creek, Nyack quadrangle	2	2		3		Argillite.
242151	42 GP	do	McDonald Creek, Glacier National Park, along Going-to-the-Sun Highway near McDonald Falls.	2	1		2		Green argillite high in formation.
242153	44 GP	do	do	2	2		3?		Dark-gray argillite below 242151.
242148a	FM 68	do	Many Glacier Hotel, Glacier National Park	2	2		3		
239378	R 2	do	Rock Creek, southeast part of Bonner quadrangle.	3	2		3		
242144	FM 69	Altyn limestone	Many Glacier Hotel, Glacier National Park			2	3		
	N 3	Chamberlain shale	Jefferson Creek near Neihart	3		1		3	

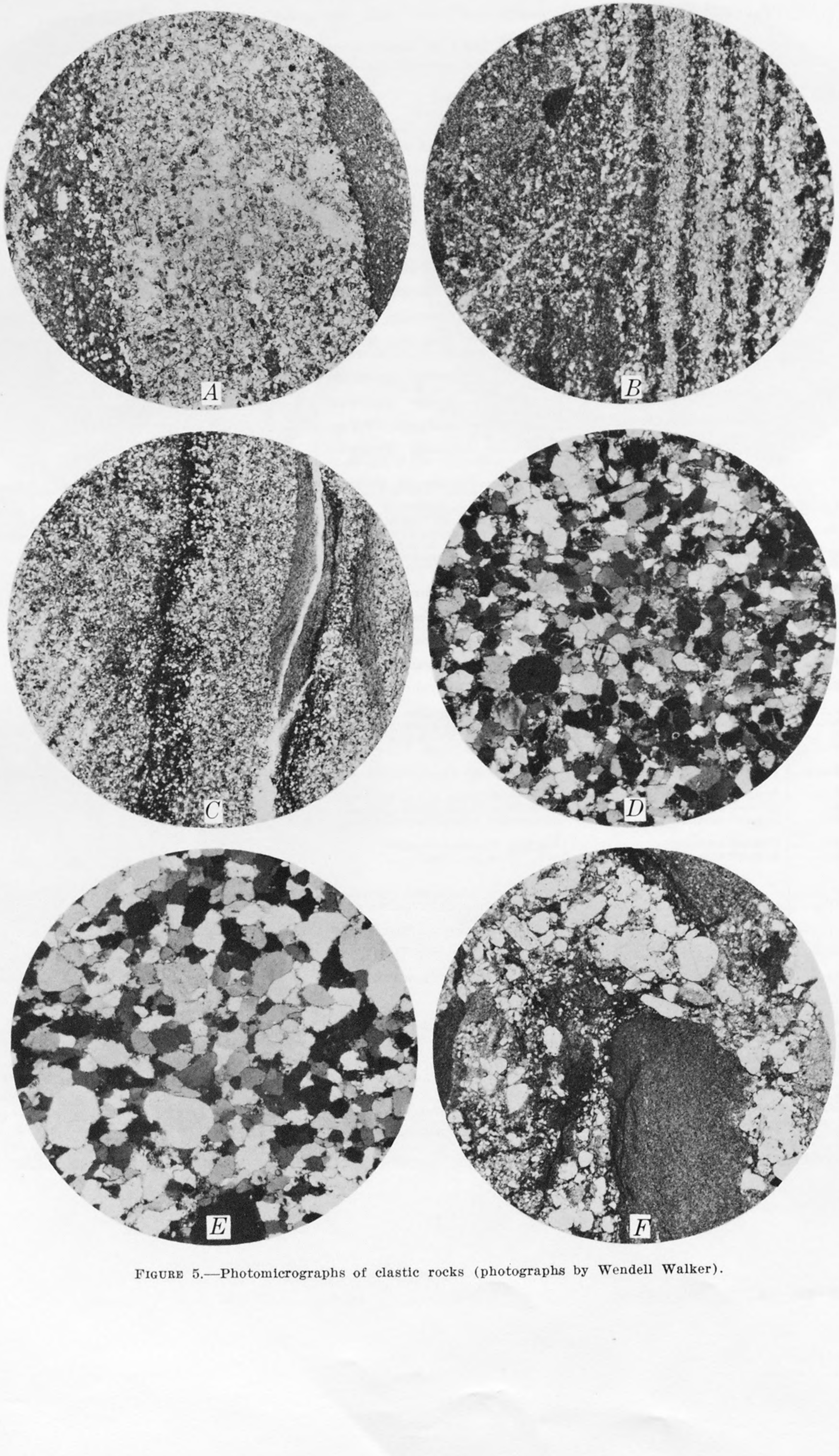

FIGURE 5.—Photomicrographs of clastic rocks (photographs by Wendell Walker).

TABLE 15.—*Analyses of rocks of the Ravalli group in and near Glacier National Park*

[All analyses by Harry Hyman, except for sample 51-2160CDSD by R. N. Eccher].

	51-2160CDSD	51-2161CDSD	ID-21350	ID-21450	ID-13650	ID-14650	Appekunny formation			
							ID-21550	ID-21650 [1]	ID-13950	ID-14450
SiO_2	79.03	67.28	63.25	70.21	70.15	79.51	66.73	65.74	69.82	75.99
Al_2O_3	3.92	14.65	16.55	14.19	15.29	9.98	15.97	14.42	11.96	10.20
Fe_2O_3	.13	1.07	5.08	1.79	3.36	2.44	.90	.71	.84	2.75
FeO	.72	1.98	2.16	2.22	.63	.90	3.70	[2] 3.47	2.25	1.98
MgO	2.44	5.05	2.57	2.58	1.58	.79	2.77	2.33	2.12	1.75
CaO	3.41	.37	.26	.17	.12	.16	.20	2.32	3.22	1.13
Na_2O	1.33	1.39	.78	1.04	1.69	2.05	1.42	1.62	2.03	2.98
K_2O	.78	3.65	5.27	3.98	4.22	2.20	4.01	3.98	2.90	1.50
H_2O-	------	.06	.06	.05	.07	.15	.07	.03	.09	.09
H_2O+	.36	3.25	3.10	2.69	2.00	1.13	3.09	2.36	1.61	.83
TiO_2	.05	.50	.63	.60	.61	.50	.57	.54	.42	.44
CO_2	5.15	.25	.02	.11	.03	.03	.10	1.77	2.39	.02
P_2O_5	.04	.15	.19	.05	.08	.07	.10	.15	.04	.15
SO_3	[3] .81	------	------	------	------	------	------	[4] .52	------	------
MnO	.06	.01	.02	.02	.01	------	.04	.08	.10	.01
BaO	1.35	------	------	------	------	------	------	------	------	------
Total	99.58	99.66	99.94	99.70	99.84	99.91	99.67	100.04	99.79	99.82
O for S	------	------	------	------	------	------	------	.26	------	------
								99.78		

[1] This sample contains organic matter.
[2] Because of the presence of acid-soluble sulfides, results for ferrous iron are not reliable.
[3] Calculated from total S.
[4] Reported as S.

NOTE.—The map in Ross (1959) will aid in fixing the location of samples from Glacier National Park and vicinity. The Rock Creek mentioned is a tributary of Clark Fork southeast of Missoula. This note applies also to most of the samples in the other tables of analyses.

Laboratory	Field	Rock specimen	Source of sample
51-2160CDSD	A-1	Quartzite	Badrock Canyon, T. 30 N., R. 20 E.
2161CDSD	A-2	Argillite	Do.
ID-21350	41B-GP	Purple argillite	Going-to-the-Sun Highway above McDonald Falls, Glacier National Park.
21450	41-GP	Green argillite	Do.
13650	FM 23	Argillite	Noisy Creek, Nyack quadrangle, T. 28 N., R. 19 W.
14650	4 of 1949	Argillite	Rock Creek in unsurveyed T. 10 N., R. 16 E. Two miles up Rock Creek from the mouth of Ranch Creek in the 30-minute Bonner quadrangle, southeast of Missoula.
21550	42-GP	Green argillite	Upper part of the formation, Going-to-the-Sun Highway above McDonald Falls, Glacier National Park.
21650	44-GP	Dark quartzitic argillite	Lower part of formation, Going-to-the-Sun Highway above McDonald Falls, Glacier National Park.
13950	FM 68	Argillite	Near Many Glacier Hotel, Glacier National Park.
14450	2 of 1949	Argillite	Rock Creek above the mouth of Ranch Creek, unsurveyed T. 10 N., R. 16 E., in the 30-minute Bonner quadrangle southeast of Missoula.

EXPLANATION OF FIGURE 5

A. Specimen N 2. Neihart quartzite. This rock is nearly all quartz with some micaceous aggregates that may be clay minerals. Some of the well-rounded grains are nearly a millimeter long. Crushed quartz grains are tightly fitted together and some have accretion rims, not visible in the picture.

B. Specimen N 3. Chamberlain shale. This is an irregularly laminated rock with subangular to angular quartz grains as much as 0.05 mm in diameter in a mat of quartz and micaceous flakes, possibly including some clay, and much brown, opaque coloring matter. The largest grains in the mat are 0.05 mm in diameter.

C. Specimen FM 68. Appekunny formation. This rock has distinct laminae 0.5 to 1.0 mm wide, consisting mostly of quartz, with about 10 percent plagioclase. The grains are as much as 0.015 mm in diameter. Small amounts of pyrite, chlorite and muscovite are also present.

D. Specimen FM 16. Grinnell formation. The main part of the rock consists of quartz, some altered plagioclase, and interstitial carbonate. Some of the grains are well rounded. Diameters are as much as 0.5 mm. The included fine-grained slabs show attrition along their borders and have contributed streaks of fine material to the main mass. These slabs are colored with disseminated specularite grains.

E. Specimen FM 9. Missoula group. A fairly well sorted and rounded clastic rock with component grains as much as 0.5 mm in diameter. Grains are mostly quartz, but some are feldspar and a little sericite and carbonate is present.

F. Specimen MM 6. Spokane shale. This is a fine-grained, indistinctly laminated shale composed largely of quartz and micaceous material. Quartz grains are as much as 0.1 mm in diameter. Specularite is abundantly diffused in the groundmass. The irregular texture of the rock suggests that the original mud was disturbed while still soft.

TABLE 16.—*Analyses of noncarbonate rocks low in the Belt series*

[Analysts: No. C–405, J. G. Fairchild *in* Wells (1937, p. 28); others, Harry Hyman]

	C–405	51-2163CDSD	51-2164CDSD
SiO_2	69.06	96.38	78.29
Al_2O_3	15.88	1.61	10.55
Fe_2O_3	1.21	.57	1.56
FeO	3.17	.18	2.52
MgO	1.09	.10	.94
CaO	.00	.00	.18
Na_2O	1.62	.00	.18
K_2O	3.96	.46	2.19
H_2O-	.00	.02	.43
H_2O+	1.98	.37	2.59
TiO_2	.75	.10	.40
CO_2	.00	.01	.01
P_2O_5	Trace	.02	.07
SO_3	-----	-----	-----
MnO	Trace	Trace	.01
BaO	.10	-----	-----
ZrO	.03	-----	-----
Total	[1] 98.85	99.82	99.92

[1] Sample contains organic carbon, probably graphitic.

NOTE.—Source and description of samples.

Laboratory	Field	Kind of rock and source
C–405		Slightly metamorphosed sandy argillite of the Prichard formation, Pend Oreille Lake, Idaho.
51-2163CDSD	N 2	Neihart quartzite, Jefferson Creek near Neihart, Mont., Northeast part of unsurveyed T. 13 N., R. 8 E.
51-2164CDSD	N 3	Chamberlain shale, Jefferson Creek near Neihart, Mont., Northeast part of unsurveyed T. 13 N., R. 8 E.

TABLE 17.—*Analyses of noncarbonate rocks of the Missoula and North Boulder groups*

[Analysts: Harry Hyman and R. N. Eccher]

	1D–14150	1D–14250	1D–14350	1D–14950	1D–15050	1D–15150	1D–15250	1D–15350	1D–14550
SiO_2	58.74	73.16	66.94	74.12	88.60	71.66	66.45	53.57	56.50
Al_2O_3	13.28	11.05	14.65	9.49	5.53	4.50	16.05	11.98	12.29
Fe_2O_3	3.37	2.81	3.92	1.95	.40	.78	1.02	1.35	1.04
FeO	.82	1.31	1.44	1.35	.52	.76	2.54	2.67	7.69
MgO	5.56	1.37	2.94	2.16	.40	4.44	3.32	6.43	4.99
CaO	3.93	1.83	.34	2.06	.03	5.73	.28	6.47	7.51
Na_2O	.72	1.67	1.35	2.23	.02	.54	1.42	.61	2.94
K_2O	3.77	2.62	5.00	2.15	3.57	1.65	4.65	3.99	1.08
H_2O-	.44	.15	.08	.09	.04	.02	.17	.10	.12
H_2O+	2.44	1.43	2.08	1.17	.40	.67	3.15	2.54	2.22
TiO_2	.47	.64	.60	.37	.06	.08	.54	.50	1.06
CO_2	5.69	1.18	.03	2.54	.03	8.63	.11	9.37	2.30
P_2O_5	.11	.13	.14	.09	.01	.09	.08	.12	.11
MnO	.13	.09	.04	.06	Tr.	.12	.01	.04	.20
Total	99.47	99.44	99.55	99.83	99.61	99.67	99.79	99.74	100.05

NOTE.—The map in Ross (1959) will aid in fixing the location of samples from Glacier National Park and vicinity. The Rock Creek mentioned is a tributary of Clark Fork southeast of Missoula. This note applies also to most of the samples in the other tables of analyses.

Laboratory	Field	Kind of rock and source
1D–14150	6–47	Argillite of the Spokane shale, Spokane Hills, east of Helena.
14250	9–47	"Miller Peak" of Clapp and Deiss (1931). Prickley Pear Creek, north of Helena.
14350	19–47	Argillite of Missoula group southwest of Missoula, T. 12 N., R. 24 W.
14950	Belt 1	Quartzite argillite in Missoula group, Rock Creek opposite mouth of Gilbert Creek in unsurveyed T. 10 N., R. 16 E., in the 30-minute Bonner quadrangle southeast of Missoula.
15050	5	Pink quartzite, Missoula group, near Bonner in Blackfoot Canyon, Bonner quadrangle.
15150	7	Red quartzite, Missoula group, Middle Fork Flathead River, Nyack quadrangle.
15250	8	Green argillite, Missoula group, Middle Fork Flathead River, Nyack quadrangle.
15350	9	Green quartzite at base of Missoula, Middle Fork Flathead River, Nyack quadrangle.
14550	7–49	Grit in North Boulder group, Jefferson Canyon near La Hood.

TABLE 18.—*Analyses of rocks of the Missoula group from near Superior, Mont.*

[Data furnished by A. B. Campbell. Analyzed by rapid methods by Harry F. Phillips, Paul L. D. Elmore, and Katrine E. White]

	U.S. Geological Survey Laboratory No.			
	140865	140867	140868	140869
SiO_2	72.8	63.2	65.1	68.2
Al_2O_3	14.2	15.8	17.2	14.2
Fe_2O_3	1.9	3.9	1.3	1.4
FeO	.38	2.3	2.5	5.9
MgO	2.2	3.2	2.6	1.5
CaO	.31	.48	.26	.28
Na_2O	1.6	2.2	3.5	2.2
K_2O	4.5	4.8	4.4	2.1
TiO_2	.22	.56	.72	.55
P_2O_5	.11	.14	.11	.11
MnO	.02	.04	.01	.04
H_2O	2.3	3.2	2.9	3.3
CO_2	.08	$<.05$	$<.05$	$<.05$
Total	101	100	101	100

Laboratory	Field	Rock specimen	Source of sample
140865	12054	Siltstone	Mouth of Trout Creek, sec. 14, T. 16 N., R. 26 W.
140867	38054	do	Trout Creek, SE¼ sec. 14, T. 16 N., R. 26 W.
140868	102054	do	Cedar Creek, sec. 1, T. 16 N., R. 27 W.
140869	108054	Micaceous quartzite	Cedar Creek, sec. 6, T. 16 N., R. 28 W.

TABLE 19.—*Analyses of rocks of the Spokane and Greyson shales from the southern Elkhorn Mountains, Mont.*

[Data furnished by M. R. Klepper. Analyzed by Paul L. D. Elmore and Paul W. Scott by rapid methods.]

	53-102SC	53-103SC	53-104SC	53-105SC	53-106SC	53-107SC	53-108SC	53-109SC	53-110SC
Acid insoluble	97.6	11.6	71.0	71.4	76.5	65.4	62.0	98.4	50.0
CaO	.16	48.1	12.2	13.6	10.4	17.0	20.0	.30	27.2
MgO	.09	.57	1.6	.85	1.0	.82	.51	.03	.17

Laboratory	Field	Location		Description of sample
		Section	Township and range	
53-102SC	48-K8	NW¼ sec. 27	T. 6 N., R. 1 E	Red argillite, Spokane shale, 20 ft below unconformity.
103SC	49-Pcb	SE¼ sec. 13	T. 4 N., R. 2 W	Limestone, Spokane shale, 400–500 ft below unconformity.
104SC	960b	NE¼ sec. 26	do	Reddish-brown argillite, 270 ft above base of Spokane.
105SC	960b	do	do	Reddish-brown argillite, same as that 2 ft below 49-960.
106SC	960c	do	do	Quartzite, same as that 235 ft above base of formation.
107SC	960d	do	do	Quartzite, same as that 8 ft below 49-960c.
108SC	960e	do	do	Quartzite, same as that 13 ft below 49-960c.
109SC	A 12	Sec. 22	do	Tan quartzite, Greyson shale within 100 ft of top.
110SC	A 29		do	Oolitic limestone. Gradational beds at base of Spokane shale.

TABLE 20.—*Analyses of carbonate rocks in the Belt series*

[Analysts: Harry Hyman, R. N. Eccher, and W. T. Schaller]

	51-2159CDSD	51-2162CDSD	1D-13550 [1]	1D-13750 [1]	1D-13850	1D-14750		1D-13950	1D-14550	(1)
SiO_2	33.30	50.59	23.73	42.16	7.53	67.66	48.81	18.85	29.34	44.80
Al_2O_3	4.47	6.82	4.25	6.17	1.23	10.42	7.51	1.90	5.98	8.96
Fe_2O_3	.22	.56	.45	.37	.25	.40	.24	.12	.35	2.85 (FeO and Fe_2O_3)
FeO	1.33	3.29	.99	1.72	1.58	1.85	1.65	.62	1.76	
MgO	6.57	10.41	2.73	3.68	17.36	2.43	3.73	16.36	3.51	5.90
CaO	25.59	9.38	36.31	22.81	28.85	5.79	17.55	23.79	29.68	16.42
Na_2O	.30	.03	.32	.66	.08	1.47	1.13	.18	.73	.43
K_2O	1.17	.74	.94	1.39	.36	3.36	1.91	1.06	1.56	2.14
H_2O-	.10	.38	.14	.16	.05	.10	.04	.02	.08	3.33 (H_2O- and H_2O+)
H_2O+	1.04	3.08	1.19	1.32	.22	1.59	1.49	.01	1.12	
TiO_2	.17	.25	.14	.22	.05	.39	.33	.06	.26	.37
CO_2	25.38	13.85	28.61	18.94	41.97	4.36	15.42	36.65	25.10	15.42
P_2O_5	.08	.08	.07	.05	.03	.08	.07	.12	.08	-------
SO_3	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------
MnO	.05	.10	.03	.04	.27	.02	.03	.05	.06	-------
BaO	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------
Total	99.77	99.56	99.90	99.69	99.83	99.92	99.91	99.79	99.61	100.62

[1] Carbonaceous matter in sample.

NOTE.—See explanation for table on page 48.

TABLE 20.—*Analyses of carbonate rocks in the Belt series*—Continued

Laboratory	Field	Kind of rock and source
51-2159CDSD	M 2	Limestone of Missoula group, U.S. Highway 2 at Devil Creek, Glacier National Park.
2162CDSD	N 1	Newland limestone, Jefferson Creek near Neihart.
1D-13550	FM 21	Siyeh limestone in unsurveyed sec. 30, T. 28 N., R. 18 W., Noisy Creek, Nyack quadrangle.
13750	FM 39	Siyeh limestone, Emery Creek, Nyack quadrangle.
13850	FM 67	Dolomite in Shepard formation, NW¼ sec. 22, T. 35 N., R. 17 W., Glacier National Park.
14750	5-49	Limestone of Missoula group, Rock Creek below Babcock Gulch in the 30-minute Bonner quadrangle.
	10-49	Siyeh limestone, Middle Fork Flathead River, Nyack quadrangle.
1D-13950	FM 69	Altyn limestone, near Many Glacier Hotel, Glacier National Park.
14550	3-49	Newland limestone with segregation structure, Rock Creek, just downstream from an igneous mass and about 8 miles above Ranch Creek, 30-minute Bonner quadrangle.
		Newland limestone, gorge below Georgetown Lake, Philipsburg quadrangle.

TABLE 21.—*Analyses of rocks from in and east of the Coeur d'Alene mining region, Shoshone County, Idaho, and Mineral County, Mont.*

[Data furnished by S. W. Hobbs. Analyses by rapid methods by Harry F. Phillips, Paul L. D. Elmore, and Katrine E. White]

	54-416SC	54-417SC	54-418SC	54-419SC	54-420SC	54-423SC	54-424SC	54-425SC	54-426SC	54-427SC	54-428SC	54-429SC
SiO_2	70.3	74.7	70.4	35.4	66.0	63.3	47.1	62.9	80.8	71.9	63.6	81.7
Al_2O_3	15.3	13.6	17.6	5.2	17.1	24.4	30.4	17.6	10.4	13.5	12.4	9.4
FeO	.62	.60	.27	2.9	1.2	.08	.78	3.8	.76	.71	2.2	.27
Fe_2O_3	3.6	2.8	2.2	.4	1.8	1.3	4.2	1.7	1.6	3.0	1.6	1.9
MgO	.71	1.3	.26	11.6	3.9	.14	2.7	2 9	.86	2.2	3.4	1.0
CaO	.52	.18	.17	16.0	.22	.20	.17	.79	.21	.34	5.8	.3
Na_2O	1.4	2.8	1.2	.80	.70	.27	.16	1.6	2.9	1.8	2.3	3.1
K_2O	4.1	3.1	4.9	1.2	5.6	6.7	10.6	4.3	1.5	4.0	1.9	2.0
TiO_2	.59	.48	.64	.14	.67	.40	.36	.69	.28	.50	.40	.34
P_2O_5	.20	.07	.06	.07	.10	.14	.14	.16	.06	.14	.13	.11
MnO	.01	.01	.01	.16	.01	---	.01	.04	.01	.02	.19	.01
CO_2	.81	<.05	<.05	24.0	<.05	<.05	<.05	.84	<.05	<.05	4.4	<.05
H_2O	1.5	1.2	2.2	.53	3.1	3.0	3.7	2.8	1.0	2.0	2.2	.51
Total	100	101	100	98	100	100	100	100	100	100	101	101

Laboratory	Field	Formation	Kind of rock	Source of sample
54-416SC	SWH-1-54	St. Regis		Portal of Rock Creek adit, Idaho.
417SC	65-48	Burke		Glidden Lake, Idaho.
418SC	ABG-121-48	Upper part of Prichard		Near Burke, Idaho.
419SC	16-48	Wallace	Argillaceous limestone	
420SC	29-48	do	Argillite	Placer Creek road, Idaho.
423SC	SWH-2-54	Prichard	do	Near first sawmill on Pine Creek, Idaho.
424SC	14-JH-52	St. Regis		1.5 miles south of Saltese on Silver Creek road, Montana.
425SC	15-JH-52	do		Do.
426SC	8-JH-52	do		St. Regis road, near Lookout Pass, Mont.
427SC	9-JH-52	do		Lookout Pass, Mont.
428SC	37-JH-52	do		7.1 miles east of Henderson on U.S. Highway 10.
429SC	13-JH-53	do		Little Baldy Mountain near Murray, Idaho.

TABLE 22.—*Average compositions of various sedimentary rocks*

	Shale [1]	Sandstone [1]	Limestone [1]	Slate [2]	Graywacke [3]	Subgraywacke [3]	Precambrian lutite [4]
SiO_2	58.10	78.33	5.19	60.64	68.1	77.8	58.32
Al_2O_3	15.40	4.77	.81	[5] 18.05	15.4	9.5	18.44
Fe_2O_3	4.02	1.07	.54	2.25	[6] 3.4	.9	3.75
FeO	2.45	.30	---	3.66	3.4	2.6	4.13
FeS_2	---	---	---	.38	---	---	.54
MgO	2.44	1.16	7.89	2.60	1.8	1.6	2.21
CaO	3.11	5.50	42.57	1.54	2.3	1.2	.78
Na_2O	1.30	.45	.05	1.19	2.6	2.0	.96
K_2O	3.24	1.31	.33	3.69	2.2	1.5	4.42
H_2O	5.00	1.63	.77	4.13	2.1	1.7	3.86
TiO_2	.65	.25	.06	---	.7	.6	.78
S	---	---	.09	---	---	.1	.03
P_2O_5	.17	.08	.04	---	.2	.2	.13
CO_2	2.63	5.03	41.54	1.47	---	.5	.74
Cl	---	---	.02	---	---	---	---
SO_3	.64	.07	.05	---	---	---	.14
MnO	---	---	.05	---	.2	.2	.08
BaO	.05	.05	---	---	---	---	---
C	.80	---	---	---	---	.2	.69
Total	100.00	100.00	100.00	99.60	102.4	100.4	100.00

[1] From Clarke (1924, p. 34).
[2] From Eckel (1904).
[3] From Pettijohn (1949, p. 250, 256).
[4] From Nanz (1953, p. 57, 58, 62).
[5] Includes TiO_2 and P_2O_5.
[6] Pettijohn (1949) notes this figure probably should be 1.4. This would make the total 100.4.

To facilitate comparison among the anaylses the data are summarized in triangular diagrams in figure 6. The form of the diagrams is adopted from one developed by W. W. Rubey and his coworkers. In turn their diagram is a modification of one by Mason (1952, p. 131). Most of the calculations necessary to permit plotting the analytical data on the diagrams were made by Thelma Hill, an assistant of Rubey's. These diagrams serve to show at a glance the relative proportions of free silica, carbonate, and those other constitutents that may be broadly classified as argillaceous. In each of the diagrams the apex that corresponds to the argillaceous constituents is labeled with the formula of a theoretical clay mineral that has been found convenient for the purpose of the diagram. Actually, as noted above, the amount of clay minerals now present in the Belt series is small. Their argillaceous components, in the sense here intended, are at the present time in the form of mica, chlorite, feldspar, and other minerals that under suitable circumstances could break down to form clays.

The analyses given include several representatives of quartzites. One of them, from the Neihart quartzite, is exceptionally pure. If the sampling had been strictly in proportion to the quantity of each lithologic type the number of quartzitic rocks sampled would have been larger. Beds of apparently pure quartzite did not seem to offer special interest from a chemical standpoint and many were not sampled. Some rocks whose general appearance and resistance to weathering led to calling them quartzite in the field proved to contain so little silica that they cannot strictly be given that name. Quartzose argillite might be a more appropriate designation.

Most of the clastic rocks sufficiently impure and fine grained so that the terms quartzite and sandstone are obviously inappropriate have been called argillite or shale by most writers who wrote about the Belt series. So far as the analyses here presented reveal, the rocks commonly called argillite and shale are more siliceous and have less alumina and iron oxides, especially the latter, than the slate and shale grouped in published tables of averages (Pettijohn, 1949, p. 82, 271, 285; Nanz, 1951, 1953, p. 57, 58). Similar tables for argillite are not given but, as this term is applied to shale and similar rock that has been only moderately metamorphosed, the range in composition of argillite would be within that in the tables just cited, excerpts from which are given in table 22.

The argillaceous rocks of the Belt series have the chemical resemblance to graywackes that would be inferred from the mineralogic resemblances already commented upon. These Belt rocks are rather high in magnesia and ferrous oxide which, according to Pettijohn (1949, p. 250, 251, 259, 271), is characteristic of graywacke, but they contain more potassium oxide than sodium oxide, which he notes is true of arkose and shale but not graywacke. The few available analyses from clastic rocks of pre-Ravalli age show less carbonate than those from higher horizons. The differences in proportion of ferric and ferrous oxides accord fairly well with differences in the colors of the rocks, assuming that reddish colors correspond to ferric iron and greenish colors to ferrous iron. The amount of iron tends to decrease with increase in silica.

The fact that most of the argillaceous rocks analyzed are similar chemically points to the probability that the rocks from which the original sediments were derived were similar throughout. It follows that the 12 major chemical constituents can be expected to give little aid in stratigraphic correlations.

Analyses of 10 carbonate-rich rocks are given in table 20 and are plotted on figure 6*A*. Two short tables (tables 23 and 24) contain partial analyses. These data agree with remarks made above in showing that carbonate-rich rocks are widespread in the Belt series in Montana. They vary in composition from magnesian limestone to dolomite and nearly all contain much noncarbonate material. Thus the chemical features are not distinctive enough to be of value in stratigraphic correlation, except perhaps locally.

All the rocks represented in table 20 contain more than enough calcium and magnesium oxides to balance their carbon dioxide content. The excess is in silicate minerals, mainly plagioclase and chlorite. On weathered surfaces the rocks are brownish and yellowish enough to suggest the presence of iron carbonate, but the analyses show that the amount must be very small. This contrasts with reports (Ransome and Calkins, 1908, p. 41–42; Hobbs and others, 1950) that siderite and ankerite are conspicuous in parts of the Belt series in Idaho. Part of the siderite in Idaho is reported to be of secondary origin (Ransome and Calkins, 1908, p. 97), which may account for its scarcity in parts of Montana where opportunity for hydrothermal alteration is less than in the mining districts of northern Idaho.

The analyses provide arguments for changing the name of the Altyn limestone to Altyn dolomite, but such a change might be more confusing than helpful at present. Similarly, if the Shepard formation were being named today, it might well be called Shepard dolomite. However, the unit contains enough argillaceous and siliceous material to give some justification for the term Shepard formation now in established use. It is not dominantly a quartzite as the original description would suggest (Willis, 1902, p. 316, 324).

The validity of the term "Helena limestone" has been questioned on the ground that most of the formation appears to be dolomite (Knopf, 1950, p. 837). This statement is based on the presence locally of magnesium-rich metamorphic minerals and on field measurements of density. Both because of long-established usage and because significant portions of the unit have the composition of limestone it seems well to retain the designation "Helena limestone." The partial analyses of selected samples given in table 23 and field observations by the writer, indicate that some beds are entirely limestone and the beds consisting of irregular mixtures of buff and gray material that are characteristic of the formation include much that has the composition of limestone. Though the buff-weathering material in these beds is mostly dolomitic, some contains an excess of calcium carbonate. Some geologists are inclined to

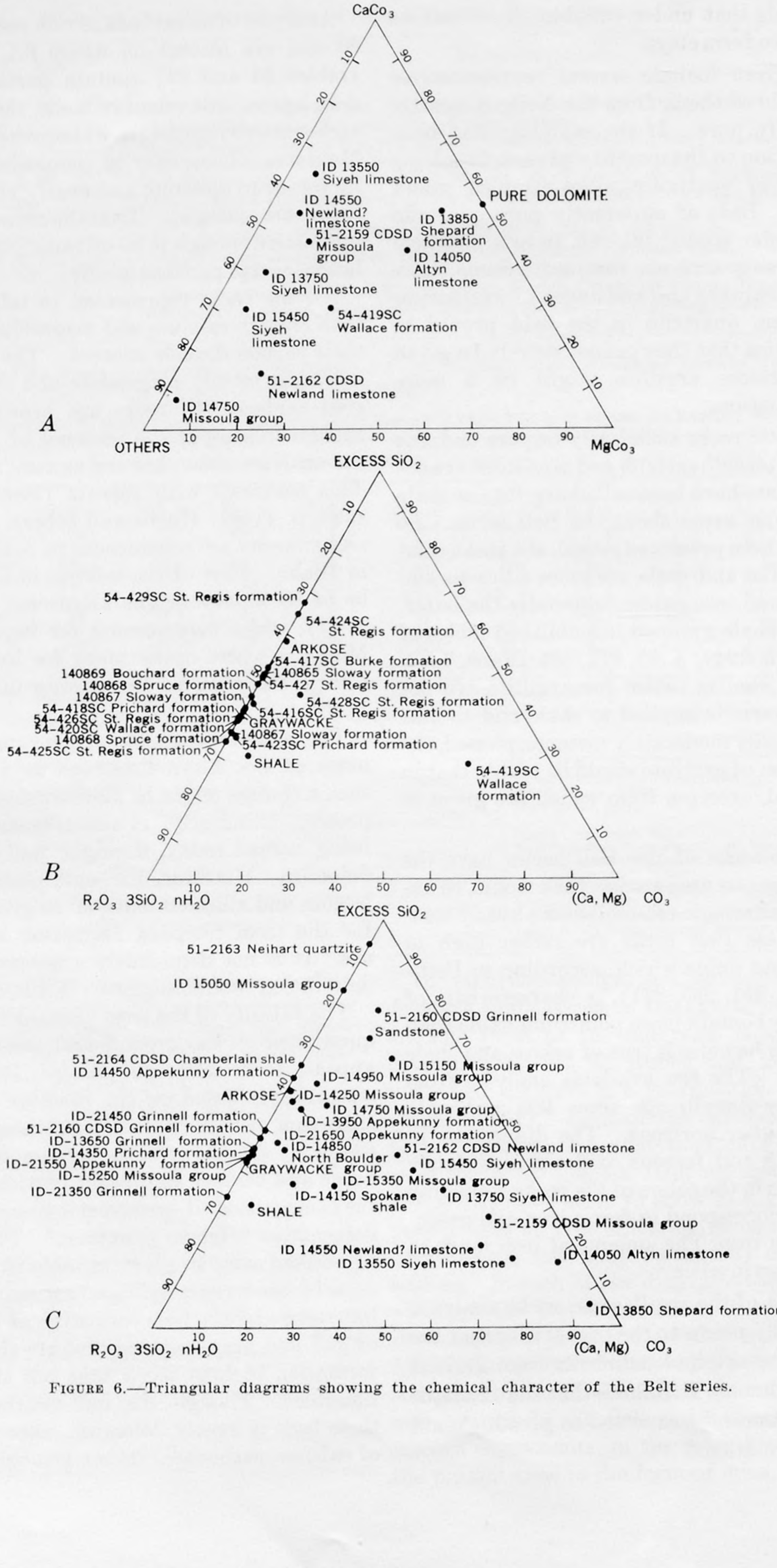

FIGURE 6.—Triangular diagrams showing the chemical character of the Belt series.

extend the term "Helena limestone" to carbonate-rich beds at a distance from Helena (Clapp and Deiss, 1931). On the basis of present knowledge, this is inadvisable. Exact equivalence is unproved and now seems improbable but broad stratigraphic similarity is clear. Samples M2 and 5 in table 20 represent beds far from Helena but for which correlation with the Helena limestone has been suggested. In both, the magnesian carbonate is less than in the dominantly dolomitic Helena limestone of the type locality. This fact suggests, inconclusively, a lack of stratigraphic identity with the Helena limestone.

In order to obtain chemical data regarding molar-tooth structures and stromatolites, samples were drilled from areas of different characteristics in such material, with the results shown in table 24. These analyses show that all samples are limestone rather than dolomite. Those partings that weather in relief are relatively high in insoluble material, mostly quartz, with small amounts of feldspar, mica, and traces of clay minerals, according to F. A. Hildebrand (written communication). These portions tend to weather a rusty color, but the form in which the iron occurs has not been established.

TABLE 23.—*Partial analyses of the Helena limestone near Helena, Mont.*

[Data furnished by Richard Rezak. Analysts: Harry F. Phillips, Joseph M. Dowd, and Katrine E. White. Results are given in percent]

	52-1875	52-1876	52-1877	52-1878	52-1879	52-1880	52-1881	52-1882	52-1883	52-1884
MgO	10.0	8.8	1.4	11.8	3.4	11.7	1.0	1.2	1.8	2.0
CaO	17.0	11.3	47.7	17.2	36.2	17.1	43.2	41.5	38.6	41.2
Insoluble	40.2	53.1	10.7	34.5	24.0	33.2	21.0	22.5	24.2	18.4
Loss on ignition	23.4	17.8	------	25.5	------	23.4	34.1	33.1	31.2	33.5
Total	90.6	91.0	------	89.0	------	85.4	99.3	98.3	95.8	95.1

Laboratory	Field	Kind of rock and source
52-1875	BhD1	Bed 31 in. thick, 17 ft above base of section.
1876	BhD2	Same bed as BhD1 50 ft away.
1877	BhC1a	Limestone curleycues in a dolomite bed 19 ft above base of section.
1878	BhC1b	Dolomite matrix around BhC1a.
1879	BhC2a	Limestone curleycues in same bed as BhC1a but 50 ft away.
1880	BhC2b	Dolomite matrix of BhC1a.
1881	BhA1	Algal limestone in 96-in. zone of alternate limestone and dolomite, 16 ft above base of section.
1882	BhA2	Same as BhA1 but 50 ft away.
1883	BhL1	Limestone bed, 8 in. thick, 48 ft above base of section.
1884	BhL2	Same as BhL1 but 50 ft away.

TABLE 24.—*Partial analyses of molar-tooth structures and stromatolite heads*

[Analyst, Edwin Goldenthal. Results are given in percent]

	51-2171	51-2172	51-2173	51-2174	51-2176	51-2178	51-2179	51-2175	51-2177	51-2180	51-2181
CaO	13.4	40.2	34.3	3.9	37.2	31.6	29.8	49.0	29.2	48.4	23.6
MgO	3.6	1.6	3.4	4.2	2.0	2.8	3.4	.62	1.2	.71	1.6
Insoluble residue	62.1	20.8	27.1	76.7	26.5	31.6	32.8	10.3	40.9	10.6	45.8

Laboratory	Field	Kind of rock and source
51-2171-2173	FM 342	Siyeh limestone, Going-to-the-Sun Highway, 2 miles north of Logan Pass, Glacier National Park. 2171. Rusty areas in limestone. 2172. Black filaments in limestone. 2173. Light-gray areas in limestone.
2174, 51-2176	M-1	Limestone of Missoula group, U.S. Highway 2 near the mouth of Devil Creek south of Glacier National Park. 2174. Rusty material that weathers in relief in limestone. 2176. Dark material that weathers in depressions in limestone.
2178-2179	FM 341	Siyeh limestone near Field No. FM 342. 2178. Thin layers that weather black in argillaceous limestone. 2179. Broad layers that weather gray.
2175, 51-2177	FM 31	Stromatolite in Siyeh limestone, Tongue Mountain, T. 28 N., R. 18 W. 2175. Gray, less-resistant layers. 2177. Hard rusty laminae.
2180-2181	FM 32	Stromatolite in Siyeh limestone, Tongue Mountain, T. 28 N., R. 18 W. 2180. Gray layers in stromatolite. 2181. Rusty layers.

EXPLANATION OF FIGURE 6

A. Carbonate rocks of the Belt series in terms of calcium carbonate, magnesium carbonate, and all other constituents grouped together.

B. Components of the Belt series in northern Idaho and the adjacent part of Montana in terms of free silica, the carbonate component, and a theoretical clay molecule.

C. Components of the Belt series in western Montana in terms of the free silica, the carbonate component, and a theoretical clay molecule.

TRACE ELEMENTS

In the hope that trace elements might furnish clues to stratigraphic correlation, samples of Belt rocks obtained in the course of fieldwork in Montana have been analyzed spectrographically and similar analyses from other areas have been assembled. The analyses are listed in tables 25 to 32. The data in them are not rigidly comparable because the elements sought and the analytical procedures used were not identical throughout. Data on the general distribution of trace elements in ordinary sedimentary rocks have been compiled by Rankama and Sahama (1950, table 5.52, p. 226) and by Krauskopf (1955, p. 416).

None of the units of the Belt series has been sampled specifically to learn the distribution of trace elements within it. Hence the analyses do not lend themselves well to statistical tests. Analysis of many more carefully selected samples would be required for any final conclusions. W. C. Krumbein (W. G. Schlecht, written communication, July 8, 1955) in consultation with statisticians of the U.S. Bureau of Standards concluded that the distribution of trace elements in the analyses in tables 25, 28, and 29 may be random.

TABLE 25.—*Quantitative spectrographic analyses, in percent, of carbonate rocks in the Belt series*

[Analyst: A. A. Chodos]

	51-2159SDCD	51-216SDCD	50-587CS	50-859CS	50-860CS	50-876CS	50-862CS	50-867CS	50-869CS
Ni	0.0006	0.002	0.0002	0.0002	--------	0.0002	--------	0.0002	0.001
Co	.0005	.001	--------	--------	--------	--------	--------	--------	.001
Cr	.002	.004	.003	.003	0.0007	.003	0.0007	.003	.005
Cu	.001	.003	.001	.0006	.006	.0005	.0007	.002	.003
V	.002	.004	.006	.003	.005	.004	--------	.007	.005
Ga	.0004	.0006	--------	--------	--------	--------	--------	--------	.001
Sc	.001	.001	--------	--------	--------	--------	--------	--------	.001
Pb	--------	--------	--------	--------	.006	--------	--------	--------	--------
Ba	.06	.02	.02	.02	.008	.06	.1	.02	.05
Sr	.05	.03	.02	.01	.01	.01	.01	.02	.006
Be	--------	--------	--------	--------	--------	--------	--------	--------	.0001
Mo	--------	--------	--------	--------	--------	--------	--------	--------	--------
Y	.001	.004	--------	--------	--------	--------	--------	--------	--------
Yb	.0001	.0002	--------	--------	--------	--------	--------	--------	--------
La	--------	--------	--------	--------	--------	--------	--------	--------	.02
Zr	.004	.008	.003	.003	.002	.03	.005	.006	.03
B	.004	.001	.002	.004	.002	.004	.002	.004	.01

Laboratory	Field	Kind of rock and source
51-2159SDCD	M 2	Limestone of Missoula group, U.S. Highway 2 at Devil Creek, Glacier National Park.
2162SDCD	N 1	Newland, Chamberlain Creek, south of Neihart.
50-587CS	FM 21	Siyeh, unsurveyed sec. 30, T. 28 N., R. 18 W., Noisy Creek, Nyack quadrangle.
859CS	FM 39	Siyeh or Missoula, Emergy Creek, Nyack quadrangle.
860CS	FM 67	Shepard, Flattop Mountain, NW¼ sec. 22, T. 35 N., R. 17 W., Glacier National Park.
876CS	10-49	Siyeh, Middle Fork along Flathead River, near Walton, Nyack quadrangle.
862CS	FM 69	Altyn, Many Glacier Hotel, Glacier National Park.
867CS	3-49	Newland, Rock Creek, just downstream from igneous mass about 8 miles above Ranch Creek, 30-minute Bonner quadrangle.
869CS	5-49	Limestone of Missoula group, Rock Creek, below Babcock Gulch in 30-minute Bonner quadrangle.

TABLE 26.—*Quantitative spectrographic analyses, in percent, of limestones from Glacier Park, drilled selectively from samples showing molar-tooth and similar structures*

[The following elements were looked for but not found: Co, Sc, Mo, Y, Yb, La]

	1	2	3	4	5	6	7	8
Ni	0.002	0.004	0.003	0.002	0.002	0.003	0.003	0.004
Cr	.002	.0006	.001	.002	.002	.0005	.002	.0005
Cu	.002	.006	.002	.004	.004	.01	.01	.02
V	.002	.002	.003	.003	.002	.002	.002	.002
Ga	.0007	0	.001	.001	.001	0	0	0
Pb	0	0	0	0	.003	0	0	0
Ba	.02	.03	.02	.02	.03	.02	.01	.008
Sr	.07	.1	.1	.09	.07	.1	.1	.1
Be	.0002	0	0	.0002	.0002	0	.0002	0
Zr	.02	.004	.005	.02	.006	.002	.01	.003
B	.01	.004	.01	.03	.009			
Ti	.06	.01	.02	.06	.04	.007	.02	.007
Ag	.0005	0	0	.0005	.0003	0	.0007	.001
Cd	0	.01	0	0	0	0	.01	.20
Mn	.05	.02	.02	.02	.08	.02	.04	.02

1. FM 342. Rusty areas in Siyeh limestone with molar-tooth structure near Glacier National Park.
2. FM 342. Black filaments in same sample.
3. FM 342. Gray areas in same sample.
4. FM 341. Siyeh limestone. Broad gray bands.
5. M-1. Hard rusty laminae in molar-tooth structure or similar structure in limestone in Missoula group from Devil Creek south of Glacier National Park.
6. M-1. Dark depressed areas in same sample.
7. FM 31. Siyeh limestone from Tongue Mountain. Hard rusty laminae in stromatolite head.
8. FM 31. Gray less resistant bands in stromatolite head.

TABLE 27.—*Semiquantitative spectrographic analyses, in percent, of selected specimens from the Helena limestone along the Helena Scenic Drive about 2 miles east of its intersection with U.S. Highway 10 N, Helena, Mont.*

[Analyst: Richard S. Harner]

	52-1875 CWSW	52-1876 CWSW	52-1877 CWSW	52-1878 CWSW	52-1879 CWSW
Mn	0.0x0	0.0x0	0.0x0	0.0x0	0.0x0
Ni	0	0	0	0	0
Fe	x	x	x	x	x
Al	x	x	x	x	x
Cr	.000x	.000x	.000x	.000x	.000x
V	0	0	.00x	.00x	.00x
Ti	.0x	.0x	.0x	.0x	.0x
Zr	.000x	.000x	0	.000x	.000x
Sr	.000x	.000x	.00x	.000x	.00x
Ba	.00x	.00x	.00x	.00x	.00x
B	.00x	.00x	.00x	.00x	.00x

	52-1880 CWSW	52-1881 CWSW	52-1882 CWSW	52-1883 CWSW	52-1884 CWSW
Mn	0.0x0	0.0x0	0.0x0	0.0x0	0.0x0
Ni	0	0	0	.000x	0
Fe	x	.x	x	x	.x
Al	x	x	x	x	x
Cr	.000x	.000x	.000x	.000x	.000x
V	0	.00x	.00x	.00x	.00x
Ti	.0x	.0x	.0x	.0x	.0x
Zr	.00x	.000x	.000x	.000x	.000x
Sr	.000x	.00x	.00x	.00x	.00x
Ba	.00x	.00x	.00x	.00x	.00x
B	.00x	.00x	.00x	.00x	.00x

Laboratory	Field	
52-1875	Bh-D-1	Dolomite 17 ft above base of section.
1876	D-2	Do.
1877	C-1-a	Limestone "curleycues" 19 ft above base of section.
1878	C-1-b	Dolomite matrix 19 ft above base of section.
1879	C-2-a	Limestone "curleycues" 19 ft above base of section.
1880	C-2-b	Dolomite matrix 19 ft above base of section.
1881	A-1	Alternate algal limestone and dolomite, 31 ft above base of section.
1882	A-2	Alternate limestone and dolomite 31 ft above base of section.
1883	L-1	Massive limestone 48.5 ft above base of section.
1884	L-2	Do.

TABLE 28.—*Quantitative spectrographic analyses of noncarbonate rocks of the Missoula group and related units plus miscellaneous ones*

[Analyst: A. A. Chodos]

	50-863CS	50-864CS	50-871CS	50-872CS	50-873CS	50-874CS
Ni	0.002	0.002	0.002	0.0002	0.0007	0.002
Co	.001	.001	.0008	.0003	.0005	.001
Cr	.007	.009	.007	.0007	.001	.008
Cu	.001	.001	.0004	.0002	----	.0004
V	.005	.009	.005	.001	.002	.01
Ga	.002	.001	.001	----	----	.002
Sc	.002	.002	.001	----	.0005	.002
Pb	.005	----	----	----	----	.003
Ba	.05	.06	.05	.05	.1	.05
Sr	.006	.01	.01	.004	.04	.006
Be	.0002	----	----	----	----	.0001
Mo	----	----	----	----	----	----
Y	.009	.01	.009	----	----	.007
Yb	.0008	.0008	.0006	----	----	.0005
La	.02	.02	.02	----	----	.02
Zr	.02	.07	.05	.009	.009	.01
B	.004	.01	.009	.004	.002	.02

	50-875CS	50-870CS	50-865CS	x	y
Ni	0.002	0.010	0.004	0.001	0.003
Co	.001	.004	.001	.0001	.0009
Cr	.008	.02	.01	.0005	.004
Cu	.0003	.007	----	.0004	.006
V	.008	.03	.01	.0003	.005
Ga	.002	.002	.002	----	.0005
Sc	.002	.004	.002	----	.001
Pb	----	----	----	----	----
Ba	.05	.04	.06	.0003	.02
Sr	.01	.04	.01	.0001	.007
Be	.0002	----	----	----	----
Mo	----	----	----	----	.0002
Y	.009	.006	.009	----	.005
Yb	.0006	.0008	.0007	----	.0002
La	.02	----	.01	----	----
Zr	.03	.02	.02	.006	.02
B	.008	----	.01	.004	.005

Laboratory	Field	Kind of rock and source
50-863CS	6-47	Spokane shale, Spokane Hills, west of Canyon Ferry quadrangle.
864CS	9-47	"Miller Peak" of Clapp and Deiss (1931). Prickly Pear Creek.
871CS	Belt 1	Missoula group, Rock Creek opposite mouth of Gilbert Creek, in the Bonner quadrangle.
872CS	5	Pink quartzite along Blackfoot Canyon north of Bonner, in the Bonner quadrangle.
873CS	7	Red quartzite, Middle Fork Flathead River, Nyack quadrangle.
874CS	8	Green argillite. Middle Fork Flathead River, Nyack quadrangle.
875CS	9	Green quartzite at base of the Missoula group, Middle Fork Flathead River, Nyack quadrangle.
870CS	7-49	Grit, North Boulder group, Jefferson Canyon near La Hood.
865CS	19-49	Missoula group, West of Missoula in T. 12 N., R. 24 W.
x	N 2	Neihart quartzite, Jefferson Creek, northeastern part of unsurveyed T. 13 N., R. 8 E.
y	N 3	Chamberlain shale, Jefferson Creek near Neihart in the northeastern part of unsurveyed T. 13 N., R. 8 E.

TABLE 29.—*Quantitative spectrographic analyses, in percent, of the Ravalli group in Glacier National Park*

[Analyst: A. A. Chodos]

	Grinnell Formation						Appekunny Formation			
	51-2160SDCD	51-2161CDSD	50-2065SC	50-2066SC	50-858CS	50-868CS	50-2067SC	50-2068SC	50-861CS	50-866CS
Ni	0.0002	0.002	0.002	0.001	0.002	0.001	0.002	0.002	0.001	0.002
Co	.0002	.001	.001	.0008	.0009	.0004	.001	.0008	.0007	.002
Cr	.0003	.004	.006	.004	.008	.006	.005	.005	.007	.009
Cu	.01	.01	.0003	.001	.01	.0003	.0006	.002	.0008	.0002
V	.0004	.006	.009	.006	.01	.004	.006	.006	.005	.005
Ga	.0001	.001	.002	.001	.002	.001	.002	.001	.002	.001
Sc	----------	.001	.001	.001	.002	.001	.0009	.001	.002	.002
Pb	----------	----------	----------	----------	----------	----------	----------	----------	----------	----------
Ba	1.0	.05	.06	.06	.06	.04	.06	.06	.04	.04
Sr	.03	.005	.01	.005	.01	.005	.01	.02	.01	.006
Be	----------	.0002	.0001	.0001	----------	----------	.0001	.0001	----------	----------
Y	.0002	----------	.01	.01	.008	.009	.007	.007	.01	.01
Yb	.001	.004	.0006	.0006	.0006	.0007	.0004	.0004	.0008	.0007
La	.0001	.0004	.01	.01	.02	.007	.01	.01	.02	.02
Zr	.006	.02	.02	.02	.03	.07	.02	.02	.03	.05
B	----------	.005	.004	.004	.005	.03	.005	.004	.007	.006

NOTE: The map in Ross (1959) will aid in fixing the location of samples from Glacier National Park and vicinity. The Rock Creek mentioned is a tributary of Clark Fork southeast of Missoula. This note applies also to most of the samples in the other tables of analyses.

Laboratory	Field	Rock specimen	Source of sample
51-2160SDCD	A-1	Quartzite	Badrock Canyon, T. 30 N., R. 20 E.
2161CDSD	A-2	Argillite	Badrock Canyon, T. 30 N., R. 20 E.
50-2065SC	41BGP	Purple argillite	Going-to-the-Sun Highway near McDonald Falls.
2066SC	41CGP	Green argillite	Going-to-the-Sun Highway near McDonald Falls.
858CS	FM 23	Argillite	Noisy Creek, Nyack quadrangle.
868CS	4-49	do	Rock Creek, unsurveyed T. 10 N., R. 16 E., 2 miles up from mouth of Ranch Creek in 30-minute Bonner quadrangle.
2067SC	42GP	Green argillite	Upper part of formation near McDonald Falls, Going-to-the-Sun Highway.
2068SC	44GP	Dark argillite	Lower part near McDonald Falls, Going-to-the-Sun Highway.
861CS	FM 68	Argillite	Many Glacier Hotel.
866CS	2-49		Rock Creek above mouth of Ranch Creek, unsurveyed T. 10 N., R. 16 E., in 30-minute Bonner quadrangle.

TABLE 30.—*Quantitative spectrographic analyses, in percent, of rocks of the Spokane and Greyson shales from the southern Elkhorn Mountains, Mont.*

[Data furnished by M. R. Klepper. Analyst: Janet D. Fletcher]

	53-102SC	53-103SC	53-104SC	53-105SC	53-106SC	53-107SC	53-108SC	53-109SC	53-110SC
Cu	0. 002	0. 002	0. 0004	0. 0008	0. 0004	0. 003	0. 001	0. 001	0. 001
Pb	0	. 02	0	0	0	0	0	0	0
Mo	0	. 009	0	0	0	0	0	0	0
Co	0	0	. 0009	0	0	0	0	0	0
Ni	. 0006	0	. 002	. 001	. 0006	. 0006	0	0	0
Ga	. 001	0	. 002	. 001	. 001	. 0004	. 0006	0	0
Cr	. 004	. 001	. 004	. 003	. 002	. 001	. 001	. 0002	. 0002
V	. 003	. 002	. 002	. 002	. 001	. 001	. 001	0	. 002
Y	. 004	0	. 004	. 004	. 003	. 002	. 002	0	0
Yb	. 0005	0	. 0005	. 0005	. 0004	. 0002	. 0003	0	0
Zr	. 02	. 003	. 02	. 02	. 02	. 004	. 003	0	0
Sr	. 002	. 03	. 003	. 008	. 002	. 002	. 03	0	. 1
Ba	. 07	. 02	. 02	. 02	. 02	. 02	. 02	. 0008	(1)
B	. 003	0	. 006	. 005	. 005	0	0	. 002	0

1 More than 1 percent.

Laboratory	Field	Location			Kind of rock
		Section	Township and range		
53-102SC	48K8	NW¼ sec. 27	T. 6 N., R. 1 E.	Limestone Hills	Spokane; red argillite within 20 ft of unconformity at top of Precambrian.
103SC	49PCb	SE¼ sec. 13	T. 4 N., R. 2 W.	North of county road	Spokane; cryptozoan limestone 400 to 500 ft below unconformity at top of Precambrian.
104SC	49-960a	NE¼ sec. 26	do		Spokane; reddish brown mudstone ("argillite") with salt casts, approximately 270 ft above base of formation.
105SC	960b	do	do		Spokane; reddish brown mudstone ("argillite") with nodular structure, 2 ft below sample 960a.
106SC	960c	do	do		Spokane; quartzite (uneven grained with clay-silt lenses), 235 ft above base.
107SC	960d	do	do		Spokane; quartzite (uneven grained with clay-silt lenses), 8 ft below sample 960c.
108SC	960e	do	do		Spokane; quartzite (uneven grained), 13 ft below sample 960c.
109SC	A12	Sec. 22	do	Near Ida mine	Greyson; tan fine-grained quartzite, within 100 ft of top of formation.
110SC	A29		do		Spokane; varicolored oolitic limestone. Gradational beds between red Spokane and nonred Greyson.

TABLE 31.—*Quantitative spectrographic analyses, in percent, of argillite rock samples from the Bonner 2 (15-minute) quadrangle near Missoula, Mont.*

[Data furnished by W. H. Nelson. Analyst: Harry Bastron]

	Miller Peak Formation				McNamara formation			
Locality 1								
	146253	146254	146255	146256	146257	146258	146259	146260
Cu	0.0005	0.0002	0.0004	0.0004	0.001	0.0003	0.004	0.002
Ag	0	0	0	0	0	0	0	0
Sn	0	0	0	0	0	0	0	0
Pb	.005	0	0	0	.003	0	0	0
Bi	0	0	0	0	0	0	0	0
Mn	.04	.06	.02	.02	.01	.02	.02	.01
Co	.0003	.0003	.0004	.0004	.0004	.0005	.0005	.0004
Ni	.0005	.001	.0006	.0005	.0006	.0008	.0006	.0004
Fe	1.7	1.4	1.0	.8	1.3	1.1	1.4	1.0
Ga	.001	.001	.001	.001	.001	.001	.001	.001
Cr	.01	.008	.003	.002	.004	.004	.005	.004
V	.006	.006	.004	.003	.004	.005	.004	.003
Sc	.002	.002	.0005	.0005	.001	.001	.001	.0005
Y	.004	.004	0	0	.003	.004	.004	.004
Yb	.0002	.0002	0	0	.0002	.0002	.0002	.0001
La	.003	.003	0	0	.003	0	.003	.003
Ti	.3	.2	.2	.1	.2	.2	.2	.1
Zr	.04	.01	.04	.01	.02	.04	.03	.03
Be	.0002	.0001	.0002	.0002	.0002	.0003	.0002	.0002
Ca	.5	>2	.2	.3	.05	.05	.05	.04
Sr	.003	.01	.001	.001	.002	.003	.003	.003
Ba	.05	.2	.02	.02	.03	.03	.03	.03
B	.02	.007	.01	.01	.003	.007	.003	.003
Locality 2								
	146261	146262	146263	146264	146265	146266	146267	146268
Cu	0.0002	0.0005	0.0008	0.0005	0.0007	0.001	0.002	0.003
Ag	.0002	0	0	0	0	0	0	0
Sn	.003	0	0	0	0	0	0	0
Pb	.004	0	0	0	0	0	0	0
Bi	.004	0	0	0	0	0	0	0
Mn	.01	.01	.01	.01	.02	.02	.02	.04
Co	.0003	.0003	.0005	.0004	.0003	0	.0003	.0005
Ni	.0008	.0009	.0008	.0006	.0004	.0004	.0004	.0008
Fe	1.4	1.7	2.0	1.9	1.2	.9	1.5	2.2
Ga	.001	.001	.001	.001	.001	.001	.001	.002
Cr	.008	.01	.005	.005	.003	.003	.003	.005
V	.005	.006	.005	.006	.004	.003	.004	.006
Sc	.002	.003	.002	.002	.001	.001	.001	.002
Y	.003	.007	.004	.004	.004	.003	.004	.004
Yb	.0001	.0004	.0003	.0003	.0003	.0002	.0003	.0003
La	.003	.003	.003	.003	.003	.003	.003	.003
Ti	.2	.3	.3	.3	.2	.2	.2	.2
Zr	.04	.05	.05	.04	.03	.03	.02	.02
Be	.0003	.0002	.0002	.0002	.0002	.0002	.0002	.0002
Ca	.05	.03	.04	.03	.06	.09	.05	.1
Sr	.001	.002	.001	.001	.002	.002	.001	.002
Ba	.02	.02	.04	.04	.06	.05	.06	.06
B	.01	.02	.02	.01	.01	.004	.009	.008

Laboratory	Field	Location		Kind of rock
		Section	Township and range	
146253	DFH6-26-1	SW¼SE¼ sec. 32	T. 12 N., R. 17 W.	Sandy argillite.
146254	2	SE¼SW¼ sec. 32	do	Argillite.
146255	8	do	do	Sandy argillite.
146256	9	SW¼SW¼ sec. 32	do	Do.
146257	5	SW¼SE¼ sec. 32	do	Do.
146258	6a	do	do	Argillite.
146259	6b	do	do	Sandy argillite.
146260	7	NW¼SE¼ sec. 32	do	Argillite.
146261	CH-10-13-9	SE¼NE¼ sec. 24	T. 12 N., R. 19 W.	Do.
146262	10	SE¼NW¼ sec. 24	do	Sandy argillite
146263	CH-10-14-37	NW¼NE¼ sec. 26	do	Argillite.
146264	38	SW¼SE¼ sec. 23	do	Do.
146265	CH-10-13-3	NE¼SW¼ sec. 18	T. 12 N., R. 18 W.	Do.
146266	3a	do	do	Do.
146267	3b	do	do	Do.
146268	4	NE½SW¼ sec. 18	do	Do.

TABLE 32.—*Semiquantitative spectrographic analyses, in percent, of sedimentary rocks from the Coeur d'Alene mining district and environs, Shoshone County, Idaho, and Mineral County, Mont.*

[Data furnished by S. W. Hobbs. Analyst: Janet D. Fletcher]

	54–414SC	54–415SC	54–416SC	54–417SC	54–418SC	54–419SC	54–420SC	54–421SC	54–422SC	54–423SC	54–424SC	54–425SC	54–426SC	54–427SC	54–428SC	54–429SC
Cu	0.000x	0.000x	0	0.000x	0.00x	0.000x	0	0	0.00x	0.00x	0.000x	0.000x	0.000x	0	0.00x	0.000x
Pb	.00x	.00x	0	.00x	.00x	.00x	0	.00x	.00x	.00x	0	.00x	.00x	.00x	.00x	.00x
Mn	.x	.x	.0x	.00x	.0x	.x	.00x	.00x	.00x	.0x	.0x	.0x	.0x	.0x	.x	.00x
Co	.000x	.000x	.000x	.000x	.000x	0	0	0	0	.00x	.000x	.00x	.00x	.00x	.00x	0
Ni	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.000x
Ca	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x
Cr	.00x	.00x	.00x	.00x	.00x	.000x	.00x	.00x	.000x	.00x	.00x	.00x	.00x	.00x	.00x	.00x
V	.00x	.00x	.0x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x
Sc	.000x	.000x	.00x	.000x	.00x	.000x	.00x	.000x	.00x	.00x	.000x	.00x	.00x	.00x	.00x	.000x
Y	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x	.00x
Yb	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x	.000x
La	0	0	0	0	.00x	0	0	.0x	.0x	.00x	0	0	0	0	.00x	0
Ti	.x	.x	.x	.x	.x	.x	.x	.x	.x	.x	.x	.x	.x	.x	.x	.x
Zr	.0x	.0x	.0x	.0x	.0x	.00x	.0x	.0x	.0x	.0x	.0x	.0x	.0x	.0x	.0x	.00x
Be	.000x	.000x	.000x	.000x	.000x	0	.000x	.000x	.000x	.000x	0	.000x	.000x	.000x	.000x	0
Sr	.000x	.000x	.00x	.00x	.00x	.0x	.00x	.0x	.000x	.0x	.00x	.00x	.00x	.00x	.0x	.00x
Ba	.0x	.0x	.x	.0x	.0x	.0x	.0x	.x	.0x	.0x	.0x	.x	.x	.0x	.0x	.0x
P	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
B	.0x	.0x	.0x	.00x	.00x	.00x	.00x	.0x	.0x	.00x	.00x	.00x	.00x	.00x	.00x	.00x

Laboratory	Field	Formation	Kind of rock	Source of sample
54–414SC	VCF–286U–53	Burke or Prichard	Sericitic quartzite	2,400-foot level of Frisco mine.
415SC	286A–53	do	do	Do.
416SC	SWH–1–54	St. Regis	Argillite	Surface at portal of Rock Creek adit.
417SC	65–48	Burke	Quartzite	Glidden Lake measured section.
418SC	ABG–121–48	Upper part of Prichard	Argillite and quartzite, laminated	
419SC	16–48	Wallace	Argillaceous limestone	
420SC	29–48	Wallace, lower part	Argillite interbedded with quartzite	Placer Creek road.
421SC	SWH–28–48	Wallace	Sericitic rock	Gold Hunter mine.
422SC	82–48	St. Regis	Phyllite	East Fork Willow Creek—Reindeer Queen structure.
423SC	2–54	Prichard	Argillite	Near first sawmill on main Pine Creek.
424SC	14–JH–52	St. Regis	Quartzite, purple	1.5 miles south of Saltese on Silver Creek road.
425SC	15–JH–52	do	Argillite, purplish-gray	Do.
426SC	8–JH–52	do	Argillite, banded	Road cut near Lookout Pass.
427SC	39–JH–52	do	Argillite	Lookout Pass.
428SC	37–JH–52	do	Argillite, green	7.1 miles east of Henderson on U.S. Highway 10.
429SC	13–JH–53	do	Argillite	Little Baldy Mountain, Murray district.

The analyses are, thus, inadequate for use in stratigraphic correlation. Even so, in view of the fact that so few trace element analyses of Belt rocks are on record it seems worthwhile to give the available ones in this report. Further, careful qualitative comparisons, aided by various attempts at representing them on graphs, leave the writer with the strong personal impression that possible distinctions between stratigraphic units are suggested even by the present analyses. Hence, the idea has sufficient merit to warrant pursuit whenever a suitable area of Belt rocks is exhaustively studied.

In general (Krauskopf, 1955) nearly all the minor metals are somewhat more concentrated in shales than in limestones or sandstones, but the amount in ordinary shales is seldom more than twice the crustal abundance. The greatest enrichment of rare metals is in rocks rich in organic matter, but the Belt series contains no rocks very rich in organic matter. The preceding analyses agree reasonably well with Krauskopf's findings. This fact, in itself, is of interest because it indicates that the relative proportions of trace elements have survived such metamorphism as the Belt series has been subjected to. The carbonate-rich rocks contain distinctly fewer trace elements than the argillaceous rocks. As most of the carbonate rocks are argillaceous, the demarcation is not sharp. The pure quartzites, such as the Neihart, contain relatively small quantities and somewhat fewer of the trace elements than the argillaceous ones.

LITHOLOGIC DESCRIPTIONS

PRICHARD FORMATION

DISTRIBUTION

The Prichard formation (or slate) is rather widely exposed in Shoshone County, Idaho, especially south of the St. Joe River (Umpleby and Jones, 1923, pl. 1), and extends into Clearwater County (Anderson, 1930b, p. 10) and northward into Boundary County (Kirkham and Ellis, 1926, p. 15–16, 20–27). In Montana, the Prichard formation is widespread from the northwest corner of the State southward almost as far as Superior. Small exposures referred to it with different degrees of certainty are found at intervals from near Lolo Pass to areas southwest of Anaconda.

CHARACTER

Ransome and Calkins (1908, p. 29–32) say that in the Coeur d'Alene area the thickness exceeds 8,000 feet. All except the uppermost part consists of dark-blue-gray argillite with shaly partings and imperfect slaty cleavage interbedded with considerable indurated sandstone. The rock is reddish brown on weathered surfaces, mainly as a result of weathering of iron sulfides. Part of the sandstone is quartzitic and part grades into the argillite. The upper few hundred feet of the formation is more sandy and, near the top, greenish-gray siliceous shale, like that in the Burke formation, appears. In these upper few hundred feet ripple marks, sun cracks or mud cracks, rain pits, intraformational conglomerate, and "pseudoconglomerate" are conspicuous. The pseudoconglomerate is accounted for by Calkins as having been formed on a sloping shore by waves that rolled up masses of water-soaked sand, which became flattened by their own weight.

Under the microscope Calkins found that the rocks consist of angular to subangular detrital grains of quartz, feldspar, and muscovite in an abundant cement of sericite and quartz mingled with minute opaque particles, probably carbonaceous. Minor constituents, in part of hydrothermal origin, include leucoxene, rutile, and zircon, and such minerals as tourmaline, siderite, pyrite, magnetite, biotite, and chlorite.

A more generalized description (Umpleby and Jones, 1923, p. 7) states that the formation consists of blue shale and slate with some gray sandstone and quartzite at least 8,000 feet thick. According to this, a massive quartzite middle member is 200 feet thick near Pine Creek, south of Kellogg, and a unit presumed to be the same member is 1,500 feet thick in the southern part of Shoshone County. More recent studies (Forrester and Nelson, 1944, p. 7) indicate that in the Pine Creek area this member is 400 feet thick, the middle part being argillite. Unpublished mapping in the same general area (A. B. Campbell, 1959, oral communication) indicates an aggregate thickness of 2,000 feet for the quartzite and argillaceous strata of the middle member. The different thicknesses thus recorded by different observers show that the member is ill-defined and may vary markedly along the strike. However, the fact that members have been mapped shows that the Prichard is susceptible of subdivision.

The Prichard (here termed Prichard-Aldridge) formation in Boundary County, Idaho, is notably different from the type section in the Coeur d'Alene area (Kirkham and Ellis, 1926, p. 15–16, 20–27). The thickness exceeds 10,000 feet, with the base not exposed. The formation consists of pure light-gray quartzite, argillaceous quartzite, and argillite. The argillaceous rocks are gray to black. Argillite is subordinate to quartzitic rocks, the reverse of the relationship in most other localities.

Specimens collected by S. Warren Hobbs from the Prichard formation and its transitional zone with the Burke in eastern Shoshone County consist of laminated argillite containing quartz and white mica with some biotite, hornblende, tourmaline, and epidote. The quartz grains are as large as 0.06 mm in diameter and some of the amphibole metacrysts are 0.3 mm long. Those rocks are more metamorphosed than most of the Belt series in Montana.

Another description (Shenon and McConnel, 1939, p. 3) gives the maximum thickness as more than 12,000 feet with the base not exposed and says that the formation consists principally of dark-gray to almost black laminated argillite, commonly with slaty cleavage. According to this description the formation contains at least two well-defined quartzitic zones, one near the top and the other about 10,000 feet lower.

In northern Bonner County (Anderson, 1930a, p. 14-15) the formation is more sandy than it is in much of Shoshone County. It consists of gray argillaceous quartzitic sandstone with interbedded shale and minor lenses of intraformational conglomerate. The thickness is in excess of 20,000 feet, of which only the upper 1,500 feet resembles the typical argillaceous rock of the Prichard in the Coeur d'Alene region.

In Kootenai and Benewah Counties, Idaho, (A. B. Campbell, 1950, written communication) only a small part of the formation is exposed. The part there visible is chiefly dark-gray to black argillite with some argillaceous sandstone and quartzite particularly near the top. The argillite is thinly laminated, with various shades of gray and black. In places the rock is reported to have a distinctly varved appearance. Slaty foliation is especially prominent low in the sequence. Mud cracks and ripple marks have been noted high in the section.

In Mineral County, Mont. (Wallace and Hosterman, 1956, p. 578-579; Campbell, 1960, p. 552-553), the Prichard formation includes fine-grained, medium-gray and greenish-gray quartzite and siliceous argillite in beds 2 inches to 2 feet thick. The rock here is similar to that in the Coeur d'Alene region, Idaho, except that it is more uniformly siliceous, with fewer argillite beds. Along the Clark Fork below its junction with the Flathead River, in Sanders County, the Prichard is exposed almost continuously and appears to be 17,000 feet thick, with the top cut off by a fault. The possibility of duplication by faulting within this thick section remains but is supported by no evidence now at hand. Further, as in all other areas, the base of the Prichard is not exposed. In this locality the Prichard formation consists mostly of fine-grained quartzite and siliceous argillite. Common colors are medium gray and greenish gray. Some of the argillite has pronounced cleavage parallel to the bedding.

Tweto (see p. 41) speaks of rock from the Prichard formation a mile north of Camas Hot Springs, Sanders County, Mont., as quartz-sericite phyllite or quartzitic argillite. The rock is steel gray, weathers drab or green, and consists of quartz, sericite, and biotite with minor amounts of graphite, tourmaline, epidote, grossularite, chlorite, titanite, and muscovite. Campbell (1960, p. 553) noted most of these and, in addition, some feldspar and pyrite.

The Prichard in the Libby quadrangle (Gibson, 1948, p. 10–11) consists mostly of dark-gray to blue-gray sandy laminated argillite, interbedded with some light-colored sandstone and quartzite and a little thin-bedded shale. Some beds contain carbonate, possibly ferruginous, and sparse sun cracks (or mud cracks) and ripple marks. Sandy beds that were studied petrographically were found by Gibson to be composed of quartz and sericite, mainly 0.01 to 0.03 mm in diameter. Some chlorite and a little detrital zircon, tourmaline, apatite, and rutile are also present. The formation is so deformed as to be difficult to measure, but at one place a minimum of 9,700 feet is exposed, with the base hidden. Presumably the rock in the Libby quadrangle is representative of most of the formation in western Montana. Most other areas there are known only on a reconnaissance basis.

East of Lolo Pass the Prichard is reported (Langton, 1935, p. 36) to be predominantly brownish sandstone, containing feldspar, much of which is highly metamorphosed to biotite-muscovite schist. Still farther southeast (Calkins and Emmons, 1915, p. 3–4) the formation consists of micaceous schist and gneiss, dark gray on fresh fractures, deep reddish brown on weathered surfaces. These metamorphic rocks were derived from clay shale with some sandy layers, now represented by quartzite. The least metamorphosed of the argillaceous rocks is reported to be laminated and to resemble the slate of the Prichard formation in Shoshone County, Idaho. The exposed thickness in this part of Montana has been estimated at about 5,000 feet.

Calkins and Emmons distinguished a unit composed of white to gray quartzite with micaceous partings and, in the upper part, layers of mica schist, about 1,000 feet thick, with the base not exposed. They called it Neihart quartzite. Outcrops of the unit are small and relations to other rocks are not clear. So many uncertainties are involved that such a correlation seems unwarranted at the present time. Neihart, the only other locality where the name has been used, is nearly 140 miles away. Possibly part of the quartzite they regarded as Neihart is interbedded with the argillaceous rocks supposed to belong to the Prichard formation. Even the correlation of the argillaceous rocks with the Prichard is far from proven as the outcrops are small, metamorphosed and far from the type locality of that formation in Idaho.

Obviously the various descriptions just summarized indicate significant differences in character and thickness of rocks assigned to the Prichard. Some of the differences are in areas close to each other. Because the formation crops out in so many mining districts, more is known about it than about some other Belt rocks. Lateral variations as marked and frequent as those indicated for the Prichard formation may exist in less well known areas that contain later parts of the Belt series.

NEIHART QUARTZITE

The Neihart quartzite is definitely known only in a small area close to Neihart in the southeast corner of Cascade County, Mont. At present the use of the name should be limited to this area, there being no means of equating the quartzite at Neihart with others. The quartzite rests in depositional contact on coarse granitoid gneiss, the only place in Montana or Idaho where such a relationship is unequivocal. In the few other localities where rocks of the Belt series are close to older rocks, faults are known or suspected.

The Neihart quartzite is described (Weed, 1900, p. 281–282) as quartzite that grades into sandstone, largely massive, but with distinct bedding locally. The color is "creamy white to gray or pink." Quartz pebbles, in part in thin lenses, occur, but well-defined conglomerate beds have not been found. The basal 300 feet of the formation is a compact, uniform body of white and pink quartzite, forming conspicuous escarpments. The higher rocks are thinner bedded with greenish mica. Still higher in the sequence, micaceous shale is interbedded with the quartzite, increasing upward to a gradational contact with the overlying Chamberlain shale. The shale in the upper part of the Neihart is reported to be blackened by carbonaceous matter. The total thickness is 700 feet.

Specimens collected from the same general locality by Richard Rezak include light-pinkish or greenish-gray quartzite consisting mainly of quartz, in part in well-rounded grains, in part in a tightly packed mosaic of grains that have been crushed and recrystallized (fig. 5*A*). Grain diameters range from 0.01 to 0.50 mm the largest being uncrushed, rounded detrital material. The quartz is surrounded by thin films of micaceous minerals, in part sericite, probably with some clay.

There are also a few grains now composed of similar aggregates of micaceous habit that may be altered feldspar. A little chlorite, biotite, epidote, apatite, and tourmaline are present.

RAVALLI GROUP

In most places in Montana the Ravalli group is undivided or is split into the Appekunny formation below and the Grinnell formation at the top. In Glacier National Park the Altyn limestone is the lowest exposed unit of the Belt series. It is provisionally included in the Ravalli group but could be older than any of the units referred to that group elsewhere. Similarly the Chamberlain shale, near Neihart, is at present assigned to the Ravalli group but may be older. In Idaho and in adjacent parts of Montana the Burke, Revett, and St. Regis formations, in ascending order, are assigned to the Ravalli group. Each is described, roughly in order from west to east. There are wide areas in western Montana where the Ravalli group has been so little studied that its components have not been distinguished, and only the most general statements as to the group have been published (Walcott, 1906, p. 7–9; Calkins, 1909, p. 37–38; Clapp, 1932, p. 22).

BURKE FORMATION

DISTRIBUTION

The Burke formation is widely distributed in Shoshone County, Idaho, especially near Burke and from there into Montana (Umpleby and Jones, 1923, pl. 1). It is also present in adjacent areas in Idaho (Sampson, 1928, p. 6–7; Anderson, 1930a, p. 15–16, 1940, p. 11). It, or a unit that resembles it, is reported to be abundant in Boundary County (Kirkham and Ellis, 1926, p. 16). Although rocks equivalent to the Burke formation are known to occur in western Montana, little information is available (pl. 1) (Gibson, 1948, p. 9–12; Wallace and Hosterman, 1956, p. 588; Campbell, 1960, p. 552–553).

CHARACTER

The rocks of the Burke formation in the Coeur d'Alene region, Idaho, range from nearly pure quartzite to siliceous shale (Ransome and Calkins, 1908, p. 32–34) and the upper and lower parts approach the Revett and Prichard respectively in lithologic character. The upper boundary, particularly, is gradational and indefinite. Many of the rocks are pale greenish gray; some are light purplish; a few beds have brighter green and purple tints; and some of the quartzite is white. Sun cracks and ripple marks are seen in almost every outcrop. The rocks differ from those of the Prichard chiefly in containing less carbonaceous dust. The coarse thick beds in the formation contain "rather angular grains of quartz and feldspar and flakes of presumably clastic mica in an abundant sericitic paste" (Ransome and Calkins, 1908). Much of the quartz and sericite is secondary. A few beds are calcareous.

Specimens from eastern Shoshone County furnished by S. Warren Hobbs represent finely laminated rocks consisting of quartz, micaceous minerals, chlorite, and some plagioclase. The chlorite occurs as large flakes. The quartz is in an intricate recrystallized mosaic of grains that range in diameter from less than 0.01 mm to 0.30 mm.

In Kootenai and Benewah Counties, Idaho, (A. B. Campbell, 1950, written communication) extensive faulting and poor exposures hinder stratigraphic studies. About 75 to 85 percent of the formation is quartzite of various degrees of purity, mostly sericitic. Most beds are from 2 inches to a foot thick, with a few as much as 3 feet thick. Extremely thin, widely spaced laminations within the beds are common. Most of the quartzite is gray to light tan, with purple tints locally and, in the more sericitic beds, green casts. Many beds have numerous brown spots resulting from weathering of small groups of ankerite crystals. Thin-bedded gray-to-green siliceous argillite is scattered throughout the formation, especially in the lower part.

Hershey (1912, p. 750) used a central purplish-gray bed as a basis of subdivision into two members near Wardner. The thickness of the formation varies somewhat from place to place but approximates 2,000 feet. Possibly the thickness increases west of the Coeur d'Alene region as a recent report on an area in Kootenai and Benewah Counties, Idaho, (A. B. Campbell, 1950, written communication) notes that the thickness there "does not exceed 3,000 feet." In western Montana the Burke and Revett formations have been mapped as a unit. Near the Idaho border the two together total at least 3,000 feet (Wallace and Hosterman, 1956, p. 560–581) and somewhat farther east (Campbell, 1960, p. 554) the estimate is 4,500 feet.

In Mineral County, Mont., a major fault zone passes south of the St. Regis River and the two formations together appear to be different on the two sides of the zone (Wallace and Hosterman, 1956, p. 580–581). South of the fault zone they consist of fine-grained greenish-gray quartzite with beds commonly 1 to 4 feet thick. White vitreous quartzite, such as characterizes the Revett farther west in Idaho, is rare. Further, as the purplish-gray color characteristic of the St. Regis formation in Idaho is absent, there is difficulty in fixing the upper limit of the Revett. North of the fault zone light-colored thick-bedded vitreous quartzite such as is typical of the Revett crops out in several localities. Likewise greenish-gray thin-bedded impure quartzite similar to that which characterizes the Burke in Idaho was noted. The deformation is so intricate that mutual relationships and thicknesses were not satisfactorily determined.

A. B. Campbell (1950, written communication) describes the Burke and Revett formations together in the St. Regis-Superior area. They vary from fine-grained argillaceous quartzite near the lower contact to purer quartzite above. The rocks are commonly medium to light gray or greenish gray, the more quartzitic beds being lighter colored, locally with purple tints. Many bedding planes are lined with sericite. Ripple marks and mud cracks are common. Some of the quartzite is crossbedded.

REVETT QUARTZITE

The Revett quartzite (Ransome and Calkins, 1908, p. 35–36) is best known in the Coeur d'Alene region in Shoshone County, Idaho, although its equivalents undoubtedly extend beyond that region. Most of it is pure rather thick-bedded white quartzite, although impure sericitic beds are included. Ripple marks are reported only in the upper and lower parts of the formation. The thickness is approximately 1,000 feet where measured by Ransome and Calkins but locally exceeds 3,400 feet (Shenon and McConnel, 1939, p. 4). Recent work in the Coeur d'Alene region (S. Warren Hobbs, Mar. 28, 1955, written communication) indicates that apparent discrepancies in measurements of thickness from place to place are due to marked variations in the thickness of the formation within short distances. As an example he cites a measurement of 1,800 feet for the Revett at one place and of 3,400 feet for the same unit less than 5 miles away.

In the purer quartzite, distinctive of the formation, the clastic grains are subangular to rounded, except for secondary quartz overgrowths. Though quartz is the principal constituent, a little feldspar and muscovite, apparently of clastic origin, are generally present. Fragments of seemingly detrital slate are also common. Accessory constituents include zircon, magnetite, rutile, and tourmaline. Secondary siderite occurs locally. The quartzite in Mineral County, Mont., believed to correspond to the Revett, has been grouped with the Burke formation in recent studies, and available data on it are summarized in the description of the Burke given previously.

In Kootenai and Benewah Counties, Idaho, the Revett is composed of medium- to thick-bedded white to tan vitreous quartzite. The more sericitic beds show fine black laminations. Some of the sericitic beds are greenish. A few quartzite beds have small groups of carbonate grains scattered through them. Argillaceous material occurs as partings and thin beds but is nowhere abundant. Ripple marks are present in the sericitic beds but are rare in the massive vitreous quartzite. The thickness is 1,800 feet at a place where a measurement was made but apparently the formation thickens westward, for Anderson (1940, p. 12) reports as much as 3,000 feet in the southern part of Kootenai County.

ST. REGIS FORMATION

Exposures of the St. Regis formation are widely distributed in Shoshone County, Idaho, and probably extend into Boundary County and into neighboring parts of Montana. The formation consists of argillaceous and quartzitic beds with some limy quartzite throughout. Most of the formation as reported by Shenon and McConnel (1939, p. 4) consists of thin fine-grained beds in rather bright tints of green and purple, especially the latter. Recent studies (S. Warren Hobbs, May 2, 1956, written communication) show that the color of the St. Regis formation is very subtle, really bright colors being present locally only. The arenaceous beds are coarser textured and thick. Many are purple but some are white. The limy beds weather to an ocherous color. Shenon and McConnel (1939, p. 4) suggested that green colors result from hydrothermal alteration. Obliteration of the characteristic purple color through alteration is a possible source of misidentification during mapping (A. B. Campbell, 1950, written communication). Shallow-water features, including intraformational conglomerate, are present throughout the formation. The composition as seen under the microscope is essentially like that of formations already described. The components include quartz, feldspar, perhaps some fine-grained slaty rock, mica, rutile, zircon, and tourmaline. The purple and green colors result from specular hematite and chlorite respectively.

In the Coeur d'Alene region the St. Regis formation is about 1,000 to 1,400 feet thick (Shenon and McConnel, 1939, p. 4). Farther west, where the formation is more quartzitic and purple is less prevalent, it is about 1,800 feet thick (A. B. Campbell, 1950, written communication), and in the north-central part of Kootenai County it has been estimated to be about 4,000 feet thick (Anderson, 1940, p. 12).

Near Mullan, Idaho (Hobbs, Wallace, and Griggs, 1950, p. 2–5), the St. Regis formation has been divided into two members. The lower member, 1,200 feet thick, is about 60 percent argillite, with subordinate amounts of quartzite, 35 percent interlaminated quartzite, and argillite containing more than 25 but less than 75 percent quartzite, and the remaining 5 percent is quartzite with subordinate amounts of argillite. Carbonate is recognizable in over 45 percent of the beds in the formation here. Most of it is ankerite but locally the composition approaches that of siderite. It rarely constitutes as much as 20 or 30 percent of the rock and is most abundant in the quartzitic beds. The carbonate is commonly in discrete crystals and clusters of crystals, generally many times larger than the grains of the rest of the rock, but is also found in the cement between detrital grains.

The upper member, 150 to 450 feet thick, is a zone of transition between the St. Regis and Wallace formations. It consists of very fine grained partly porcellaneous light-apple-green thinly laminated siliceous argillite. Rock of this character was formerly grouped with the Wallace formation but is now regarded as more closely allied with the St. Regis.

Near Pend Oreille Lake, Bonner County, the Revett and St. Regis have been grouped together as the Blacktail formation (Sampson, 1928, p. 7). The lower part of the Blacktail is mainly pink quartzite with red shale partings and the upper part consists of alternating red and green argillite, with the red beds in the majority. The aggregate thickness is 8,300 feet.

In Mineral County, Mont. (Wallace and Hosterman, 1956, p. 582–583), the beds of the St. Regis formation differ to the north and south of a major fault zone, in a manner analogous to the Burke and Revett formations previously described. North of the fault zone the St. Regis has essentially the characteristics it has near Mullan, Idaho. To the south a marked facies change begins almost immediately east of the State line. Purplish colors, which in most places are characteristic of the St. Regis formation, become uncommon. They are limited to quartzitic beds, whereas thinly laminated greenish-gray argillite becomes the dominant component of the formation. This argillite has some similarities to that of the upper member in the Mullan area but is far thicker. Five miles east of the State boundary the thickness of the argillite is about 3,000 feet, 3 miles farther east it is over 5,000 feet, and 15 miles still farther east it is approximately 1,700 feet. Wallace and Hosterman say that the facies difference might correspond to deposition near a shore in Idaho and offshore farther east in Montana. They add that the thick part of the St. Regis in Montana might include beds that are time equivalents of parts of the Revett below and of the Wallace above, as these formations are mapped in Idaho.

In the St. Regis-Superior area the formation is composed of 1,900–2,200 feet of thin-bedded quartzose argillite, with a few beds of sericitic quartzite. South of the Osborn fault the predominant color is greenish gray varying to dusky yellow green or medium light gray. North of the fault the formation is greenish-gray and purplish-gray. Purple rocks like those in the Coeur d'Alene area are present in sec. 31, T. 18 N., R. 25 W. (unsurveyed).

ALTYN LIMESTONE

The Altyn limestone is known within the United States only in Glacier National Park (Ross, 1959; Ross and Rezak, 1959, p. 409, 436). Even there exposures are not widespread, and the base is not visible. Most of the formation is a very light gray magnesian limestone or dolomite that weathers a grayish orange, rendering it conspicuous on distant cliffs. Chemical and petrographic data now available indicate that so much of the formation is dolomitic that it would be more appropriately called Altyn dolomite rather than Altyn limestone. The latter name has, however, been in general use for so long that no change is here proposed.

The light-gray dolomitic rock that makes up the bulk of the formation is mostly in beds ½ to 4 feet thick. Some is siliceous and sandy, and there are a few beds of grit and conglomerate with a carbonate-rich groundmass. Thin beds of green argillite are intercalated locally, especially near the top of the formation. Some of the carbonate-rich beds weather brown and brownish red rather than the light buff that characterizes much of the unit.

Rock representative of the major part of the formation has a groundmass of crystalline dolomite with grain diameters ranging from a few hundredths of a millimeter to about 0.2 mm. Rounded bodies, at least as much as 2 mm in maximum dimension, are scattered through this groundmass and locally concentrated in layers. Some of them may be pebbles, but the majority are probably oolites, even though many have lost their internal structure. Rounded to subangular clastic grains, several millimeters in maximum diameter, are also present. Most of these grains are of quartz but many are alkalic plagioclase and some are microcline (fig. 3*A*). Some of the grains appear to be bits of rather coarse-grained granite. Others are fine quartz aggregate that may include silicified oolites. Some of the feldspar is sericitized but much is remarkably fresh. Small amounts of montmorillonite and illite are present. Intraformational conglomerate (edgewise mud breccia) occurs locally. In some specimens Tweto (p. 41) notes the presence of carbonaceous dust. Otherwise his description approximates that just given.

The Altyn limestone contains stromatolites at several localities and apparently at several horizons. None seem sufficiently persistent to be of value in stratigraphic correlations. It also contains carbonaceous films that may be fossil remnants. One thin section (Ross, 1959) shows a structure that resembles a piece of a spine of a trilobite or of a chitinous brachiopod (fig. 1*C*). This single fragment is indeterminate, but it does suggest that some living things in addition to those from which the stromatolites originated existed in Altyn time.

The Fentons (1937b, p. 1881–1885) divide their Altyn formation into three members called, in ascending order, the Waterton, Hell Roaring, and Carthew. The lowest of the members, reported to be 280 feet thick, is exposed only in Canada, where it has been regarded as a separate formation underlying the Altyn, and with its base not exposed. The Carthew member, 700–900 feet thick, is likewise known only in Canada. Hence the Hell Roaring member of the Fentons corresponds essentially to the Altyn limestone as originally described (Willis, 1902, p. 316, 321). Willis spoke of an upper member of argillaceous ferruginous limestone, 600 feet thick, and a lower member of massive siliceous limestone with concretions, 800 feet thick. His total thickness is close to the 1,200–1,300 feet that the Fentons assign to their Hell Roaring member of the Altyn but is less than the total thickness of the formation within Glacier National Park which has been estimated (Ross, 1959, p. 19) at 2,000 to 2,300 feet.

APPEKUNNY FORMATION

DISTRIBUTION

The Appekunny formation (or argillite) is best known in and near Glacier National Park but can be traced with considerable confidence southward past Flathead Lake to Ravalli. It is also present near Missoula (Langton, 1935, p. 36, fig. 3) and has been recognized as far east as northeastern Powell County (Clapp and Deiss, 1931, chart opposite p. 691, p. 693). Probably a considerable part of the Ravalli group in Montana west of long 114° belongs to the Appekunny formation.

CHARACTER

The original name given these rocks is Appekunny argillite (Willis, 1902, p. 316, 322). However, some of the argillaceous rocks contain considerable carbonate, quartzite in varying degrees of purity, and, locally in Glacier National Park, some slate is included. The diversity in lithologic character justifies changing the name from Appekunny argillite to Appekunny formation. A large part of the formation contains more than 70 percent silica and originated as sandstone rather than mudstone, so that argillite is an inappropriate name for many, probably most, of the beds. The somber colors of many outcrops give a deceptive impression of the proportion of argillaceous matter present. The descriptions here given are based mainly on data assembled by the writer and his coworkers in and near Glacier National Park.

Most of the rocks are thick bedded and weather in joint blocks with dimensions of one to several feet. Thin-bedded members occur locally and in areas of marked deformation show slaty cleavage. Most beds are gray to black with various tints of green and blue; a few are reddish. The comparatively pure quartzite beds are nearly white. Many beds contain thin indistinct laminae, some of which appear to have been broken since the consolidation of the rock. Ripple marks, mud cracks, and intraformational conglomerate are common, and rain and sleet impressions have been reported.

The principal minerals include quartz, feldspar (mainly sodic plagioclase), sericite, bleached biotite, chlorite, and carbonate, largely magnesian. The more quartzitic beds have not been studied microscopically. In many of the other beds, original, detrital rounded to subangular grains of quartz and feldspar are fairly common, but many rocks consist mainly of quartz mosaics in which the original texture has been obliterated. Few grains exceed 0.1 mm in maximum dimension and in many beds the grain size is much less. The micaceous minerals and the carbonate are probably secondary. Part of the feldspar, also, appears to have been recrystallized. None of the rocks appear to contain more than about 10 percent feldspar, and in some feldspar is scarce or absent. Very small amounts of montmorillonite are present, and some of the rock contains illite. One sample appears to contain kaolinite. Figure 5*C* is fairly representative of the formation in Glacier National Park.

The Fentons (1937b, p. 1885, 1887), who saw the Appekunny in the northern part of Glacier National Park and in Waterton Lakes Park across the border in Canada, consider it to be made up of three members. The lowest, the Singleshot, includes dolomite and dolomitic rocks interbedded with argillite and quartzite. It is allied to and has probably been included with the underlying Altyn limestone by most workers. The second and thickest of the Fentons' members, the Appistoki, corresponds in character and thickness to the Appekunny as originally defined (Willis, 1902, p. 322). The Fentons (1937b) describe it as containing "gray, green, olive-brown, and rusty-gray argillite in thin minor but thick major beds, interbedded with thickly stratified greenish, white, or pink quartzite." Intraformational conglomerate and other features of shallow-water deposition are prevalent. The thickness ranges from 2,000 to 2,200 feet. The uppermost or Scenic Point member of the Fentons' appears, from their description, to resemble parts of the Grinnell. Perhaps it should be included with that formation. They describe it as argillite, sandstone, and "gravelly conglomerate," green, purplish, buff, brown, and dull brownish red at the type locality, which is in the southeastern part of Glacier National Park. Presumably the term "gravelly conglomerate" is intended to indicate that intraformational conglomerate is not meant. To the north and south the member is reported by the Fentons to grade into thickly bedded, coarsely mud cracked argillite, which gives way to thick quartzite and subordinate gray and iron-stained argillite. The member is 200 to 700 feet thick. The Fentons' subdivisions of the Appekunny formation cannot be traced far enough to be valid as formally named units (Ross, 1959).

The Fentons (1937b, p. 1885) estimate the Appekunny formation in "its eastern phases" as 2,500 to 5,300 feet thick. South of Glacier National Park the thickness is surely as much as 2,000 feet, and in the Swan Range it may exceed 5,000, as estimated from mapping by the writer (Ross, 1959, p. 24). Clapp's (1932, p. 22) estimate of thickness is 3,500 to 10,000 feet, but he cites no specific localities. Langton (1935, p. 36), presumably for the area south of Missoula, speaks of the formation as predominantly massive bluish-gray and light-gray quartzite grading downward into massive gray quartzite, the total thickness being over 6,000 feet.

The rock along Rock Creek southeast of Missoula, believed to belong to the Appekunny formation, is an argillaceous quartzite. It is a laminated rock predominantly medium gray on fresh surfaces, with some pinkish-gray laminae. It contains some nearly black clay spalls, and some surfaces are ripple marked. It has a groundmass consisting mainly of interlocking quartz grains as much as about 0.06 mm in diameter, with a few plagioclase grains and some others that may be altered feldspar. Green biotite, chlorite, sericite, and montmorillonite are present and carbonate grains are locally segregated.

GRINNELL FORMATION

The distribution of the Grinnell is much the same as that of the Appekunny. It may not extend as far west as the Appekunny probably does but neither formation has been much studied west of long. 114°. Like the Appekunny, the Grinnell seems more satisfactorily described as a formation than as an argillite; a modification that is here proposed. One of the samples analyzed contains about 69 percent silica, less than is recorded for beds of the Appekunny formation. Conversely, one analyzed sample contains nearly 84 percent silica and some beds in the formation appear distinctly more siliceous than this. Carbonate rocks have not been recorded, except in the transition zone at the top of the formation. If a single lithologic term had to be used to describe the Grinnell formation, quartzite would be as much or more accurate than argillite.

Most of the formation is rather thin and even bedded, the characteristic color is red purple, with more distinctly purplish hues than most of the reddish beds in the Belt series. Greenish beds are locally common, and the purer quartzite beds are nearly white. The colors noted in and south of Glacier National Park include pale and grayish blue green, grayish purple and grayish red purple. South of Missoula the rocks have purplish tints but are far less conspicuously colored than in Glacier National Park. Ripple marks, mud cracks, and intraformational conglomerate are fairly plentiful. The surfaces of some quartzite beds have bulbous forms of unknown origin. Thin laminae of reddish argillite are interleaved with quartzite. Some were broken soon after deposition and slabs from them are incorporated in the quartzite (fig. 5*D*). In some of the argillaceous rocks, also, similar evidence of disturbance while unconsolidated is observable in thin section.

Microscopic examination (fig. 5*D*) indicates that most of the Grinnell formation originally ranged from a siliceous mudstone to a arkosic sandstone. In the more argillaceous rocks individual grains are only a few hundredths of a millimeter in diameter, but in the coarser layers the grain diameter ranges from 0.4 mm to over a millimeter. Locally large grains are irregularly scattered through a fine-grained matrix. The argillaceous rocks now consist largely of quartz and fine-grained mica, with some feldspar and, locally, carbonate minerals. Montmorillonite is common and some rocks contain illite (as determined by X-ray). The coarser rocks are similar except that micaceous minerals are less abundant. Feldspar, largely alkalic plagioclase, is more conspicuous in these rocks but some of it may have escaped recognition in the finer rocks. Some grains are themselves fragments of fine-grained sedimentary rocks and some of the quartz grains had strain-shadows before they were incorporated in the rock. Original textures seem better preserved in the Grinnell formation than in the Appekunny formation, but this impression may be based in part on the coarser grain of the rocks of the Grinnell that were examined petrographically.

Locally the Grinnell formation might be divided into members but it is doubtful if these could be traced far. In the northern Swan Range three poorly defined members have been noted. The lowest and thickest is dominantly pale and grayish-red-purple argillite. In the next higher member, the red-purple beds decrease upwards and much of the rock is quartzitic argillite and argillaceous quartzite, with thin argillite partings, generally rather dark red purple. Some partings more nearly resemble parts of the argillite of the Missoula group and of the purple-red argillite in the Grinnell farther north than any of the thicker beds in the Swan Range. The uppermost member commonly consists of grayish-blue-green calcareous argillite and argillaceous limestone, constituting a transition zone below the Siyeh limestone of the Piegan group. This member contains a few red-purple beds, and the unit below it contains some green beds. Nevertheless, the distinction is sufficiently definite so that the zone was mapped in the Swan Range whereas the two members beneath it, mostly poorly exposed, were not mapped. Even the transition zone is so poorly defined farther north in Glacier National Park that it was not mapped (Ross, 1959). The presence of limestone raises a question as to whether the transition zone should not be regarded as part of the Piegan group, but the zone seems more closely allied to the Grinnell formation and is included therewith largely on the basis of content of argillaceous matter.

The Fentons (1937b, p. 1887–1890) studied the Grinnell formation mainly in and north of northern Glacier National Park. They proposed three members that differ from those just described. The lowest, which they call the Rising Wolf member, contains white and pink quartzite interbedded with red argillite in layers that range from mere laminae to beds 5 feet thick. Some green beds are present. The member is reported to be 200 to 700 feet thick and not everywhere clearly distinguishable. The middle member is called the Red Gap member by the Fentons and is of varied character. It consists of argillite in thin minor and thick major beds, dominantly red but incidentally brownish or green, interbedded with pink, white, or greenish-white quartzite, brown sandstone, and sandy argillite. The maximum reported thickness is 2,800 feet, but in places it thins to as little as 650 feet.

The upper part of the formation is called the Rising Bull member by the Fentons and is reported to contain argillite, quartzite, and mud breccia (intraformational conglomerate) forming the initial transition between the Grinnell and the Siyeh. This member is gray, red, green, pink, and white and is 600 to 1,100 feet. In and west of Waterton Lakes Park in Canada, immediately north of Glacier National Park, a thin flow of amygdaloidal lava is intercalated in the upper part of this upper member but no lava has been found in the Grinnell formation anywhere south of the international boundary. As suggested above, the Scenic Point member of the Appekunny as recognized by the Fentons may belong to the Grinnell rather than the Appekunny.

Note that the colors mentioned above are those of the Fentons and do not correspond to the usage of the Rock-Color Chart (Goddard and others, 1948).

The wide range in estimated thickness and lithologic character of the Grinnell formation in and near Glacier National Park reflects much lateral variation. The poor exposures in the northern part of the Swan Range prevent accurate measurement but the total thickness there is roughly 5,000 feet. In the southern part of Glacier National Park, some 25 miles to the northeast, the thickness is close to 2,000 feet. Willis (1902, p. 322–323) estimated the thickness in the northern part of the park as 1,000 to 1,800 feet. The Fentons (1937b, p. 1887) give a range in thickness of 1,500 feet to 3,500 feet. Dyson (1949, p. 7) says the thickness varies considerably but is greater than 3,000 feet in several localities in the park.

Clapp (1932, p. 22) estimated the thickness of the Grinnell in the part of northwestern Montana shown on his generalized map as 2,000 to 3,500 feet. His map includes most of the definitely known exposures of the formation. Langton (1935, p. 36), whose work was south of Missoula, describes the Grinnell as composed of bluish-gray fine-grained thin-bedded argillite, grading downward into more massive-bedded gray quartzitic argillite, commonly ripple marked and with layers of reddish-gray quartzite; the aggregate thickness being about 7,000 feet.

On Rock Creek southeast of Missoula, rock believed to belong to the Grinnell formation is light grayish red purple except in the immediate vicinity of intrusive masses, where it may be green. It is an indistinctly laminated quartzite composed of quartz with some plagioclase, minor amounts of bleached biotite, and a little pyrite, hematite, and leucoxene. Some grains consist of felted mats of sericite, montmorillonite, and illite. The quartz and feldspar are in interlocking grains, some exceeding 0.1 mm in length.

CHAMBERLAIN SHALE

The Chamberlain shale is known only in the immediate vicinity of Neihart although possible equivalents have been reported farther south (Verrall, 1955). It grades into the Newland limestone (Piegan group) above and the Neihart quartzite below, but otherwise no evidence is available as to its stratigraphic relations. It is here assumed to belong to the Ravalli group.

The formation is described (Weed, 1900, p. 282) as composed of dark-gray, almost black shale, largely arenaceous, with some ripple marks. The rocks are slaty and jointed. At the base, transition into the underlying Neihart is shown by admixture of arenaceous and micaceous material. The middle part of the formation is mainly black shale. In the upper part calcareous shale appears, increasing upward until the unit merges with the overlying Newland limestone.

Tweto (p. 41) describes a specimen from Sawmill Creek, southeast of Neihart, as dark-green-black, thin-bedded, very fine grained fissile shale consisting essentially of colloidal clay minerals and sericite with small amounts of pyrite, dolomite, quartz, feldspar, and graphite.

Material collected by Richard Rezak (fig. 5*B*) from the Chamberlain shale south of Neihart is nearly black quartzitic shale, irregularly laminated and with clay spalls on bedding planes. It contains subangular to angular quartz grains, as much as 0.05 mm in diameter, with a few mica flakes in a mat of quartz grains and micaceous shreds, as much as 0.005 mm in maximum dimension, which include some illite.

PIEGAN GROUP

The Piegan group of the present report comprises the Wallace, Siyeh, and Newland formations of most recent writers. Inasmuch as these three terms are roughly synonymous, the area in Montana within which each is to be applied must be decided arbitrarily. The basis here adopted is to speak of the Wallace formation in the area from the Idaho boundary eastward to the vicinity of long 114°30′. Where masses of carbonate rock believed to belong to the Piegan group cross this longitude they are mapped with the noncommittal group name (pl. 1). Similarly in the mountains north of the Flathead Valley and west of the northern reaches of the Flathead River as well as in the Mission Range the term "Piegan group" is employed. Here "Siyeh limestone" is an alternative designation, but so little study has been given the rocks that as yet the group term is more appropriate. In and near Glacier National Park and from there to the southern part of the Swan Range the name Siyeh limestone is the most suitable. South of here and east of long 114°30′ Newland limestone is the preferred designation, although usage has not been strictly uniform.

The foregoing remarks suggest that at any particular locality the Piegan group is represented at present by a single named map unit. Each of these thick units, now termed "formations," is probably divisible, when detailed work is done, into subdivisions worthy of being ranked as formations, rather than members. This was done partly in early maps of the southern part of Shoshone County, Idaho (Pardee, 1911, pl. 2; Umpleby and Jones, 1923, pl. 1). The descriptions that follow show that additional subdivisions of the Wallace formation are known to be possible and that major subdivisions of the Siyeh limestone have already been suggested. Rocks now called Newland limestone are neither as widespread nor as thick as the Wallace and Siyeh but are not uniform and may be susceptible of subdivision into units of the rank of formations rather than mere members.

WALLACE FORMATION

DISTRIBUTION

The Wallace formation is exposed at intervals in Idaho from south of lat 47° to the Canadian border. It is also fairly widespread in Mineral, Sanders, and Lincoln Counties, Mont. Another mass that has been so designated (Ross, Andrews, and Witkind, 1955) extends, with interruptions, from the northeast corner of Lincoln County southeastward to the vicinity of Kalispell. This mass is so little known that it might best be referred to merely as belonging to the Piegan group. Farther south near the Clark Fork between Rivulet and Missoula the rocks of the Piegan group have been called Wallace (Ross, Andrews, and Witkind, 1955) but this decision was arbitrary. At one time all the rocks now called Wallace were given the name Newland (Calkins, 1909, p. 38).

CHARACTER

The Wallace formation is the most heterogeneous in the Coeur d'Alene region (Ransome and Calkins, 1908, p. 40–43). It includes thin-bedded fine-grained calcareous quartzite, impure limestone, and shale, largely calcareous and with slaty cleavage. All these rocks are marked by mud cracks, ripple marks, and related indications of deposition in shallow water. The conspicuous lamination in many beds is one of the diagnostic features of the formation. According to Calkins (Ransome and Calkins, 1908, p. 40–43) the lowermost part consists mainly of gray-green slaty rock, mostly calcareous or dolomitic. The middle part of the unit includes bluish and greenish argillite, limestone, and calcareous quartzite and presents a banded appearance in weathered outcrops because of the alteration of white quartzite, yellow-weathered limestone, and greenish and bluish argillite. In the uppermost part of the formation argillite predominates, and the white beds of calcareous quartzite prominent in the middle part are absent. The argillite in the upper member is especially distinctly laminated. Clusters of weathered siderite grains such as are prominent in the formation below are present in argillaceous beds in the Wallace but absent in the quartzitic beds. Calkins thought most of the carbonate in the formation was distinctly magnesian but reported some calcite. Slaty cleavage is better developed in the lower than in the higher beds, and Calkins felt that this could not be explained entirely on lithologic grounds. The composition of the rocks of the Wallace formation, as seen under the microscope, is very similar to that of rocks already described. Quartz, feldspar, sericite, muscovite, chlorite, carbonate, and such minor constituents as zircon, tourmaline, magnetite, leucoxene, and carbonaceous dust are reported. The calcareous quartzite of the middle member is arkosic.

In five specimens from the Wallace formation in eastern Shoshone County furnished by S. Warren Hobbs no carbonate is present. The specimens include argillite and quartzitic rocks, mostly with feldspar. They are conspicuously laminated in different shades of gray. The component grains range from 0.01 to 0.10 mm in diameter.

Calkins emphasizes the difficulty in measuring the thickness of the Wallace both because it is broken by many faults and because the argillite is overthickened because of minute folds. He concludes that the formation can hardly be less than 4,000 feet thick. Umpleby and Jones (1923, p. 8) estimate 4,000 to 6,000 feet. Anderson (1930b, p. 18) says the thickness north of the Coeur d'Alene region is more than 4,000 feet. Shenon and McConnel (1939, p. 5) say that locally the Wallace can be divided into four members. They estimate the total thickness in the Coeur d'Alene area as between 4,500 and 6,000 feet.

Between the southern Coeur d'Alene region and the St. Joe River five subdivisions of the Wallace formation have been mapped (Wagner, 1949, p. 12–13, pl. 1). The lowest is a heterogeneous sequence of thin-bedded light-gray to dark-gray laminated impure and mostly calcareous quartzite, shale, and sandstone with some limestone. The next consists of alternating calcareous or magnesian sandstone, with a few limestone beds, and black mostly noncalcareous shale. The purer limestone beds in this unit show molar-tooth structure. Weathered surfaces of the unit are conspicuously banded. The third member consists largely of noncalcareous black shale, in part metamorphosed to phyllite or slate. The fourth member is thin-bedded light-gray-green to dark-gray sandy, calcareous, or dolomitic argillite. The uppermost member is largely black noncalcareous shale or argillite with paper-thin laminae. These beds grade upward into more arenaceous rocks in the transition zone at the base of the Striped Peak formation. Wagner estimates the entire thickness of the formation as less than 4,500 feet.

In Kootenai County (Anderson, 1940, p. 12–13) the Wallace has three principal members much like those described by Calkins for the Coeur d'Alene region and is estimated to be not less than 5,000 feet thick. In the area in Kootenai and Benewah Counties studied by A. B. Campbell (1950, written communication) the Wallace formation is much as it is where described by Wagner in the area immediately to the east. It is, however, composed only of the lower three of the five members described by Wagner.

Near Mullan, Idaho (Hobbs, Wallace, and Griggs, 1950, p. 5–8), the part of the Wallace formation studied is thought to correspond to the two lowest members described by Wagner and to the lowest of the four described by Shenon and McConnel. In the Mullan area the rocks are very fine grained quartzite and argillite and intermediate rocks. Ankerite and related carbonates are abundant throughout. Limestone is scarce. In most carbonate-bearing beds the relatively pure carbonate occupies the cores of aggregates, 1 to 2 inches in diameter, that form layers along one or more bedding planes. The part of the formation exposed in the Mullan area is approximately 2,600 feet thick.

In the Libby and Trout Creek quadrangles (pl. 2) (Gibson, 1948, p. 13-16; Gibson, Jenks, and Campbell, 1941, p. 371–374) the Wallace formation is widespread, thick, and, as usual, lithologically heterogeneous. The principal component is gray to greenish-gray sandy argillite, largely calcareous or dolomitic and slightly ferruginous, with intercalated thick- to thin-bedded gray, greenish-gray, and brownish calcareous sandstone. A very thinly laminated soft sericitic light-gray to buff slightly calcareous shale occurs at several horizons. This shale is readily recognized whereas some of the other components of the Wallace might be confused with parts of other formations.

Stromatolitic dolomite or dolomitic limestone is reported to be the most characteristic rock in the Wallace formation of the Libby area and is present at several horizons there. This rock is light gray, pale greenish gray, or nearly white on fresh fracture. The unweathered rock is seemingly homogeneous, without visible lamination or hint of organic origin. Where weathered, the rock is dark gray or buff, and at the exposed surface, brown of various shades. On this surface curved laminae, with the shapes typical of stromatolites, are emphasized through slight differences in color, hardness, and mineral composition. Richard Rezak, after his studies of stromatolites in Glacier National

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Park, found many similar fossils in the Libby area. As noted earlier it is not now possible to use stromatolites as index fossils over distances comparable to those that separate Glacier National Park and Libby, but this remains among the possibilities to be borne in mind. The carbonate rocks in the Wallace in areas south of Libby have molar-tooth structures (Gibson, Jenks, and Campbell, 1941, p. 373). Some beds near Libby are oolitic (fig. 3*D*).

Gibson locally included some dominantly red beds with especially well formed ripple marks in the Wallace formation near Libby, noting that these beds resemble the overlying Striped Peak formation, here regarded as a component of the Missoula group. Presumably Gibson's red beds in the Wallace are approximately analogous to Clapp's (1932) "red band in the Siyeh." According to the definitions of this report, Gibson's red beds in the Wallace and any carbonate beds above them would be regarded as belonging to the Missoula group. Gibson notes that the Wallace formation in the Libby area is thick, possibly reaching a maximum of 17,000 feet. The red beds he thought of as in the upper third of the formation are a few hundred to as much as 1,000 feet thick, but it is not clear how much of the total thickness is stratigraphically above the red-bed horizon and therefore presumably to be correlated with the Missoula group as here used. Gibson indicates that the Wallace formation thins southward to only 7,000 feet in the southern Trout Creek quadrangle.

To the southeast, in Mineral County, Mont., the Wallace formation is widespread (Wallace and Hosterman, 1956, p. 19–21). It is similar to the equivalent rocks near Mullan, Idaho, except for a somewhat greater degree of metamorphism. In Mineral County the rocks are termed phyllite, crystalline limestone, and limy quartzite. Four members were noted but not mapped separately. The thickness in the western part of the county appears to be at least 6,500 feet, with the top missing.

In the St. Regis-Superior area (Campbell, 1960, p. 557–560) the Wallace formation comprises at least two lithologic units. The lower one is composed of 6,000 feet of thin-bedded dark-gray argillite or phyllite interbedded with thin-bedded light-gray quartzite, all more or less dolomitic. Impure limestone and dolomite are interspersed throughout the unit, and show molar-tooth structures locally. The upper unit is about 4,000 feet thick and includes 65–70 percent of thin-bedded light-gray quartzite that is slightly sericitic and in part dolomitic. Black argillite accounts for 10–15 percent of the unit. It is interbedded with light-gray quartzite to form a conspicuously layered rock, characteristic of the upper part of the Wallace formation. The remaining 20–25 percent of the upper unit is composed of beds of medium- to dark-gray silty dolomitic limestone to limy siltstone.

Little information is available in regard to rocks commonly called Wallace east of Superior. Near and east of Missoula their approximate stratigraphic equivalents are termed Newland formation or limestone and, to the north, Siyeh limestone. They are described under these names.

NEWLAND LIMESTONE

DISTRIBUTION

Comparatively small parts of the Piegan group are now termed "Newland limestone." The only masses incontrovertibly so named are in and near the Big and Little Belt Mountains. For some distance westward the only components of the Belt series exposed are post-Newland. Near Philipsburg and Missoula the name Newland has been applied in the past and is retained in the present report for calcareous rocks stratigraphically beneath those here assigned to the Missoula group. Near Philipsburg the unit has been called Newland formation rather than Newland limestone (Calkins and Emmons, 1915, p. 4).

CHARACTER

Near Missoula few details of the western exposures of the Newland limestone are available. Clapp and Deiss (1931, p. 693) speak of the unit near Missoula as the "Lower Wallace limestone" and, in their figure 3, "Correlation table of the Belt rocks of Montana," list it as Wallace limestone 7,800 feet thick, not divided in the table into upper and lower components. The reasons for their usage have already been discussed. Langton (1935, p. 36) cites the unit in a table as the Newland, 4,000± feet thick, comprising

> impure argillitic, dolomitic, and sideritic limestones, with some blue, thin-bedded argillites grading downward into more massive-bedded gray quartzitic argillites commonly ripple-marked and with layers of reddish-gray quartzites.

The chemical composition of a specimen obtained from the east side of Rock Creek just outside of the area mapped by Langton is shown in table 20. It is a highly impure magnesian limestone, with irregular structures that may be stromatolitic. Similar structures are common in limestone near Missoula. The stromatolite heads, however, are not as distinctly formed as those in Glacier National Park. The rock analyzed and others nearby consist mainly of carbonate and quartz with some plagioclase and microcline, with lesser amounts of muscovite, chlorite, zircon, epidote, and rutile, of which the last three appear to be detrital. A little clay may be present. Grain diameters range from 0.005 to 0.06 mm. Some of the feldspar has crystal form but with rounded and broken corners and edges.

Near Philipsburg (Calkins and Emmons, 1915, p. 4; Emmons and Calkins, 1913, p. 41–45) the unit is called the Newland formation. Calkins entertained the possibility that the Newland as mapped by him includes the Greyson. This suggestion appears to have been made in an attempt to bring the stratigraphic succession near Philipsburg into harmony with that in the Big Belt Mountains. As the two localities are some 80 miles apart this does not now seem necessary. Calkins says that, in the Philipsburg quadrangle, the basal part of the Newland is dark argillaceous rock with thin beds of impure limestone, constituting a transition zone with the Ravalli formation below. Most of the Newland consists of fine-grained thin-bedded slabby calcareous rock, with a little interbedded calcareous quartzite or sandstone. The uppermost few hundred feet of the formation consists of drab- to greenish-gray, fissile calcareous shale. The total thickness of the formation is about 4,000 feet. In the second publication on the Philipsburg area, Calkins notes that the formation, while distinctly more calcareous than those above and below, "as a whole cannot very accurately be called limestone." The analysis quoted in table 20 shows that carbonates constitute much less than half the rock. However, Calkins says that as viewed under the microscope the rocks of the Newland formation consist mainly of quartz and somewhat more abundant carbonate, thus an impure limestone. Feldspar can generally be identified in these rocks. Sericite is invariably present. Biotite, chlorite, zircon, carbonaceous dust, and tourmaline are subordinate constituents. The grains are measurable in hundredths or thousandths of a millimeter. In the Philipsburg area the Newland formation has undergone contact metamorphism locally.

Newland limestone is exposed northeast of Philipsburg, particularly along the upper Blackfoot River, but available details are meager. Tweto (p. 41) has described a rock from this formation near the upper reaches of the Blackfoot River, east of Ovando, Powell County, as a pebbly, oolitic, siliceous limestone. It consists essentially of calcite and quartz with minor amounts of dolomite, ankerite, feldspar, magnetite, and argillaceous matter. The oolites are deformed, and the concentric structure is poorly preserved. They are partly or wholly surrounded by coronas of comb-structured quartz. The matrix in which the oolites lie is argillaceous limestone with abundant quartz. The quartz is in part secondary and in part in silt-size clastic grains.

Farther east and south in the Little and Big Belt Mountains, exposures are somewhat larger and information is more complete. The following summary is based largely on work in the Canyon Ferry quadrangle (Mertie, Fischer, and Hobbs, 1951) supplemented by personal observations.

In the Big Belt Mountains the Newland limestone is reported to be a uniformly dense dark-gray dolomitic limestone, thinly and evenly bedded, except in its upper part. Much of it is more impure than this description suggests. Material from a gulch that reaches Lake Sewall near Canyon Ferry is only about half carbonate, the rest of the rock being mainly quartz plus some feldspar and mica. A little illite and probably montmorillonite are present. Much of it is in grains as much as 0.03 mm but a few elongate feldspar grains are more than 0.10 mm long. As viewed under a microscope this rock is seen to be indistinctly laminated. The limestone forms abundant float of brown platy fragments that superficially resemble a hard fissile shale. This feature is more conspicuous in the Big Belt Mountains, especially their western flank, than in any other locality familiar to the author. Thin beds of limestone and shale alternate in the upper part of the Newland, constituting a transition zone between that formation and the overlying Greyson shale. The Newland has been estimated to be about 2,000 feet thick in the Big Belt Mountains (Mertie, Fischer, and Hobbs, 1951, p. 18; Walcott, 1899, p. 206).

Pardee and Schrader (1933, p. 124–125) described the unit in much the same area as consisting mainly of fine-grained to dense dark-bluish-gray beds that weather buff or yellow. The bedding is closely spaced and commonly obscured by slaty cleavage, which is locally very pronounced. The thickness is estimated as 4,500 feet.

In the Little Belt Mountains the Newland limestone has been studied in some detail near Neihart (Weed, 1900, p. 282–283). The lower part consists of alternating limestone and shale with the amount of limestone increasing upward. The central part consists of massive, dense, dark-blue limestone with streaks and irregular markings of white calcite, and carbonaceous stains. The rock weathers buff. Near Neihart the formation is about 560 feet thick but along Newland Creek, the type locality, the thickness is reported to be much greater.

Here Tweto (p. 41) describes the rock as faded black rusty-weathering thick-bedded fine-grained argillaceous limestone. The principal component is calcite, but quartz, feldspar, graphite, sericite, muscovite, garnet, and chlorite are also present. The limestone is microcrystalline and contains rare silt-size clastic grains, which are presumably composed of quartz and feldspar. A specimen collected by Richard Rezak is medium-gray massive limestone. It consists of an even-granular aggregate of carbonate (mostly calcite) with some quartz. The grains are about 0.005 mm in average diameter.

Exposures of the Belt series east and northeast of Logan have been described by Verrall (1955, p. 15–30), who correlated much of the sequence with the Newland limestone and part with the Chamberlain shale. The beds he regarded as correlative with the Chamberlain include buff and gray argillite with minor amounts of reddish argillite, greenish coarse-grained micaceous arkose, and gray dense thin-bedded limestone, totaling more than 3,388 feet. As Verrall noted, this assemblage differs markedly in character and thickness from the Chamberlain shale of the type locality more than 60 miles to the north. Above this unit Verrall reported supposed Newland limestone made up of greenish-brown coarse-grained micaceous arkose, interbedded with buff, pink, and gray beds of argillite and medium- to thin-bedded dense limestone with layers of dense black calcareous concretions and some dolomitized limestone with features resembling molar-tooth structures. At the base is a stromatolite reef, that Verrall regarded as similar to the *Collenia symmetrica* zone of the Fentons (1937b, p. 1894–1895, 1942–1947), a suggestion of doubtful validity in the present state of knowledge relative to stromatolites as stratigraphic indicators. The thickness of the supposed Newland limestone aggregates over 1,560 feet. Klemme (1949, p. 6–7) had previously regarded the unit as Spokane shale, an assignment now in serious doubt as Robinson (1963) has found that the Spokane pinches out west of Verrall's area. All of the beds described by Verrall include so much arkose that they are regarded here (pl. 1) as belonging to the North Boulder group; presumably stratigraphically low in that group.

SIYEH LIMESTONE

DISTRIBUTION

The Siyeh limestone of the Piegan group is recognized in the Lewis and Livingstone Ranges of Glacier National Park, and the Flathead and Swan Ranges farther south. Carbonate rocks both west and south of these ranges are so similar to the Siyeh that they are likely to be assigned to that formation when further work is done. On plate 1 these rocks are indicated merely as belonging to the Piegan group. However, the term "Siyeh" has been applied as far south as the north side of the Blackfoot River, a short distance south of lat 47° (Clapp and Deiss, 1931, p. 683–684).

CHARACTER

In the Glacier National Park region (Ross, 1959) the limestone differs in many details from place to place, sufficiently so that local subdivisions will be required when detailed studies are undertaken. Lateral variation is so marked that many of the subdivisions will prove to be valid only over a few score or a few hundred square miles. Within Glacier National Park the Siyeh limestone is mainly an impure and somewhat magnesian limestone, as indicated by the analyses in table 20 and table 24. Some strata are more argillaceous than those represented in the two tables and thin beds of argillite exist. However, the Siyeh limestone as limited in the present report (p. 28–29) contains no assemblages of argillite or of distinctly argillaceous limestone.

Among the subdivisions of the Siyeh limestone that may be expected are the three suggested by Rezak (1957) on the basis of stromatolite content. His lowest subdivision or zone in the Siyeh is the *Collenia symmetrica* zone 1. This zone constitutes the lower two-thirds of the formation and was divided into two unmapped members by the Fentons (1937b, p. 1892–1897). Rezak's second zone is called *Conophyton* zone 1 and corresponds essentially to the *Collenia frequens* zone of the Fentons. This zone forms massive gray ledges that were noted by the men of M. R. Campbell's parties even before they came to realize the probable organic nature of the material. It is the most conspicuous stromatolite zone in Glacier National Park and the only one that has been mapped (Ross, 1959). This zone has three components, of which the lowest is characterized by *Collenia frequens* Walcott, the middle part by *Conophyton inclinatum* Rezak, and the uppermost by *Collenia multiflabella* Rezak and *Cryptozoon occidentale* Dawson. The colonies of *Conophyton inclinatum* in the middle part are the most conspicuous features of the zones and are identical with those previously called *Collenia frequens* in this same zone. Stromatolite names in the present report are those of Rezak (1957) and in many instances differ drastically from those of previous writers.

Nearly all the Siyeh limestone is thick bedded or massive as viewed from a distance but close examination shows it to be thinly laminated. Some is oolitic. Fresh surfaces are dusky blue, or more rarely greenish gray, with variations related to the composition of the rock. Weathered surfaces show various orange and brownish tones and commonly have irregular etched marking that corresponds to differences in the calcium carbonate content of the rock. They include the forms termed "molar-tooth" structures by Daly (1912, p. 72–76) but with infinite variety in the details so that many have no resemblance to molar-tooth markings. The rock consists of carbonate, and quartz with some sericite. In most beds individual grains of the major constituents are a few thousandths to a few hundredths of a millimeter in diameter, and accessory minerals are in such minute grains and small amounts that they were not determined. Microscopic irregularities in texture, related to the molar-tooth and similar structures so conspicuous in outcrops, are plentiful (fig. 4*A* and *C*). Well-defined oolites are not common, but many of the rocks seem to contain recrystallized oolites. Where oolites are preserved some are roughly elliptical in section, 1 to 2 mm long, and retain no internal structure. Other oolites are circular in section and have well-preserved concentric structure within them (fig. 3*C*). The two varieties may occur in the same rock. Some of the round oolites are broken and others are invaded in their outer layers by grains of clastic quartz. Most of the round oolites are less than a millimeter in diameter and are themselves components of pebblelike masses embedded in the fine-grained limestone (fig. 3*B*). While the predominant constituent of the Siyeh limestone is carbonate, cryptocrystalline silica is common, clastic grains of quartz are discernible locally and some strata contain minerals of micaceous habit. The last mentioned are largely sericite but include some montmorillonite and illite.

Stromatolites are present throughout the Siyeh limestone of Glacier National Park and may be more abundant and well preserved there than in other parts of western Montana. However, they have been found in all exposures of the Piegan group in which they have been searched for and therefore may be more widespread and abundant than is now realized.

MISSOULA GROUP

In the present paper the definition of the Missoula group has been more drastically revised than those of other groups of the Belt series. As a consequence the group extends lower stratigraphically than earlier usage would imply. The change results largely from work in the Glacier National Park region (Ross, 1959; Ross and Rezak, 1959, p. 436). The group is more widespread than might have been supposed from earlier reports, a fact that seems implied in the paper by Clapp and Deiss (1931). The best known exposures specifically assigned to the Missoula group extend from Missoula northeastward, but broadly equivalent rocks are known from northern Idaho eastward into the Little Belt Mountains and from near Dillon to beyond the Canadian boundary. The group, as now defined, includes a large part of the Belt series in the Big and Little Belt Mountains, where the series name originated. The revision of the definition transfers much of the material in these mountains that was included in the Piegan group on the Montana State Geologic Map into the Missoula. The revised definition and the reasons for it were discussed on pages 29–30, but the essential features may be summed up in the statement that the Missoula group includes all the Belt series stratigraphically higher than the great thicknesses of carbonate-rich rocks of the Piegan group. Thus, the Missoula group is characterized by an abundance of red and green clastic rocks, although other rocks, including some limestone, occur within it.

STRIPED PEAK FORMATION

DISTRIBUTION

The Striped Peak formation was first named and is best known in the Coeur d'Alene region, Shoshone County, Idaho. In Shoshone County it occurs only in scattered remnants, being less widely distributed than any other formation in the Belt series (Umpleby and Jones, 1923, p. 9). It has been recognized farther west in Kootenai (Anderson, 1940, p. 13) and Bonner (Sampson, 1928, p. 8) Counties, Idaho. Strata that resemble the Striped Peak formation have been noted at intervals from the Coeur d'Alene region northward to the international boundary (Calkins, 1909, p. 38–39, 41). Apparently the localities are mainly in the northwestern part of Montana (Calkins, 1909, p. 41, pl. 1). No rocks that appear to correspond to the Striped Peak formation are reported in Boundary County, Idaho, (Kirkham and Ellis, 1926, p. 17, 18, 27). The unit has been mapped in the Libby and Trout Creek quadrangles (pl. 2), Montana (Gibson, 1948, p. 16–17, pl. 1; Gibson, Jenks, and Campbell, 1941, p. 374–376). Presumably rocks equivalent in age to the Striped Peak formation are elsewhere in Montana but, where they have been described at all, they are given other names. The distances between outcrops are such that direct correlation is impossible.

CHARACTER

In the Coeur d'Alene region (Ransome and Calkins, 1908, p. 44) the Striped Peak formation is composed of thin-bedded shale and quartzitic sandstone with ripple marks and sun cracks. Most of it is reddish purple and green. The thickness at the type locality is about 1,000 feet, with the top removed by erosion, here and in all exposures of the formation. Immediately to the south (Wagner, 1949, p. 13–16) the formation is reported to consist of alternating beds of pinkish-gray sandstone and greenish-gray shale in the lower 500 feet of the unit, and lavender sandstone and deep-purple shale above this. The argillaceous beds in the lower part are thicker bedded and show fewer shaly partings than those higher in the sequence. Ripple marks and mud cracks are present. The sandstone, or perhaps more properly the impure quartzite, consists of grains of quartz with sericite and a little feldspar embedded in a dark-gray fine-grained matrix. Minor amounts of tourmaline, zircon, magnetite, hematite, and chlorite are present. Some of the sandstone contains carbonate minerals. The maximum thickness reported is 2,000 feet.

Representatives of the formation collected by S. Warren Hobbs in eastern Shoshone County are grayish-purple to medium-gray impure quartzite. The rock is microscopically laminated. The principal component is quartz with subordinate amounts of alkalic plagioclase, microcline, and micropegmatite, and small subparallel flakes of colorless mica abundant in the narrower laminae and present between the clastic grains in the coarser laminae. Long flakes of bleached biotite are scattered throughout the rock. Hematite is sparsely distributed. In the coarser grained parts of the rock the quartz grains have diameters of 0.10 to 0.25 mm, although most in the finer grained parts average about 0.01 mm.

In Kootenai County (Anderson, 1940, p. 13), the components are similar but vary markedly in proportions at different localities. Also some of the shale is made up of alternating dark-gray to black and yellow layers, only small fractions of an inch in thickness. The lower part includes calcareous beds difficult to distinguish from the underlying Wallace formation. The thickness of the Striped Peak formation here may locally be as much as 4,000 feet. In southern Bonner County, immediately north of Kootenai County, the rocks assigned to the Striped Peak formation (Sampson, 1928, p. 9) appear to lack the characteristic reddish beds, perhaps as a result of contact metamorphism. They are reported to include an olive-green sandstone or quartzite and a laminated argillite with very thin alternating dark-gray and yellowish layers. Ripple marks and mud cracks are common. The exposed thickness is 9,000 feet.

Farther north, along the Clark Fork (Anderson, 1930a, p. 18–19), the formation has many reddish beds and thus resembles the same unit in the Coeur d'Alene region. Anderson notes that near Pend Oreille Lake the formation has lost its red color and is predominantly olive green like the color reported by Sampson farther south. Anderson found the formation to be more than 4,000 feet thick, but he notes that the characteristic reddish beds are absent in the upper part so that this thickness may include some beds that in Montana have been assigned to the Libby formation.

Near Libby, Mont. (Gibson, 1948, p. 9, 16–19), the Missoula group comprises only the Striped Peak and Libby formations. The former consists mainly of dark-red to purplish feldspathic sandstone and quartzite, locally shaly, and with some interbedded sandy shale. Most beds are about a foot thick. Ripple marks, sun cracks, and crossbedding are common. A few light-green to olive-green argillite beds are present, especially in the upper part of the formation. Some of them grade laterally into red beds. Some of the sandy beds contain a little carbonate, and stromatolitic dolomite occurs about 200 feet above the base of the formation. The thickness is about 2,000 feet, with a maximum of 2,500 feet at once place. Farther south in the Trout Creek quadrangle (Gibson and others, 1941, p. 374–377) all the rocks of the Missoula group have been included in the Striped Peak formation, similar in general to the Striped Peak in the Libby area. Either the Libby formation is absent or it has been mapped with the Striped Peak formation. The thickness is reported as 3,000 to 3,700 feet which is much greater than that near Libby. Oolitic dolomite and dolomitic limestone containing stromatolites are interbedded at two or more horizons.

LIBBY FORMATION

The term "Libby formation" has been applied in only two small areas northwest and southwest of Libby, Mont. (Gibson, 1948, p. 17–19). It consists largely of light-gray, dark-gray, and greenish-gray argillite, commonly sandy and less commonly sericitic and calcareous, in beds 1 to 3 feet thick. Some beds resemble the laminated, brown-weathering argillite in the Prichard formation. Thin-bedded gray sandstone and dark-gray oolitic, ferruginous, magnesian limestone and a few beds of white ferruginous dolomite, in part stromatolitic, are also present. Along the Kootenai River there is a little greenish-gray argillite with calcareous layers that contain spherical and podlike argillaceous nodules. The Libby formation is at least 6,000 feet thick.

BEDS OF THE MISSOULA GROUP NEAR SUPERIOR, MONT.

Beds of the Missoula group near Superior, Mont., comprise five formations (Campbell, 1960, p. 560–569). As already noted, some or all of the formations near Superior may be closely akin to formations close to Missoula.

The lowest formation of the Missoula group near Superior rests on the Wallace formation and has been named the Spruce formation by Campbell. It is predominantly an extremely fine grained argillaceous quartzite composed of alternating light- and dark-greenish-gray laminae 1/16 to 1/4 inch thick. The light-colored layers are so fine grained that they may be called siltstone. At intervals throughout the sequence pale-purplish-gray vitreous to subvitreous quartzite appears. Small grains of magnetite are present throughout. The siltstone and quartzite near the base of the formation are slightly calcareous. In the upper 500 to 600 feet of the unit many of the beds are slightly calcareous and a few show "pronounced calcite segregation structures." The total thickness is estimated to be 3,500 feet.

The formation consists of three ill-defined gradational units. The lowest part is grayish-yellow-green quartzose argillite interlaminated with argillaceous quartzite. The basal part is slightly dolomitic. The beds in the middle unit are thicker and more quartzitic than those in the lower unit and include many beds of pale-purplish-gray quartzite. Many of the strata contain enough magnetite to deflect a compass needle. The upper unit is similar to the lowest of the three and also to parts of the Wallace. It consists mainly of quartzite, largely dolomitic, and includes a few thin beds of siliceous limestone with segregation structures.

Above the Spruce formation is the Lupine quartzite. This consists mainly of pale-pinkish-brown (or gray) to grayish-red-purple fine- to medium-grained vitreous to subvitreous somewhat feldspathic quartzite. Thin, widely spaced dark-purple or red-purple laminae and partings occur at intervals throughout the quartzite. Thin fragments of dark-purple argillite are present in many of the quartzite beds. The more argillaceous beds show mud cracks. A feature considered diagnostic is the presence of pale-pink thin, elongate lenses of dolomitic quartzite within some of the beds of noncalcareous quartzite. The Lupine quartzite is 2,750 feet thick.

The next unit upward is the Sloway formation, which is quartzitic near the bottom and becomes progressively more argillaceous upward. Some beds are rich in hematite. Most of the beds low in the sequence are light-purple or light-green medium-bedded fine-grained to very fine grained argillaceous quartzite with some massive pink or white vitreous quartzite and numerous dark laminae and thin beds of reddish-purple argillite with micaceous partings between the quartzite beds. Some of the quartzite is mottled with purple and green. Here and there calcareous nodules 6 to 12 inches long, of possible organic origin, were noted. Higher in the sequence the rocks include thin-bedded argillite or quartzose argillite with conspicuous red, red-purple, and green colors of medium chroma. Mud cracks are plentiful and some salt casts are present. Some beds are of dense chertlike green or, less commonly, purple argillite. Small-scale scour-and-fill structures are common. As the Sloway formation has been seen only in fault blocks its thickness cannot be given with accuracy, but it is thought to be fully 5,000 feet.

The Bouchard formation, as described by Campbell, immediately overlies the Sloway and is composed principally of interbedded micaceous quartzite and quartzose argillite. The prevailing colors are greenish gray with low chroma. Most of the rocks weather buff, olive gray or reddish brown. There are a few beds of white quartzite and others of dark fissile shale. Quartz and sericite predominate in the formation, but some feldspar and chlorite are present. The thickness is variable with a maximum of at least 4,000 feet. The differences in thickness are regarded by Campbell as resulting from erosion prior to deposition of quartzite of Middle Cambrian age. Numerous individual calcareous stromatolites are present in the Bouchard formation in sec. 23, T. 16 N., R. 26 W.

An unnamed quartzite, inferred by Campbell to be the highest part of the group in the Superior area, has been observed only in a single small fault block with neither top nor bottom exposed. It consists of thick-bedded medium- to coarse-grained vitreous pink, white, and reddish-purple slightly feldspathic quartzite, locally crossbedded. The observed thickness of 700 feet is obviously less than the original maximum.

Campbell (1959; 1960, p. 571–573) thinks that in the vicinity of Superior, Mont., the Flathead quartzite (Middle Cambrian) rests with apparent conformity on the Missoula group, but on different units in different places. Southeast of Superior, along the Clark Fork, fault blocks of strata regarded as of Cambrian age rest against masses of reddish rocks that are assigned to the Missoula group. The area was visited by the writer during the reconnaissance for the geologic map of Montana (Ross, Andrews, and Witkind, 1955). The rocks mapped as Cambrian are all rich in carbonate, largely dolomite. Langton (1935, p. 34–35) notes the presence of petroliferous dolomitic limestone disconformably on white siliceous limestone with some chert and shale. Those nearby mapped as components of the Missoula group include much pinkish quartzite, perhaps correlative with the Pilcher quartzite, which is described below. If any quartzite of Cambrian age is present in the area it has not been recognized in the hasty reconnaissance studies so far done.

ROCKS OF THE MISSOULA GROUP NEAR MISSOULA, MONT.

DISTRIBUTION

The subdivisions of the Missoula group established by Clapp and Deiss (1931, p. 677-688) and modified by Nelson (1959) and Nelson and Dobell (1961) are based on exposures in an area extending nearly 10 miles southeast of Missoula and eastward past Bonner, thence north and northeast along and near the lower reaches of the Blackfoot River. Similar subdivisions are present (Langton, 1935) from Missoula southward to beyond lat 46°40′.

CHARACTER

The table below shows the relations between the components of the Missoula group mapped by Nelson and Dobell (1959; 1961) and Nelson (1959) and those earlier described by Clapp and Deiss. The name Sheep Mountain formation used by Clapp and Deiss was found to be preoccupied and Pilcher quartzite was substituted. The Hellgate quartzite at its type locality was found to be such a local unit that it is best regarded as a member within the Miller Peak argillite. The mapping showed that another quartzitic unit (the Bonner) needed to be established higher in the sequence.

Langton (1935, p. 36, fig. 3) has mapped units south of Missoula that he speaks of as the Spokane and Helena formations, which he regards as stratigraphically above the Newland limestone. If they are actually above the Newland limestone, they belong to the Missolua group of the present report. Further work is needed on this point. From reconnaissance by the writer, from Langton's data and from Nelson's work it is believed that the rocks called Spokane and Helena by Langton, belong to Nelson's Miller Peak argillite and the underlying Newland limestone. Clapp and Deiss (1931, fig. 3) show their basal unit in the Missoula group at Missoula resting on the Wallace (Newland) limestone, and do not employ the term "Helena limestone" in the area. They do, however, regard the upper part of their Wallace as correlative with the Helena limestone (Clapp and Deiss, 1931, p. 69).

Langton (1935, p. 36) describes his Spokane as consisting of a lower part, 2,000 feet thick, of greenish-gray, in part calcareous argillite, with ripple marks, and an upper massive- to thin-bedded lavender quartzite of unstated thickness. Sample ID–14950 in table 17 is considered to represent the Spokane of Langton. It also resembles the Miller Peak argillite, being a laminated grayish-red-purple argillite with ripple-marked bedding planes and with inconspicuous, small mud spalls. It consists largely of interlocking quartz grains 0.03–0.06 mm in diameter, most of which are clearly recrystallized. Carbonate grains are irregularly distributed and may constitute 10 percent of the whole. Mica and chlorite are scattered throughout. A little zircon is present. Hematite is fairly abundant.

Langton (1935, p. 36) describes the rock he called Helena as thin-bedded argillaceous and sideritic limestone, locally dolomitic. The beds weather buff or cream and many show molar-tooth structure. Such a description would fit much of the Wallace, Newland, and Siyeh limestones of the surrounding region and does not accord at all closely with the Helena limestone of Helena. The sample supposed to represent Langton's Helena limestone, ID–14750 in table 20, is a decidedly impure carbonate rock not at all like the Helena limestone near Helena. Langton assigns a thickness of approximately 4,000 feet to his Helena, whereas the Helena limestone was originally described (Walcott, 1899, p. 199–215) as 2,400 feet thick and in parts of the Big Belt Mountains is far thinner (Mertie and others, 1951, p. 20). Thus, Langton's correlation is of doubtful validity.

The lowest unit in the Missoula area that is recognized by all as a component of the Missoula group is the Miller

Subdivisions of the Missoula group near Missoula

According to Nelson and Dobell—	According to Clapp and Deiss—
Pilcher quartzite	Sheep Mountain quartzite
Garnet Range quartzite	Garnet Range formation
McNamara argillite	McNamara formation
Bonner quartzite	
Hellgate quartzite member	Hellgate formation
Miller Peak argillite	Miller Peak formation

Peak argillite. Nelson and Dobell (1961) have redefined the unit so named to include the Miller Peak and Hellgate formations of Clapp and Deiss and the lowest of the three members of their McNamara formation, because he found it impossible to trace the Hellgate formation throughout his map area.

The Miller Peak argillite of Nelson is composed of reddish and greenish silty argillite, and argillaceous quartzite. Silty argillite is somewhat more abundant than argillaceous quartzite, and reddish hues more common than greenish ones. Locally pale-red, tan, and light-gray quartzite and light-gray impure limestone are present. Much of the argillite, irrespective of color, is laminated. Some of the laminae are rich in newly crystallized very fine grained chlorite and sericite, giving them a characteristic sheen. Detrital mica, oriented parallel to bedding, occurs in varying amounts, especially at the tops of the finer, more argillaceous layers. Small-scale (as much as 20 mm amplitude) scour-and-fill structure is common. Mud cracks and ripple marks occur throughout the formation. Bedding planes in the quartzose rocks may be lined with thin layers of argillite, commonly containing ripple marks and mud cracks.

Nelson describes the lower few hundred feet of his Miller Peak argillite as light-greenish-gray laminated argillite. The laminae are 0.8 to 12 mm thick and locally show small-scale scour-and-fill structures. The rocks weather readily and have a light-yellowish-brown rind on weathered surfaces. The argillite is slightly calcareous locally and contains a few intercalated lenses of carbonate rock.

The Hellgate quartzite member is composed of grayish-orange-pink, pale-red, and grayish-yellow-green medium-grained somewhat argillaceous quartzite. The beds are 8 to 36 inches thick and are locally marked by thin layers of dark minerals, probably mostly hematite. Thin beds of reddish-gray argillite and siliceous argillite are scattered throughout the member. Two of the rocks Tweto (p. 41) studied are from the type locality of the Hellgate formation of Clapp and Deiss in the canyon of the Clark Fork near Missoula. One is green-gray massive dolomite and arkosic quartzite. It contains quartz, dolomite, sericite, and feldspar with the usual minor constituents. Sericite is the principal component of the groundmass, and the dolomite forms distinct grains and masses. The clastic grains are angular. About two-thirds are silt, and the remainder are the size of very fine sand. The other rock from this locality is lavender thinly laminated argillite. It contains dolomite, sericite, quartz, feldspar, and argillaceous material and minor constituents.

The Hellgate formation of Clapp and Deiss, according to Nelson, cannot be mapped far from the railroad cut near Missoula where it was first recognized. It differs from the rest of the Miller Peak argillite of Nelson and Dobell (1959; 1961) mainly in its higher ratio of quartzite to argillite. It is but one of several similar zones, probably all local, at various horizons with the Miller Peak argillite. Nelson and Dobell report no discernible differences between the rocks above and below the Hellgate unit. Thus, the Hellgate is best regarded as a member, mapped in one locality only, rather than as a formation.

According to Clapp and Deiss the lower 1,100 feet of their Miller Peak formation is overlain by 1,500 feet of mixed purple and green-gray sandy mud-cracked and ripple-marked argillite, interbedded with some massive argillaceous sandstone and a few thin beds of fine-grained purple-gray argillite. The purple beds weather to a dull red lavender and the greenish ones to a dull gray. Above these is 300 feet of massive- to thin-bedded argillaceous sandstone, becoming increasingly sandy upward.

Nelson and Dobell estimate the thickness of their redefined Miller Peak argillite as 5,200 feet, of which about 1,200 feet is represented by the Hellgate member where that has been mapped with 3,800 feet of beds below that member and 1,000 feet above it. Clapp and Deiss thought of their Hellgate formation as 2,200 feet thick and the Miller Peak argillite below it as 2,900 feet thick.

The descriptions of the Miller Peak and Hellgate formations by Clapp and Deiss (1931, p. 678–680) seem significantly different from the summaries by Nelson and Dobell (1961). Some of the differences may well result from lateral variations so that the two field parties did not pass over the same rocks. Clapp and Deiss say that the lower 1,100 feet of the Miller Peak formation is composed of deep red-purple sandy argillite with some siliceous sandy massive- to thin-bedded purple argillite and a few thin beds of fissile gray sandy mud-cracked argillite. Near the top of this lower part of the formation beds of micaceous argillite appear. Most of the rocks weather to a characteristic dull purple lavender. This description does not mention the greenish-gray beds Nelson and Dobell report at the base of the formation. Either Clapp and Deiss did not find these rocks along their lines of traverse or included them in the underlying limestone unit. Attention may be called to the fact that, as noted above, the Spruce formation of Campbell near Superior, 3,000 feet thick, closely resembles the green argillite beds, which Nelson and Dobell describe, in stratigraphic position and character.

Nelson and Dobell have introduced a new formation name, the Bonner quartzite, for the rocks immediately overlying their Miller Peak argillite. The rocks are somewhat arkosic, pink or red quartzite in beds 4 inches to 6 feet thick. Crossbedding is common and readily discernible because of slight color differences. Bedding planes throughout tend to be marked by indistinct light-green layers. Thin beds of red argillite are interspersed among the quartzite beds. The mineral grains in the quartzite average about a quarter of a millimeter but some are as much as 12 mm across. A little montmorillonite and illite is present.

Clapp and Deiss describe their McNamara formation as composed of three members. The lowest consists of drab-green-gray to purple and maroon micaceous and sandy argillite. The proportion of sandy material increases upward, and beds of massive sandy quartzite appear in the upper part of the member. This part of the original McNamara formation should correspond to the part of the Miller Peak argillite of Nelson and Dobell above the horizon of the Hellgate member. However, Clapp and Deiss credit it with a thickness of only 400 feet whereas the part of Nelson and Dobell's Miller Peak above the Hellgate member is about 1,000 feet thick.

The middle member of the McNamara formation of Clapp and Deiss corresponds approximately to the Bonner quartzite of Nelson and Dobell. Clapp and Deiss say the lower 40 feet consists of massive gray-green coarse-grained ripple-marked and crossbedded sandy quartzite. The rest of the member includes massive beds of pink-white and red-gray, pure coarse- to fine-grained quartzite. This rock is crossbedded and ripple marked and weathers to a dull rusty buff. Toward the top several thin beds of sandy argillite appear. Clapp and Deiss report the middle member of the McNamara formation, where measured, as 810 feet thick. Nelson and Dobell (1961) and Nelson (1959) agree with their figure for that locality but think that the best average thickness is 1,500 feet. Variations in thickness are due in part to original differences and in part to the results of deformation.

Nelson and Dobell propose to restrict the name McNamara argillite to include only the upper member of that formation as defined by Clapp and Deiss. They describe their restricted McNamara argillite as composed largely of reddish and greenish argillaceous siltstone and argillite with minor amounts of more quartzose rock, all similar to some of the rocks comprising the Miller Peak argillite except that they contain less detrital mica, and new chlorite is less coarsely recrystallized. A little illite is present. Some of the beds of green argillite in Nelson and Dobell's McNamara argillite contain thin layers and rounded bodies of dense sericite so fine grained as to resemble chert and probably represent metamorphosed clay layers and galls. Other bodies consisting of recrystallized, fine silt have a similar appearance. Nelson and Dobell estimate their McNamara argillite to be 4,000 feet thick.

Clapp and Deiss say the corresponding rocks consist mainly of bright green and red fine-grained mud-cracked and ripple-marked argillite. Dense, fine-grained, chertlike green argillite is irregularly interbedded with the more massive strata. Clay galls occur in the upper part of the unit. They measured the upper member of their McNamara formation as 1,780 feet thick, in contrast to the 4,000 feet estimated by Nelson and Dobell. Clapp and Deiss emphasize the gradational upper contact of the McNamara formation, which may account in part for the apparent discrepancy.

Tweto (p. 41) describes a rock from the McNamara argillite (as restricted by Nelson and Dobell) from the canyon of the Blackfoot River above Bonner. It is a dark-green, thin-bedded, ripple-marked chloritic and arkosic quartzite containing chlorite, quartz, feldspar, and the common accessories. The chlorite constitutes the cement and also masses resulting from the decomposition of mafic grains. The quartz grains are rounded to subangular and those of feldspar are subrounded to angular and poorly sorted. The grain size is 7 percent silt, 33 percent very fine sand, and 61 percent fine sand.

The Garnet Range quartzite is next above the McNamara. Nelson and Dobell say this formation is composed predominantly of greenish-gray quartzite, mixed with slightly less dark-greenish and dark-grayish argillite, shale, and sandy argillite, with a notable content of detrital mica, and a little montmorillonite. Where observed, the Garnet Range quartzite is intricately folded and contorted, in large part isoclinally. Crumpling of this sort is so much more marked in this formation than in the others as to constitute a characteristic feature and to interfere with estimates of thickness. Nelson and Dobell estimate the Garnet Range quartzite to be only 1,800 to 3,800 feet thick. Clapp and Deiss give an estimate of 7,600 feet but their measurement was in a different locality and one in which folding and faulting are so intricate that true estimates of thickness would be impossible.

Clapp and Deiss say that the lower 1,600 feet of their Garnet Range formation is made up of brown, green-gray to gray thin-bedded micaceous coarse-grained quartzite, with argillaceous and coarse-grained quartzite sandstone near the base. Above this member is 600 feet of black-gray to dark-blue-gray sandy micaceous argillite. Next in sucession is 1,000 feet of brown to green-gray argillaceous, micaceous sandstone and sandy quartzite, overlain in turn by 1,600 feet of thin-bedded brown-gray and green-gray quartzitic sandstone.

The next unit listed by Clapp and Deiss is composed of massive coarse-grained pink-white crossbedded pure quartzite about 300 feet thick. Nelson and Dobell interpret this assemblage as an infaulted block belonging to the Pilcher quartzite (the Sheep Mountain quartzite of Clapp and Deiss), the next unit above the Garnet Range quartzite. If the beds below the supposedly faulted quartzite block correspond to the entire Garnet Range quartzite, crumbling must have thickened them 1,000 feet to correspond to Nelson and Dobell's estimate of the Garnet Range quartzite as 3,800 feet. Clapp and Deiss report a thickness of 3,050 feet of brown, green-brown, and red-gray-brown micaceous and sandy thin-bedded, locally argillaceous, quartzite above the quartzite block. According to Nelson and Dobell's interpretation this would constitute a repetition of part of the beds below the quartzite block. The summary here given is taken from the text of the report by Clapp and Deiss. It does not agree, particularly in the thicknesses of the components, with the measured section tabulated in the same report. Perhaps the thicknesses cited in the text are without allowance for overthickening because of crumpling.

Clapp and Deiss (1931, p. 682) speak of a lens of limestone, 110 feet thick in the Garnet Range quartzite at one locality. Nelson and Dobell have found fossils of early Paleozoic age in this limestone. These fossils, and the fact that lithologic character of the limestone is closely similar to that of Middle Cambrian limestones in other parts of Montana, make a Middle Cambrian age probable. The rocks are decidedly dissimilar to the Devonian rocks of the region and all known pre-Devonian Paleozoic carbonate rocks in northwestern Montana are believed to be of Middle Cambrian age (Deiss, 1933, 1936, 1938) and not a part of the Belt series. Thus the assignment by Clapp and Deiss must be in error. The limestone is not, as they supposed, a lens in their Garnet Range quartzite. Instead it is a separate body of post-Belt strata.

The highest part of the Missoula group in the Missoula area was named the Sheep Mountain quartzite by Clapp and Deiss. As that name is preoccupied, Nelson and Dobell called the formation the Pilcher quartzite. They describe it as composed of coarse-grained, crossbedded, light-pink to dark-purplish-red vitreous quartzite. The beds range in thickness from an inch to 10 feet. The most striking feature is the crossbedding, rendered conspicuous by color contrasts. The bottom of one stratum may be moderate red and the top of the underlying stratum may be pale orange pink. Some beds, especially high in the sequence, are uniformly deep red and certain of these have spots and stringers bleached to a cream color. Near the top of the formation some beds are reddish black. Most strata contain 90 percent quartz. The remainder of the rock includes chlorite, sericite, and bits of quartzite, argillite, and other rocks. No feldspar was noted. The coloring matter is hematite dust. A few thin layers of reddish and greenish argillite occur in the uppermost 10 to 20 feet of the Pilcher quartzite. The formation is thought by Nelson and Dobell to vary in thickness in different localities but to average close to 1,000 feet. They found no place where the thickness was as great as the 2,300 feet reported by Clapp and Deiss for their Sheep Mountain quartzite, a fact that may be in part accounted for by differences in the limits placed on the Pilcher and Sheep Mountain units during the two investigations. Some question remains as to the character of the contact of the Pilcher quartzite with overlying shaly beds but Nelson and Dobell favor the concept that the contact is gradational.

Nelson and Dobell assign shale and limestone that are stratigraphically above the Pilcher quartzite to the Cambrian(?), in part on the evidence of poorly preserved fossils. Direct correlation with Cambrian formations in other parts of Montana is regarded by Nelson and Dobell as impossible on present evidence but they suggest a rough correspondence between quartzite at the base of their Cambrian(?) assemblage and the Flathead quartzite, and note resemblances between the shale and the Wolsey shale, and between the limestone and the Meagher limestone, well-known units of Middle Cambrian age in southwestern Montana. Nelson and Dobell also entertain the possibility that the Pilcher quartzite is correlatable with the Flathead quartzite. The major differnce between the two formations is that the Pilcher is much the thicker unit. If this interpretation should prove to be correct, the Pilcher quartzite, or part of it, would be of Cambrian rather than Precambrian age.

ROCKS OF THE BELT SERIES IN AND NEAR THE OVANDO AND SILVERTIP QUADRANGLES

DISTRIBUTION

In the broad area between Bonner and Helena and north as far as lat 48° rocks of the Belt series are abundant. Most belong to the Missoula group, but older units of the series are included, especially near Helmsville and Lincoln and in the Swan Range. The general distribution of the various units is shown on plate 1, based largely on unpublished maps by C. F. Deiss and C. H. Clapp made available through the courtesy of C. F. Deiss, coupled with published papers (1931; Deiss, 1935, 1943a, 1943b). Large areas of the Belt series, especially south of lat 47°, have as yet been mapped in reconnaissance only, but measured sections by Clapp and Deiss are available. Names for formations of the Missoula group in areas from near lat 47° to lat 48° have been proposed, and the distribution of these formations is shown so far as practicable on plate 1. Recent work near Bonner (Nelson and Dobell, 1961; Nelson, 1959) has resulted in minor modifications of the nomenclature of Clapp and Deiss there. This and the absence of published maps has made it impracticable to show on plate 1 all the named formations in the region here considered, or to fit them into the classification of the Belt series offered in the present paper. The principal difficulty in doing this arises from the fact that Clapp and Deiss carried some well-known names farther from their type localities than seems advisable.

The Miller Peak argillite, Cayuse limestone, Hoadley formation, and Ahorn quartzite of Clapp and Deiss are recorded in the Ovando, Coopers Lake, Silvertip, and Saypo quadrangles.

CHARACTER

For areas near and south of lat. 47° available lithologic descriptions are meager. In general the rocks have the characteristics of those near Bonner, already summarized. Tweto (p. 41) has described in detail many rocks from areas near this latitude. Of these, eight are regarded by him as belonging to the Miller Peak argillite of Clapp and Deiss. Six of them came from a ridge a couple of miles southwest of Cottonwood Lake in T. 16 N., R. 14 W. One of this group is described as light-blue-green laminated arkosic argillite containing quartz, orthoclase, plagioclase, and argillaceous matter with minor amounts of chlorite, sericite, calcite, muscovite, epidote, apatite, zircon, green biotite, magnetite, and sillimanite. Another rock is pinkish-gray massive coarse mud-cracked ferruginous quartzite with interlaminated thin flakes of dark-maroon argillite. It consists of quartz and hematite with minor amounts of orthoclase, sericite, chlorite, magnetite, titanite, apatite, zircon, and calcite. The original quartz grains were rounded and subrounded but addition of later quartz has produced an interpenetrant mosaic. Grain sizes are: silt size, 1 percent; very fine sand, 4 percent; fine sand, 21 percent; medium sand, 61 percent; and coarse sand, 19 percent. A third rock is a laminated light- and dark-maroon argillite composed of quartz, orthoclase, plagioclase, and hematite with minor quantities of muscovite, tourmaline, calcite, chlorite, sericite, magnetite, ilmanite, apatite, green and brown biotite, epitote grossularite, and zircon. The feldspar grains are much altered. Elongate hematite masses give a banded appearance to the thin section. Over 70 percent of the grains are silt size and the rest the size of very fine sand. Another of these rocks is a greenish-gray thin-bedded calcareous quartzite showing mud cracks and ripple marks. Thin laminae of white calcareous quartzite alternate with light- and dark-green argillite. The essential minerals are quartz, orthoclase, plagioclase, calcite, and the minor constituents include argillaceous material, muscovite, sericite, chlorite, magnetite, hematite, titanite, zircon, grossularite, rutile, and tourmaline. The feldspar grains are altered. Calcite is in grains in the quartz mosaic. Some of the chlorite aggregates may correspond to original mafic minerals. The grains are subrounded and subangular. In the coarser laminae 3 percent are silt-size, 15 percent the size of very fine sand, and the remainder that of fine sand. Another rock is a greenish-gray brown-weathering quartzitic argillite with laminae and flakes of dull purple micaceous argillite. The constituent minerals are similar to those in rocks described above. The component grains are angular and subangular and 74 percent of them are silt size, with the rest a little coarser. Still another rock from the same locality is light-green, buff-weathering thin-bedded argillite that consists mostly of argillaceous matter with a little quartz, sericite, feldspar, and sillimanite. The last of the group is a red and green calcareous quartzitic argillite with interspersed flakes and laminae of fine-grained ferruginous red and green argillite. The components are about the same as those listed above. The grains are poorly sorted and mostly subangular. They range in size from 60 percent silt through 30 percent very fine sand to 11 percent fine sand. Some specimens from the Miller Peak argillite contain small amounts of illite and montmorillonite but much of the rock has no detectable clay mineral.

The general description published by Deiss (1943a, p. 213, 215–216) shows that the Cayuse limestone includes dull gray, crystalline, tan-weathering dolomite interbedded with pale-maroon and greenish-gray, fissile argillite which grades upward into alternating zones of shaly calcareous argillite and sideritic and dolomitic marble. The upper part of the unit is blue and dull gray thick-bedded finely oolitic and coarsely crystalline algal limestone. The total thickness is about 1,000 feet.

Details have been provided by Tweto, (p. 41) who has described 14 specimens thought to belong to the Cayuse limestone, four that are from T. 16 N., R. 14 W., and the rest from T. 17 N., R. 14 W. One of the first four specimens is green arenaceous and calcareous argillite composed of calcite, quartz, argillaceous material, and minor amounts of orthoclase, microcline, plagioclase, muscovite, chlorite, sericite, biotite, graphite, dolomite, and epidote. It contains angular silt grains of quartz and feldspar in a matrix of calcite and argillaceous material. Another is a laminated light- and dark-green and buff quartzitic argillite. A third is a green calcareous argillite with much the same composition as the first two. The feldspar is much altered. The grain size includes 32 percent silt size, 47 percent very fine sand and 21 percent fine sand. The fourth rock is pink vitreous arkosic quartzite with sparse flakes of dull-maroon and purple laminated argillite. The rock consists of quartz, orthoclase, microcline, and plagioclase, with minor quantities of chlorite, biotite, hematite, ilmenite, graphite, epidote, titanite, apatite, tourmaline, zircon, and rutile. The grains are 65 percent silt size, 30 percent very fine sand, and the rest fine sand.

One specimen is light-green-gray impure cryptocrystalline laminated limestone. It consists dominantly of calcite but contains some pyrite, chlorite, muscovite, orthoclase, dolomite, plagioclase, quartz, and ilmenite. Another specimen is a dark-green thin-bedded argillaceous, calcareous quartzite of the usual composition. Chlorite is dominant in the groundmass. The feldspar grains are angular cleavage fragments, commonly much altered. The grain size is 45 percent silt, 24 percent very fine sand, and 30 percent fine sand. A third rock specimen is light-gray-green and green laminated calcareous quartzite with thin laminae of green fine-grained argillite. It contains 51 percent quartz, 38 percent feldspar, 9 percent calcite and some muscovite, sericite, zircon, biotite, chlorite, monazite, rutile, tourmaline, grossularite, apatite, magnetite and ilmenite(?). The grain size is 14 percent silt, 37 percent very fine sand, 41 percent fine sand, and 8 percent medium sand. Nearby a green impure dolomite was obtained. This consists mainly of dolomite with a little pyrite, quartz, sericite, and graphite. A fifth specimen is a light-gray-green vitreous quartzite with thin laminae of green argillite. It contains 57 percent quartz, 31 percent feldspar, and some calcite, chlorite, sericite, tourmaline, epidote, and siderite. The grain size is 20 percent fine sand, 43 percent medium sand, and 39 percent coarse sand. Another rock specimen is a green-gray, thin-bedded impure limestone consisting of calcite, quartz, orthoclase and plagioclase with some dolomite, muscovite, sericite, zircon, and chlorite. The rock is crossbedded on a microscopic scale. Another rock specimen is a light-green siliceous limestone, consisting of calcite, quartz, and feldspar with some dolomite, muscovite, zircon, rutile, sericite, argillaceous matter, chlorite, and graphite. Like the one just mentioned, it is microscopically crossbedded. An arkosic rock nearby contains quartz, orthoclase, plagioclase, perthite, and microcline, with the usual minor constituents. Half of the grains are silt size, the rest slightly larger. Another specimen is green-gray thin-banded impure limestone with structures that may be stromatolitic and veinlets of graphite bordered by argillaceous material.

The Cayuse limestone is one of the various limestone bodies intercalated in the Missoula group. Like many of the others it includes much clastic material. As already noted, some such units have been called Helena limestone, but the correlation thus implied rests on an insecure foundation. The unpublished map of the Ovando quadrangle by Clapp and Deiss shows a unit termed "Helena limestone" below the Miller Peak argillite and one called Spokane quartzite below that. In accord with the usage of Clapp and Deiss, Tweto (p. 41) assigned four of the specimens he described to the Helena limestone. One is from about 8 miles north of Ovando.

This rock is a dark-blue-gray oolitic limestone with scattered pebbles as much as 25 mm long. It consists of calcite and quartz with some dolomite, and argillaceous material. The oolites show a little concentric structure and all show radial structure emphasized by limonite. The centers commonly consist of limonite-stained dolomite. The oolites are elliptical with their long axes parallel. Each oolite is partly or wholly surrounded by a corona of prismatic quartz. Two other specimens are from north of the upper reaches of the Dearborn River. One is gray oolitic limestone consisting of calcite and some argillaceous cryptocrystalline limestone. The second rock specimen described is a portion of a stromatolite. Tweto speaks of this as a brown-gray argillaceous limestone. It consists of calcite and argillaceous material with rare collophane fragments and some limonite. It is finely laminated as viewed under the microscope. Another sample from north of Lincoln is far enough east to be included in rocks assigned to Helena limestone on plate 1, and is described as dark-blue-gray dense argillaceous limestone containing a little quartz and carbonaceous matter.

Deiss (1943a, p. 213, 216-218) has described the Hoadley formation and the Ahorn quartzite above his Cayuse limestone. The Hoadley formation is dark-greenish-gray and maroon thick-bedded siliceous argillite with some vitreous quartzite, overlain by red and buff arenaceous calcareous argillite and fine-grained sandstone that grades upward into calcareous sandstone that weathers greenish buff, pink sandstone, and argillite. The upper part of the formation is a brilliant dark-red soft thin- and thick-bedded platy ripple-marked sandstone, with some beds of pale-red argillite. The estimated thickness of the formation is 4,100 feet.

The Ahorn quartzite (Deiss, 1943a, p. 213, 217-218) is coarse- and fine-grained quartzite with scattered beds of red quartzite in the lower 1,700 feet, overlain by green and reddish-gray thin-bedded argillite and a few thin beds of fine-grained sandstone in the upper 400 feet of the formation. It is the uppermost formation of the Belt series in and near the Ovando quadrangle.

Rocks from near Ovando and presumably from the Hoadley and Ahorn formations, or nearby strata, have been described by Tweto (p. 41). Six of his specimens come from a ridge in T. 16 N., R. 14 W. Of these, one is a light-purple and maroon coarse vitreous quartzite consisting essentially of hematite, quartz, and feldspar, with minor amounts of magnetite, muscovite, chlorite, dolomite, green biotite, epidote, titanite, sericite, and zircon. The rock is laminated: the fine-grained laminae are rich in hematite and the coarser ones consist of a mosaic of angular grains of quartz and feldspar intergrown with abundant chlorite and hematite. Magnetite is common and is derived from biotite. Microcrystalline dolomite is also abundant. The grain size is 52 percent silt size, 40 percent very fine sand size, and 9 percent fine sand size. A second rock specimen is impure laminated maroon and pink dolomite with hematite laminae and with minor amounts of quartz, feldspar, tourmaline, chlorite, muscovite, and magnetite. Angular quartz and feldspar grains are abundant in a microcrystalline to cryptocrystalline groundmass of dolomite. The grain size is about 29 percent silt, 31 percent very fine sand, and 20 percent fine sand. Another rock is buff to salmon soft thin-bedded arkosic sandstone, containing quartz, feldspar, and sericite, with minor amounts of ilmenite, tourmaline, titanite, zircon, and muscovite. The cement is mainly sericite. The grain size is approximately 9 percent silt, 35 percent very fine sand, 30 percent fine sand, and 5 percent medium sand. A fourth rock specimen is tan-pink arkosic quartzite containing quartz, sericite, and feldspars, with the usual accessory minerals. The grains are sharply angular and are arranged with their long axes parallel. The feldspars are sericitized in varying degree. A fifth specimen is bright-red argillite containing flakes or laminae of argillaceous quartzite. It consists of hematite and sericite with minor amounts of quartz, feldspar, dolomite, magnetite, muscovite, and titanite. The grain size is about 33 percent silt, 46 percent very fine sand, and 21 percent fine sand. The sixth specimen rock described is buff-salmon, coarse, arkosic quartzite, consisting of quartz and feldspar with minor amounts of sericite, ilmenite, hematite, epidote, chert(?), and alteration products. The feldspar is fresher and the grains of quartz and feldspar are better rounded than in most specimens described. The grain size is about 7 percent silt, 3 percent very fine sand, 38 percent fine sand, 45 percent medium sand, and 10 percent coarse sand, so this is an exceptionally coarse rock by comparison with others of similar stratigraphic position.

Four specimens from about the same horizon as those just described but from T. 17 N., R. 14 W., were studied by Tweto (p. 41). One is laminated, maroon, purple, and light-green argillaceous dolomite with minor quantities of quartz, feldspar, muscovite, titanite, sericite, zircon, hematite, and magnetite. Another rock specimen from the same general locality is a laminated red, pink, and salmon vitreous dolomitic arkose. It contains dolomite, quartz, and feldspar with minor amounts of muscovite, hematite, magnetite, chlorite, tourmaline, and epidote. The dolomite constitutes the microcrystalline matrix of the rock. The feldspar, mostly orthoclase, is in sharply angular cleavage fragments. The grain size is 35 percent silt, 41 percent very fine sand, and 24 percent fine sand. A third rock specimen is thinly laminated red and pink mud-cracked dolomitic and arkosic quartzite. It contains dolomite, hematite, quartz, and feldspar with minor amounts of magnetite, muscovite, zircon, and chlorite. The dolomite forms a crystalline cement for angular, silt-size grains of quartz and feldspar. A fourth rock specimen is green chloritic graywacke whose essential minerals are quartz, chlorite, and feldspars. Subordinate constituents include muscovite, green biotite, epidote, glauconite, grossularite, titanite, and sericite. The reported presence of glauconite is of interest in a rock of this character and reported age.

ROCKS OF THE MISSOULA GROUP IN AND SOUTH OF GLACIER NATIONAL PARK

DISTRIBUTION

Only fragments of the Missoula group remain, mostly on the higher peaks, within Glacier National Park. West of the upper reaches of the Flathead River, it is extensively exposed, although little known. From the southern border of the park southward to the area near Missoula just described and east of Flathead Lake it forms broad expanses, especially on the northeastern flanks of the mountain ranges. It is, however, absent in the closely spaced ridges of the border of the mountains east of the North Fork of Sun River. The rocks of the Missoula group south of lat 48°00′ and in the general vicinity of Ovando and north of Lincoln have already been described.

CHARACTER

Only a few of the components of the Missoula group in this broad region are sufficiently well-known to be assigned formal names. Even some of the formations named by Willis (1902) are inadequately or incompletely mapped. The rocks of the Missoula group in this region are broadly similar to those near Missoula except that carbonate rocks similar lithologically but not stratigraphically to the Siyeh limestone of the Piegan group are more plentiful.

In and close to Glacier National Park the Purcell basalt and various sedimentary components of the Missoula group have been mapped (Ross, 1959). These include a stromatolite zone, an unnamed unit of greenish calcareous argillite, one of green and another of pinkish argillite, as well as several limestone masses. Each of these, and probably others are expected to become formations when detailed studies are undertaken in the region.

The greenish calcareous argillite locally distinguished at the base of the Missoula group near Glacier National Park is transitional with the limestone of the Piegan group below, but is included in the Missoula because it is dominantly argillaceous. Also the lithologic change is sharper at the base than at the top of this argillite. The unit is nearly continuous in the northern Swan Range, and has been noted in the Flathead Range and locally farther north although it thins in that direction. Where it is especially characteristic, the thickness is 500 to 800 feet. This unit may correspond somewhat to the basal part of the Miller Peak argillite near Missoula.

The Purcell basalt, that in places within the park is conspicuously intercalated in the lower part of the Missoula group, is an irregularly amygdaloidal, effusive basalt with pillow structure. Fresh surfaces are greenish gray to almost black, locally purplish. Weathered surfaces are stained brownish. The rock is too highly altered for precise mineral determination, but it originated as a basalt containing calcic plagioclase, augite, and some olivine. Traces of diabase texture remain. The feldspar now approximates oligoclase or albite-oligoclase but may originally have been labradorite. Much of the rock consists of fine-grained, nearly opaque aggregate of secondary minerals that include chlorite, sulfides, calcite, quartz, and possibly serpentine.

The basalt in the park is 100 to 275 feet thick and in different localities consists of two to eight or more flows. In some places argillite separates the flows.

The Shepard formation is at present recognized only within Glacier National Park, where it overlies the Purcell basalt. The principal component is fairly pure dolomite, but some beds are argillaceous and siliceous and a few are conglomeratic with pebbles of lava. The dolomite weathers to distinctive colors that range from pale yellowish brown to grayish orange. The matrix consists of dolomite grains about 0.01 mm in diameter, but embedded in this matrix are numerous oval masses as long as 4.0 mm that consist of carbonate grains as much as 0.2 mm in diameter. Perhaps these ovals are thoroughly recrystallized oolites (fig. 3*F*). Some beds contain stromatolites. Some illite and montmorillonite are present. The maximum reported thickness of the formation is 700 feet.

Unnamed lenticular bodies of carbonate rocks of mappable proportions are present in the Missoula group of the region at various horizons. Perhaps some of them may prove to be essentially equivalent stratigraphically to the Shepard formation. Most of them closely resemble the Siyeh limestone in appearance and composition except that they contain more argillaceous material especially in the thinner masses. Segregation structures of one kind or another are plentiful. Some are identical in appearance with the molar-tooth structure for which the Siyeh limestone is noted; others are so irregular as to bear no resemblance to the markings on teeth.

Stromatolite zones are fairly plentiful in the limestone bodies of the Missoula group (Rezak, 1957). One of these zones, approximately 6,000 feet above the base of the group, has been mapped locally as the *Conophyton* zone 2 (Ross, 1959, p. 54). This zone is about 100 feet thick, has five subdivisions and contains only two stromatolite species: *Conophyton inclinatum* and *Collenia frequens*. Another zone, about 500 feet lower in the section, is 35 feet thick. It contains heads of *Collenia symmetrica* as much as 5 feet high and 6 feet in diameter. *Cryptozoon occidentale* and *Collenia frequens* are also present.

The dimensions of the limestone masses vary widely. Here and there throughout the Missoula group they occupy a few inches or a few feet of strata. Most of the bodies that have been mapped are hundreds of feet in maximum thickness and one is as much as 3,000 feet thick.

These different units are mostly local features of the Missoula group in and near Glacier National Park. The main body of that group consists largely of argillite that is softer and less siliceous than most of the Grinnell formation. In most places purplish-red beds predominate, but yellow-green ones are rarely entirely absent and some beds are gray. Others are red to pink or, rarely, almost white. One mass of green argillite was mapped separately, but most are individually too small to be thus distinguished. Distinctly purplish rocks are almost entirely absent, a feature that aids in discriminating between argillite beds of Missoula age and those of the Grinnell. Quartzite, of varying degrees of purity, is intercalated in the argillite in many localities. In places, near the top of the group, the quartzite is the dominant rock. Ripple marks are very common, and mud cracks and intraformational conglomerate are fairly so.

Much of the argillite in the main body of the Missoula is composed of subangular grains with maximum dimensions of 0.05 mm or less. Some of the quartzite beds are made up of grains as much as 0.5 mm in diameter but most are distinctly finer grained. The components of the coarser beds are more perfectly rounded than those of the finer layers. Some of the rounded grains are coated with thin films of sericite. Some recrystallization has taken place, but most of the grains have retained their original, clastic shapes.

The principal mineral in all the beds is quartz, but most contain some feldspar, mainly rather alkalic plagioclase. Sericite is plentiful; chlorite is locally present; and hematite dust is diffused through most of the rocks. Clay minerals also occur. Carbonate, mostly calcite with some magnesium carbonate therein, is present in many of the beds and conspicuous in some where they weather rusty in color.

THE MISSOULA GROUP NEAR AND EAST OF HELENA

DISTRIBUTION

The Belt series was named from the mountains east of Helena where most of it belongs to formations here correlated with the Missoula group. These include the Greyson, Spokane, and Empire shales, the Helena limestone, the Marsh shale, and also the Greenhorn Mountain quartzite named by Knopf (1950, p. 839). These formations extend fully 50 miles north and south of Helena and occupy large areas in the Big Belt Mountains to the east of Helena. Essentially the same formations are widespread in the Little Belt Mountains, also, but mapping there is inadequate to delineate them. Some of the units extend at least 25 miles northwest of Helena also.

GREYSON SHALE

In the Little Belt Mountains the Greyson shale has been described (Weed, 1900, p. 282) as composed of dark-gray or black, fine- and coarse-grained siliceous shale. The lower part of the formation is pearly gray because of the abundant sericite. The upper part is more siliceous and includes beds of sandstone a foot thick. The formation is 955 feet thick.

In the Big Belt Mountains the Greyson is prominent, thick, and, judging by descriptions, is variable. According to the original description (Walcott, 1899, p. 206) the lower part is dark coarse siliceous and arenaceous shale that passes upward into bluish-gray almost fissile shale, resembling a poor quality of porcelain. This is succeeded upward by dark-gray siliceous and arenaceous shale with beds of buff sandy shale and sporadic layers of hard compact greenish-gray and drab siliceous rock. In the northern part of the Big Belt Mountains (Tweto, p. 41) the formation includes dark-blue-gray quartzitic argillite containing sericite, quartz, dolomite, graphite, feldspar, zircon, muscovite, and tourmaline. Along Deep Creek, east of Townsend, the base of the formation includes quartzite interbedded with shale and with a basal conglomerate, composed of sand and pebbles as much as 8 inches in diameter, derived from the underlying rocks of the Belt series. The formation is about 3,000 feet thick, and the coarse beds along Deep Creek aggregate 100 feet in thickness. These beds must record a local unconformity between the Newland limestone and the Greyson shale and may represent the North Boulder group of the present report or an isolated occurrence of similar rock. Walcott calls attention to the absence of the coarse beds at the base of the formation in the part of the Little Belt Mountains described by Weed, but the Fentons (1936, p. 610) report corresponding beds about 8 miles west of White Sulphur Springs. There they found the Greyson to consist of blue-green to olive-brown argillite and metaargillite, weathering to brown on shaly slopes, with, near the base, pink to brownish quartzite or arkosic sandstone of variable thickness and containing conglomerate lenses. Walcott (1899, p. 206, 235-238) reported trails and fragments of crustaceans in calcareous shale about 100 feet above the base of the Greyson shale both along Deep Creek in the Big Belt Mountains and near Neihart in the Little Belt Mountains. The Fentons (1936, p. 616) did not find the trails and note doubts as to the crustaceans. White (1929, p. 393) shared these doubts.

East of White Sulphur Springs the formation consists (Tanner, 1949, p. 12-14) predominantly of yellowish-gray to greenish-gray thin-bedded sericitic fissile shale with few beds of hard argillite and some lenses of soft calcite-cemented fine-grained sandstone. Immediately north of White Sulphur Springs thin beds of black limestone are common in the formation. The transition zone with the Spokane shale is 40 feet thick and contains buff sandy ripple-marked shale with lenses of deep red argillite and silty shale. The formation is somewhat over 2,000 feet thick.

According to Pardee and Schrader (1933, p. 125) along the Missouri River near the Spokane Hills, the formation is chiefly dark-gray shale that weathers rusty brown. It has markedly gradational contacts both at top and at bottom. The basal beds include much slate-gray shale in which some of the lighter colored beds are thickly peppered with brown specks regarded by Pardee as oxidized grains of siderite. Next above the gray beds is soft fissile light-gray to pale-buff shale whose layers have a shimmering pearl-gray coating. The rusty brown specks are abundant in these beds also. Commonly slaty cleavage is so conspicuous as to obscure the bedding. The middle part of the Greyson is mostly dark-gray shale, with which many thin quartzitic layers are interbedded. Above this is several hundred feet of dark-bluish-gray to brownish-gray thinly laminated shale that weathers brown. The transition zone at the top is 500 feet thick, but the dividing line with the overlying Spokane shale is so placed that practically all the red layers are included with the latter.

The Greyson in the Big Belt Mountains is thought by Pardee to be 3,000 feet thick. In one place the apparent thickness is 5,000 feet but repetition by faulting is possible.

Near Canyon Ferry, east of Helena, the Greyson shale differs in detail from that in the localities described previously. The formation there consists mainly of shale, siltstone, and fine-grained sandstone, all thinly and evenly bedded (Mertie, Fischer, and Hobbs, 1951, p. 18–19). These rocks are dark gray to dark brown where fresh but weather to somber shades of brown or red. Medium- to coarse-grained sandstone, locally conglomeratic, is present in the upper few hundred feet of the formation. These beds are dominantly light brown or light gray but partly white or red. Some are cross-bedded and ripple marked. The 200-foot transition zone between the Greyson and the overlying Spokane shale consists of red shale interbedded with light-colored sandstone. The contact was mapped at the top of the uppermost recognizable bed of sandstone. The thickness of the formation near Canyon Ferry is 2,000 to 3,000 feet.

One small exposure of rock supposed to belong to the Belt series is known in the Big Snowy Mountains in or near T. 11 N., R. 16 E., east of the area shown on plate 1. This was assigned to the Piegan group on the geologic map of Montana (Ross, Andrews, and Witkind, 1955) on the basis of the scanty descriptions on record (Calvert, W. R., quoted *in* Walcott, 1916b, p. 273–274; Reeves, 1931, p. 145; and Tweto, see p. 41). These rocks are mainly highly indurated limy shale, approximating a slate or, according to Tweto, shaly dolomite. The observed thickness is only 300 feet. Calvert suggested that the carbonate may have been derived from meteoric water percolating downward from Paleozoic limestone above. Tweto assigned the rock to the Greyson shale on the basis of petrographic character. With the limits of the groups adopted here, the most probable assignment of these isolated exposures is to the Missoula group rather than to the restricted Piegan group.

Near Marysville, north of Helena, (Barrell, 1907, p. 30) the Greyson shale consists of dark-gray to nearly black siliceous and arenaceous shale. Some portions are almost fissile; other consists of hard, cherty strata with cubical jointing. Deep-dark-red or purplish shaly beds are present locally.

South of Helena and east of the North Boulder River, the Greyson is plentiful (Klepper, Weeks, and Ruppel, 1957).

Thick uninterrupted sequences of rocks assigned to the Greyson are reported (G. D. Robinson, 1958, written communication) just east of Toston and in the northern Horseshoe Hills. At the northern end of the Horseshoe Hills near Sixteenmile Creek the formation is not less than 6,000 feet thick where it is overlain conformably by the Spokane shale. Farther north on Greyson Creek in the Duck Creek Pass quadrangle the Greyson measures 10,064 feet in thickness (W. H. Nelson, 1959, written communication). To the south the Spokane and the upper part of the Greyson are missing owing to pre-Flathead erosion, and at Big Davis Creek 10 miles south of Sixteenmile Creek the Greyson is only about 3,000 feet thick. East of Toston, where the top of the formation is present but the lower part missing, owing to thrust faulting, the formation appears to be much thicker, not less than 8,000 feet thick. Far more than half of the formation is dark-yellowish-brown and dark-olive-gray finely laminated platy argillite, partly derived from clay shale and partly from siliceous siltstone. In the basal 1,000 feet are scattered layers of coarse reddish-brown arkose and micaceous shale. At the top is more than 2,000 feet of lighter colored rocks, yellowish brown and yellowish gray, mainly argillite but with many thin beds of quartzite. Beginning about 3,500 feet below the top of the formation in the Horseshoe Hills, is a sequence, about 400 feet thick, of thin-bedded dark-gray limestone interbedded with dark platy argillite. Only a little limestone appears at a similar stratigraphic position east of Toston.

SPOKANE SHALE

Near Neihart in the Little Belt Mountains the Spokane is represented only by 210 feet of red shale (Weed, 1900, p. 283) and at one place in this area is missing entirely (Walcott, 1899, p. 207). Farther south in the Little Belt Mountains and also in the Big Belt Mountains the formation is widespread and conspicuous but available details are scanty (Walcott, 1899, p. 207; Pardee and Schrader, 1933, p. 125–126; Mertie, Fischer, and Hobbs, 1951, p. 19). It consists of siliceous, argillaceous, and arenaceous, noncalcareous shale, mostly deep purple red. Parts are altered to greenish in spots, layers, or along zones several tens of feet thick. The green material appears to be in part related to fractures or to igneous intrusions but locally may be original. Distinction between the Spokane and the overlying Empire shale is based mainly on the criterion of red versus green color. Insofar as the green may result from local alteration it is a weak stratigraphic criterion. Much of the Spokane shale in the Big Belt Mountains is soft and poorly bedded, but some has slaty cleavage. The thickness there is 1,500 to 2,000 feet.

North of Helena the Spokane shale is widespread and thick. Along Little Prickly Pear Creek, south of the settlement of Wolf Creek, the formation (Clapp and Deiss, 1931, p. 691) is 4,500 feet thick.

Tweto (p. 41) studied a rock from Arrastra Creek, west of Lincoln, that he, following Clapp and Deiss, regarded as a component of the Spokane shale. This rock is purple-gray, thin-bedded arkosic quartzite containing quartz and feldspar with the usual accessory minerals. The clastic grains are subangular and subrounded, in siliceous cement. The grain size is 22 percent silt, 63 percent very fine sand, and 15 percent fine sand.

Near White Sulphur Springs (Tanner, 1949, p. 14–17) the basal part of the formation is argillite with numerous lenses of thin-bedded silty shale and some fine-grained sandstone, all deep red to chocolate except for scattered streaks of bright green, which do not everywhere accord with the bedding. The next few hundred feet consists largely of argillite and shale in beds a fraction of an inch to several inches thick. Throughout this part there are thin zones of grayish shale indistinguishable from that of the Greyson. Near the top of the formation silty shale and fine-grained sandstone are more common than argillite. The shale and sandstone are thin bedded, ripple marked, and mud cracked. Locally there are thin beds of buff impure limestone and at one place a stromatolite reef was noted.

In the southern Big Belt Mountains the Spokane is reported (G. D. Robinson, 1958, written communication) to thin steadily from about 2,000 feet at Dry Creek in the southwest part of T. 6 N., R. 4 E., to 0 near Sixteenmile Creek, 9 miles to the south. Farther west, along the Missouri River, the Spokane seems to thin similarly southward, from something like 2,000 feet in secs. 12 and 13, T. 5 N., R. 3 E., northeast of Toston to no more than a few tens of feet 10 miles southwest in the Lombard Hills, but the situation is complicated by thrust-faulting.

In and near T. 6 N., R. 6 E., beds that have been correlated with the Spokane shale (Klemme, 1949, p. 6–7) are about 3,000 feet thick; in sec. 27, T. 4 N., R. 3 E., they are 800 feet thick; and in sec. 31, T. 5 N., R. 4 E., they are less than 100 feet. Where the strata correlated with the Spokane shale are thin they are underlain by reddish-brown argillite, pale-green shale, green-to-pale-brown quartzite and argillite, and some siliceous limestone, the whole being about 1,000 feet thick. This sequence is correlated (Klemme, 1949, p. 7) with the Greyson shale.

EMPIRE SHALE

In the Big Belt Mountains (Mertie and others, 1951, p. 19–20) the Empire shale is commonly a hard dense siliceous shale or argillite that is thinly bedded and laminated and includes some thin beds of limestone in its upper part. The shale is light to dark greenish gray, except that red shale similar to the Spokane has been mapped with the Empire in its lower part. The lower part of the Empire shale is so gradational and variable that the base cannot be fixed within an interval closer than 200 feet stratigraphically. The upper contact is also gradational but only within a range of 25 to 50 feet. The formation has a maximum thickness of about 1,000 feet but in places pinches out entirely.

In the general vicinity of Marysville, where the Empire shale received its name (Walcott, 1899, p. 207) most of the formation is entirely contact metamorphosed. There the unmetamorphosed portion has been described (Barrell, 1907, p. 31) as finely laminated, soft limy shale, grayish green or buff, with a few reddish bands. Its lower members include yellow and reddish sandstone, in part shaly and in part calcareous. Barrell measured 520 feet of the Empire shale cut off by a fault but thought the maximum thickness in the vicinity of Marysville was much greater.

The Empire shale is well exposed just northwest of Helena (Knopf, 1950, p. 837). There it contains pale-green argillite, brick-red, maroon, and lavender argillite beds identical with those in the underlying Spokane shale, and fine-grained light-green and white quartzite. Knopf regards the quartzite beds as the best means of distinguishing the Empire from the Spokane. He places the base of the Empire at the first place where quartzite appears. He regards the Empire as about 1,000 feet thick at this locality.

HELENA LIMESTONE

The Helena limestone covers large areas (pl. 1) from near Helena northwest a short distance past long. 112°30′ north of Little Prickly Pear Creek and has also been recognized in the Big Belt Mountains east of Helena. West of 112°30′, as already noted, the name has been used by some, but the correlations implied thereby do not seem warranted on the basis of existing data. Some of the carbonate bodies west of 112°30′ are shown on plate 1, without formation names.

Walcott (1899, p. 207) applied the name to rocks near Helena that consist of bluish-gray and gray thick-bedded limestone of varied purity. Weathered surfaces are colored in part buff, in part light gray. Irregular layers of broken oolitic and concretionary limestone occur at various horizons. Gray siliceous shale and greenish and purplish shale are intercalated in beds from half an inch to several feet thick. The formation at Helena was estimated by Walcott to be 2,400 feet thick.

In the Big Belt Mountains (Mertie and others, 1951, p. 20) relations to other formations justify assignment to the Helena limestone, but the rock differs in detail from that near Helena. It is composed of thinly but unevenly bedded, laminated fine-grained to dense dolomitic limestone. Thin partings of shaly limestone or limy shale are locally present. The maximum observed thickness is at least 500 feet, but the formation thins westward and disappears. Mertie and his coworkers attribute this to an erosional unconformity prior to deposition of the Flathead quartzite.

Observations by the writer show that the limestone is oolitic. Both near Helena and in the Big Belt Mountains it contains stromatolites. The rather wide variations in chemical content of the Helena limestone recorded in table 22 are, in part, related to the stromatolites or to apparently kindred features. These variations are reflected in various etched and differently colored features on weathered surfaces that give a distinctive appearance to the rock. The etched figures are reminiscent of the molar-tooth markings in the Siyeh limestone but are more broadly rounded, irregular, and of larger scale than these markings.

Knopf (1950, p. 837–838) notes the presence of siliceous oolites, edgewise conglomerate and stromatolites. He supports Barrell's (1907, p. 32) estimate of 4,000 feet for the thickness of the Helena near Marysville. Knopf also thinks that the composition of the formation justifies changing the name from Helena limestone to Helena dolomite. The proportions of dolomite and calcite in the formation vary from place to place, and the name change is not advocated here.

MARSH SHALE

Walcott (1899, p. 207–208) speaks of the Marsh shale as occurring near Helena and at intervals for some 20 miles to the north of the city. He describes the unit merely as reddish shale and gives thicknesses in different localities that range from 75 to 300 feet.

In the Marysville area the Marsh shale has been reported (Barrell, 1907, p. 33) to consist of red shale in which the lower beds are calcareous in some places and quartzitic in others. The thickness in that vicinity was estimated to exceed 1,000 feet.

Knopf (1950, p. 838) says deep-red argillite, in part maroon, purplish, and violet, constitutes much of the Marsh, but quartzite, in members as much as 500 feet thick, and cross-laminated siltstone are interbedded with the argillite. Some laminae contain casts of salt crystals. The top of the formation south of Marysville is deep-maroon mud-cracked and ripple-marked argillite with thin quartzite beds. The reddish argillite in the Marsh is regarded by Knopf as indistinguishable from the similar argillite in both the Spokane and Empire formations. He thinks that the most distinctive member of the Marsh is dark-red quartzite that weathers to a dark blackish red. Knopf reports that the Marsh shale thickens from 250 feet at Helena to 3,000 feet south of Marysville.

Tweto (p. 41) describes an impure mottled light-purple and light-green thin-bedded argillite from the Marsh shale in the city of Helena. This rock consists of quartz, feldspar, and minor amounts of calcite, chlorite, hematite, biotite, muscovite, limonite, sericite, epidote, and rutile. The feldspar grains are altered, and some of the chlorite masses have shapes that suggest they may have been derived from amphibole. The grain size is chiefly silt and the rest very fine sand.

GREENHORN MOUNTAIN QUARTZITE

The Greenhorn Mountain quartzite of Knopf (1950) is exposed on the mountain of that name south of Marysville and may extend as far west as Nevada Mountain (Eleanora B. Knopf, written communication). The Greenhorn Mountain quartzite has been grouped with Paleozoic rocks by earlier workers (Pardee and Schrader, 1933, pl. 2) and, hence, may be more widespread than is now known. The lower part is massive quartzite, in places showing festoon cross-lamination; the upper third is well stratified in beds 1 to 2 inches thick. The Greenhorn Mountain quartzite contains notable quantities of clear microcline, which distinguishes it from the nonfeldspathic Middle Cambrian Flathead quartzite. The Greenhorn Mountain quartzite is reported to be 1,800 feet thick, with the top bounded by the erosion surface on which the Flathead rests.

NORTH BOULDER GROUP

DISTRIBUTION

The rocks to which the name North Boulder group was originally applied (Ross, 1949, p. 113) are in T. 2 N., Rs. 3–4 W.; T. 1 N., Rs. 1–4 S.; T. 1 S., R. 4 W., and in and near T. 2 N., Rs. 3–6 E., and northward. Similar material is exposed east of Divide in T. 1 S., Rs. 7–8 W. (Sahinen, 1939, 1950). Rocks so coarse that they seem best regarded as belonging to the group are now known to occur west and south of Divide (written communication by P. A. Guttormsen, 1952; S. D. Theodosis, 1956; W. B. Myers, 1957). Other masses are expected to be recognized when further mapping is done in and near Jefferson and Gallatin Counties, and perhaps elsewhere in southwestern Montana.

CHARACTER

E. S. Perry (1950; written communication, 1947) reports that the lower part of the North Boulder group in its type locality consists of conglomerate with boulders, as much as 2 feet in diameter, of schist, gneiss, pegmatite, marble, and quartzite, typical of the Cherry Creek group (pre-Belt). That group is extensively exposed farther south. The conglomerate is estimated by Perry to be 2,000 to 4,000 feet thick . The rocks next above the conglomerate consist of well-bedded greenish to gray-green coarse sandstone (graywacke or arkose) composed of angular to subangular grains of quartz, feldspar, and other silicate minerals embedded in a dark matrix. These beds were estimated by Perry in 1947 to be 2,000 feet thick, but later in his guidebook Perry (1950, p. 40–43) stated that "the thickness is in the order of one mile." The beds are overlain by arkosic shale. The shale is thin bedded and in different parts is red, pink, purple, black, green, and gray, the last being dominant. The thickness of the shale unit varies from zero to about 1,000 feet. The Flathead quartzite (Middle Cambrian) high on the north side of the valley of the Jefferson River rests on the shaly beds. Field inspection by M. R. Klepper and the writer shows that the shaly unit is largely composed of beds that resemble the Greyson and Spokane shales farther north. A large expanse covered by Tertiary sedimentary rocks intervenes between the exposures of the Belt series near North Boulder River and those south of Helena to which formational names have been applied.

The beds in the area just cited have also been studied by Alexander (1955). He named them the La Hood formation, taking the name from a small settlement along the Jefferson River east of Whitehall. A composite section measured by Alexander in disconnected exposures in the SE¼ sec. 12, T. 1 N., R. 3 W., and SW¼ sec. 7, T. 1 N., R. 2 W., has three major subdivisions. The lowest consists of 1,121 feet of olive-gray to dark-greenish-gray arkosic sandstone with irregular lenses of conglomerate and minor amounts of silty shale. The largest conglomerate lens noted is 240 feet long and 30 feet thick. The conglomerate grades both laterally and vertically into sandstone. Many of the boulders in the thickest lenses are more than 2 feet in diameter, enclosed in a sandy matrix. Most of the boulders are well rounded but one lens contains angular fragments of light-colored quartz diorite gneiss 2½ feet long. The boulders consist of milky quartz, granite gneiss, garnet-hornblende-plagioclase schist, plagioclase, clear quartz, orthoclase, gabbro, diorite, and light-gray marble with tremolite needles. The middle unit in the measured section, 775 feet thick, consists of notably hard light-olive-gray sandstone with beds of silty shale as much as 2 feet thick in its lower part. The upper unit, 820 feet thick, consists mainly of soft-gray to brown sandstone with a few erratic boulders and a little silty shale. The sandstone beds throughout the section are similar and contain 15 to 40 percent quartz, 35 to 60 percent feldspar, 1 to 17 percent pyribole, 12 to 32 percent micaceous matrix, and such accessory constituents as hematite, apatite, garnet sphene, muscovite, calcite, zircon, and magnetite. The sand grains are poorly sorted and subangular to angular.

Throughout the unit as described by Alexander the bedding is regular in resistant beds and highly irregular in soft beds. In places sandstone lenses 8 feet thick may disappear within 50 feet laterally. The beds vary from a few inches to 20 feet in thickness. Crossbedding, ripple marks, and mud cracks are absent but evidence of intraformational scour is abundant.

From Alexander's description, the proportion of shaly beds in the North Boulder group appears to increase northward as a result of facies changes. About 10 to 15 miles to the northwest of this section Alexander records three units, which he regards as Greyson shale that are separated from each other and from the overlying Flathead quartzite by coarse beds such as characterize the North Boulder group. He does not mention the red shaly and sandy beds that resemble the Spokane shale. They crop out east of the lower reaches of the North Boulder River. Alexander could not measure the full thickness of the North Boulder group (his La Hood formation) but regarded it as between 2,700 and 5,000 feet with the larger figure approximating the total near Whitehall. Tansley and Schafer (1933, p. 11, 12), who worked mainly south of Whitehall, found similar rocks in the northern end of their area. They estimated the maximum total thickness as 5,000 feet, but the unit thinned abruptly to the east and south.

Four miles northwest of Sappington, coarse rocks here assigned to the North Boulder group have been reported (Berry, 1943, p. 5–8). They are dark-greenish-gray coarse arkosic sandstone, in part conglomeratic. The pebbles are of gneiss and white vein quartz. Micaceous shale is interbedded with the sandstone in the middle part of the sequence. The total recorded thickness is 4,000 feet.

The northern end of the Tobacco Root Mountains contains rocks whose character (as reported by J. V. Barnes, in 1954, in an unpublished report, "North end of the Tobacco Root Mountains, Madison County, Montana," for Indiana Univ.) warrants assignment to the North Boulder group. The rocks consist mainly of tan and gray conglomerate, arkosic sandstone, and argillite.

In an area of 3 square miles in T. 1 N., R. 1 W., which includes part of that described by Berry, and in a smaller area in sec. 5, T. 1 N., R. 1 E., Robinson (1963) has described rocks of the North Boulder group. These consist mostly of thick-bedded dark-yellowish-brown coarse micaceous, feldspathic sandstone. Some layers contain many flat pebbles. Thin laminae of micaceous olive-gray siltstone separate many of the sandstone beds, especially in the upper part of the sequence. In the lowest, southernmost exposures, lenses of cobble conglomerate appear. A distinctive reddish-brown to grayish-red coarse arkosic sandstone, 100 to 250 feet thick, is at the top of the exposed sequence. The average grain size throughout the assemblage decreases northward, both stratigraphically upward and down the dip. The exposed thickness is 4,000 feet.

The rocks are rich in mica, chlorite, microcline, and sodic plagioclase. The coarse sandstone, here the dominant component, contains 40 percent quartz, 30 percent white or colorless oligoclase-andesine, 15 percent orange-pink microcline, 5 percent mica, mostly biotite, and 10 percent matrix, largely chlorite, clay minerals, limonite and hematite, and, locally, carbonates. Minor constituents include zircon, apatite, sphene, and opaque oxides. The siltstone laminae contain much the same minerals but with chlorite, mica, and clay minerals dominant, and quartz and feldspar rare. The cobbles in the conglomerate include vein quartz, quartzite, granite pegmatite, diorite, gneiss, amphibolite, and dolerite. The red sandstone at the top of the sequence locally has as much as 25 percent of clay-size material. It is low in mica and high in chlorite, limonite and hematite. The plagioclase is thoroughly argillized though the microcline is fresh.

The Belt series in areas east of Three Forks and extending as far north as White Sulphur Springs has recently been described in three doctoral dissertations (Klemme, 1949, p. 6–8; Tanner, 1949, p. 12–17; Verrall, 1955, p. 15–30). The descriptions are summarized in preceding parts of the present report. They indicate that representatives of the Spokane shale, Greyson shale, Newland limestone, and possibly the Chamberlain shale, may be present. Toward the south, mostly in the area studied by Verrall, these rocks interfinger with coarse beds regarded by Verrall as stratigraphically lower than the La Hood formation of Alexander (1955). The North Boulder group of the present report includes all the coarse components of the Belt series in this part of Montana.

Coarse sedimentary rocks of the Belt series have long been known in the mountains in and near T. 2 N., R. 6 E. (J. P. Merrill, *in* Peale, 1893, p. 49; Peale, 1896, p. 2). These rocks have recently been studied in some detail by McMannis (1952, p. 11–14; 1955, p. 1390–1392).[2] His description resembles that of the North Boulder group in its type locality so closely that the correlation can be made with confidence. The rocks include coarse conglomerate, coarse arkosic sandstone and, in the northern part, dark-gray argillite and a few beds of siliceous limestone. The conglomerate contains cobbles and boulders of gneiss, quartz, and schist in a sandy matrix. The sandstone is so massive that bedding is difficult to find. The sandstone is moderately dark gray green where fresh and weathers dark brownish green, locally reddish. It contains quartz, microcline, and orthoclase, with some plagioclase in a matrix composed largely of chlorite and altered biotite and amphibole. The assemblage of coarse sedimentary rocks is over 6,000 feet thick.

The rocks of the Belt series east of Three Forks and north of the lower reaches of the East Gallatin River were originally described (Peale, 1893, p. 16–20) as consisting of coarse micaceous sandstone and conglomerate, with beds of hard, brown, greenish-black, and red argillaceous slate and thin-bedded dark-blue siliceous limestone. The sandstone is prominent in both the upper and the lower parts of the sequence. The limestone is mainly low in a sequence, 2,300 feet thick, that was measured by Peale. He quotes two analyses of the limestone, given in table 33, below.

TABLE 33.—*Analyses, in percent, of limestone from the north side of the East Gallatin River near its mouth*

[Charles Catlett, analyst]

	I	II
Calcium carbonate ($CaCO_3$)	67.85	59.11
Magnesium carbonate ($MgCO_3$)	6.18	1.96
Iron oxide (Fe_2O_3) } Aluminum oxide (Al_2O_3) }	2.50	1.92
Insoluble (silica)	23.50	35.26
Total	100.03	98.25

Exposures here tentatively correlated with the North Boulder group are in and near T. 1 S., Rs. 7–8 W., east of Divide. These rocks are described (Sahinen, 1950, p. 13–15) as composed of a thick conglomerate

[2] Attention is called to McMannis' report entitled LaHood Formation—A coarse facies of the Belt Series in southwestern Montana; Geol. Soc. America Bull., 74, p. 407–436, 1963. His LaHood is essentially the North Boulder Group of this report.

overlain by argillite, slate, quartzite, and quartzitic conglomerate. The basal conglomerate, which overlies schist unconformably (Sahinen, 1939, p. 14), is greenish gray and contains subangular to well-rounded pebbles and small boulders of quartz, quartzite, slate, and schist in a matrix of fine quartz sand and argillaceous material, with appreciable amounts of white mica. The pebbles range from less than one-quarter to more than 6 inches in diameter. The conglomerate grades upward into coarse-grained dark-gray arkose containing feldspar, quartz, and white mica. The argillite, the predominant rock, is gray green, fine grained, and irregularly laminated and weathers dark red. It contains quartz, feldspars, sericite, and clay minerals in grains 0.25 mm and less in diameter. Much of the argillite is still finer grained; in some the maximum observed diameter of quartz grains is 0.04 mm. Most of the quartzite is in pink and black layers and composed of grains 0.5 mm and less in diameter. Some of the quartz has been recrystallized into a mosaic and shows minute needle-shaped crystals and shreds of sericite. The conglomeratic beds contain pebbles as much as 4 inches in diameter in a pink quartzite matrix streaked with black. The quartzite can generally be distinguished from that of Paleozoic age in the vicinity by the dark irregular layers or streaks.

S. D. Theodosis (unpublished Ph. D. dissertation, Indiana Univ., 1956) mapped rocks that presumably belong to the North Boulder group in T. 1 S., R. 9 W., in the Melrose area southeast of Divide. He estimated the thickness as over 4,000 feet and recorded black, gray, green, and red silty shale and argillite with conglomeratic quartzite in the lower parts of the sequence. The upper few hundred feet include purplish-red, white, yellow, and gray conglomeratic quartzite, intercalated with brown siliceous argillite.

Exposures west of Divide seem to be, at least in part, correlatable with the North Boulder group (written communications by P. A. Guttormsen, 1952; W. B. Myers, 1958). These beds include maroon sandy quartzite, buff to white quartzite and quartzitic sandstone, siltstone, argillite, and conglomerate. Unlike much of the conglomerate in the group, the contained pebbles and cobbles are well rounded and rather well sorted, with long axes tending to be parallel to the bedding planes. Diameters range from one-eighth of an inch to 8 inches, commonly about 2 inches. The pebbles and cobbles are mostly quartz and quartzite, with some jasper. The matrix is quartzitic and contains feldspar, sericite, and hematite. The sandstone above is cross-bedded and locally shows ripple marks and mud cracks. Some beds contain garnet. The Hasmark formation (Upper Cambrian) is reported to rest directly on the rocks just described, but the upper part of the Belt sequence here is quartzitic sandstone, leaving the possibility that it may include representatives of the Flathead quartzite, which in most of the region is the lowest of the Cambrian formations.

THE BELT SERIES IN AND NEAR NORTHEASTERN WASHINGTON

DISTRIBUTION

The Belt series is present in and near the northeastern corner of Washington, but its components cannot at present be correlated with named units farther east. Rocks of the Belt series crop out in Boundary County, Idaho, west of the valley of the Kootenai River, in and near Pend Oreille County, Wash., and equivalent strata reach over large areas in British Columbia northward at least as far as the vicinity of lat 52°. In British Columbia, and to some extent in Washington, the Shuswap complex (Canada Dept. Mines and Research, 1948) of metamorphic rocks probably includes rocks of the Belt series, that cannot be separately delineated on the basis of available data. The Priest River group, or terrane, once regarded as pre-Belt (Daly, 1912, p. 258–271), is believed to be part of the Belt series for reasons outlined on page 89. Rocks in and north of southern Stevens County, Wash. (Weis, 1959), seem likely to be correlatable with the Belt series.

CHARACTER

In northeastern Washington (Park and Cannon, 1943, p. 5–15) the stratigraphic sequence includes, in ascending order, the Priest River group, Shedroof conglomerate, Leola volcanics, and the Monk formation. In Daly's (1912, p. 258–271) nomenclature the Priest River group is the Priest River terrane, the Shedroof conglomerate is the Irene conglomerate, and the Leola volcanics is the Irene volcanic formation. In a later report (Little, 1950, p. 5–6) applicable only in Canada the Shedroof conglomerate becomes the Toby conglomerate, and the Leola volcanics are called the Irene volcanic formation. Little regards the Toby and Irene as belonging to the Windermere series, whose Precambrian part corresponds to the upper part of the Belt series of the United States. The Priest River group is a complex sequence of metamorphic rocks that includes phyllite, schist, limestone, dolomite, quartzite, and volcanic rocks, regarded as unconformably below the Shedroof conglomerate. The conglomerate is coarse very poorly sorted dingy-gray-brown rock. Most of the contained fragments are 1 to 8 inches across. Many are smaller, but a few irregular limestone blocks have diameters of 5 feet or more. Some sandy phyllitic and dolomitic beds are intercalated in the conglomerate. The pebbles in the Shedroof conglomerate have probably been derived from the Priest River group, but the bedding in the latter appears to parallel the contact with the conglomerate. The thickness of the Shedroof conglomerate varies widely but is several thousand feet. The rock may be a fanglomerate, but its origin is not clear. "The wide deposition of such immense quantities of uniformly coarse debris is difficult to explain" (Park and Cannon, 1943, p. 9). The Shedroof conglomerate grades upward into the Leola volcanics. The latter includes greenstone and green schist, regarded as metamorphosed lava and tuff. The thickness is variable and difficult to determine but is supposed to be over 5,000 feet.

The Monk formation overlies the Leola volcanics, apparently unconformably. It consists mainly of fine-grained phyllite but includes quartzite, grit, and quartzitic limestone, with conglomerate at the base. Park and Cannon tentatively assigned the Monk formation to the Cambrian, as it is conformable with and grades upward into the Gypsy quartzite, which has yielded Cambrian fossils. Two other formations of Cambrian age (Maitlen phyllite and Metaline limestone) lie above the Gypsy quartzite. In British Columbia the contact between the Precambrian and Cambrian has been placed tentatively somewhat higher on the basis of more recent work (Little, 1960, p. 10–11). The change is made mainly because the base of the Monk formation appears to be 12,000 to 14,500 feet below a horizon at which abundant pleosponges have been found. The latter are regarded as probably equivalent to the *Olenellus* zone of Early Cambrian age. The fossils collected by Park and Cannon in their Gypsy quartzite are thought by Little to come from a horizon several hundred feet below the horizon that yielded the pleosponges. Little tentatively places the Precambrian-Cambrian boundary at the base of his Quartzite Range formation, which appears to correspond to a horizon within the Gypsy quartzite of Park and Cannon. The latter is 5,300 to 8,500 feet thick. In any event, no hiatus has been suggested anywhere between the base of the Monk formation and the horizons that have yielded Lower Cambrian fossils. Hence the horizon selected as the base of the Cambrian system in the locality is necessarily arbitrary, chosen mainly on the basis of ready recognition in mapping.

In the northwest corner of Idaho (Kirkham and Ellis, 1926, p. 17–27, 39–40) the Irene conglomerate and Irene volcanic formation of Daly have been grouped as Purcell lava, although obviously sedimentary rocks make up a large part of the unit. At the base of the so-called Purcell lava is 5,000 feet of conglomerate in which the sand and pebbles are chiefly made up of sericitic quartzite and white quartzite supposed to have been derived from the Revett quartzite of Kirkham and Ellis, part of an assemblage they thought equivalent to the Priest River terrane of Daly. The conglomerate interfingers at the top with 6,000 feet of altered basaltic and andesitic lava flows, greenstone schist, tuff, conglomerate, breccia, and dolomite.

The rocks included in the Priest River terrane are metamorphosed, and Daly (1912) thought that they were bounded at the top by an unconformity. In northeastern Washington these rocks have been called the Priest River group (Park and Cannon, 1943, p. 6) and include phyllite, schist, limestone, dolomite, quartzite, and volcanic rocks. The unconformity at the top is noted. In British Columbia (Canada Dept. Mines and Research, 1948) the rocks Daly included in the Priest River terrane are mapped as belonging to the upper part of the Purcell series and the Windermere series.

In and north of southern Stevens County, Wash. (Weis, 1959), Precambrian rocks are overlain unconformably by conglomerate with volcanic rocks over this, and both are unconformably overlain by Lower Cambrian quartzite. These various rocks extend northward into the Nelson map area in British Columbia. Weis suggests that the three kinds of rock below the strata of Early Cambrian age represent a facies of the Belt series. He says they differ in that shallow water features are scarce and the fine-grained clastic rocks in the southwesternmost exposures include carbonate strata. Both these features exist in parts of the Belt series in Montana and they, therefore, are not sufficient to prevent correlation of the rocks in Stevens County with the Belt series in a general way.

GENETIC PROBLEMS

It is obvious from the above descriptions that the Belt series covers broad expanses of country and was deposited in shallow water. Each geologist who has discussed its genesis has had his own concepts as to the form and boundaries of the area of deposition, the relations to the open oceans, marine versus lacustrine deposition, sources of the material, and climatic conditions during deposition. Many of the facts necessary to solve such problems are still lacking. Some are obscured by an extensive cover of later deposits; others are to be sought in regions so remote that adequate geologic studies cannot be expected for a long time to come. Even within western Montana, the region of immediate interest in the present investigation, there are wide areas for which available data are scanty. Thus, any ideas as to the origin of the Belt series must necessarily be tentative.

The Belt series in and west of Montana was once thought of as a representative example of the Algonkian system. It was laid down in an essentially continuous wide, shallow basin, whose floor and borders are nowhere well exposed. Towards the west the old shores are now either obliterated by intrusions or covered by lava flows. Northward the basin may have extended to the Arctic Ocean. Sediments were deposited as far south as northern Mexico but whether in a single basin or partly or wholly separate basins cannot be proved. The sediments composing the series are products of the erosion of gneiss and schist and sedimentary rocks derived from these rocks, probably carried mainly by streams of low gradient under rather arid conditions, but possibly accompanied by glaciation locally. They were laid down under marine rather than lacustrine conditions but in water that, at least for long intervals, was more shallow than is common in either environment. The carbonate rocks, like those of younger seas, derived their constituents largely from organisms that grew in the water in which those rocks formed. In some localities deposition may have been essentially continuous from Belt time into later periods. Elsewhere the series was eroded and deformed before being covered by later deposits.

THE TERMS "ARCHEAN" AND "ALGONKIAN" IN RELATION TO THE BELT PROBLEM

The Belt series constitutes the assemblage of dominantly sedimentary rocks to which the name "Algonkian" was originally intended to be applied in Montana (Powell, 1890, p. 19–20, 59–61, 66). As studies throughout North America progressed it became evident that sedimentary rocks were included in the "basement complex" to which the name "Archean" had been restricted (Wilmarth, 1925, p. 103–104). Consequently the definition was modified so that the Algonkian rocks came to include all those below the Cambrian to which "the ordinary methods of stratigraphy could be applied" (Walcott, 1903, p. 26), a qualification that has been reiterated more recently (Leith, 1934, p. 171). The rocks of the Algonkian system as defined at that time are chiefly shales, sandstones, and limestones and their metamorphic equivalents, but it was recognized that some igneous rocks were so situated that they could not be excluded from the Algonkian. Fossils were reputed to be present in strata then assigned to the Algonkian, but it was not possible to use them as tools of stratigraphy. In this respect the ordinary methods of stratigraphy, called for in the definition, could not be applied to components of the Algonkian system. Under both the original and revised definitions, the Belt series over most of its area of outcrop constitutes the sole representative of the Algonkian system. Geologists in the first quarter of the 20th century tended to use Algonkian and Belt as essentially synonymous.

It was natural to assume the presence of a great unconformity at the base of the Belt series everywhere, scanty though the direct evidence was. The surface on which the series rests was spoken of as the Ep-Archean surface or peneplain, a term originally proposed by Lawson (1902, 1914, p. 361–363) in a different region and by Hinds (1936a, p. 5; 1938). Controversy as to nomenclature in the Lake Superior region (Van Hise and Leith, 1909, p. 38, 322–323, 371, 382; James, 1958, p. 28–30) has beclouded the issue somewhat. As far as the Belt series is concerned, the matter is complicated by the fact that near Neihart the series rests on gneiss that by any definition can be safely thought of as Archean or of Archean type, whereas farther south it probably rests in large part on highly metamorphosed but indisputably sedimentary rocks. These include the Cherry Creek group of Peale (1896) and the Pony series of Tansley, Schafer, and Hart (1933, p. 8–11), both having igneous rocks associated with them. The gneiss near and south of Neihart and even the Cherry Creek group and Pony series has, in the past, been regarded as of Archean age (Weed, 1896, 1899; Tansley, Schafer, and Hart, 1933; McMannis, 1952, 1955) but the Cherry Creek group has been referred to, more or less tentatively, as Algonkian (Tansley, Schafer, and Hart, 1933; Van Hise and Leith, 1909, p. 38; Peale, 1896).

Parts of the rocks of the Cherry Creek group and the Pony series would yield, presumably, to attack by the ordinary methods of stratigraphy and might, therefore, be thought of as of Algonkian type; but much of the laminated gneiss in this part of Montana, whether or not originally sedimentary, probably will be impossible to study effectively by such methods. On this basis they might well be referred to as of Archean type, but the term would have little significance. On the basis of present data, the age relations between the various rocks of igneous appearance, markedly laminated or not, and those rocks that are still obviously sedimentary are uncertain. Further, reconnaissance in connection with the compilation of the geologic map of Montana (Ross, Andrews, and Witkind, 1955) disclosed localities where granitic rock reputed to be of pre-Belt age seems to intrude Paleozoic limestone. The uncertainty as to the age of granitic rocks in southwestern Montana has been referred to by Klepper (1950, p. 72). Heinrich (1949a, p. 7–11; 1949b, p. 309–322) pointed out that pegmatites of three different ages occur in the area of metamorphic rocks and difficulties arise in dating specific pegmatites and determining their parent rocks. His data serve to reinforce the concept that granitic rocks of more than one age are present.

All this serves as an illustration of the difficulties and complexities of stratigraphic correlations in Precambrian rocks, even where some are relatively little metamorphosed. Because of situations of this sort, and others more complex, Leith (1934, p. 171–172) has proposed to apply the terms "Algonkian type" and "Archean type" to rocks that are or appear to be of the kind formerly assigned to the Algonkian and Archean systems. His terms would imply that the rocks had the lithologic characteristics on which former usage was based, but would carry no age significance. His proposal has found favor with many geologists and is virtually at present adopted as the official usage of the U.S. Geological Survey (Geologic Names Committee, 1953). The conveniences afforded by usage of this kind are obvious but it is hoped that the expedient is a temporary one. If progress is to be made in working out the history of Precambrian rocks, age relations need to be considered. The difficulties are many and have been pointed out by all who have touched on the problem. They are not necessarily insuperable. As detailed mapping and study continue, some of the specific problems will be solved. Where exposures of old rocks are separated from each other by broad expanses of younger rocks some other means of study is essential. Perhaps eventually radioactive methods of age determination will progress far enough to be of greater assistance than they are at the moment.

THE LOWER LIMIT OF THE BELT SERIES

According to general usage, the Belt series consists dominantly of sedimentary rocks so little metamorphosed as to be easily recognized as such. By implication, at least, pre-Belt rocks are more thoroughly metamorphosed or otherwise so different from those typical of the Belt series that no confusion arises. On the whole, this implication seems justified. Here and there rocks now regarded as belonging to the Belt series are so much metamorphosed as a result, directly or indirectly, of igneous activity, that they have in the past been supposed to be of pre-Belt age. Their present assignment to the Belt derives from tracing progressive metamorphic changes from undoubted Belt rocks into more severely altered units. The Shuswap series in British Columbia and gneissic rocks near Shoup, Idaho, are cases in point. Another case may be the small patch near Lost Creek beyond the southeast corner of Granite County once mapped as pre-Belt gneiss, schist, and related rocks (Ross, Andrews, and Witkind, 1955). Reconsideration of the scanty evidence has resulted in revision of opinion, and this small area is shown on plate 1 as belonging in part to the Missoula, in part to the Piegan group.

South of the region in which the Belt series is known, definitely stratified rocks, mainly of sedimentary origin but regarded as of pre-Belt age, are plentiful in several localities. The Cherry Creek group and Pony series in southern Montana, the stratified rocks of Precambrian age in northwestern Wyoming, and the Uncompahgre formation in southern Colorado, are examples. Some of these units are enough like the Belt series so that, as noted below, uncertainties as to correlation have arisen. Within the region occupied by the Belt series itself no uncertainty of this particular kind is known to exist. The age relations between that series and the Hyndman and East Fork formations north of Hailey, Idaho, and the Albion Range group in Cassia County, Idaho, are not conclusively established but these units are so limited in distribution and so far away from rocks generally accepted as components of the Belt series that the problem is not acute.

Some beds low in the Belt series in and near the Coeur d'Alene region, Idaho, contain subordinate amounts of fragments of argillaceous rocks. These fragments might have come from unknown pre-Belt strata rather than from the series itself. Thus, where relations between rocks now thought of as near the base of the series and those at higher horizons are obscure, it is conceivable that the stratigraphically low rocks are of pre-Belt age. The metamorphosed rocks east of Lolo Hot Springs at present assigned to the Prichard formation may be an example. These rocks show marked effects of the intrusion of the nearby Idaho batholith, but the possibility of prior metamorphism from other causes remains. Conceivably they belong not to the Prichard formation but to some unrecognized unit of pre-Belt age.

FORM AND EXTENT OF THE BASIN

In Montana and northern Idaho, the basin within which the Belt series was laid down is believed to correspond fairly closely to the region containing outcrops of that series. The eastern boundary of these outcrops starts at the Canadian border a dozen miles east of long 114° (fig. 7). It extends southeast to a point near long 110°20′, lat 46°40′, swings southwestward to near long 111°, lat 45°52,′ thence almost due west and then southwest to the extreme southwest corner of Montana. In Idaho the boundary is less adequately defined but extends southwest roughly to long 114°, lat 44°. West of that point, the boundary of the Belt series is complicated by the presence of the Idaho batholith, but it may extend to about long 116°, lat 45° and thence to long 117°, lat 46°. Probably all Idaho north of this poorly defined boundary was originally underlain by rocks of the Belt series. To the south, small areas of sedimentary rocks that have been supposed to be of Precambrian age are known, but none of them has been correlated with the Belt series. The Snake River Plain with its lava flows and associated sedimentary rocks effectively eliminates the possibility of direct tracing of relationships among the old rocks. In Washington there are rocks allied to the Belt series, at least in the northeastern corner (Park and Cannon, 1943; Weis, 1959), but the cover of later rocks is so extensive that no western boundary can be drawn on the basis of present knowledge. Taking into account the Precambrian rocks north of Washington and their absence in most of eastern Washington and in eastern Oregon the best guess is that this boundary is not far west of the border between Washington and Idaho. Thus, roughly 65,000 square miles in Montana, Idaho, and Washington contain exposures of the Belt series.

Though deposits of some sort may be assumed to have covered all of the submerged basin surface, those along the shores may have been thin enough in places to have been readily removed by erosion. As most slopes were gentle and currents slow, such thin marginal deposits may have covered fairly large areas. Because of the thick cover of Paleozoic and later rocks along the eastern boundary of the Belt series in Montana, little direct evidence bearing on this point is to be expected there. The position of the border of the submerged area must have shifted from time to time during the long interval in which the Belt series was laid down and may have been modified locally by erosion before it was covered by Paleozoic sediments. As the deposits were at times mud flats on which algal heads projected above the water, the shore was ill defined even while deposition was in progress. For long stretches no distinctive near-shore sediments such as might be expected in deeper basins of deposition were formed. The thrusts that have broken the rocks along the mountain front have shoved masses of strata of the Belt series east or northeast of their original position. However, it seems unlikely that thrusting has extended the area underlain by Belt rocks more than roughly 50 miles eastward beyond its original limit. When the other uncertainties as to the shape of the Belt basin are considered, the error introduced as a result of thrusts becomes of minor consequence. The scanty data available as to extensions of the Belt rocks in Montana east of the main mass are as follows:

Near Neihart the floor of crystalline rocks is high enough so that rocks of early Paleozoic age rest on the surface of the northern part of a small mass of gneiss, whereas rocks of the Belt series lie on the southern part of the same mass. The locality is only a short distance north of the easternmost point of exposures of the main body of the Belt series as outlined above (long. 110°20′, lat. 46°40′) and just inside that border. Small exposures of rocks believed to belong to the Belt series are known in the Big Snowy Mountains, some 50 miles east of that point (Reeves, 1931, p. 145). Perhaps these are remnants of the thin and fine-grained marginal deposits.

East of the map area (pl. 1), near the Canadian border, a minette sill in strata of Cretaceous age west of Whitlash (Ross, 1950, p. 187) contains inclusions of gneiss, granite, quartzite, and schist believed to have been derived from the complex of crystalline rocks of pre-Belt age that underlies the region. This has been taken to imply that no representatives of the Belt series were present at this locality when Paleozoic beds were laid down. The exposure is roughly 120 miles due east of the present eastern boundary of the Belt series. The Belt shore near the present international boundary, therefore, may have been somewhere within the 120-mile interval.

A few of the many wells drilled in search of oil east of the mountains (Dobbin and Erdmann, 1955) have passed through the Paleozoic assemblage and reached igneous rocks interpreted as of Precambrian age. None of these wells penetrated any rocks allied to the Belt series. Alpha (1955), who describes the igneous rocks drilled, gives convincing evidence that they belong to what he terms the Precambrian basement, and, on the basis of well data and correlative information, argues for an old structural high, called the Genou Trend. Uplift along this trend zone began so long ago that it is reflected in a topographic ridge composed of rocks of the Precambrian basement (pre-Belt) and now covered by Paleozoic and later rocks. The Genou Trend extends some 35 miles northeast from south of Collins (east of map area, pl. 1), which is in T. 25 N., R. 2 W.

The evidence for areas near Neihart, Collins, and Whitlash favors the idea that the shore of the Belt basin in Montana and north of Neihart may have trended northwest and been approximately 50 miles from the main mountain front. Subsurface data in Alberta (Burwash, 1957) reveal no evidence of the Purcell or Windermere series east of the foothills. The eastern shore of the original Belt basin may have been even closer to the present mountain front in Alberta than it was in Montana, a suggestion that is in harmony with ideas expressed by Warren (1951, p. 3–5) and by Webb (1954, p. 3–7). Farther north data are scanty, but the ancient shore may have continued to trend northwest.

Deiss (1935, p. 104–105) has suggested that the eastern shore of the Belt sea lay somewhat to the east of central Montana and extended from the site of Yellowstone Park to beyond the present Highwood Mountains. On this basis the Belt sea would have reached farther eastward in the area north and south of the Big Snowy Mountains than was suggested. Deiss apparently thought of this shore as irregular, with a series of bays or estuaries separated either by low westward-extending peninsulas or by low islands. He assumed that the islands (or peninsulas) "were situated essentially in the areas now held by the isolated mountain masses"—such as the Highwood, Big Snowy, Castle, and Big and Little Belt Mountains. The area near Neihart, where rock of the Belt series rests against granitic gneiss, might well be the site of one such an island but evidence for others is scanty. The structural deformation with which the present mountain masses are associated is far younger than, and apparently unrelated to, events of Belt time.

The southern border of the exposures of the Belt series in Montana has exceptional features. Here the series consists almost exclusively of the North Boulder group, whose coarse graywacke, or arkose, and conglomerate are without apparent counterparts elsewhere in Montana. Although the Belt series here is 5,000 to 8,000 feet thick it stops abruptly along a line close to lat 45°45′, roughly between long 111° and 113°.

Lowell (1956) mentions Belt rocks, so coarse that they can be correlated safely with the North Boulder group, resting on pre-Belt metamorphic rocks in the southwest corner of T. 1 N., R. 3 W., which is close to this line. South of this line the Flathead quartzite rests on the metamorphic rocks of pre-Belt age. East of Divide, rocks that at least in part belong to the North Boulder group exist but faulting makes their relations to pre-Belt rocks obscure. In much of this part of Montana faults of supposed Tertiary age are plentiful and published geologic information is scanty.

Speculation as to the signficance of the features just summarized began as soon as geologic exploration in the area got underway (Peale, 1893, p. 14, 16–20). At that time a westward-trending Precambrian shore was recognized, but orogenic movement near it was deduced to be of post-Belt age. Deiss (1935, p. 104-106) has inferred that in Belt time a large island lay in the area now occupied by the Tobacco Root, Madison, Gallatin, Beartooth, and Absaroka Ranges south of present lat. 45°45′. This implies that the Belt sea reached south of that latitude so as to surround the island, but deposits that would support the concept have not been recorded. Erosion in Early Cambrian time would have removed part of any rocks of the Belt series that may have existed, but the apparent absence of remnants of the Belt series beneath the Flathead quartzite (Middle Cambrian) anywhere south of this line is surprising. The alternative explanation is that the present contact of the Belt series near lat 45°45′ corresponds fairly closely to the original southern shore of the Belt basin and that few, if any, deposits of Belt age ever existed south of that line. Such an explanation calls for relatively deep water north of the straight eastward-trending shore. The sediments of the North Boulder group would have been brought by streams from relatively high ground to the south and dumped abruptly close to shore.

The straight eastward-trending shore may (Klemme, 1949, p. 8; Alexander, 1955, p. 7–36; McMannis, 1955, p. 139–192) correspond essentially to a fault scarp. The inferred fault would have been in existence prior to Belt time but, as the character of sediments of the Belt series here suggests deposition during rapid subsidence, fault activity may have been resumed or persisted during Belt time. The Belt basin on the whole formed during slow rather than rapid subsidence. The postulated fault scarp might have separated high ground to the south from the Belt basin to the north. The coarse conglomerate in parts of the North Boulder group suggests that the high land south of the scarp was mountainous, whereas it is commonly believed that the land bordering most of the Belt basin in Montana was low.

In Montana west of long 113° and in most of Idaho there is no basis for tracing the border of the Belt basin. In south-central Idaho particularly, outcrops of the Belt series are widely scattered and no pre-Belt rocks may be exposed. The largest continuous bodies of Belt rocks are in Lemhi County, and parts of them are known only in reconnaissance. It was formerly supposed that the base of the Belt series is exposed near Shoup, Idaho (Umpleby, 1913a, p. 41–42, 48, pl. 6), but more recent work by Davidson (1939) and his coworker Anton Gray has disproved this. This weakens Umpleby's (1913a, p. 31-32) correlation of the rocks near Shoup with the Prichard slate of northern Idaho. In those few areas in Idaho south of lat 46° where enough has been done to warrant formation names (Ross, 1934, 1947) no correlation with units in Montana and northern Idaho has been made although suggestions are offered.

The great mass of coarse conglomerate in northeastern Washington, northwestern Idaho, and large areas in adjacent parts of British Columbia, variously called Shedroof conglomerate, Irene conglomerate, Toby conglomerate, and the basal part of the Purcell lava, has been postulated to be either fanglomerate (Walker, 1926, p. 15) or a "deposit along the shoreline of a sea that spread rapidly over the land, roughly reworking the loose, surficial deposits" (Rice, 1941, p.

23). The conglomerate is the basal part of the Windermere series of the Canadian geologists and rests unconformably on their Purcell series. In one area the angular discordance is locally as much as 45°, and the Purcell series was folded markedly before the Windermere was laid down (Walker, 1926, p. 12).

These data seem to establish the presence of a rugged land mass in northwestern Idaho, western British Columbia, and adjacent areas in late Belt time, a concept long held in Canada (Rice, 1937, p. 36; Schofield, 1941, p. 710; Warren, 1951, p. 2–3). Land may have been present in eastern Washington and Oregon, also, although direct evidence of it is lacking there. The land appears to be genetically related to diastrophism in the interval between deposition of the Purcell series and that of the Windermere series of Canadian writers. No comparable diastrophic event has yet been proved in central Idaho or in Montana. For most of that area the gradational contacts throughout the Belt series demonstrate that no such an event has occurred. In central Idaho south of lat 45°30′, rocks correlated with the Belt series are deformed and metamorphosed. It has been supposed that these changes occurred near the end of Precambrian time (Ross, 1934; Leonard, 1957), but recent studies (Leonard, 1958, oral communication) raise the possibility that they are related to the Idaho batholith during the late Mesozoic. If the diastrophism in central Idaho was, however, in part of Precambrian age, it may have some relation to that just cited in Canada. If so, some support is afforded to the concept that the border of the Belt basin, at least in late Belt time, was not far west of Idaho.

A metamorphic complex near the Canadian border known as the Shuswap terrane, once called Archean, was more recently believed to be composed mainly of parts of the Belt series metamorphosed much later than Belt time (Dawson, 1890; Daly, 1912, p. 258–271, 1913, p. 122–132, 1915, p. 10, 1950, p. 741–743; Cairnes, 1940; Brock, 1934; Hinds, 1940, p. 290). Some Paleozoic rocks may be included among the metamorphic rocks. A recent review of the evidence (Reesor, 1957, p. 172–175) shows that doubt remains as to the dating of the rocks included in the Shuswap terrane but leaves the possibility that some may be late Precambrian. If so, sedimentation of Belt age may have extended in northern Washington at least as far west as long 118°30′, well to the west of the coarse rocks of the Windermere series and its equivalents. These rocks suggest that they were deposited near or, conceivably, on dry land. As already noted, it has been postulated that some of this land was farther east, near long 116°. The evidence is insufficient to show whether the shoreline shifted 50 miles or more during late Precambrian time or whether the land west of the Belt basin was sufficiently irregular in outline to permit deposition far enough west to reach the Shuswap area. There may have been a group of islands rather than a single land mass.

The extent of the land westward from Idaho is uncertain because of the extensive cover of Paleozoic and later rocks in Washington and Oregon. Some paleogeographers postulate that early in the Precambrian (Schuchert, 1923, p. 214) land lay west of the present coast of the Pacific Ocean and that later the eastern shore was near long 118° (Schuchert, 1923, p. 213; Schofield, 1923, p. 93–94, 1941, p. 710). Such a land mass would merge with that suggested in the preceding paragraph. Schofield (1941) has assembled data that give some support to the concept of a large land mass in western Canada late in Precambrian time. His evidence consists largely in the presence of conglomerates that include cobbles of granitic and metamorphic rocks of pre-Mesozoic age, inferred to have been derived from the postulated land mass.

Another, more tenuous, hypothesis is that the eastern border shown on figure 7 corresponds essentially to the eastern shore of the Pacific Ocean in late Precambrian time. There is little direct evidence in the rocks now exposed in the western United States and Canada to compel belief in land areas in the late Precambrian west of the depression sketched on figure 7. The hypothesis would not be invalidated by such features as an island near or west of the northern end of Idaho. Indeed, there might well have been a string of islands west of the geosynclinal area in which the Belt series was laid down. Some support for the hypothesis is afforded by the presence in Stevens County, Wash. (Weis, 1959), of rocks without distinctive shallow-water features.

SOURCE OF SEDIMENTS

The relatively small part of the Belt series that is coarse conglomerate derived its material from close at hand in pebbles large enough to be recognized. The North Boulder group is largely composed of material from the Cherry Creek group and Pony series (Tansley, Schafer, and Hart, 1933, p. 7–12; Heinrich, 1949b, p. 308–319; Reed, 1957, p. 3–8). The relatively few pebbles and cobbles in the North Boulder group not readily recognizable as derived from the two units named above probably came from rocks of igneous aspects in slightly more distant parts of southwestern Montana.

Near Neihart, conglomerate is absent. The Neihart quartzite contains 96 percent silica. It consists dominantly of quartz with only subordinate amounts of feldspar, mica, and other minerals. If this rock has been derived from the gneiss of pre-Belt age, abundant near Neihart, more thorough disintegration of the parent rock and longer transportation of the resulting detritus is implied than is probable for the conglomerate in the North Boulder group. This inference tends to weaken the assumption, cited above, that the gneiss formed a part of an island in the Belt sea.

The clastic beds in the Altyn limestone in Glacier National Park contain many quartz grains, less plentiful grains of feldspar, and some grains that contain both quartz and feldspar and are probably bits of coarse-grained granite. Most of the feldspar is alkali plagioclase, but some is microcline. Most feldspar grains are strikingly fresh, but some are sericitized. Some grains are very fine aggregates of quartz, and some include rounded masses that may be silicified oolites. These grains might have come from silicified limestone, but most of the clastic grains in the Altyn limestone probably were derived by erosion from a granitic rock similar to that exposed near Neihart or to those in the southwestern part of the State. Quartzite recurs at intervals throughout the Belt series although it is a dominant constituent of few of the formations. The quartzite beds consist mainly of quartz, some with strain shadows, but commonly contain some feldspar and subordinate quantities of other minerals, largely micaceous. The feldspar includes sodic plagioclase, microcline, and rare microperthite. Grains composed of fine-grained sedimentary rocks are exceptional. Insofar as diagnostic minerals have survived in the quartzite, they appear to have come from rather alkalic granitic rocks. Some of the quartz may have come from gneissic rocks. The proportion of feldspar in the quartzite of the Belt series is so low that these rocks were derived either from older quartzite or their component grains had been transported so far from the original sources that most minerals less resistant than quartz had disappeared. The prevalence of minerals of micaceous habit, including some clay, in much of the series favors the second alternative. Neither the Cherry Creek group nor the Pony series is so quartzose as to have yielded quartzite directly, without the winnowing out that results from weathering and long transportation. Rocks supposedly of pre-Belt age in Cassia County, Idaho, in parts of northern Utah, and in northwestern Wyoming contain quartzite. They may have been among the sources of material in the quartzite of the Belt series.

The clastic grains of feldspar that are widespread, though only abundant in a few spots throughout the rocks of all grain sizes in the Belt series, were derived from rocks like the granitic gneiss near Neihart (Weed, 1899, p. 3, 1900, p. 278–279; Schafer, 1935, p. 6–10) and the granitoid metamorphic rocks in Madison and Park Counties (Iddings and Weed, 1894, p. 3; Peale, 1896, p. 2, 4; Tansley, Schafer, and Hart, 1933, p. 8–11, 13; and personal observations) with their associated pegmatites (Heinrich, 1949b, p. 310–319; 1949c).

The fine-grained clastic rocks that compose by far the greater part of the Belt series differ from normal argillaceous rocks in that they are more siliceous than the average, and contain clastic grains of unweathered alkalic feldspar. The impure quartzite beds are not essentially different from those that have been called argillite and similar names. They are merely more siliceous and somewhat coarser grained.

Some of the silica in the argillite and argillaceous quartzite may be derived from opaline material chemically precipitated during deposition of the sediments. This explanation has been offered for siliceous shales in general (Pettijohn, 1949, p. 286–287). In the Belt series it can be only partly true. Most of the quartz is in clastic grains, locally with overgrowths added during diagenesis. The quartz particles of the matrix are too small to determine whether they may be clastic also. While it would be difficult to determine quantitatively the amount of silica present in overgrowths and matrix and hence, not positively known to be of clastic origin, inspection under the microscope indicates that it is not large. Rubey (1929, p. 153–170) has studied the siliceous Mowry shale of the Black Hills region and shows that the origin of the Mowry shale is intimately connected with the occurrence of volcanic ash. Silica leached from the ash was introduced into the sediment as a gel. But many features of the Mowry shale are not found in the argillaceous rocks of the Belt series. The latter lack the resistance to weathering and the conchoidal fracture characteristic of cherty rocks. Most of the Belt series shows no volcanic ash, and the volcanic strata present are not high in silica.

If no significant part of the silica in the fine-grained rocks was introduced chemically, it follows that the source rocks were very rich in quartz. They were also so free from weathering that the alkalic feldspar in them had not been decomposed. Weathering conditions in late Precambrian time may have differed from those of the present, in part because the early plants can hardly have had the same weathering effects as modern ones. It seems possible, even probable, that some land plants existed. Presumably, however, the vegetation around the borders of the Belt basin was insufficient either to provide a protective cover for the rocks or to aid greatly in chemical destruction of the component minerals. Mosses and lichens would have furnished some protection to the rock although not as much as turf and bushes. The disruptive effect of deep roots would have been absent. Mechanical disintegration, as a result of temperature changes and similar factors, should have been rapid as compared to those of the present day, whereas chemical breakdown of feldspar and other minerals may well have been slow. Thus the presence of unaltered feldspar grains and of clastic flakes of mica does not necessarily imply that the site of deposition was as close to the source as might be deduced from younger rocks.

Some chemical decay does seem to have occurred. The rarity of calcic feldspar may result in part from its greater susceptibility to such destruction. The same may be true of pyroxene and amphibole. The fine micaceous flakes that are abundant in the argillaceous rocks may have been in part detrital, derived from schists of pre-Belt age, but they are in part the product of the breakdown of other minerals during weathering and transportation. The fine-grained rocks that compose much of the Belt series must originally have contained much clay. The greater part of the clay has recrystallized into such minerals as sericite, and secondary feldspar, but small amounts of clay minerals remain.

Present exposures are sources for the granitic rocks that contributed directly to parts of the Belt series. Most of the series contains no evidence to suggest derivation by erosion from granitic rocks; it seems to represent a second generation whose immediate sources were sedimentary rocks and is less readily accounted for without looking to exposures far away.

As an unconformity has been postulated in Canada at the base of the Windermere series, it is possible that parts of the Belt series came from erosion of pre-Windermere rocks. The pre-Windermere rocks would be the whole of the Purcell series, including the Priest River group and, thus, may correspond to all of the representatives of the Belt series below the Missoula group that are exposed in and north of areas in Idaho and northeastern Washington along the Canadian border. A possibility worthy of consideration in the future is that there may have been sufficient interruptions in deposition in the Belt basin in Montana, also, so that some of the younger Belt units represent reworking of eroded older ones. A few unconformities within the Belt series are on record and others could have been overlooked easily in the many areas where intensive studies have not yet been attempted.

In the carbonate rocks the noncarbonate grains come from the same sources as those of the clastic rocks commented on above. These grains are widely distributed throughout the carbonate rocks, which helps to indicate that the carbonate rocks were in part deposits in currents of water much as the siliceous and argillaceous sediments were. The carbonate in these and other Precambrian rocks has been much discussed. In part because of the absence of the kinds of fossil remains prevalent in younger carbonate rocks, the suggestion has been offered that the Precambrian carbonate rocks are chemical precipitates (Daly, 1912, p. 670–675). It was supposed that chemical conditions in the atmosphere and, hence, in the seas of Precambrian time were significantly different from those of later eras. Carbonates brought to the sea by streams were assumed to be promptly precipitated, mainly by ammonium carbonate from decaying organic matter on the sea bottom. The scavenging animals that later kept the amount of such decaying matter to a minimum were assumed not to have been in existence. Increase in the supply of carbonates resulting from increase in land areas produced by uplift in late Precambrian time and decrease in accumulations of decaying matter in proportion to the development of fishes and other scavengers were supposed to have proceeded slowly. The Precambrian and early Paleozoic limestones and dolomites were, on this hypothesis, largely precipitates, leaving the sea water with so little calcium content that none was available for shells and skeletons. After very early Paleozoic times the marine limestones were supposed to have been formed largely from the debris of the hard parts of animals and plants. Leith (1934, p. 154–155) has raised doubts as to the validity of the hypothesis although he grants that "the concept of limeless pre-Cambrian seas remains" a working hypothesis. Later, however, Rubey (1951, p. 1113–1114) came to the conclusion that "the composition of sea water and atmosphere has varied surprisingly little, at least since early in geologic times." His conclusion, supported by a variety of data, removes the need for postulating a limeless sea and wholesale precipitation of carbonates in Precambrian time.

Rubey was concerned with what might be termed "normal sea water." The Belt basin was so shallow that the water in it may not have been strictly comparable in the concentration and perhaps in the proportions of its dissolved solids. Local conditions may have permitted precipitation of carbonates at times through such processes as the warming of shallow water to the point where enough carbon dioxide was driven off so that carbonates would precipitate. Some of the fine- and even-grained thinly laminated carbonate rocks in the Belt series may have such an origin. Most of the carbonate rocks, however, are believed to be stromatolites, detritus derived from stromatolites, or else detrital beds whose carbonate is derived in some way from organisms. Fairly large parts of the carbonate-rich beds throughout the Belt series contain either stromatolite biostromes and bioherms or material intimately related to these. An even larger part of these beds contains irregular widely variable structures that have not been demonstrated to be stromatolites, but that may well have originated through the work of organisms and been modified by later solutions. Under suitable conditions ordinary ground water can profoundly alter and replace carbonate rocks with other substances.

Most of the carbonate rocks include oolitic beds, or beds that preserve vague features suggestive of former oolites. The oolites may have originated as coprolites as in some modern deposits.

Thus, the carbonate rocks are essentially no different in origin from most carbonate rocks of Paleozoic and later eras. Much or all of their carbonate content was derived directly or indirectly from organisms. One difference from later deposits is that most of the organisms were plants. Those carbonate beds that include clastic grains of quartz and other noncarbonate minerals are dominantly detrital, whatever the origin of the carbonate.

MARINE VERSUS LACUSTRINE DEPOSITION

The problem of whether the Belt series is dominantly of marine or lacustrine origin has engaged attention almost since the series was first noted by geologists. Walcott (1899, p. 200–244) seems originally to have assumed a marine origin but for a time (Walcott, 1914a, p. 1–16) thought that the series was of continental origin and regarded this as a major reason for the paucity of recognized fossils in the rocks of the series. In this he followed Brooks (1894) who supposed that life first developed in the open ocean and that only at about the end of Precambrian time did organisms discover and colonize the shores and bottoms. Brooks believed that the ocean bottom, far enough from shore to be free from sediment and deep enough to permit a pelagic fauna to be well developed above it, was first colonized. Shallow water and that along shores had dangers of various kinds and were colonized later. His hypothesis would help to explain the paucity of organic remains in the Belt series, for an outstanding characteristic of that series is the shallowness of the water in which most of it must have been laid down. The character of sedimentation and perhaps also the development of organisms probably depend more on this than on whether or not the water was marine. In a later paper Walcott (1916a, p. 236–237) amplified his ideas. He thought that the North American continent was larger at the beginning of Cambrian time than at any subsequent period, except possibly at the ends of the Paleozoic and Mesozoic eras. He thought that this "presupposes that the great series of pre-Cambrian Algonkian sediments in the Rocky Mountain region were deposited in an inland mediterranean, or a series of great lakes and flood plains such as existed in Tertiary times." The comparison with continental deposits of Tertiary age is one of the weak points in his argument. The differences between those deposits and the Belt series are many and varied.

Apparently Walcott eventually returned to the concept of a marine origin for the Belt series because of the abundance of limestone and dolomite in it, and because of the rare casts of salt crystals in its clastic beds. This reversal of opinion is recorded by the Fentons (1937b, p. 1935) on the testimony of Mrs. Walcott. Actually Walcott's change in viewpoint was not great. In his original paper (1914a, p. 3) he spoke of the Belt series, and similar rocks, as deposited in "fresh or brackish-water seas." Such seas would not be vastly different from the seas in which many Paleozoic and later rocks, commonly called marine, were deposited. A major difference, as originally conceived by Walcott, seems to be that he thought that the Belt seas were less commonly connected directly with "extra-continental seas," or oceans, than later seas that overspread parts of the continents are supposed to have been.

The argillaceous beds and probably some of the quartzite beds were deposited in shallow water. The ubiquitous ripple marks are among the bits of evidence pointing to this conclusion. They are mainly of types commonly interpreted as originating in shallow to very shallow water (Fenton and Fenton, 1937b, p. 1913–1918). The Fentons recorded amplitudes of symmetrical ripple marks that range from 5 to 45 mm, the larger ones being rare. They noted wavelengths ranging from 2.5 to 40 cm. In the Bonner quadrangle (Nelson and Dobell, 1959) symmetrical ripple marks are similarly abundant. Most wavelengths range from 25 to 50 mm but some are several times this. Ripple marks preserved in rocks of the Belt series have not been intensively studied. As the Fentons remark (1937b, p. 1913–1918), various kinds are present, but the oscillation type seems to be the most prevalent. Most students (Kindle, 1917, p. 29; Twenhofel, 1939, p. 520–521; Bucher, 1919, p. 186–188; Pettijohn, 1957, p. 184–187; Evans, 1949) would regard the most common kinds of ripple marks in the Belt series as products of wave action in very shallow water, although the more recent reports increase the emphasis on the uncertainties of interpretation of fossil ripple marks. This uncertainty has been added to by proof that ripples can form at considerable depths in the ocean by processes not yet understood (Ericson and others, 1955, p. 209; Menard, 1952). As far as the Belt series is concerned, the abundance of small oscillation ripple marks is merely one of several bits of evidence that point to deposition in very shallow water.

Mud cracks, raindrop impressions, and similar features show that at times muds were exposed to the air, with essentially no cover of water. These features can hardly be ascribed to tidal fluctuations. Whether the water body or bodies in the Belt basin are thought of as lakes or as seas, they must have been so nearly landlocked that tides had little effect. It seems a fair guess that the argillaceous and quartzitic rocks were deposited in water that was rarely more than a few hundred feet deep. Perhaps the depth was commonly less than a hundred feet. For protracted periods the sediments were awash or even above water.

Insofar as the carbonate rocks are genetically related to stromatolites they too must have been deposited in shallow water. It is commonly assumed that algae cannot grow at depths below that to which sunlight penetrates, say 90 feet. Recent studies by Richard Rezak have led him to the conclusion that the stromatolites of the Belt series, particularly in Glacier National Park, are of forms akin to those produced at present by algae, mainly marine, that grow in water so shallow that for notable stretches of time the algal heads are not covered. Certainly none of the fossils now recognized in the Belt series are opposed to the concept that the water in which that series was deposited was, on the whole, distinctly shallower than that of most of the epicontinental seas of later eras. Even the latter were surely far shallower than the major expanses of the open oceans of today.

If any of the carbonate rocks of the Belt series are chemical precipitates of the kind suggested by Daly, this might be taken as evidence opposed to the concept of deposition at shallow depths. He (Daly, 1912, p. 660) speaks of calcium carbonate precipitates in the Black Sea, whose greatest depth he cites as 1,227 fathoms. Arkhangelskii (1927) confirms the presence there of calcium carbonate deposits. However, he attributes the relative abundance of calcareous mud to lack of clastic deposits rather than to precipitation of carbonate. The calcareous tufa of modern lakes and ephemeral streams is not strictly analogous to the precipitates Daly had in mind, but it is a precipitate of calcium carbonate in shallow water. Thus, even if some part of the carbonate in rocks of the Belt series originated as a chemical precipitate, it need not have formed in deep water.

Casts of salt crystals have been reported from various localities in rocks of the Belt series. As far as known, they are nowhere abundant. None of the available evidence suggests that sediments of the Belt series anywhere included beds of any composition, such as would be expected as deposits from highly saline water. One can merely speculate that the water was saline enough so that minor amounts of salt were incorporated in the muds and some of this aggregated into salt crystals, whose casts are now preserved. The few salt casts that have been found are, at the least, not opposed to the idea that the water in the Belt basin may have had a composition broadly similar to that of the ocean, but was not as concentrated as that in many modern saline lakes. As noted above, the character of the stromatolites is compatible with this idea.

The fact that many of the formations in the Belt series have not been traced far beyond their type localities might be taken as proof that the water bodies in which they originated were of limited extent. Marked differences in the details of stratigraphic succession within formations in many of the relatively few places where details of that sort have been obtained might also be taken to support a concept of deposition in an assemblage of lakes and swamps, rather than a single basin. However, the broadly uniform character of each of the groups into which the greater part of the series is divided is opposed to any such an interpretation. Further, throughout all of the region underlain by the Belt series, except its southeastern and northwestern margins, no evidence has been found of shores or of masses of rock of pre-Belt age rising above the level of the basin of deposition. As far as can be judged, the various major components of the series were deposited in essential continuity throughout most of the region of outcrop. Minor components, even some of considerable thickness, were limited in original extent, thereby furnishing the basis for some of the problems in stratigraphic correlation that have arisen. Lateral differences of this sort seem inevitable and inherent in deposits in water as shallow as that of the Belt basin must have been. Slight changes in currents within the basin, fluctuations in the character, and amount of the sediment brought in by tributary streams would have been among the reasons for variations in the deposits. Algal reefs may have been large enough in places to have had much the effect of islands. The weather may not have been uniform over a basin hundreds of miles in width, and of unknown, but large, longitudinal dimensions.

The data above outlined appear to indicate that the Belt series was laid down in shallow, saline, brackish water in a basin that in Montana and Idaho, at least, was essentially an uninterrupted expanse of water and mud flats. A geographic feature as large as the water body in the Belt basin seems more appropriately called a sea than a lake. The fact that it was so shallow one must suppose that at any given time mud flats interrupted the sweep of the water does not seem to nullify this. If, as suggested below, the Belt basin was connected with the open oceans, the resemblance to epeiric seas of later periods becomes pronounced.

THE UPPER LIMIT OF THE BELT SERIES

Fixing the boundary between strata of Precambrian and Cambrian ages invariably presents problems wherever the Precambrian rocks are sedimentary and not of such a character and relation as to be generally accepted as very old (Rogers, 1956). The boundary has been placed at the first clearly evident unconformity below unquestioned Paleozoic rocks; at the base of strata containing an Olenellus fauna or its equivalent; and, where the Olenellus fauna is lacking or not regarded as offering a good basis for division, some horizon between rocks of accepted Precambrian age and those of accepted Paleozoic age has been selected. In Montana, Lower Cambrian fossils are lacking and the lowermost of the recognized Paleozoic units (the Flathead quartzite of Middle Cambrian age) is unfossiliferous. Hence paleontologic criteria for fixing the upper limit of the Belt series are inadequate. The top of the series is not everywhere marked by a clearly demonstrable unconformity, making this means of fixing the upper limit uncertain. Where the Flathead quartzite is well represented, and especially where fossiliferous beds supercumbent on the Flathead are preserved, the top of the Belt series can be mapped with a fair degree of assurance. In places in Montana the position of the top is in doubt, and when correlation is carried into areas in Canada where fossiliferous Lower Cambrian rocks exist complications increase. Age relations between parts of the Belt series in and near Montana and rocks of demonstrably Early Cambrian age have the uncertainties that arise wherever diagnostic fossils are absent in significant parts of a given sequence. In situations of this kind transgressions in time can be postulated but are often difficult to prove. The Belt series is one of those for which this difficulty has arisen.

One of the concepts involved is that of the Lipalian interval (Walcott, 1914a, p. 1–15). According to this concept the continents were above sea level during that interval, which was in late Precambrian time, and, in consequence, the oceans were unable to spread over the continents and marine deposits could not be formed within the continental areas. The concept was advanced as an explanation of the fact that hard-shelled fossils, usable for stratigraphic correlations, are not known in Precambrian rocks. Walcott postulated that evolution was not as abrupt as the sudden appearance of hard-shelled fossils seemed to imply. He supposed that life similar to that recorded in the Cambrian rocks developed along the shores and shelves of the Precambrian continents, but that the rocks there formed are buried beneath the present oceans and hence not now accessible for study. He defined the Lipalian interval as "the era of unknown marine sedimentation between the adjustment of pelagic life to littoral conditions and the appearance of the Lower Cambrian fauna." His definition implied that all of the upper Precambrian strata now accessible, including the Belt series, are nonmarine and that the Paleozoic era began with the deposition of beds of recognizable marine character. The basal Paleozoic strata would thus either contain marine fossils of Cambrian age or would be related clearly to unfossiliferous quartzite or conglomerate below. Snyder (1947, p. 152) points out that the Lipalian interval and the unconformity at the top of the Precambrian rocks should not be confused with each other, as Hinds (1936a) has done.

In North America and elsewhere such great thicknesses of strata now regarded as marine lie beneath the earliest fossiliferous Paleozoic rocks that Walcott's idea of a Lipalian interval no longer seems valid, a conclusion already announced by Snyder (1947). Changes in viewpoint as to what constitutes marine deposits have contributed toward weakening the Lipalian concept and reconsideration of the significant paleontological data (Raymond, 1947, p. 19–28) shows that Cambrian fossils do not represent such an abrupt step in evolution as Walcott supposed. Recently (Gussow, 1957) Lipalian has been revived in a new sense, that of a system equivalent to certain components of the Windermere series of Canadian geologists, which is the upper part of the Belt series of the present paper. Gussow proposes that beds of the Purcell series, below those of his Lipalian system, be assigned to his Beltian system. His proposals do not seem to have received acceptance in Canada and have not been adopted in the United States.

Most geologists in the United States assume that the Belt series is bounded at the top by an unconformity. Deiss (1935) has assembled data favoring this concept, drawn largely from Montana. In some places in that state unconformable relations do exist, but in others the concept is open to argument. Few exposures in Montana show large angular discordances. Further, as already noted, Okulitch (1949) believes that, along the boundary between British Columbia and Washington, a single lithologic unit, the Hamill series, itself a component of the Windermere series, is of Precambrian age in one locality and Cambrian in another, a situation analogous to that in and near southern Nevada (Wheeler, 1947). In some of the Canadian reports already cited the base of the Cambrian rocks is spoken of as an unconformity but in several districts the strata on both sides of the contact are parallel and unconformity is not demonstrated clearly by the published data. Nowhere is a sharp angular unconformity recorded. In central Idaho discordance between the Belt series and the oldest Paleozoic rocks is evident (Ross and Forrester, 1958, p. 29–30). Even here, however, the angular discordance is rarely great.

In most places in Montana the earliest known Paleozoic rocks are of Middle Cambrian date (Flathead quartzite) and in parts of southwestern Montana and south-central Idaho strata of Ordovician age (Kinnikinic quartzite) rest on rock assigned to the Belt series. Where the contact is a visible unconformity the matter is simple, but in many places the evidence is inconclusive. The doubt arises largely from the fact that commonly the beds above and below the supposed top of the Belt series are nearly parallel to each other. In any place where an unconformity is absent or minor, the uppermost parts of the rocks that have been assigned to the Belt series may have actually been deposited at the same time as strata in Canada that contain Cambrian fossils. The upper part of the Belt sequence in Montana contains quartzite and argillaceous quartzite, not well adapted to the preservation of fossils, and the lowest unit of generally acknowledged Paleozoic age is the unfossiliferous Flathead quartzite of Middle Cambrian age. In areas such as the Canyon Ferry quadrangle (Mertie, Fischer, and Hobbs, 1951) the Three Forks quadrangle (G. D. Robinson, 1958, written communication) and the Duck Creek quadrangle (W. H. Nelson, 1958, written communication) the Flathead rests on different components of the Belt series in different places in such a manner as to imply unconformity by overlap. The idea of unconformity by overlap has been advanced (Deiss, 1935) for areas near and north of Missoula also, but without enough map data to be convincing. There the Missoula group is far thicker than it is known to be anywhere else. This in itself suggests that some or all of the upper beds in the Missoula region may have no equivalents in the Belt series farther east and southeast. As Deiss has indicated, Calkins' assignment of the top of the series in the Philipsburg quadrangle to the Spokane is no longer valid. The stratigraphic details of the Missoula group vary markedly from place to place, which is to be expected of an assemblage of strata formed in exceedingly shallow water under conditions that must have shifted often. Sedimentation in the central part of the Belt basin in Missoula time was different from that in the site of the present Big Belt Mountains, where overlap is believed to have been demonstrated. In the central part of the basin the suggestion of overlap is unproved. In the Flathead region differences in lithologic character between beds regarded as Paleozoic and those regarded as belonging to the Belt series exist locally but no proof of unconformity has been found. Furthermore, the recorded lithologic differences between these two are so minor that they require close observation to recognize. Differences between rocks immediately above the top of the Belt series as mapped and those they rest on are no greater than those between some of the quartzite now regarded as Cambrian and quartzite strata within the Missoula group in the same region.

Eight localities where angular discordance between beds of Belt and post-Belt age have been reported are recorded by Deiss (1935, p. 96–103, 119–122). Two are in the Philipsburg quadrangle, two near Helena, two in the Coopers Lake quadrangle, one in the Little Belt Mountains, and one in the Big Snowy Mountains. The brief descriptions that have been published indicate that in each of these localities the angular discordance has been observed only in a single outcrop. The unconformity in the southwest part of the Philipsburg quadrangle (Emmons and Calkins, 1913, p. 50–51), although apparently visible only in a single cliff face, has the largest angular discordance so far recorded. At this spot the Flathead quartzite (Cambrian) strikes nearly north and dips 25° W., whereas the underlying strata of the Belt series strike north-northeast and dip 50°–60° W. Poulter (1957) recorded angular discordances in the Philipsburg quadrangle, notably one of 15° half a mile west of Rainbow Mountain and about 3 miles west of Mount Tiny in unsurveyed sec. 34, T. 4 N., R. 14 W. Poulter stresses the close similarity between the Flathead quartzite and the Belt rocks underlying it where he worked.

The locality in the southeastern part of the Philipsburg quadrangle where Calkins recorded the presence of an unconformity was visited in 1955 by W. H. Nelson, J. P. Dobell, and the writer. The angular discordance here is slight and is better appreciated in views from a distance than it is on closer inspection. At one place where it is advantageously exposed, the Flathead quartzite above the unconformity strikes N. 60° E. and dips 21° W., whereas the strata of the Missoula group below strike N. 58° E. and dip 25° W. The unconformity is visible for a distance of more than a mile along a valley wall. It surely records a time gap in the sedimentary record, but the extent of the gap is indeterminate. In rocks of the character involved here it need not be large. The conglomerate above the unconformity contains pebbles that appear to have been derived from the underlying rocks of the Belt series after the latter had been compacted and cemented. Perhaps this is the most significant bit of evidence pointing to a time gap of some duration. However, conglomerate occurs at many places and at various horizons in and above the Belt series, so that its presence here constitutes no proof of a major unconformity.

The two unconformities near Helena (Walcott, 1899, p. 212) are so minor, according to the published diagrams, as to be of no material regional significance. The same can be said of the angular discordance at one place in the Little Belt Mountains (Walcott, 1899, p. 216) and one reported from the Coopers Lake quadrangle (Deiss, 1935, p. 121).

Verrall (1955, p. 26, 27, 33) reports two localities northeast of Logan where angular discordance between the Belt series and the overlying Flathead quartzite is visible. He says that the contact cuts down through the Belt rocks at an angle of 10° in such fashion that the top of the Belt sequence at one locality is 5,000 feet stratigraphically lower than at a locality some 6 miles away.

Another instance of angular unconformity at the top of the Belt series remains to be mentioned. This is in the Big Snowy Mountains so far to the east of most outcrops of the Belt series that assignment to the Belt series would be open to argument if any more probable assignment could be offered. In at least one place in those mountains (Calvert, W. R., quoted in Walcott, 1916b, p. 273–274; also Reeves, 1931, p. 145) rocks that seem best regarded as Belt show an angular discordance of 19° with crossbedded quartz sandstone, correlated by Calvert with the Flathead quartzite, resting horizontally on them. In the Little Rocky Mountains also (Knechtel and others, 1944) an unconformity at the base of the Paleozoic section is recorded, but the Precambrian rocks below are regarded as older than the Belt series. Both the Big Snowy Mountains and the Little Rocky Mountains are too far east to be shown in the maps accompanying the present report.

It may be significant that the outcrops in, near, and east of the Big and Little Belt Mountains cited above as showing, in limited areas, distinct angular unconformities are near the eastern border of the Belt basin and those in the southern part of the Philipsburg quadrangle are not far from the probable position of another part of the shore of that basin.

Walcott and later Deiss relied mainly on the evidence of regional overlap rather than the scattered exposures of angular discordances to establish the postulated break between the Belt series and the Cambrian strata. More recent work in and near the Big Belt Mountains has furnished support for the concept in that area but in the rest of western Montana corroborative data are scanty and, as noted below, studies since Deiss' summary was published cast doubt on his inferences. His deductions are illustrated and supported largely by a series of diagrammatic cross sections (Deiss, 1935, pl. 8, p. 106–113), which Nelson and Dobell (1961) show to be based in part on erroneous data.

Near Missoula, Mont. (Nelson and Dobell, 1961), rocks formerly supposed to belong to the Belt series recently have yielded fossils of Early Paleozoic age. In the Bonner quadrangle near Missoula, where exposures have been visited by this author in company with Nelson and Dobell, a formation named by them, the Pilcher quartzite, seems to pass downward through a gradation zone 50 to 100 feet thick into the Garnet Range quartzite below. They point out the possibility that the strata interpreted as a gradation zone may instead be products of reworking of the upper part of the Garnet Range quartzite along an unconformity. This interpretation is a conceivable one but has little to support it. Even if the supposed gradation zone incorporates material derived from the immediately underlying Garnet Range quartzite and, as is possible, that quartzite is of variable thickness, no age break other than could be expected from some minor change in conditions of deposition in the shallow Belt basin need be required. The argument based on variations in thickness might be applied with equal validity to many of the divisions of the Belt series. In the case of the Garnet Range quartzite it has the additional weakness that intricate deformation has prevented accurate thickness measurements.

The Pilcher quartzite is the uppermost unit assigned to the Belt series in the Bonner quadrangle. About 55 miles west of Missoula, rocks that resemble the Pilcher quartzite are exposed in a small area and are bounded by faults and by Quaternary deposits so that their relations to other rocks are indeterminate (Campbell, 1960, p. 569–572). However, Campbell (1960, p. 571–573), on the basis of regional relations, assumes an unconformity. Nelson and Dobell (1961) show that the Pilcher quartzite has so many resemblances to the Flathead quartzite farther east that correlation is possible. This possibility of correlation is strengthened by the fact that the Pilcher quartzite is overlain successively by shale that resembles the Wolsey shale and by limestone that resembles the Meagher limestone. Such a succession would be a normal one if the Pilcher quartzite is equivalent to the Flathead quartzite. In the Bonner quadrangle the shale and limestone just mentioned are assigned a Cambrian(?) age. The Flathead, Wolsey, and Meagher are all generally regarded as of Middle Cambrian age. Correlation of the Pilcher with the Flathead would imply a Middle Cambrian age for the Pilcher and suggest the interesting possibility that some part of the rocks now assigned to the Missoula group may be of Early Cambrian age. This would accord with Okulitch's ideas, summarized below.

In Montana northeast of Missoula and Superior data on the Flathead quartzite and its relations to the Belt series are fragmentary. Deiss (1943a, p. 219; 1939, p. 6–34) applied the name "Flathead sandstone" to rocks in the Coopers Lake quadrangle, unconformably on the Missoula group; some in the southwestern part of the Saypo quadrangle, resting "slightly unconformably on" green argillite of the Belt series; rocks on Pentagon Mountain and Pagoda Mountain in the Silvertip quadrangle; rocks in Nannie Basin and Kid Mountain in the Ovando quadrangle, referred to as in apparent conformity on the upper beds of the Missoula group; also to rocks near the head of Dearborn River, where the basal contact is conformable. In his unpublished geologic map of the Silvertip quadrangle, Deiss mapped his Flathead sandstone and Gordon shale together but along many miles of contact did not distinguish this unit from the Cambrian rocks resting on the Missoula group. Deiss used local names for Cambrian units above the Flathead, not those in use in areas to the south and east, mentioned in the preceding paragraph. In the Flathead region (Ross, 1959), north of the areas where Deiss worked, the base of the Paleozoic sequence is rarely exposed. On Cruiser Mountain in the southern part of the region, white quartzite conformable on strata of the Missoula group is regarded as assignable to the Flathead quartzite. On Gable Peaks, nearby, Flathead quartzite was noted also.

Okulitch (1949, p. 16–20), whose field of activity has been north of lat. 50°15′ and east of long. 118°, has presented a summary of his concepts as to the Precambrian-Cambrian boundary. He states with reference to the Cordilleran geosyncline during latest Precambrian and early Cambrian times: "No great and widespread orogenic movements are known to have occurred, and truly regional unconformities, with the exception of one at the base of the Windermere, are lacking." He thinks that an eastward marine transgression began in very late Precambrian (Windermere) time and that, in consequence, each of the formations, which are based on lithologic characteristics, becomes progressively younger eastward. This idea is similar to that expressed by Wheeler (1947), for southwestern Nevada.

Farther west, where British Columbia borders on northeastern Washington, the Cambrian and Precambrian rocks differ materially from those in Montana. Special problems are raised by the Windermere series. In that region the boundary between rocks assigned to the Cambrian and the Precambrian has fluctuated within rather wide limits with the judgment of each of the investigators (Daly, 1912, p. 174–194; Walker, 1926, p. 17–19; 1934, p. 2–10; Park and Cannon, 1943, p. 11–15; Little, 1950, p. 10–11). Whatever the correct position of the boundary may be, it is clear that no discernible break exists between beds of Early Cambrian age and those that Canadian geologists regard as Precambrian. There are, however, unconformities within the latter.

In southeastern British Columbia, unlike Montana, fossiliferous beds of Early Cambrian age rest on strata that are correlated with the Belt series (Schofield, 1922). No angular discordance has been recognized but the beds above and below the contact are reported to differ in degree of metamorphism. Conglomerate in the Cambrian strata contains pebbles that appear to have been derived from the Precambrian rocks after the latter had been metamorphosed. In different localities the Cambrian rocks rest on different units in the Precambrian assemblage, but this bit of evidence is weakened by the fact that the lowest units of Cambrian age likewise differ in the different localities, which might imply local variations in conditions rather than a major time gap. Schofield lays some stress on the fact that the rocks of Cambrian age appear to be of marine origin, whereas the underlying strata differ in such a way as to indicate deposition in shallower water. This difference argues for a change in conditions, but not necessarily for such a time gap as would be implied by an unconformity of any consequence.

In northern Idaho Paleozoic strata have been recognized only in small fault blocks near Pend Oreille Lake. In that locality the three sedimentary formations in these fault blocks have been referred to the Cambrian (Sampson, 1928, p. 9–10). The upper two of these contain abundant Middle Cambrian fossils (Resser, 1938b). Sampson assumes that the Cambrian rocks are separated from the Belt series by a major unconformity because of "evidence from nearby districts" but says that within the area he examined, no useful information on that point was obtained because of poor exposures of the critical contact. Thus, his assumption of unconformity merely reflects prevailing opinion at the time he wrote.

In south-central Idaho, strata tentatively regarded as Cambrian have been mapped in one locality (Ross, 1938, p. 12–14) but their relation to Precambrian rocks is unknown. In other parts of that region Ordovician strata rest on the Belt rocks, and in the Lemhi Range the contact is an angular unconformity, at least locally (Ross, 1947, p. 1099; 1959). In most places in south-central Idaho, especially in its eastern part, available evidence as to the date of initial deformation of the Belt series is scanty, but in most areas the assumption has been that pre-Paleozoic deformation and metamorphism have occurred (Umpleby, 1913a, p. 49; 1913b, p. 33; 1917, p. 38; Ross, 1934, p. 72–74). Recent unpublished work by B. F. Leonard and the author in areas north of Yellow Pine tends to confirm this, but the question is still under study.

In those reports on areas in Montana that refer to the degree of metamorphism, the Belt rocks are commonly stated to be more metamorphosed than the Paleozoic strata, but few details are given to substantiate this. From microscopic examination of specimens of the Belt series from near Missoula and from areas south of Glacier National Park, the writer has the impression that these rocks are thoroughly indurated and somewhat recrystallized, but, except in some places close to intrusive bodies, contain no minerals suggesting high or even moderate metamorphism. Specimens from the Missoula group, the youngest in the series, seem to show less evidence of recrystallization than those from older units. In this respect the quartzite and argillite of the Missoula group are almost or quite indistinguishable from Paleozoic rocks of similar original compositions.

The summary just given shows that data bearing on problems relative to the boundary between the Belt series and later beds are scanty and in part contradictory. In some areas there is demonstrable unconformity; in others there is conformity or, at the least, absence of angular disconformity. The writer's observations in many parts of the area of exposure of the Belt series leave him with the strong impression that lateral variation and intertonguing of lenticular assemblages of varied appearance are the rule, rather than the exception. This, coupled with crossbedding and other irregularities, combines with local complexities of structure to cast doubt on theories of unconformity except in those few places where good exposures have been studied in detail. Where variations in thickness and distribution of Belt units, particularly the Missoula group, have been regarded as proof of erosion and, hence, of unconformity, it should be remembered that differences in original thicknesses and limitations in the areas covered by particular units would produce similar effects. Variations of these kinds are inevitable under the conditions that governed the deposition of the Belt series. Also, as the Belt sea was so shallow that its deposits were not continuously submerged, erosion may well have taken place locally at various times. Variations of the sort here commented on are abundant in the Missoula group and are present elsewhere in the series, even in those parts, such as the Ravalli group, where broad uniformity in lithology persists over broad areas.

The best supported instances of unconformity in and adjacent to Montana are close to the postulated border of the original Belt basin. Breaks in erosional history and, even, diastrophic disturbances near the borders of the Belt basin would not necessarily affect the broad, central part of the basin. This idea cannot be adequately tested because large areas within the basin no longer retain remnants of recognized Paleozoic strata. The exceptional thicknesses of the Missoula group near Missoula and Superior and thence northeastward to Glacier National Park suggest that deposition of strata now assigned to the Belt series may have persisted longer in the central area than it did in marginal parts of the Belt basin, a fact that accords with the idea that some of these strata may be of Paleozoic, rather than Precambrian, age.

The Flathead quartzite has so many similarities to quartzitic bodies at various horizons in the Missoula group that misinterpretation and confusion are easily possible. Both in the writer's field notes and in many of the published reports lithologic similarities of this sort are stressed repeatedly. The Pilcher quartzite is an example of the sort of confusion that can result. Even where the Flathead quartzite has been correctly identified and mapped, the presence within the Missoula group of similar rocks indicates repetition of conditions of deposition.

The problems resulting from similarities between the Flathead quartzite and quartzite in the upper part of the Belt series are illustrated northwest of Argenta (Myers, 1952, p. 4–6; Hanson, 1952, p. 12–13). Here, rock originally regarded as an usually thick representative of the Flathead quartzite is best regarded as part of the Belt series (W. B. Myers, 1958, written communication), already known to include several quartzitic components. On this basis the Flathead quartzite is absent over a large area in Beaverhead County, Mont. Another example is reported (P. A. Guttormsen, 1952, written communication) in T. 11 S., R. 11 W., where the Hasmark formation is thought to lie directly on quartzitic Belt rocks, the Silver Hill formation and Flathead quartzite being absent. A possible alternative interpretation here is that the quartzite assigned to the Belt series, or some part of it, is actually Flathead quartzite.

NORTH AMERICAN RELATIVES OF THE BELT SERIES

A basin as large and as deeply filled with sediments as the Belt basin was, must be supposed to have played a part in the development of the North American continent. From this viewpoint, data on sedimentary rocks in the Rocky Mountain region now regarded as of

Precambrian age and in varying degree resembling the Belt series are summarized below. Figure 7 shows their distribution. This map also shows rocks farther east that have stratigraphic characteristics similar to those of the Belt series. These rocks have been referred to on page 8, but are not discussed here as they are not believed to have been connected with the Belt series during deposition. As an aid in consideration of Belt paleogeography, figure 7 also shows places where strata supposed to be of Early Cambrian age are known to be close to strata that have some kinship with the Belt series. Reasons have been given in previous parts of this report for the belief that, in places, sedimentation may have been continuous from Belt into Cambrian time. Thus, the distribution of sedimentary rocks of Cambrian age needs to be taken into consideration in discussion of Belt paleogeography. The data for fig. 7 come from various sources. In the United States and Mexico reliance was placed mainly on the original reports on the areas involved and on state geologic maps, where available. The geologic map of Alaska (Dutro and Payne, 1954) was similarly used. In Canada the 1955 map was used and interpretations are based largely on Reesor's compliations of data (Reesor, 1957). Papers on Greenland (Troelsen, 1949, 1950; Koch, 1929b, 1933) were of much value. An advance copy of the geologic map of North America that is in preparation for publication was used as an overall guide. Interpretations of the age relations of upper Precambrian rocks to each other and to strata of Early Cambrian age have differed and fluctuated so much that it has been necessary to decide among the various possible interpretations. Figure 7 shows areas within which the different rocks are thought to be present, rather than the individual localities where present outcrops are mapped.

IDAHO

In south-central Idaho north of Hailey two formations, the Hyndman and East Fork, have been tentatively regarded as of late Precambrian age (Umpleby, Westgate, and Ross, 1930, p. 9–17). The Hyndman and East Fork formations are far south of any of the rocks so far assigned to the Belt series. They occupy a small area close to intrusive masses and are intricately faulted and highly metamorphosed. Many of their metamorphic features are related to the intrusive activity and, hence, have no bearing on the age of the beds.

Field inspection by the present writer leaves the impression that the Hyndman and East Fork formations are, as has been assumed in previous studies, older than any of the Paleozoic rocks nearby but furnishes no basis for decision as to their relationship to the Belt series. For the present, Anderson's assumption (1931, p. 24–29) that they are of pre-Belt age may be accepted as logical, although unproven. No exposures assumed to be older than the Belt series are known farther north in central Idaho. Some have argued for a pre-Belt age for certain metamorphic rocks near Shoup, Lemhi County (Umpleby, 1913a, p. 30–32), but personal observations and the conclusions reached in a recent study by Kaiser (1956) favor the concept that these rocks are of Belt age, although in part so permeated under stress by material related to the Idaho batholith as to be gneissic.

Southern Idaho contains two areas of strata of supposed Precambrian age. One, near Pocatello, has been divided (Ludlum, 1942, 1943) into the Bannock volcanic formation at the base, the Pocatello formation, supposed to be largely of glacial origin, and the Blackrock limestone at the top, reportedly unconformably overlain by the Brigham quartzite (Middle and Lower(?) Cambrian). Ludlum's assignment to the Precambrian is based on his belief that the three formations are more metamorphosed than the Paleozoic rocks of the region and that the Blackrock limestone is unconformably under the Brigham quartzite. Anderson (1928, p. 3–4) thought the contact was conformable, so the break cannot be a marked one. Originally the rocks now termed Precambrian were supposed to be of Ordovician age (Weeks and Heikes, 1908, p. 178–179) and later they were regarded tentatively as Early Cambrian (Anderson, 1928, p. 3–4) although the possibility of a Precambrian age was recognized.

Another area underlain by Precambrian(?) rocks is far to the west in Cassia County. Here Anderson (1931, p. 24–29) has named and described a thick sequence of quartzite beds, with some schist, originally termed the Harrison series and later the Albion Range group as the first-chosen name is preoccupied. He postulated that these rocks were of pre-Belt age because they are more metamorphosed than most of the Belt series and because he thought they had resemblances to the Hyndman and East Fork formations, which are also metamorphosed. The metamorphism in all three units is believed to be related to igneous activity of Mesozoic or later date (Anderson, 1934, p. 376–392). The evidence of greater metamorphism thus seems irrelevant. However, Anderson's remarks that the Albion Range group is coextensive with rocks in the Raft River Range in Utah, believed to be Precambrian (Butler and others, 1920, p. 78) and to resemble the lower part of the Precambrian sequence in the Wasatch Range (Blackwelder, 1910, 1935), may be more significant. The last-cited unit is presumably of pre-Belt age.

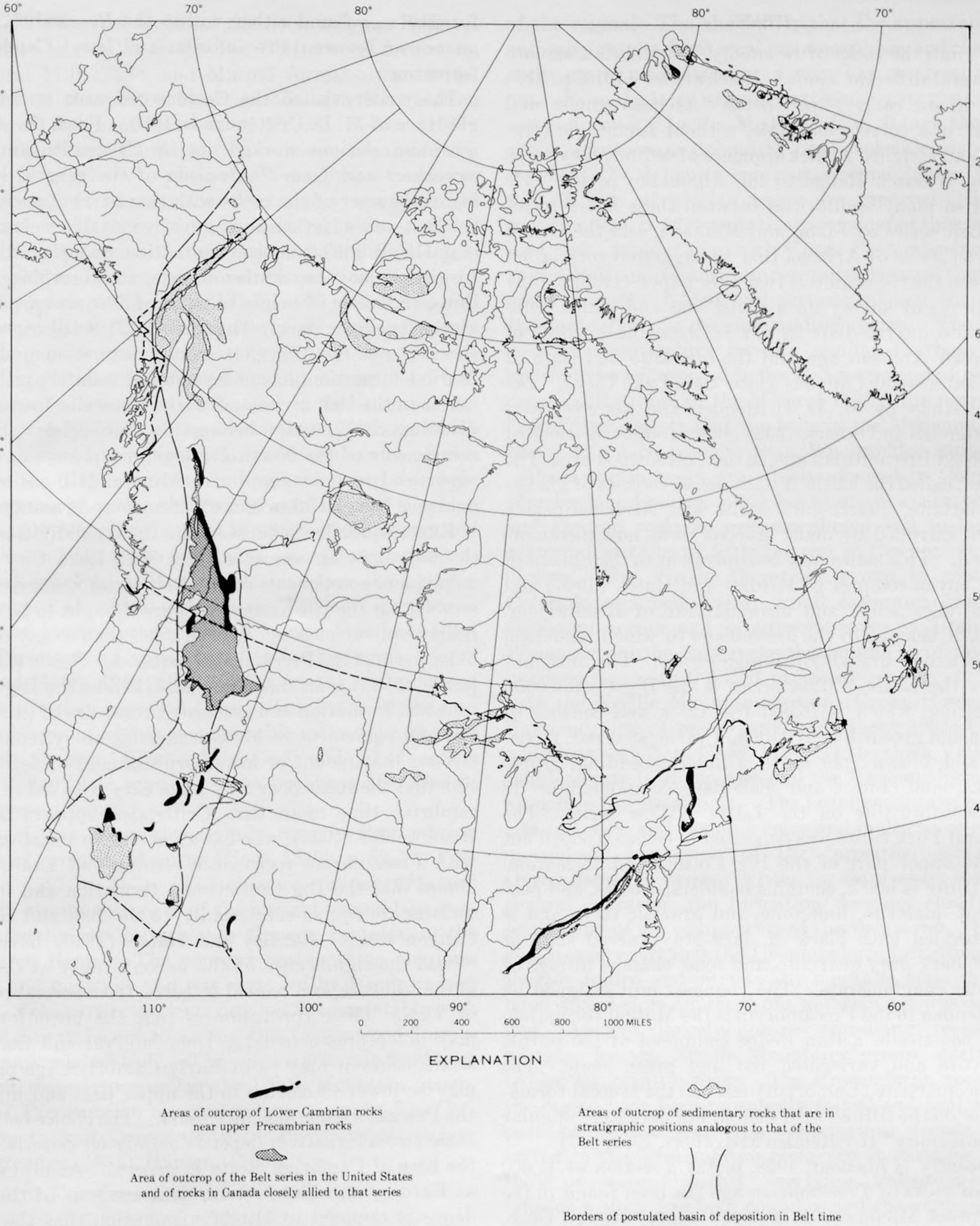

FIGURE 7.—Map of North America showing the distribution of the Belt series and broadly related rocks, and the inferred zone of sedimentation in the western part of the continent in Belt time. The distribution of Lower Cambrian rocks is also indicated.

UTAH

In Utah the rocks of reputedly Precambrian age are of several different kinds. Blackwelder (1910, p. 518–526) in an early paper listed "Archean gneiss and schist" in a correlation table, without further description, and assigned a thick sequence of sedimentary rocks in the Wasatch Range to the Algonkian period. He noted so many similarities between these latter rocks, the Belt series of Idaho and Montana, and the Grand Canyon series of Arizona that he suggested correlation between the three units. In a later paper (1925, p. 132) he speaks of silvery mica schist and and white metaquartzite intermediate in age between the gneisses of supposed Archean age and the quartzite and slate of the Cottonwood Canyons (near Salt Lake City). Detailed study by M. D. Crittenden and his coworkers (Crittenden and others, 1952) shows four formations of supposed Precambrian age, in the Cottonwood area. The lowest, called the Little Willow series, consists of gneissic quartzite, quartz-mica schist and stretched-pebble schist, intruded by mafic igneous rocks now metamorphosed. This series may be equivalent to the Farmington Canyon complex of Eardley and Hatch (1940) and to the mica schist and metaquartzite of Blackwelder. If so, it belongs to the assemblage to which Anderson would assign his Albion Range group. The next unit above the Little Willow series is the Big Cottonwood formation, which is 16,000 feet thick and consists of white and greenish quartzite and variegated red, greenish, and blue-purple shale with crossbedding, ripple marks, mud cracks, and shale-flake conglomerate. It is unconformable on the Little Willow series. The Mineral Fork tillite was deposited in basins scooped out of the upper part of the Big Cottonwood formation. The tillite is black, contains boulders, cobbles, and pebbles of quartzite, limestone, and granitic rocks and is interbedded with black or dark-gray varved slate or shale, dark-gray quartzite, and some channel fillings of boulder conglomerate. The youngest unit assigned by Crittenden to the Precambrian is the Mutual formation. This constitutes a thin wedge composed of red-purple quartzite and variegated red and green shale. The Tintic quartzite (Cambrian) rests on the Mutual formation or on the tillite "with a small but persistent angular unconformity" (Crittenden and others, 1952, p. 7).

Recently (Cohenour, 1958, p. 12) a section of 11,000 feet of rocks of Precambrian age has been found in the Sheeprock Mountains in Tooele County, western Utah. These rocks have resemblances to those in the Cottonwood area, including the presence of rocks regarded as tillite. The rocks in the Sheeprock Mountains are reported to be of continental origin with lagoonal and estuarine deposits plentiful. No marked angular conformity was found either within the Precambrian sequence or between Precambrian and basal Cambrian formations.

The writer visited the Cottonwood area under the guidance of M. D. Crittenden in 1950. From the stratigraphic relations worked out by Crittenden and his coworkers and from comparison of the general lithologic character of the rocks with that of components of the Belt series farther north, it is reasonable to suppose that the Big Cottonwood formation and the Mutual formation may be similar in age, stratigraphic relations, and mode of origin to parts of that series, particularly its upper part. The Mineral Hill tillite, which lies between the Big Cottonwood formation and the Mutual formation, has no recognized counterpart anywhere in the Belt series, unless the Pocatello formation (Ludlum, 1942, 1943) belongs to that series. Some components of the North Boulder group have at least superficial resemblances to the Mineral Hill tillite but evidence for glacial origin of that group is scanty.

Exact equivalence between the three subdivisions of the Precambrian sequence near Salt Lake City just named and components of the Belt series is not demonstrable, but the two assemblages are thought to be more nearly equivalent in age to each other than to any of the older parts of the Precambrian sequence. Hinds (1936b, p. 88, 120-121), on the other hand, thinks the Big Cottonwood formation and associated rocks are of pre-Belt age and equivalent to his Uncompahgran system. He stresses the amount of metamorphism and of deformation that the rocks near Salt Lake City received in Precambrian time more than Crittenden appears to do. Blackwelder (1935), who has an extensive acquaintance with Precambrian rocks in Wyoming and Utah, concludes that the Big Cottonwood formation and its associates are equivalent, broadly, to the Belt and Grand Canyon series. Eardley and Hatch (1940) have discussed the significance of the unconformity at the top of the sequence, and related features, from several points of view. Their conclusion is that the unconformity may not represent a large time interval and that the strata below it may be of Early Cambrian age or else may be lower Cambrian in the upper part and high in the Precambrian in the lower part. The choice between these two alternatives depends mainly on criteria as to the base of Cambrian strata in general. At all events, as Eardley and Hatch state, their analysis of the evidence is opposed to Hinds' assumption that the beds under consideration are of pre-Belt age.

Numerous other sedimentary rocks of known or probable Precambrian age are known in Utah but many have been so little studied that significant data are scanty. Hinds' (1935, 1936a, 1936b) summaries furnish the principal basis for the notes below. Beds reputed to be similar to the Big Cottonwood formation, Mineral Hill tillite, and Mutual formation are widespread on the west side of the Wasatch Range from north of Brigham to south of Provo. Baker (1947) calls the tillite Cambrian(?), raising once more the query as to where to place the Precambrian-Cambrian boundary. Hinds says these rocks have been identified in northern Utah and may be included in the so-called Cambrian quartzite of the Tintic and Oquirrh Ranges, a suggestion that remains open to question (Lindgren and Loughlin, 1919, p. 29; Gilluly, 1932, p. 7–9, and oral communication, 1957). Distinctly older rocks, largely schist, gneissic quartzite, and granitic gneiss, underlie the Big Cottonwood formation and its supposed correlatives in several localities and in addition are exposed widely in localities in northern Utah in which equivalents of the Big Cottonwood seem to be absent. There is a distinct unconformity at the top of the highly metamorphosed rocks and, at least in most places, a less easily recognized one at the base of the recognized Cambrian beds. The highly metamorphosed rocks appear themselves to be divisible into two assemblages, the upper of which includes the Little Willow series (Crittenden and others, 1952, p. 3) near Salt Lake City, the Farmington Canyon complex farther north (Eardley and Hatch, 1940; Bell, 1952, p. 38–43) and unnamed rocks in other parts of Utah. The lower of the two assemblages of highly metamorphic rocks appears to be largely gneissic (Blackwelder, 1925) and old enough so that it has little bearing on the problems discussed in the present paper.

The Uinta Mountains seem to contain Precambrian rocks of two general sorts (Hinds, 1936b, p. 82–92; Forrester, 1937, p. 637). The older one is quartzite and schist, presumably part of the ancient assemblage represented by the Farmington Canyon complex in the Wasatch Range. The younger one contains conglomerate and quartzite and was regarded by Hinds and by most of the earlier students he cited as correlatable with rocks they referred to as the Big Cottonwood formation and related beds. The view commonly held now is that these latter, called the Uinta Mountain group, are of Precambrian age and as young as the Big Cottonwood and supercumbent Precambrian strata (Crittenden, 1955; Hansen, 1955).

WYOMING AND COLORADO

In the mountains of western Wyoming rocks of probable Precambrian age are extensively exposed (Love, Weitz, and Hose, 1955), but few details are available (Blackwelder, 1926, 1935). A highly metamorphosed, largely gneissic, complex may represent the oldest rocks of the region. The alternative that the appearance of antiquity results from local metamorphism and that the rocks grade into less altered ones has not been ruled out. Even the later rocks are extensively recrystallized and deformed. In the Medicine Bow Range they include arenaceous, dolomitic, and volcanic deposits. Some of the dolomite contains stromatolites. Three sets of beds of glacial origin are reported. Blackwelder (1926, p. 621) tentatively considered these stratified rocks as older than such Upper Precambrian units as the Belt and Grand Canyon series because they are more intensely folded and metamorphosed than the Paleozoic strata in the surrounding region. The presence of stromatolites broadly similar to those now known to be widespread in the Belt series and of glacial deposits like those of the Mineral Hill tillite near Salt Lake City suggests that the beds in the Medicine Bow Range might be younger than Blackwelder supposed.

Precambrian rocks are widely exposed in Colorado (Burbank and others, 1935) but most are gneiss, schist, and kindred rocks so metamorphosed and so largely granitoid as to be of little interest in the present connection. The Needle Mountains group in southern Colorado and little-known small exposures near the northwest corner and in adjacent areas in Utah and Wyoming are the only rocks that might be of interest in connection with the Belt problem.

In the Needle Mountains group (Cross, Howe, and Ransome, 1905; Cross, Howe, and others, 1905; Cross, and Larsen, 1935, p. 17–32, 110–111) the thick Vallecito conglomerate is at the base and the Uncompahgre formation in the upper part. The latter consists of quartzite and slate, and varies in color and in degree of metamorphism. Inspection of the formation south of Ouray gave the present writer the impression that, in general character, the formation does not closely resemble any of the components of the Belt series. Cross and Howe report that the Uncompahgre formation was sharply folded and intruded by a granite mass and related dikes before quartzite of Cambrian age was deposited. Ancient, largely igneous appearing, rocks are overlain by the Needle Mountains group, itself intruded by granitic rocks. The Precambrian rocks of the region have been repeatedly folded and faulted, in part before the Cambrian. Structural trends preserved in the Precambrian rocks are at variance with those in unconformably overlying Paleozoic. The igneous history has been complex and the youngest of the intrusive masses of inferred Precambrian age cut the Needle Mountains group.

Hinds (1936b, p. 59–73, 119–123) concluded that the Uncompahgre formation is older than the Belt series. Clearly the complex history intervening between the deposition of the Uncompahgre and the Cambrian sedimentary rocks implies a long time interval in harmony with his conclusion.

ARIZONA AND WESTWARD

Arizona contains diverse and widely distributed rock units of Precambrian age. Many, including the Vishnu, Pinal, and Yavapai schists, are highly metamorphosed and are commonly regarded as formed early in the era. The unit regarded as next younger is the Mazatzal quartzite (Wilson, 1922, p. 299–312; Darton, 1925, p. 234–235) which rests with marked unconformity on the older rocks but has been recognized only in a rather small area in central Arizona (Darton and others, 1924). In its type locality the Mazatzal is overlain by beds of Cambrian age. However, Darton reports that the Apache group (Precambrian) overlies the Mazatzal unconformably near Payson, a short distance northeast of the type locality of the Mazatzal.

Strata commonly regarded as of late Precambrian age have been reported from numerous localities in Arizona. One such unit is the Grand Canyon series, known only in the Grand Canyon of the Colorado River (Noble, 1914, p. 37–61; Hinds, 1936b, p. 105–110). This series rests in decided unconformity on the Vishnu schist and is unconformably overlain by beds of the Tonto group (Middle Cambrian). The series includes red and green sandy and argillaceous beds, with some oolitic and stromatolitic limestone in the upper part. Lava is intercalated in the low part of the sequence among beds of red, gray and green sandstone and shale. Still lower in the section the rocks are mainly red, pink, and purple quartzite and shale. The lowest beds are stromatolitic limestone with local conglomerate, arkose, shale and sandstone. The maximum thickness appears to be fully 10,000 feet

The second unit regarded as of late Precambrian age is the Apache group, as restricted by Darton (1932). The group was originally set up by Ransome (1903, p. 38–39) and was then tentatively supposed to be Cambrian. Later, after Ransome (1911, 1915) had made minor revision, Darton (1932) decided that the uppermost formation in the group as defined by Ransome was Cambrian, because it contains Upper and Middle Cambrian fossils, and the rest was of Precambrian age. Darton restricted the name Apache to the beds he assigned to the Precambrian, and pointed out their resemblance to the Grand Canyon series. The Apache group, as restricted, consists, in descending order, of the Mescal limestone (with stromatolites), the Dripping Spring quartzite, Barnes conglomerate, Pioneer shale, and Scanlan conglomerate. Darton reports an unconformity at the top of his Apache group. In its type locality the group is only about 1,400 feet thick. The Apache group, as restricted, is fairly widespread in central and southern Arizona and Darton's ideas as to its age appear widely accepted (McKee, 1952). Resemblances between the Grand Canyon series (particularly its lower part) and the Apache group on the one hand and the Belt series on the other are obvious. In the future, comparisons among the stromatolites in the three groups may yield further data on this point.

Old sedimentary rocks are exposed at intervals from the Grand Canyon west to the vicinity of Death Valley, Calif., and are regarded by Hinds (1936b, p. 105–107) as correlatable with the Belt series. Wheeler (1947, p. 147–159) has discussed the rocks in this region (especially those of the Nopah Range, Nev.) in relation to the general problem of the position of the Precambrian-Cambrian boundary. He states that certain lithologic units become progressively younger eastward, passing from Precambrian, through Lower Cambrian to Middle Cambrian. Conversely, in his view, the base of the Cambrian rocks courses downward through these lithologic units until it meets an upwardly transgressing regional unconformity. This unconformity would be regarded as marking the base of the Cambrian rocks by those who emphasize unconformities as limiting major stratigraphic units. The validity of Wheeler's concepts and nomenclature depend on the emphasis to be placed on known fossils versus unconformities and other criteria in making stratigraphic correlations over broad regions. Longwell (1950, p. 414–415) considers the entire sequence just referred to as of Cambrian age, although he acknowledges that the lower beds in the sequence do not contain known fossils. He regards the record of essentially continuous deposition as the significant feature on which to base the age assignment.

In Death Valley (Noble, 1934, p. 173–174; McCulloh, 1954, p. 13–15) rocks reputed to be Precambrian include a highly metamorphosed assemblage and in addition comparatively unmetamorphosed beds as much as 11,000 feet thick. Although the two appear generally separated by a marked unconformity, in places none is recognized. The relatively metamorphosed beds include limestone, shale, sandstone, and conglomerate, intruded by sills and dikes of basic igneous rock. The limestone contains algal remains. Noble says these strata are "as a whole roughly equivalent to the Algonkian Grand Canyon series (Unkar and Chuar groups)" and are separated from overlying beds of Early Cambrian age by an angular unconformity.

TEXAS AND VICINITY

Precambrian rocks crop out in various isolated localities in Texas. Those near Van Horn have recently been thoroughly discussed (King and Flawn, 1953,

p. 18–20, 27–34, 39, 41–43, 45–49, 51–67, 73–97, 125–133). Here the Carrizo Mountain group consists of metamorphic, presumably relatively old, rocks. The Allamore formation and Hazel formation (formerly grouped in the Millican formation) are below the Van Horn sandstone, now regarded as Precambrian(?). The Allamore consists of interbedded cherty limestone, phyllite, and volcanic rocks. Some of the limestone contains structures that may be stromatolites. The Hazel formation has a thick conglomerate made up mainly of angular fragments at the base, interbedded with and succeeded upward by silty, red sandstone, overlain unconformably by the coarse, red Van Horn sandstone, which itself is unconformably below beds of definitely Cambrian age. King agrees with Darton (1932) in calling attention to the resemblances between the Allamore and Hazel formations in Texas and the Grand Canyon series in Arizona. Judging by published accounts, the resemblances are sufficient to suggest strongly approximate age equivalence. Perhaps the Van Horn sandstone, in spite of the unconformity at its base, is not greatly different in age from the youngest Precambrian strata in Arizona. The three formations in Texas did, however, have a more tumultuous structural history prior to the beginning of the Cambrian period than those in Arizona did. The suggested correlation between the units in Texas and those in Arizona implies a similar relationship with the Belt series, or part of it, but the difference in history raises doubts as to the correlation.

Other areas of Precambrian rocks are known in Texas and New Mexico. Most, however, consist mainly of metamorphic and igneous rocks and are thus not of interest here (Darton, Stephenson, and Gardner, 1937, Darton, 1928).

Present information suggests that sedimentary rocks of late Precambrian age are rare in northern Mexico (Garfias and Chapin, 1949, Lochman-Balk, 1956, p. 530; Mejorada, 1956) and absent in the rest of the republic. Near Caborca in the state of Sonora, however, (Arellano, 1956, p. 509, 515–526) a thick sequence of limestone, dolomite, shale, and quartzite, containing *Collenia*, has been referred to the Sinian system, a designation that indicates that these rocks are regarded as essentially equivalent to the Belt series of the northern United States. Fossiliferous rocks of Early Cambrian age are present in the area but their relations to the late Precambrian beds are not definitely known.

CANADA AND ALASKA

Rocks in British Columbia and Alberta that are correlated with the Belt series (Reesor, 1957, Gunning, 1957) are widely exposed and diversely named (p. 21–25). They stretch northward perhaps as far as the vicinity of lat 54°. The age relations of the northernmost are imperfectly known. Some may be of Early Cambrian age. Similar rocks are known at intervals past the southern boundary of Yukon Territory, which is lat 60°. Farther northwest into Alaska uncertainties increase (Gunning, 1957, p. 180–182; Poole, 1955), but late Precambrian strata are probably present. Some fossils of Early Cambrian age have been found in rocks "for which stratigraphic details are not yet available" (Gunning, 1957, quoting Okulitch).

In Alaska the Tindir group (Mertie, 1933, p. 392; Smith, 1939, p. 8–9) includes sedimentary rocks thought to be of late Precambrian age. The extent westward of this group is not known and the area assigned to it on figure 7 may be too liberal. The latest geologic map of Alaska (Dutro and Payne, 1954) shows lower Precambrian rocks west to near long 148° but it seems likely that these may include younger rocks. No Precambrian stratified rocks are known close to the present Alaskan coasts.

The Northwest Territories contain several exposures of rocks that are regarded, with varying degrees of certainty, as of late Precambrian age (Gunning, 1957, p. 182; Brown and Wright, 1957, p. 79–92). Some of these have been tentatively correlated with the Belt series (Williams, 1923, p. 69b–76b).

SUMMARY

The data outlined above show that rocks similar in various ways to the Belt series in Montana are present in many places from Alaska to northern Mexico. In the Rocky Mountains region in the United States the Bannock volcanic formation and the Pocatello formation, the Big Cottonwood formation, with the Mineral Fork tillite and Mutual formation above it, comparable rocks in other parts of Utah, possibly some in Wyoming, the Grand Canyon series and Apache group, the upper part of the Precambrian sequence in southern Nevada and nearby parts of California, possibly the Allamore formation, Hazel formation and perhaps also the Van Horn sandstone, rocks near Caborca, Mexico, and less well-known units elsewhere may be all of the same age as part or all of the Belt series. Throughout the Rocky Mountains region some parts of the rock units listed are sufficiently similar to the Belt series as to suggest close similarity in genesis. These include stromatolitic limestones and probably some other strata that in colors and textures remind one of parts of the Belt series. It may be possible eventually to establish correlations between the volcanic rocks in the Belt series and those in certain of the units listed above. A glacial origin has been advocated by some for certain of the rocks in Utah, Wyoming, and southern Idaho. If parts of the Belt series are found in future studies to be related to glaciation, an age tie may be indicated. On the other hand, in a few instances, such as the Van Horn area in Texas, the rocks have had a complex pre-Paleozoic history and thus contrast so much with the Belt series in Montana as to suggest either that they are not correlatable in age or that, in these instances, the structural history was exceptional.

BELT PALEOGEOGRAPHY

The basin in which the Belt series was deposited was a geosyncline. If the various rocks mentioned in the previous section were deposited in depressions linked with the Belt basin, the composite geosyncline that resulted was of major dimensions, comparable to the younger troughs whose crumpled sedimentary deposits have given rise to the world's great mountain ranges. It is, therefore, surprising to note that much of the Belt series in Montana is almost undisturbed or, at most, warped. Folding and faulting have affected the series in certain places but both appear everywhere to be of late Mesozoic or early Tertiary date, and the old rocks are less intricataely crumpled than Paleozoic and Mesozoic strata in the same part of the state. Even where marked discordance between the Belt series and Paleozoic rocks is best displayed, the Precambrian rocks are not as intensely deformed as to imply major orogeny. Outside of Montana the difference in degree of deformation between the late Precambrian and supercumbent strata varies but in most places the contrast is perceptible. This raises two questions. Why did the thick prism of sedimentary rocks of late Precambrian escape orogenic movements, and by what means did the Belt basin subside so greatly that roughly 40,000 feet of strata were deposited in it. Similar queries could be raised in regard to rocks throughout the world that resemble the Belt series, but satisfactory answers are not available. The first query implies relative quiesence in certain localities throughout the world at about the end of Precambrian time but supplies no reason for the lack of major orogeny. The subsidence can be assumed to result from isostasy but such an explanation does not seem completely adequate. In a sea as shallow as this one appears to have been the accumulation of beds of mud and lime at any one time must have been thin and light. The resulting weight on the earth's crust would have exceeded that where no deposition was occuring but the difference would not seem to have been great enough to produce the result recorded. To argue that flexing of the crust by some other process aided isostasy seems to contradict the evidence of orogenic quiescence. Perhaps the seeming inconsistency results from our inability to evaluate the character and magnitude of the crustal movements involved and to fix their timing accurately enough.

The summaries just given, while demonstrating the scanty, patchy character of current knowledge of the Precambrian rocks of the Rocky Mountains within the United States, justify the idea that at least three major components exist. The Belt series belongs to the youngest of the three. It is much thicker and seems originally to have been continuous over a far larger area than any of the other units that make up the youngest component of the Precambrian sequence here discussed. The next older component, also mainly sedimentary, is in demonstrable stratigraphic relations to the others in only a few localities. It may be more widespread than is at present realized. Conceivably the oldest of the beds now placed in the component to which the Belt series belongs may belong to this older component. The third and oldest of the three components comprises the gneisses and schists, often spoken of as the basal complex, on which rocks of clearly sedimentary origin rest with marked unconformity. The tendency during the past half century has been to shrink the areas of exposed basal complex by showing that certain metamorphosed rocks are of sedimentary origin and far younger than originally supposed. This tendency may continue. Nevertheless the basal complex of ancient rocks is doubtless widespread throughout the Rocky Mountains.

Paleogeographic maps of late Precambrian time indicate seaways extending from the present position of Montana and Idaho northward to the ocean. Some show southward connections also. These maps are based necessarily on the distribution of present outcrops of rocks of supposed late Precambrian age. Various uncertainties as to correlations are inevitable, and much vital information is buried under later deposits. The principal reports containing pertinent data have been reviewed above.

Figure 7 shows that south of Montana outcrops of rocks that can be thought of as allied, in any way, with the Belt series are small and scattered. In southern Canada the area is larger. Here the Shuswap complex or series is included on figure 7 with rocks allied to the Belt series. This seems well established for much of the Shuswap, but if strata of pre- and post-Belt age could be excluded the western part of the area in Canada might be reduced significantly. In the Yukon and Northwest Territories many gaps in the data remain. Here also the map may show areas that would be eliminated if the rocks were better known. Obviously the greater the distance from the type locality of the Belt series, the greater the uncertainity as to relationship.

Among the published paleogeographic maps, that of the Fentons (1937b, fig. 13, p. 1439) deals most specifically and completely with the present problem. According to this map the Belt basin in Canada was narrow along the southwest border of Alberta and in eastern British Columbia, and widened somewhat along and west of the western boundary of the Northwest Territories. This part of the basin was largely in the present Yukon Province. It passed into Alaska near the northeast end of the boundary between that State and Canada and thence reached the Arctic Ocean. An embayment swung eastward and southward near the point where the western border of the Northwest Territories reaches the ocean. This embayment occupied much of what is now the District of Mackenzie in the Northwest Territories.

The Fentons note that their map is modified from one by Cooke (1933, p. 468). That map is drawn in simpler style than the one by the Fentons. It shows no connection between the belt of late Precambrian rocks that extends from Idaho and Montana into Alaska and the other body in the Northwest Territories that corresponds to the embayment shown by the Fentons. Cooke's map represents his concept of conditions in the Keweenawan period. Its caption says: "The oldest deposits of the Rocky Mountain trough (Belt, Purcell, and Windermere, Tindir) are doubtfully of this period." The Fentons map the Belt basin as being within a geosynclinal area whose margins are coincident with those of the basin in northern Canada. Farther south, especially within the United States, the Fentons represent the geosyncline as wider than the Belt basin and as having a broad eastern extension through North and South Dakota. Cooke's map shows little within the United States and does not have the distinction between basins and geosynclines. It does, however, indicate a basin in the area of the Great Lakes that extends westward into the Dakotas and would thus, if continued, correspond to the eastward extension of the geosyncline indicated by the Fentons. Figure 7 shows a small area of late Precambrian strata in South Dakota but, wherever exposed in Montana east of the Belt basin and in Wyoming, earlier Precambrian rocks are overlain by Paleozoic and later strata (U.S. Geol. Survey, 1932, Geologic map of United States). This is opposed to the idea of any connection between Belt and Keweenawan deposits.

Schuchert's map (1923, fig. 1, p. 168–170) portraying his idea of conditions in late Precambrian time is similar to that of the Fentons in and near Montana and Idaho. He shows the northward-trending belt as swinging abruptly westward to reach the Pacific across southeastern Alaska but notes that it may extend to the Arctic Ocean instead. Schuchert shows an extension southwestward from Montana that reaches the Pacific across southern California that is not shown on the Fenton's map. Schuchert's idea of a link with the ocean across southeastern Alaska receives no support from available geologic maps.

In a recently published series of paleogeographic maps by Schuchert (1955) no reflection of the Belt basin is shown for early Cambrian time. The second, third, and fourth in the series of maps of Cambrian time record the presence of the parts of the Windermere series and related rocks that are of Cambrian age. The fourth map shows the Flathead quartzite (there termed "sandstone") in Montana but with a more restricted distribution than is now known. The fifth map, which records conditions in Upper Acadian time, shows a seaway broadly identical to that shown on the paleogeographic maps of late Precambrian time already cited. As no major orogenic movements are on record for most of the area containing the Belt series in the interval between the end of Belt sedimentation and the deposition of the Flathead quartzite, the marked hiatus indicated by this series of maps is strange. It would be explained by the concepts advanced in the present paper that any break in deposition near the beginning of the Paleozoic may not have been a major one and that part of the uppermost Belt series may be of Early Cambrian age. On these assumptions, paleographic maps for Belt through Middle Cambrian time would be similar throughout.

In the light of the generalized paleogeographic maps just cited and the compendium of present data represented in figure 7 it seems reasonable to conclude that the Belt basin was part of a depression that extended from the site of Montana to the ocean somewhere near where the Arctic Ocean now borders Alaska. In the Northwest Territories and Saskatchewan and thence northeastward into Greenland there are scattered exposures of rocks that may not differ greatly in age from the Belt series. One interpretation might be that these correspond to an eastern embayment similar to but perhaps even larger than that inferred in the Fentons' (1937b, fig. 13) paleogeographic map. An alternative and perhaps more probable interpretation would be that there was a break in sedimentation roughly in the region along the present valley of the Mackenzie River and the present Great Bear Lake, Great Slave Lake, and Lake Athabasca. Rocks farther east would have been deposited in one or more depressions independent of that of the Belt series and its associates. The area indicated constitutes a zone several hundred miles wide in which no rocks of late Precambrian age are known to be exposed. The rocks of supposed late Precambrian age east of this zone shown on figure 7 rest on the old rocks of the Precambrian shield (Brown and Wright, 1957, fig. 1).

The extension of the Belt basin eastward through North and South Dakota indicated by the Fentons seems unlikely because, as indicated on geologic maps of Montana and Wyoming, rocks of early Paleozoic age rest on reputably old Precambrian rocks in several localities (Ross, Andrews, and Witkind, 1955; Love, Weitz, and Hose, 1955). There is a small exposure of Precambrian rocks in western South Dakota (Darton, 1951), but this mass has been inferred to be older than Keweenawan (Darton and Paige, 1925, p. 4). If so, it is too old to be significant in the present connection. It is overlain by rocks of Late Cambrian age.

South of Montana there are enough exposures of rock with supposed affinities to the Belt series to justify the idea that the depression referred to above persisted southward across Utah and Nevada into southern California. Here, as farther north, uncertainties of various kinds exist and the outline of the postulated depression shown on figure 7 is subject to modification as new data become available. Rocks along the Grand Canyon of the Colorado and farther south in Arizona have so many resemblances to the Belt series that they are included. The small exposures in northern Mexico likewise seem best included. Those in Texas are excluded because to swing the eastern border far enough east to include them would introduce a complexity in form not supported by other evidence, especially as the rocks in Texas are not very closely allied to the Belt series in other respects.

The postulated depression indicated on figure 7 corresponds broadly to the basins or geosynclines shown on the published maps cited above, although differing in details. While the idea that some such a depression was a major feature of the geography of late Precambrian time is regarded with confidence, it is not necessary to assume that throughout its existence this depression was occupied by a single, continuous water body. Perhaps this was true at times, but it is at least equally possible that at times the depression was a series of basins of varying depth and characteristics. Connections between them may have been slight or absent entirely in response to variations in local conditions.

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